The ocean's role in polar climate change: asymmetric Arctic and Antarctic responses to greenhouse gas and ozone forcing

John Marshall, Kyle Armour, Jeffery Scott and Yavor Kostov (MIT)

David Ferreira, Theodore G. Shepherd (University of Reading)

Cecilia M. Bitz (University of Washington)

October 1, 2013

Abstract

In recent decades the Arctic has been warming with sea ice disappearing. But the Southern Ocean around Antarctica has been (mainly) cooling and sea ice extent growing. We argue here that interhemispheric asymmetries in the mean ocean circulation, with sinking in the northern North Atlantic and upwelling around Antarctica, strongly influences the Sea Surface Temperature (SST) response to Green House Gas (GHG) forcing, accelerating warming in the Arctic and delaying it in the Antarctic. Moreover, while the amplitude of GHG forcing has been similar at the poles, significant ozone depletion only occurs in the Antarctic. We argue that the response of SST around Antarctica to ozone depletion is initially one of cooling and only later adds to the GHG-induced warming trend as upwelling of warm water associated with stronger surface westerlies imprints itself on surface properties.

We organize our discussion around 'Climate Response Functions' (CRFs) i.e. the response of the climate to 'step' changes in anthropogenic forcing in which GHG and/or ozone hole forcing is abruptly turned on and the transient response of the climate revealed and studied. Convolutions of known or postulated GHG and ozone-hole forcing functions with their respective CRF's then yield the SST response, providing a context for discussion of the differing warming/cooling trends in the Arctic and Antarctic. It is suggested that the period through which we are now passing may be one in which the delayed warming of SST associated with GHG forcing around Antarctica is largely cancelled by the cooling effects associated with the ozone hole. By mid-century, however, ozone-hole effects may instead be adding to GHG warming around Antarctica but with diminished amplitude as the ozone hole heals. The Arctic, meanwhile, responding to GHG forcing but in a manner amplified by ocean heat transport, may continue to warm at an accelerating rate.

1 Introduction

Over the last few decades, the two polar regions of our planet have exhibited strikingly different behaviors. The Arctic has warmed, much more than the Northern Hemisphere as a whole, primarily in winter (ACIA 2004), while Arctic sea-ice extent has decreased dramatically (Comiso and Nishio 2008). In contrast, the eastern Antarctic and Antarctic plateau has cooled, primarily in summer, with warming over the Antarctic Peninsula and Patagonia (Thompson and Solomon 2002). Moreover, sea ice extent around Antarctica has modestly increased (Comiso and Nishio 2008). The largest observed changes in the Antarctic atmospheric circulation are associated with a strengthening and poleward shift of the summertime surface westerly jet (Fogt et al. 2009).

Observed and modeled surface temperature trends over 1979 to 2005 are shown in Fig.1. It is clear that there are very different observed Arctic and Antarctic temperature trends and each differs from the global trend. Moreover there is significant spread between models: the ensemble-mean tends to be biased warm in the Antarctic, and the model spread is particularly large in the Arctic. The annually averaged temperature in the Arctic has increased by over twice that of the global mean (a phenomenon known as Arctic amplification). Since 1979, the beginning of the reliable satellite record, Arctic summer sea-ice extent has decreased by order 12% per decade, with smaller reductions in winter. Coupled models suggest that under greenhouse-gas induced warming, the Arctic will warm the most: models generally exhibit enhanced warming and sea-ice loss in the Arctic in response to increasing greenhouse gases, but the observed changes over the last decade lie at the upper limit of the model projections (Stroeve et al. 2007; 2012). According to predictions of the Fourth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC, 2007), by the end of the 21st century the annual average temperature in the Arctic will increase by 2.8 to 7.8°C, with more warming in winter (4.3 to 11.4°C) than in summer. Decreases in sea-ice extent and thickness are projected to continue, and some models suggest that the Arctic Ocean will be free of sea ice in late summer by mid-century (see the discussion in Wang and Overland, 2009).

Many mechanisms are at work in 'Arctic amplification' (see, e.g., Serreze and Barry 2011, and references therein). A positive snow and sea-ice albedo feedback plays a significant role in amplifying the warming signal (Holland and Bitz 2003). The albedo feedback operates in summer when solar radiation is maximal. Where sea ice is lost and water is exposed, warming due to absorbed shortwave can be large and enhance sea ice loss through lateral melt (e.g., Perovich et al 2008). In addition to these processes, the warmed ocean mixed layer delays sea ice growth (Blanchard-Wrigglesworth et al 2011) and thus influences wintertime surface temperatures through a thinner ice pack. Because the Arctic atmosphere is stably stratified by thermal inversion at the surface, any warming that occurs there does not reach high up into the troposphere. Moreover, the surface energy balance is very sensitive to processes going on in the planetary boundary layer and cloud radiative processes (see, e.g. Hall, 2004). And, as is emphasized in our own work presented here, the climate of the polar caps is determined by more than regional and vertical energy balance, with lateral advection by ocean circulation playing a significant role.

