Ocean-atmosphere coupling in midlatitudes: does it invigorate or damp the storm track?

Arnaud Czaja[1]

Department of Physics, Imperial College, London & Grantham Institute for Climate Change at Imperial College

Abstract

A few ideas regarding the impact of the extra-tropical oceans on the storm track are discussed. The first refers to the "anchoring effect" of the ocean circulation on the storm track through the thermodynamic impact of warmer and more spatially varying sea surface temperatures on the low level atmospheric temperature field. The second refers to the thermal damping of weather disturbances through air-sea heat exchanges and the idea that the ocean circulation acts as a brake on the growth of baroclinic waves. It is emphasized that the two paradigms are in conflict, one being a source of instability and growth, while the other is a source of damping and decay. The impact of the ocean on the atmosphere through low Richardson number processes occurring at atmospheric fronts is also highlighted as a promising new area of research.

1 Motivation

There is currently a very exciting resurgence of interest in ocean-atmosphere coupling in midlatitudes. Indeed, although the natural climate variability simulated in the extra-tropics by climate models has previously been interpreted as predominantly reflecting the passive response of the ocean to the atmosphere on timescales ranging from weeks to decades (e.g., Kushnir et al., 2002), recent results suggest that this interpretation might be an oversimplification.

Firstly, it is clear that when the SST anomalies are large enough and maintained over a few decades, the atmosphere in climate models is significantly influenced. This has emerged in the context of the response to anthropogenic forcing which is mediated by changes in the North Atlantic ocean circulation (Woollings et al., 2011). It has also been shown that biases in the simulation of Atlantic blocking can be corrected by a better representation of North Atlantic currents in the Hadley Centre's HadGEM3 climate model (Scaife et al., 2011). Interestingly, these results are not tied to the latest generation of climate models. For example, the study by Smith et al. (2010), using the older HadCM3 model (Gordon et al., 2000), showed predictive skill of tropical cyclones' activity arising from extra-tropical SST anomalies with a lead time of a few years. Rather than improvements in models, it appears that the oceanic influence can now be

[1] Corresponding author address: Dr. A. Czaja, Imperial College, Prince Consort Road, Huxley Building, Room 726; London SW7 2AZ, UK. tel: +44 (0)20 7594 1789 email: a.czaja@imperial.ac.uk

better isolated from intrinsic atmospheric variability as a result of the use of longer simulations, ensemble of multi-decadal simulations, and multi-model ensemble averaging. Physically, it is likely that this reflects the larger SST anomalies developing in the extra-tropics as timescales increase. These results comfort statistical studies which have shown the existence of oceanic precursors to large scale changes in atmospheric circulation over the North Atlantic (Czaja and Frankignoul, 2002) and the North Pacific (Frankignoul et al., 2011) in reanalysis datasets, as well as studies which have sug- gested a relationship between warm North Atlantic ocean and higher frequency of blocking in historical observations (Deser and Blackmon, 1993; Hakkinen et al., 2011).

In addition, there is indication from recent observations and high resolution atmospheric simulations that the strength of the oceanic influence on climate might simply be underestimated in the current generation of climate models. Indeed, satellite-based surface winds and sea surface temperature measurements have indicated the presence of a strong coupling between midlatitude oceanic fronts and winds in the maritime boundary layer on spatial scales of O(100km), the so-called mesoscale (Chelton et al., 2004). Numerical experiments with atmospheric models run at higher spatial resolution than IPCC-class models have suggested that the influence of ocean fronts can extend well above the boundary layer over the Gulf Stream (Minobe et al., 2008) and even shift the Atlantic storm-track significantly (Woollings et al.,2009).