The area poleward of the 70 °N latitude circle receives more energy due to atmospheric transport than from the Sun. Moreover, this lateral heat-flux convergence is largely balanced by outgoing infrared radiation, with surface fluxes contributing a relatively small amount to the energy budget (see Winton, 2008). The sensitivity of poleward heat transports to climate change is currently under debate (see, e.g. Hwang et al, 2011): polar amplification reduces meridional temperature gradients, which might be expected to reduce meridional atmospheric heat transport from lower latitudes, thus counteracting a portion of the amplification. Some studies argue that anomalous atmospheric heat transport, mainly due to increased moisture, have given rise to greater atmospheric warming above the surface (see Graverson et al, 2008; Thorne, 2008). However the validity of the analyzed atmospheric trends on which such studies are based is disputed [Grant et al, 2008; Bitz and Fu, 2008, Graverson (reply) 2008]. Beyond atmospheric heat transports, the high-latitude response to greenhouse forcing may involve anomalous ocean-heat transport into the Arctic; as we shall see this occurs even if a weakened meridional overturning circulation (MOC) diminishes the heat transport at lower latitudes (Holland and Bitz 2003; Bitz et al, 2008). In addition, the ocean can act as a reservoir for the heat gained in summer while the sea ice retreats, and this heat is possibly stored through the winter months (Blanchard-Wrigglesworth et al; 2011).

The mix of processes going on in the Antarctic is rather different than those of the Arctic. By far the largest anthropogenic perturbation of the atmospheric circulation over the last few decades has been the Antarctic ozone hole. Although the ozone hole is a stratospheric phenomenon, the polar cooling resulting from reduced solar heating has delayed the late-spring breakdown of the stratospheric polar vortex, which model studies show has been the principal driver of the observed strengthening and poleward shift of the summertime surface westerly jet (Arblaster and Meehl 2006; Fogt et al. 2009). In contrast to greenhouse-gas induced changes, which can be expected to continue, the Antarctic changes associated with the ozone hole are likely to reverse as it recovers through the remainder of this century.

In the Antarctic, models that include a representation of the ozone hole quantitatively recover the observed summertime trends in the surface westerly jet (Son et al. 2010), and reproduce the observed summertime warming of the Antarctic Peninsula and the cooling over East Antarctica (McLandress et al. 2011). However, coupled models consistently predict a decrease in sea-ice extent in response to ozone depletion (Lefebvre et al, 2004, Lefebvre and Goosse 2008, Sigmond and Fyfe 2010, Bitz and Polvani 2012, Smith et al 2012), which is at odds with the observed increase. Ferreira et al (2013) argue that there are two timescales involved: an initial cooling around Antarctica promoting sea-ice growth, and a longer-term warming which ultimately, perhaps after several decades, results in loss of sea-ice. On top of the anthropogenic changes are modes of polar climate variability (e.g. the Southern Annular Mode and the North Atlantic Oscillation) involving the atmosphere, ocean, land, and sea ice, which operate on decadal timescales (Rigor et al. 2002). Moreover, as we will discuss, anthropogenic forcing projects on to these natural modes of variability (see Fyfe et al, 1999).

As briefly reviewed in our introductory remarks above, many competing effects are at work in modulating the response of polar climates to anthropogenic forcing. The main goal of the present paper is to set out a framework for thinking about and quantifying the different responses of the Arctic and Antarctic. We organize our discussion around 'Climate Response Functions' (CRFs) i.e. the response of the climate to 'step' changes in anthropogenic forcing in which greenhouse gas (GHG) and/or ozone hole forcing is abruptly turned on and the transient response of the climate revealed and studied. We will use these to probe the role of the ocean in shaping the asymmetric response of polar climates to anthropogenic forcing.

In Section 2 we consider CRFs associated with GHG forcing and in Section 3 CRFs associated with ozone hole forcing. In Section 4 we convolve time histories and projections of GHG and ozone hole forcing with these CRFs to contrast the response of the high-latitude climate. Our calculations suggest that the ocean plays a central role in delaying warming around Antarctica relative to the Arctic in response to GHG forcing. Furthermore ozone hole forcing initially acts (for a few decades or so) as a cooling influence around Antarctica as the ozone hole grows and only later adds to the GHG-induced warming trend. In Section 5 we conclude.

2 Modulation of the surface response to GHG forcing by ocean circulation

2.1 Asymmetric response of Arctic and Antarctic surface climates to GHG forcing

Coupled climate models agree that under GHG forcing the Arctic warms more rapidly than the Antarctic. For example, Fig.2a shows the ensemble-average response of sea-surface temperature (SST) after 100 y in CO_2 quadrupling experiments computed from 17 general circulation models participating in the Coupled Model Intercomparison Projects phase 5 (CMIP5;

Taylor et al 2009). In such experiments coupled models were integrated out to (quasi) equilibrium forced with pre-industrial GHG concentrations. The CO_2 concentration was then abruptly quadrupled to study how the coupled climate evolved toward a new equilibrium. There is a rich spatial structure in the SST response after 100 y. In some regions of the globe SST increases by more than 4 °C whereas in others, particularly in the circumpolar band around Antarctica, SST increases by less than 1 °C. Marked hemispheric and polar asymmetries are evident with SSTs in the NH being generally considerably warmer than in the SH. Fig.3a documents the time evolution of SST as a function of latitude by zonally averaging over selected meridional bands.[1] These are our GHG CRFs. Delayed warming is evident in the Southern Ocean around Antarctica, Arctic amplification is clearly present with SSTs rising much more rapidly than around Antarctica.

What is the essential 'physics' behind the amplitude and timing of such differing polar responses to GHG forcing? Are they due simply to the interaction between GHG forcing and local radiative feedback processes (e.g., due to cloud changes) that perturb the energy budget in divergent ways over the two poles? Are they driven by different responses in atmospheric circulations and energy transports? Do they reflect different patterns of storage of anthropogenically-induced temperature signals in the deep ocean?

Here we suggest that, independent of the above mechanisms, the patterns and timing of warming evident in Figs. 1-3 can be largely explained in terms of the advection of anthropogenic temperature anomalies by the underlying ocean circulation.

2.2 Role of ocean circulation

To isolate the role of ocean circulation and expose the essential processes at play in setting the patterns in Fig.2a, we take away details of the atmospheric component of the coupled system by running an ocean-only model driven by an atmosphere that is represented in a

[1] In Fig.3 we choose the Arctic/Antarctic band between 50 ° and 70 ° to avoid persistently covered sea ice regions and to compare equivalent latitudes in both hemispheres. We also choose SSTs because here we focus on the role of the ocean.

highly parameterized, schematic way. Precise details of the methodology are described in Marshall et al (2013).

Briefly:

1. we spin up a global version of the MITgcm (Marshall et al 1997a,b) ocean model from Levitus/PHC for 300 years, using CORE normal year forcing (see Griffies et al, 2009) to compute, via bulk formulae, air-sea heat, freshwater and momentum fluxes. The run is continued for 10 additional years and all fluxes into the top of the ocean (including below the prognostic sea ice) are diagnosed daily; a 10-yr mean of 'daily forcing' and SST's is computed and stored as 'data'. The reference solution is then continued on, driven by a repeating annual cycle of this 'daily forcing' data.

2. the effect of warming due to GHG forcing is then parameterized by imposing a spatially uniform and constant in time surface downwelling longwave flux of $\mathcal{H} = 4\mathrm{W\,m^{-2}}$. The anomalous flux is only applied to the ice-free ocean which therefore warms so that its T is typically greater than that of the reference solution T_{ref}. We will call the difference 'anthropogenic temperature': $T_{anthro} = T - T_{ref}$. Note that the wind field and the freshwater fluxes are held constant and the role of sea-ice is considered passive, all of which are considerable simplifications. Likewise the anthropogenic sea surface temperature is given by $SST_{anthro} = SST - SST_{ref}$.

3. climate 'feedbacks' are parameterized by introducing a damping term, $-\lambda SST_{anthro}$ where λ, our 'radiative feedback parameter', is chosen to be spatially uniform and have a value of $1\mathrm{W\,m^{-2}\,K^{-1}}$.

One can question the simplicity and validity of these assumptions but, in the present context, we can turn them to our advantage. In particular, because $\mathcal{H}$ and λ are constant in space and time, any spatial patterns that emerge in the resulting temperature perturbations must be directly caused by, and hence attributable to, ocean circulation.

Fig.2b shows the SST perturbations after 100 y and can be directly compared to Fig,2a from the ensemble coupled climate models. Not only are gross patterns captured but also subtle details. Indeed, the similarity in spatial patterns is so striking, particularly at higher latitudes, that it tells us that they are largely a consequence of the underlying ocean circulation rather than (much more complex and uncertain) processes going on in the atmosphere under global change. Climate response functions from the ocean only model (not shown but discussed in Marshall et al, 2013) also have a very similar form to Fig.3a.

A glimpse at the interior structure of the anthropogenic temperature signal (T_{anthro}) is given in Fig.4(middle) where the zonal-average temperature perturbation is plotted. There is a clear interhemispheric asymmetry with T_{anthro} being much larger in the Arctic than in the Antarctic. The time integrated anomalous air-sea fluxes over 100 years (energy accumulation) is plotted in Fig.4(top) and reveals that most of the energy is fluxed in to the ocean around Antarctica due to the delayed warming there. However, it is not stored around Antarctica. Instead, as can be seen in Fig.4 (bottom), there is anomalous ocean heat transport northward away from Antarctica, keeping the waters around Antarctica cool. The reverse is true in the Arctic. We see that the ocean carries heat in to the Arctic (bottom) panel, increasing its temperature to such an extent that heat is actually lost to the atmosphere over the Arctic (top panel).