A reason why the strength of extra-tropical ocean atmosphere coupling could conceivably increase with resolution is proposed in Fig. 1. In a coarse climate model (Fig. 1, left panel), the bulk of isentropes in midlatitudes have intersected the sea surface in the subtropics. Changes in subtropical SSTs are felt locally at low levels through air-sea heat exchange but the modified air masses are subsequently advected approximately along isentropes further poleward and upward. The midlatitude Jet Stream is thus influenced remotely from the subtropics. At higher resolution however (Fig. 1, right panel), more extra-tropical isentropes intersect the sea surface locally

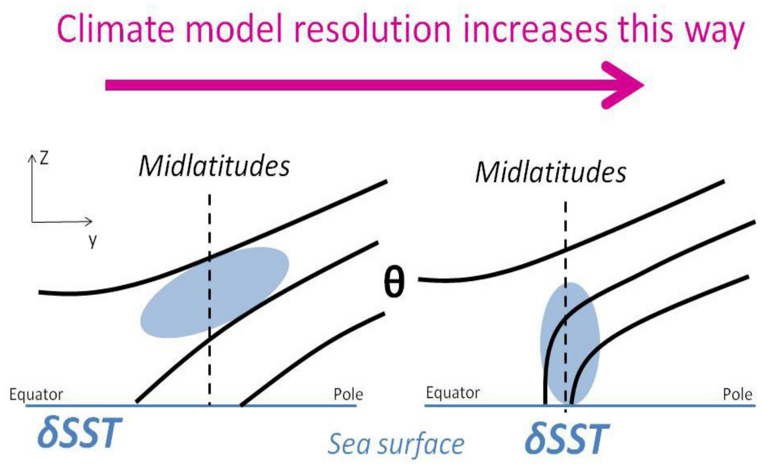

Figure 1: Schematic illustrating how the strength of extra-tropical ocean atmosphere coupling can increase with spatial resolution. In each panel a change δ SST in sea surface temperature is indicated in blue and the part of the atmosphere that it affects is indicated by a blue patch. Black lines denote surfaces of constant potential temperature θ . See text for details.

because of the better representation of frontal structures embedded in extra-tropical cyclones (steeper isentropic slopes). As a result, the mid-latitude atmosphere can now be influenced locally by changes in extra-tropical SST following a "quasi upward" isentropic pathway, in addition to the SST forcing from the subtropics.

2 Scope of the note

The following discussion focuses on the impact of the ocean circulation on the storm track. Its starting point is that by moving warm and cold waters around, the ocean circulation modifies the SST distribution in two ways:

1. It enhances its gradients at the latitude of the maximum surface westerlies (as a result of the confluence of isotherms at the boundary between subtropical and subpolar gyres in the Northern Hemisphere and, in the Southern Hemisphere, by strengthening the jets and associated temperature fronts making up the Antarctic Circumpolar Current).
2. It raises the SST at western ocean boundaries in the subtropics (as a result of warm advection by western boundary currents such as the Gulf Stream or the East Australian current in the Southern Hemisphere).

These features act together as a source of instability for the atmosphere, helping to lower its Richardson number at low levels over western boundary currents. This is sometimes referred to as the "anchoring effect" of the ocean on the midlatitude storm track (e.g., Hoskins and Valdes, 1990; Nakamura et al. 2008; Minobe et al., 2008; Czaja and Blunt, 2011). In addition, these SST features also enhance the thermal damping of baroclinic waves at low levels, shaping the spatial and temporal structure of the weather systems (e.g., Valdes and Hoskins, 1988; Hall and Sardeshmukh, 1998; Zhang and Stone, 2011). Support for these two broad classes of ideas is provided in sections 3 and 4, respectively.

Before doing so, it is worth emphasizing that although the SST features highlighted in points 1 and 2 above were introduced in the context of the time mean state, similar SST anomalies develop on seasonal to interannual timescales as a result of large scale atmospheric forcing (air sea heat fluxes and anomalous Ekman advection). The North Atlantic Oscillation for example creates, in its positive phase, a tripolar SST pattern having both features 1 and 2. Thus, the paradigms discussed below are relevant to timescales ranging from seasonal to inter-decadal.