The advective process shaping the response is the upper cell of the ocean's meridional circulation with sinking in northern polar regions and upwelling in the southern ocean around Antarctica (see Marshall and Speer, 2012). This is the primary interhemispheric asymmetry of the global climate, a consequence of differing geometrical constraints on ocean circulation in the Arctic relative to the Antarctic: this promotes sinking of surface waters to depth in the northern North Atlantic and upwelling of waters around Antarctic.

As discussed in Marshall et al (2013), in the Southern Ocean T_{anthro} evolves very much like a passive tracer, 'injected' at the sea surface, weakly damped at the surface by climate feedbacks but governed by an advection-diffusion equation in the interior. Here, to a good approximation, T_{anthro} remains sufficiently ‘small’ that even after 100 years or so it does not significantly affect ocean currents. However, this is not true in the North Atlantic where changes in ocean currents (and the Atlantic Meridional Overturning Circulation — AMOC) contribute significantly to changes in ocean heat transport.

Before going on important caveats should be mentioned. Our ocean-only strategy permits sea-ice a role in establishing the mean stratification, but not in stratification changes. Moreover, freshwater surface fluxes are not allowed to change. Changes in precipitation, ice sheet/shelf runoff, and sea ice freshwater exchange may all play a significant under GHG forcing. Our calculations suggest that anthropogenic warming effects play the dominant role (compare Figs.2, top and bottom) but other effects must also be at play.

3 Response of the Antarctic to ozone hole forcing

3.1 Perturbation of Antarctic climate by the ozone hole

The dramatic depletion of the Antarctic ozone since the late 1970s has introduced a major perturbation to the radiative balance of the stratosphere with a wide range of consequences for climate. There is strong evidence that ozone loss has significantly altered the climate of the southern hemisphere troposphere, including the surface, with implications for ocean circulation, the cryosphere and coupled carbon cycle. Observations indicate a poleward shift of the southern hemisphere atmospheric circulation over the past few decades, predominantly in late spring and summer. Thompson and Solomon (2002) ascribed this shift to polar ozone depletion in the Antarctic lower stratosphere. The observed changes have the structural form of the Southern Annular Mode (SAM) in its positive phase: the surface wind maximum, the storm tracks, and the edge of the Hadley cell all shift poleward. While similar changes, with the same sign, have been reproduced in models under greenhouse warming scenarios (e.g., Fyfe et al., 1999; Kushner et al., 2001) they are also found in response to imposed ozone depletion (e.g., Gillett and Thompson, 2003; Shindell and Schmidt, 2004). In fact, on the basis of GCM studies in which both forcings were included, separately and together, McLandress et al. (2011) and Polvani et al. (2011) argued that thus far, ozone depletion has been the primary cause of the observed changes. In the future, assuming ozone depletion weakens as expected, the effects of greenhouse gas and ozone forcings will oppose each other.

Strengthening of SAM has been invoked to postulate enhancement in the strength of the upwelling branch of the MOC, and increases in the slope of density surfaces and eddy heat fluxes of the ACC (Hall and Visbeck, 2002; Hogg et al, 2008). Enhanced communication of the interior ocean with the surface could have marked effects on Earth's climate through changes in rates of heat and carbon sequestration as well as consequences for ice shelves around Antarctica which may be vulnerable to enhanced upwelling of warm water from depth (Martinson et al, 2008; Holland et al, 2010, Goldberg et al, a,b. 2012, Little et al, 2012). The stratification of the SO is also delicately poised and sensitive to changes in the freshwater balance, Gordon (1990), Wong et al (1999).

The links between the upwelling of deep water in the SO and the SH westerly winds and consequences for climate have been examined in observations and models (Russell et al, 2006). Although changes in the slope of density surfaces in the ACC cannot yet be detected (Boening et al 2008), ocean observations do indicate a freshening of Antarctic Intermediate Water (Wong et al, 1999; Durack and Wiffels, 2010) and a substantial warming of the SO equatorward of the ACC at all depths (Gille, 2008; Purkey and Johnson, 2010) which may be linked to atmospheric forcing (Fyfe et al, 2007). Modeling studies and theory, however, suggest that eddy transport in the ACC can readily compensate for changes in Ekman transport leading to little change in the strength of the MOC (Henning and Vallis, 2005; Hallberg and Gnanadesikan, 2006; Abernathey et al, 2011).