3 Anchoring of the storm-track

Storm tracks can be singled out as regions of maximum growth for baroclinic waves in the midlatitudes (e.g., Hoskins and Valdes, 1990). As a result, any air-sea interactions whose effect is to enhance the growth rate will in effect act to localize the storm track. A standard measure of growth is the Eady growth rate σ_{Eady}:

$$\sigma_{\text{Eady}} \propto f_o / \sqrt{R_i} \text{ with } R_i \equiv \frac{N^2}{|\upsilon_z|^2} \tag{1}$$

in which f_o is the the Coriolis parameter and R_i is the Richardson number, the ratio of the stabilizing effect of buoyancy (N^2) over the destabilizing effect of windshear ($|\boldsymbol{v}_z|^2$). Assuming that spatial variations in SST are imprinted on the low level atmospheric temperature field, enhanced SST gradients will, through the thermal wind relationship, enhance $|\boldsymbol{v}_z|^2$ and as a result lower Ri and enhance σ_{Eady}. Likewise, assuming that warmer SSTs lead to a local reduction in low level static stability, warmer SSTs are expected to lower the Richardson number and enhance σ_{Eady}.

As the previous discussion made it clear, the anchoring effect of the ocean requires a mechanism to “communicate upward” variations in SST from the sea surface to low levels. The growth rate σ_{Eady} is typically estimated over the 700 – 800mb layer, i.e., most of the time above the marine boundary layer. It is however likely that the boundary layer itself is enough to anchor the storm track. One line of thought in support of this statement is that surface temperatures are dynamically active in a rapidly rotating fluid: horizontal SST variations on the scale of the deformation radius create a potential vorticity sheet whose circulation occupy the entire troposphere (Bretherton, 1966; Gill, 1982 –see his section 13.2). As a result, intuition based on non-rotating fluid dynamics might mislead in putting too much emphasis on mechanisms allowing a “deep penetration” of SST signals into the atmosphere such as convection. A more pragmatic line of thought in support of this view is that atmospheric models perturbed over a shallow boundary layer do indeed seem to be affected over a deeper layer. In simulations of air-sea interactions occurring over a stable boundary layer (idealized aquaplanet type geometry or Southern Ocean - like, as opposed to the more unstable situations encountered in the Northern Hemisphere in winter as a result of large land sea temperature contrasts) the synoptic eddy activity is clearly modulated up to the 250mb level (e.g., Nakamura et al., 2008; Brayshaw et al. 2008). Surface effects are also key to the numerical experiments of Hall and Sardeshmukh (1998) discussed further in section 4.

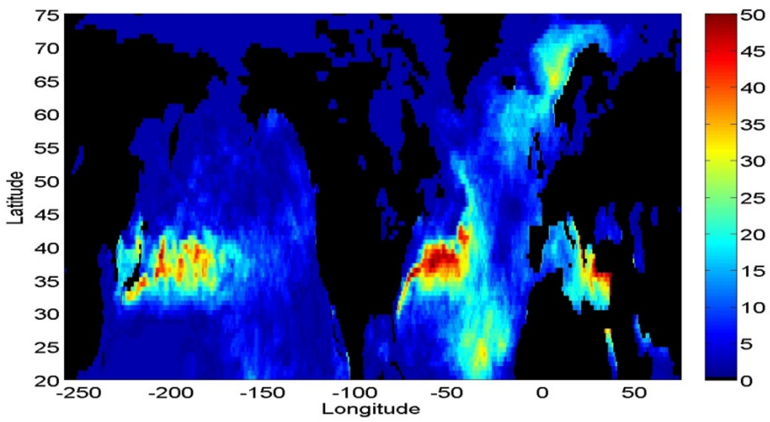

Figure 2: Fraction of the time (in %) during the boreal winter 2003/2004 when the moist surface entropy s_o is greater than the entropy at the tropopause. This is a simple measure of convective instability over a deep atmospheric layer. Note that s_o is taken as the entropy of moist air at the same temperature as the sea surface and with a relative humidity of 80 %. See Czaja and Blunt (2011) for more details on the calculation.

Communication of spatial or temporal variations in SST beyond a shallow 1000mb – 900mb is nevertheless expected in the Northern Hemisphere in winter because of the unstable conditions caused by the simultaneous occurrence of westerly winds, cold eastern continental boundaries and warm western boundary currents at midlatitudes. For example, Korty and Schneider (2007) show a clear oceanic signal at 775mb in January over the Gulf Stream and the Kuroshio extension in their study of the occurrence of near zero moist potential vorticity events in the NCEP-NCAR reanalysis data (their Fig. 9e). This point is further illustrated in Fig. 2 which displays the fraction of time during the 2003-04 boreal winter in which conditions are favourable for a moist convective instability from the sea surface to the tropopause in the ERA interim dataset (Berrisford et al., 2009). The Gulf Stream and Kuroshio emerge as regions of high occurence (typically 30 % of the time) of such conditions.

4 Thermal damping of weather systems

Weather disturbances reflect the interaction between two propagating edge waves, one near the Earth's surface, the other near the tropopause (Bretherton, 1966; Eady, 1949; Gill, 1982). When temperature anomalies are created at low levels through horizontal stirring of the background temperature gradient, an air-sea temperature difference results which restores thermodynamic equilibrium (in effect restoring air temperature to SST owing to the much larger thermal inertia of the oceans) –see Fig. 1. As a result, the lower boundary edge wave is more rapidly damped than the upper boundary one, their interaction is made less efficient, and the atmosphere becomes overall less unstable. This effect was clearly seen in the severe damping of baroclinic waves that occurs in life cycle experiments coupled to a slab ocean mixed layer, in comparison to the life cycle simulated without this thermal coupling (Hoskins, personal communication).

Because of the enhanced SST gradient caused by oceanic advection near the latitude of maximum surface westerlies (point 1 in section 2), the air-sea temperature contrast resulting from low level stirring is even more pronounced (Fig. 3, green arrow) than it would be in absence of ocean circulation (Fig. 3, blue arrow –as occurs in slab ocean mixed layer experiments). As a result, baroclinic waves will be more strongly damped when the ocean circulation strengthens. Because of the dynamical effect of this damping, emphasized in the previous paragraph, it is the overall temporal (growth rate) and spatial (wavenumber) structure of baroclinic waves which can then be affected. For example, Hall and Sardeshmukh (2008) found an increase in wavelength and period (wavenumber 7 → 6, period 6 → 6.5 days), a severe decrease in growth rate ($0.4\text{day}^{-1} \rightarrow \approx 0$), and a deeper vertical structure for the meridional heat flux, when increasing low level thermal damping[2] from zero to a more realistic value of about 1 day^{-1} in a linearized primitive equation model.

The schematics in Fig. 3 provides a rough quantitative estimate of the change in low level thermal damping brought about by changes in the strength of the ocean

[2] Strictly speaking, by increasing the damping on both vorticity and temperature at low level. Nevertheless, Hall and Sardeshmukh showed that the same behaviour is found when increasing thermal damping at fixed vorticity damping.

circulation. As discussed in Fig. 3's caption, the damping of a temperature anomaly is proportional to the air-sea temperature difference ΔT . In an ocean at rest ΔT equals the distance BC between points B and C in Fig. 3. Conversely, with active ocean circulation, ΔT equals the distance AB between points A and C. The relative change in thermal damping brought about by a spin up of the ocean circulation is thus on the order of AB/BC. From the results presented by Wilson et al. (2009), $AB \approx 3K$ (their Fig. 9e) while $BC \approx l dT_o/dy \approx 1000km\ 1/100km = 10K$ in which l is the typical meridional displacement of air parcels and dT_o/dy is a typical meridional temperature gradient (see Fig. 3's caption). Switching on and off ocean currents should thus modulate the thermal damping of baroclinic waves by approximately 30%. In other words, the ocean circulation is expected to act as a significant brake on the growth of weather systems.

Support for this idea comes from the numerical experiments conducted by Wilson et al. (2009). By building complexity in a climate model, from a ground state with no orography and a motionless slab ocean to a realistic state with orography and a full dynamical ocean, they were able to isolate the effect of the ocean circulation on various measures of storm activity. The results for eddy kinetic energy at 250mb and low level heat flux for example support clearly the damping effect of the ocean circulation (compare Fig. 5b with 5e and Fig. 6b with 6.e), with a magnitude comparable to the rough scaling above. Note that the damping effect of the ocean circulation is also entirely consistent with the hypothesis put forward by Bjerknes (1964) regarding compensation between oceanic and atmospheric poleward heat transports.