Changes in the SH westerlies (and SAM) have also been linked to changes in SSTs and sea-ice extent (SIE) around Antarctica, at least on interannual time scales (e.g., Hall and Visbeck 2002, Thompson and Solomon 2002, Lefebvre et al. 2004, Turner et al. 2009) where a positive SAM induces an overall cooling through the enhanced Ekman transport of cold surface waters northward. There is, however, debate about the cause of the observed decadal-scale trends, which show a small positive trend in total SIE but large regional trends of opposing sign. The expected sign of the sea-ice trend is still a key research question and almost certainly depends on timescale, as discussed in Ferreira et al, (2013). Simulations with coupled models, including one with an eddy-resolving ocean, all indicate that ultimately Antarctic ozone depletion causes a decrease in SIE (Sigmond and Fyfe 2010, Bitz and Polvani 2012, Smith et al 2012). Enhanced upwelling of warm water from depth around Antarctica associated with strengthening winds, likely result in sea-ice retreating on long timescales. It could also indicate that other factors, including natural variability, may be playing a role in the observed trends (Zunz et al 2012; Polvani and Smith 2013).

3.2 Response of the ocean to SAM forcing

Here we illustrate some of the ideas discussed above in an ocean-only context, in the same spirit as for the GHG response functions outlined in Section 2.

The direct effect of ozone hole forcing on the ocean's surface is essentially mechanical through its projection on to the surface winds associated with SAM (and thence, of course, air-sea heat and freshwater fluxes). This should be contrasted with GHG effects considered in Section 2 which are primarily felt through thermodynamic processes[2]. To further explore effects of anomalous winds we use the same ocean model described in Section 2 but now instead of perturbing it with a downwelling longwave flux mimicking GHG warming, we perturb it through an anomaly in the wind field around Antarctica mimicking ozone hole forcing. The procedure is as follows.

1. we take the model described in point 1. in Section 2 and perturb the forcing by introducing a SAM anomaly wind stress field calculated using the same bulk formula function as CORE. This fixed pattern is then multiplied by a global function that

[2]It is important to note that GHGs also project on to surface westerlies. Such effects are implicit in the CRF's shown in Fig.3 top.

varies between zero and one at one cycle per year peaking at the end of November. The resulting SAM anomaly field then added to the normal, daily forcing. This crudely represents the forcing of SAM by the ozone hole which peaks in the summertime. Note that only the wind stress is perturbed and here we do not try to represent the effect of wind anomalies on air-sea latent and sensible heat fluxes.

2. climate feedbacks are again parameterized as describe in point 3. in Section 2.

This simple procedure sidesteps complex (but nevertheless important) issues about how and to what extent ozone hole forcing projects on to SAM. Nevertheless when we interpret our results in Section 4 below, it will be implicitly assumed that trends in SAM over the past few decades are primarily due to ozone hole forcing, as argued by, for example, Polvani et al. (2011) and McLandress et al (2011).

The SST_{anthro} field after 1 and 50 years is shown in Fig.5. Initially we see a broadly axi-symmetric but dipolar SST anomaly pattern with cooling around Antarctica and warming further north. As noted above, this can readily be understood as the direct response of SST to anomalous advection by Ekman currents induced by (positive) SAM forcing. But over time, widespread subsurface warming (top few hundred meters) of the ocean appears which ultimately imprints itself on surface temperatures. This is revealed in the evolution of the SST index obtained by averaging between 50 and 70 °S and plotted as a function of time in Fig.5c. Initially we see a cooling and then a prolonged warming trend, much as plotted in Fig.3(bottom). Two timescales are at work, a 'fast' cooling period (several years) followed by a 'slow' warming trend (over decades), as discussed in detail in the coupled model results discussed in Ferreira et al (2013) and in Sigmond and Fyfe 2010, Bitz and Polvani 2012, Smith et al 2012.

The mechanism of the slow warming trend involves the response of the ocean to SAM forcing. As sketched in Fig.6, when the summertime SAM is in its positive phase, upwelling is induced around Antarctica with downwelling further north. In the region of upwelling there is a temperature inversion (the surface is colder than waters below), an inversion made possible by the melting/freezing and export of ice and resulting freshening of the surface waters. Thus upwelling in response to SAM brings warm water up toward the surface in the band of seasonal sea ice. In the region of downwelling to the north, away from the region of seasonal ice, warm water is brought down from the surface. Thus in response to a positive SAM forcing, we expect to see, and indeed observe in Fig.5, widespread warming of the ocean just below the mixed layer. Over time this warming signal becomes entrained in to the mixed layer leading to a warming of SST.

As discussed in Ferreira et al (2013), the warming trend is $T'_t = -w'\overline{T}_z$ where w' is the anomalous upwelling induced by SAM forcing acting on the mean stratification $\overline{T}_z$ and the overbar is an average over the seasonal cycle.