Additional support for thermal damping of baroclinic waves by air-sea interactions is provided, on seasonal to interannual timescales, by an investigation of the ERA interim dataset (Fig. 4). In this analysis, the alignment of sea surface isotherms with the local windshear vector ($\partial \boldsymbol{v}_{750}/\partial p$ in which $\boldsymbol{v}_{750}$ denotes the wind vector at 750mb and p is pressure) was measured at daily intervals by an angle denoted by θ (e.g., $\theta = 0°$ means perfect alignment and so, because of thermal wind relationship, warm air over warm water / cold air over cold water). The plot shows that for all ocean basins considered (different colours, see Fig. 4's caption for definition) strong windshear vectors are always more aligned with the underlying sea surface isotherms than weak windshear vectors (i.e., the bin centered on $\theta = 0°$ is more populated for strong windshear events). In other words, situations in which warm air does not overlay warm waters are more strongly damped, as expected from Fig. 3.

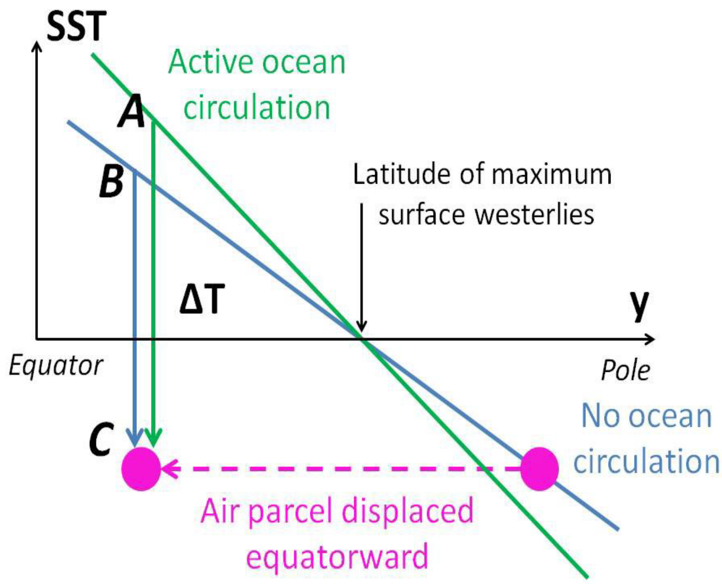

Figure 3: Schematic illustrating how thermal damping of a low level baroclinic wave depends upon the ocean circulation. The temperature T of an air parcel displaced equatorward varies approximately as $dT/dt = Q$ in which d/dt is the rate of change following the horizontal flow and Q is proportional to the air sea heat flux. Decomposing T as the sum of an environmental or background component $T_o = T_o(y)$ where y denotes latitude, and a perturbation T', this can be rewritten as $dT'/dt = -v dT_o/dy + Q$ in which v is the meridional velocity. As expected, an equatorward displaced parcel creates locally a cold anomaly (the 1st term on the r.h.s. is negative since both $v < 0$ and $dT_o/dy < 0$). This advective source is opposed by the air-sea heat flux $Q \propto (SST - T) \equiv \Delta T > 0$. In presence of an active circulation, ΔT (colored arrow) is increased near the latitude of maximum surface winds (green arrow) and, as a result, the growth of T' l is reduced compared to the case of a motionless ocean (blue arrow). The points labeled A, B and C are discussed in the text.