Key modeling uncertainties include:

- how the MOC responds to impulsive wind forcing as a function of timescale: Meredith and Hogg (2006), Screen et al. (2009). The Ekman response to a change in the wind is essentially instantaneous, but eddy contributions to the residual overturning circulation become increasingly important as time progresses. These processes are crudely parameterized in and/or resolved in models but it is not at all clear that they can adequately capture the heat budget of the mixed layer which involve the parameterization of both skew and residual fluxes.

- processes that set the near-surface stratification of the ocean in the region of seasonal sea-ice. The stratification under ice is typically delicately balanced with both T and S playing a role (see Gordon, 1991). This is very challenging to observe and capture in models.

- the spatial and temporal patterns of response, not just of (ozone-hole forcing $\longrightarrow$ surface winds) but also (surface winds $\longrightarrow$ SST and sea-ice cover). The ozone hole CRF function integrates over this detail, but that detail is central to setting the regional patterns of response — see Turner et al, 2009.

4 The combined effect of GHG and Ozone Hole forcing on polar climates

If one knows the Climate Response Functions and the respective GHG and ozone hole forcing functions, convolving one with the other yields the predicted response. We present such calculations here for plausible CRFs and forcing functions and contrast the evolution of SST over the Arctic relative to the Antarctic.

More precisely we may write:

$$SST(t) = \int_{0}^{t} CRF\,(t - t')\,\frac{\partial F}{\partial t}(t')dt', \tag{1}$$

where CRF is the response function and F is the prescribed forcing.

In Sections 2 and 3 we have discussed the contrasting forms of CRF's for GHG and ozone hole forcing for the Antarctic and Arctic. Typical examples are plotted in Fig.3. It is useful to express these as the sum of two exponential functions corresponding to a 'fast' and 'slow' process thus:

$$CRF = T_f\left(1 - e^{-\frac{t}{\tau_f}}\right) + T_s\left(1 - e^{-\frac{t}{\tau_s}}\right). \tag{2}$$

The coefficients T_f, T_s, τ_f and τ_s depend on whether GHG of ozone hole forcing is being considered, and whether we are in the Arctic or Antarctic, as set out in Table 1. The GHG coefficients are estimated from fitting Eq.(2) to the curves shown in Fig. 3 (top); the ozone hole coefficients are consistent with the ensemble-average spread of the CRFs reported in Ferreira et al (2013). It should be noted that there are considerable uncertainties in all of these parameters, particularly those associated with the ozone hole forcing. The family of curves indicated by the thin colored lines in Fig.3 are computed from Eq.(2) and the parameters set out in Table 1. Note that a critical difference between the ozone hole and GHG CRFs is that T_f^{Ozone} is negative whereas T_f^{GHG} is positive: ozone hole forcing promotes cooling of

SST around Antarctica on fast timescales, whereas GHG forcing promotes warming. Note T_s^{Ozone} is positive because on long timescales the effect of ozone hole forcing is a warming, as discussed in Section 3 and is clear in Fig.5.

Forcing	Region	T_f (K)	T_s (K)	τ_f(y)	τ_s(y)
GHG	Arctic	2.5 ± 0.7	5.5 ± 0.5	5 ± 1.5	520 ± 180
GHG	Antarctic	1.0 ± 0.6	7.0 ± 0.5	12 ± 3.0	900 ± 200
O_3	Antarctic	-0.44 ± 0.15	2.0 ± 0.3	3 ± 2.0	100 ± 30

Table 1. Timescales (in years) and amplitudes (in K) of GHG and ozone hole Climate Response Functions (CRFs), Eq.(2), in the region indicated. Families of CRF curves spanning the range of parameters tabulated are plotted in Fig.3.

As discussed in detail in Ferreira et al (2013), there is a large uncertainty in the processes that set the time-scale of the cross-over from cooling to warming evident in Fig.3(bottom). One of the difficulties is that, as of writing, the necessary coupled calculations for ozone-hole-like impulse forcing functions have yet to be carried out within the sufficiently large spectrum of CMIP5 models. In Table 1, therefore, and as plotted in Fig.3, we consider a rather wide range of parameters which imply zero-crossing timescales from several years to several decades.

Our assumed GHG and Ozone hole forcing functions are shown in Fig.7(left). Appropriately scaled, their convolutions with the CRFs in Fig.3 yield the SST time series plotted in Fig.7(right). The adjusted GHG forcing function is familiar and available from GISS (see Hansen et al, 2011, for a discussion). The total forcing trend is dominated by GHGs, but modified by volcanoes and anthropogenic aerosols. Note the downward spikes in the historical period due to volcanic activity. Projections in the future assume that the forcing increases smoothly to 4.5 $\mathrm{W\,m^{-2}}$ by 2100. We also assume that the same CRF's as calculated from abrupt CO_2 forcing apply to 20th and 21st century forcing. This is indeed a rough approximation, since forcings other than CO_2 (e.g., tropospheric aerosols, black carbon, volcanic aerosols) probably affect the Arctic differently than the Antarctic. Specifically, we are ignoring the 'efficacy' of individual climate forcings (Hansen et al 2005), and assuming they all drive a similar response as that to CO_2. To carry out the ozone hole convolutions we have scaled the ozone forcing between zero in 1920 (no ozone hole) and unity in 2000 (maximum of the ozone hole) with a steady recovery thereafter, so that by 2060 it is imagined to have completely healed. Moreover we assume a linear scaling between ozone hole forcing and SAM.

The SST timeseries shown in Fig.7(right) clearly reveals the differing responses to GHGs with the Arctic warming up more than twice as rapidly as the Antarctic: by 2050 the Arctic signal exceeds 1 °C compared to the Antarctic rise of 0.4 °C or so. The family of SST response curves to Antarctic ozone hole forcing results in a cooling of order 0.2 °C between 1980 and 2000 or so (but note the large spread between the individual curves). From roughly 2010 onwards, however, the ozone-induced response adds to the warming induced by GHGs. The sum of the GHG and ozone-hole responses delays the warming trend by perhaps 20 to 30 years. It is tempting to suggest that this is the period through which we are now passing. By mid century, however, ozone-hole effects are adding to GHG warming but its contribution diminishes in the latter half of the century as the ozone hole heals.

5 Conclusions

We have presented a framework in which to consider the asymmetric response of the Arctic and Antarctic to GHG and ozone hole forcing. The centerpiece of the framework are the respective GHG and ozone hole CRFs i.e. functions that quantify in a suitably integral sense the transient response of the climate to 'step' changes in anthropogenic forcing.

Green house gas CRFs are familiar to the community and have a long history in considering the global response to anthropogenic forcing. Here we have applied the approach regionally to contrast Arctic vs Antarctic responses. The central role of the ocean circulation in setting the SST response to GHG forcing is illustrated by comparing Fig.2(top) to

Fig.2(bottom). Clearly an ocean-only model can capture the broad spatial maps and timing of the response. Delayed (accelerated) warming in the Antarctic (Arctic) is a consequence of anomalous advection of heat out of (in to) the Antarctic (Arctic).

Ozone CRFs have only just recently been computed and in very few models. The first preliminary experiments with a highly idealized coupled model and a very sophisticated one, are described in Ferreira et al (2013). The two models yield quite different timescales for the onset of the slow processes. As of writing we are discussing the possibility of other coupled modeling groups carrying out similar CRF calculations for an Antarctic ozone hole in which an 'impulse' ozone hole forcing with a repeating seasonal cycle is used to perturb the coupled atmosphere, ocean, ice system. This would expose the elemental processes, patterns and timescales at work and the differences across models. Indeed perhaps the biggest uncertainty is in the response of the surface climate around Antarctica to ozone hole forcing. In the context of our framework this involves understanding and quantifying the form of the ozone hole CRF, Fig.3(bottom). Here we have considered a range of parameters (Table 1) and expressed the CRF in terms of a simple analytical expression, Eq.(2). Further study is required to understand what processes control its shape, whether it is well represented in models, and how we might constrain its form from observations.

Once the CRFs are quantified we can use them to consider what might, or might not happen, for plausible anthropogenic forcing functions, as in Fig.7. It is tempting to suggest that the current slight cooling of the climate around Antarctica might be a consequence of the cooling effects of the ozone hole which peaked around the turn of the century, offsetting the delayed warming tendencies of GHGs. But as the century proceeds GHG and ozone hole forcing are likely to both contribute to warming around Antarctica. However, as we have seen, such warming effects are mitigated by advection of heat by ocean circulation away from Antarctica. The opposite happens over the Arctic where warming is accelerated by ocean heat transport across the Arctic circle. Finally an important unresolved question is the extent to which natural variability confounds attempts to rationalize the problem. Perhaps nature is following one ensemble member of a plethora of other, equally plausible/possible trajectories.

6 Acknowledgements

JM and YK would like to acknowledge support from the NASA MAP program and the NSF FESD program. KA was supported by a James S. McDonnell Foundation Postdoctoral Fellowship.

.....

7 References

AICA: (2004) Impacts of a Warming Arctic: Arctic Climate Impact Assessment, Cambridge University Press (2004)

Abernathey, R., Marshall, J., and Ferreira, D. (2011) The Dependence of Southern Ocean Meridional Overturning on Wind Stress, J. Phys Oceanogr, vol. 41, no. 12, 2261–2278.

Arblaster JM, GA Meehl, (2006): Contributions of external forcing s to Southern Annular Mode changes, J Clim, 19, 2896-2095

Bitz, C.M. and Q. Fu, (2008): Arctic warming aloft is data set dependent. Nature, 455, 10.1038/nature07258

Bitz, C.M., (2008): Some aspects of uncertainty in predicting sea ice thinning, in Arctic Sea Ice Decline: observations, projections, mechanisms, and implications, AGU Geophysical Monograph Series, vol , edited by E. deWeaver, C. M. Bitz, and B. Tremblay, pp. 63-76 , American Geophysical Union.