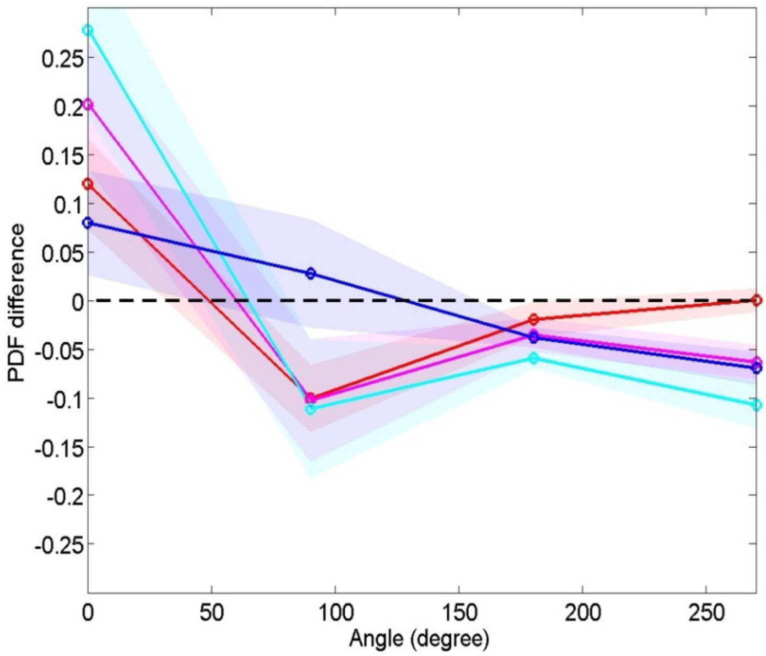

Figure 4: Difference in the probability distribution function of θ between strong (upper 10 %) and weak (lower 10 %) windshear vectors $\partial v_{750}/\partial p$ (see text for definition of mathematical symbols). The calculation was based on 5 winters (2001-2005) using domains centred on the Gulf Stream (red), the Kuroshio (magenta), the East Autralian current (dark blue) and the Brazil-Malvinas confluence region (light blue). Four bins of θ were considered (centred on $\theta = 0°$, $\theta = 90°$, $\theta = 180°$ and $\theta = 270°$) and the distributions were normalized (i.e, a uniform distribution has a value 0.25 for each bin). For a given bin, a positive value on the $y-$ axis means that this bin is more populated in strong windshear events than in weak windshear events. The shading around each curve gives ± one standard deviation within the 5-yr (i.e., 5-member) ensemble.

5 "Frontal" ocean-atmosphere coupling

The prospect of a meso-scale pathway to communicate changes in ocean circulation to the atmosphere, hinted at in Fig. 1 and Fig. 2, is extremely challenging because the associated mechanisms have been mostly studied within the context of the "weather" rather than the climate. Figure 5 for example displays the frequency of occurrence of low Richardson numbers at 700mb over the period 01/12/2003 -28/02/2004 in the ERA interim dataset. Convective events are typically highlighted as events with $R_i = 0$ (left panel) and, as expected, these occur frequently at low latitudes (in excess of 30 % of the time over land and 10 – 20 % of the time over the tropical oceans). More surprising is that similar or even larger occurrences are found for events with small but non zero Richardson numbers ($0 < R_i \leq 3$, right panel) over the Kuroshio extension, the Gulf Stream and throughout the Southern extra-tropics. These low R_i events reflect the passage of synoptic fronts and the instabilities developing on them.

An example of a relevant situation for ocean-atmosphere coupling is that occurring over western boundary currents such as the Gulf Stream when air parcels are lifted in the warm sector of a low pressure system. The moistening occurring over the warm ocean current helps to reduce the moist stability of the air column, possibly leading to an inertial instability along the steep (moist) isentropes in the right panel of Fig. 1 (see for example the discussion in Hoskins and Bennetts, 1979). In effect the interaction of the cold front with the ocean acts to further enhance the upward motion and the condensational heating present in the cyclone.

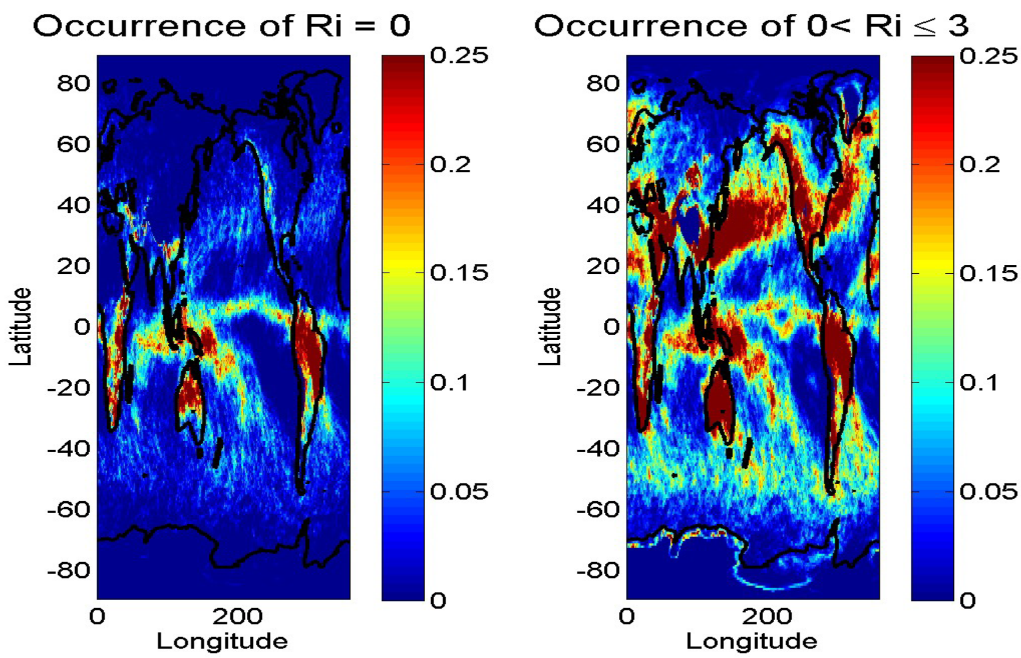

Figure 5: Fraction of the time during the boreal winter 2003/2004 when the Richardson number (R_i) at 700mb is (left) zero and (right) non zero but lower than 3. R_i was computed using a moist buoyancy frequency when the relative humidity was greater than 80 %. The strong features over the Kuroshio and the Gulf Stream in the right panel reflects mostly the time variability of the windshear (i.e., they are not affected when the local median is used instead of the actual value for the buoyancy frequency in the calculation of R_i –not shown).

6 Conclusion

The “anchoring” paradigm is somewhat simpler and more intuitive than the “thermal damping” paradigm because it relies on a thermodynamic modulation of baroclinic growth rates by the SST field. In contrast, the thermal damping paradigm relies on the non intuitive role played by boundary conditions in low Rossby number flows. Although sections 3 and 4 have highlighted where observations and climate models provide support for both paradigms, it is important to acknowledge that they are not always compatible with each other. For example, as a result of a stronger circulation of the subtropical and subpolar gyres, the anchoring paradigm predicts faster growth of weather disturbances and a larger poleward atmospheric heat transport, while the thermal damping paradigm predicts exactly the opposite. This conflict might actually be what shapes weather disturbances themselves: Nature’s solution to the problem of tapping into the enhanced baroclinicity provided at low levels by the ocean circulation while minimizing the thermal damping resulting from the large heat capacity of the oceans.

Thermal damping and anchoring are associated with large Richardson number mechanisms ($R_i \gg 1$), but it must be emphasized that changes in ocean circulation might also be communicated to the atmosphere in the fronts embedded in extra-tropical weather systems through a range of instabilities occurring at low Richardson numbers ($R_i \simeq 1$). These instabilities are currently not parameterized in climate nor weather forecast models and they might open an untapped source of predictability for the low pressure systems themselves on timescales of seasons and longer.

Acknowledgements: It is a pleasure to acknowledge inspiring discussions with Brian Hoskins and Raymond Hide, as well as with the students of the Space and Atmospheric Physics Group at Imperial College.

7 References

Bennetts, D. A., and B. J. Hoskins, 1979: Conditional symmetric instability - a possible explanation for frontal rainbands, Quart. J. Roy. Met. Soc.,105, 945-962.

Berrisford, P., et al., 2009: The ERA interim archive, ERA Report series 1.

Bjerknes, J., 1964: Atlantic air-sea interactions, Advances in Geophysics, Vol. 10, Academic Press, 1–82.

Brayshaw, D. J., B. J. Hoskins and M. Blackburn, 2008: The Storm-Track Response to Idealized SST Perturbations in an Aquaplanet GCM, J. Atm. Sci., 65, 2842–2860.

Bretherton, F., 1966: Baroclinic instability and the short wavelength cut-off in terms of potential vorticity, Quart. J. Roy. Met. Soc., 92, 335-345.

Chelton, D. B., et al., 2004: Satellite measurements reveal persistent small-scale Features in Ocean Winds, Science,303, 978-983.

Czaja, A., and C. Frankignoul, 2002: Observed impact of Atlantic SST anomalies on the North Atlantic Oscillation. J. Climate, 15, 606–623.

Czaja, A. and N. Blunt, 2011: A new mechanism for ocean atmosphere coupling in midlatitudes, Quart. J. Roy. Met. Soc., 137, 1095-1101.

Deser, C. and M. L. Blackmon, 1993: Surface climate variations over the North Atlantic Ocean during winter: 1900-1989, J. Clim., 6, 1743-1753.

Eady, E. T., 1949: Long waves and cyclone waves, Tellus, 1, 33-52.

Frankignoul, C., N. Sennechael, T. Kwon, and M. A. Alexander, 2011: Influence of the Meridional Shifts of the Kuroshio and the Oyashio, J. Clim., 24, 762–777.

Gill, A. E. J., 1982: Atmosphere-ocean dynamics, Academic Press, 182 pp.

Gordon, C., C. Cooper, C. A. Senior, H. Banks, J. M. Gregory, T. C. Johns, J. F. B. Mitchell and R. A. Wood, 2000: The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments, Clim. Dyn., 16, 147-168.

Hakkinen, S., P. B. Rhines, and D. L. Worthern, 2012: Atmospheric Blocking and Atlantic Multi- decadal Ocean Variability, Science, 334, DOI: 10.1126/science.1205683.

Hall, N. M. J., and P. D. Sardeshmukh, 1998: Is the time mean Northern Hemisphere flow baroclin- ically unstable? J. Atm. Sci., 55, 41-56.

Hoskins, B. J., and P. J. Valdes, 1990: On the existence of storm-tracks, J. Atm. Sci., 47, 1854-1864.

Korty, R. L., and T. Schneider, 2007: A climatology of the tropospheric thermal stratification using saturation potential vorticity, J. Clim., 20, 5977-5991.

Kushnir, Y. et al., 2002: Atmospheric GCM response to extratropical SST anomalies: synthesis and evaluation, J. Clim., 15, 2233-2256.

Minobe, S., et al., 2008: Influence of the Gulf Stream in the troposphere, Nature, 452, 206-210.

Nakamura, H, T. Sampe, A. Goto, W. Ohfuchi and S. Xie, 2008: On the importance of mid-latitude oceanic frontal zones for the mean state and dominant variability in the tropospheric circulation, Geo- phys. Res. Let., 35, L15709, doi:10.1029/2008GL034010.

Scaife, A. A., D. Copsey, C. Gordon, C. Harris, T. Hinton, S. Keeley, A. O'Neill, M. Roberts and K. Williams, 2011: Improved Atlantic Blocking in a Climate Model, Geophys. Res. Let., in press.

Smith, D. M., R. Eade, N. J. Dunstone, D. Fereday, J. M. Murphy, H. Pohlmann and A. A. Scaife, 2010: Skilful multi-year predictions of Atlantic hurricane frequency, Nature, 846-849.

Valdes, P., and B. J. Hoskins, 1988: Baroclinic instability of the zonally averaged flow with bound- ary layer damping, J. Atm. Sci., 45, 1584–1593.

Wilson, C., B. Sinha, and R. G. Williams, 2009: The Effect of Ocean Dynamics and Orography on Atmospheric Storm Tracks, J. Clim., 22, 3689–3702.

Woollings, T., B. J. Hoskins, M. Blackburn, D. Hassel, and K. Hodges, 2009: Storm track sen- sitivity to sea surface temperature resolution in a regional atmosphere model, Clim. Dyn., DOI 10.1007/s00382-009-0554-3.

Woollings, T., J. Gregory, M. Reyers, and J. G. Pinto, 2011: Ocean-atmosphere interaction in the Atlantic storm track response to climate change, Geophys. Res. Let., in press.

Zhang, Y., and P. Stone, 2011: Baroclinic Adjustment in an Atmosphere–Ocean Thermally Coupled Model: The Role of the Boundary Layer Processes, J. Atm. Sci., 68, 2710–2730.