Bitz, C.M., and L.M. Polvani, (2012): Antarctic climate response to stratospheric ozone depletion in a fine resolution ocean climate model, Geophys. Res. Lett., 39, L20705, doi:10.1029/2012GL053393

Blanchard-Wrigglesworth E, KC Armour, CM Bitz and E DeWeaver (2011) Persistence and inherent predictability of Arctic sea ice in a GCM ensemble and observations, Journal of Climate, 24, 231-250, doi: 10.1175/2010JCLI3775.1

Boening, C.W., Dispert, A. and Visbeck, M. (2008): The response of the Antarctic Circumpolar Current to recent climate change, *Nature Geo.*, 1, 864-869

Comiso, J.C., F. Nishio, (2008): Trends in the sea ice cover using enhanced and compatible AMSR-E, SSM/I, and SMMR data, J. Geophys. Res. 113, C02S07, 10.1029/2007JC004257

Durack, P.J. and S.E. Wijffels: (2010) Fifty-year trends in global ocean salinities and their relationship to broad-scale warming. *J. Clim*, **23**, 4342–4362

Ferreira, D, J. Marshall, S. Solomon, R. A. Plumb and C. Bitz et al (2013): Ocean and sea ice response to Ozone depletion: a two timescale problem. In preparation

Fogt, R. L., J. Perlwitz, A. J. Monaghan, D.H. Bromwich, J. M. Jones, and G. J. Marshall, (2009): Historical SAM Variability. Part II: 20th Century Variability and Trends from Reconstructions, Observations, and the IPCC AR4 Models. J. Climate, 22, 5346-5365

Fyfe J.C., Saenko O.A., Zickfeld K., Eby M, Weaver A.J. (2007): The role of poleward-intensifying winds on Southern Ocean warming, *J. Clim*, 20, 5391-5400

Fyfe J.C.„ G.J. Boer and G.M. Flato. (1999): The Arctic and Antarctic Oscillations and their Projected Changes Under Global Warming, Geo Res Lett, 26, No.11, 1601-1604

Gille, S. Decadal-Scale Temperature Trends in the Southern Hemisphere Ocean. (2008): *J. Clim*, **21**, 2749-2765.

Gillett, N.P. and D. W. J. Thompson, (2003): Simulation of Recent Southern Hemisphere Climate Change Science, 10, Vol. 302 no. 5643 pp. 273-275, DOI: 10.1126/science.1087440

Goldberg, D.N., C.M. Little, O. Sergienko, A. Gnanadesikan, R.W. Hallberg, and M. Oppenheimer (2012a) Simulation of ocean-land ice interactions through a strongly thermally-forced ice shelf, Part I: Model description and behavior, in press, J. Geophys. Res.-Earth Surface.

Goldberg, D.N., C.M. Little, O. Sergienko, A. Gnanadesikan, R.W. Hallberg, and M. Oppenheimer (2012b) Simulation of ocean-land ice interactions through a strongly thermally-forced ice shelf, Part 2: Sensitivity to external forcings, in press, J. Geophys. Res.-Earth Surface.

Grant, A.N., S. Brönnimann, & L. Haimberger, (2008): Recent Arctic warming vertical structure contested, Nature 455, E2-E3, doi:10.1038/nature07257

Graversen, R.G., T. Mauritsen, M. Tjernström, E. Källén, & G. Svensson (2008): "Graversen et al. reply, " Nature 455, E4-E5, doi:10.1038/nature07259

Graversen, R.G., T. Mauritsen, M., E. Källén & G. Svensson (2008): Vertical structure of recent Arctic warming, Nature 451, 53-56, doi:10.1038/nature06502

Gordon, A. (1991): The role of thermohaline circulation in global climate change, in *Lamont-Doherty Geological Observatory 1990 & 1991 Report*, Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York, 44-51

Griffies S. et al, (2009): Coordinated ocean-ice reference experiments (COREs), Ocean Modeling, 26,1-46

Hall, A., (2004): The role of surface albedo feedback in climate. J. Climate, 17, 1550–1568

Hall A, Visbeck M (2002): Synchronous Variability in the Southern Hemisphere Atmosphere, Sea Ice, and Ocean Resulting from the Annular Mode. J Clim. 15: 3043-3057.

Hallberg, R. W., and A. Gnanadesikan, (2006): The role of eddies in determining the structure and response of the wind-driven Southern Hemisphere overturning: Results from the modeling eddies in the Southern Ocean (MESO) project. Journal of Physical Oceanography, 36(12), 2232-2252.

Hansen, J., R. Ruedy, J. Glascoe, and M. Sato (1999, updated 2012): GISS analysis of surface temperature change. JGR Atmos., 104, 30997-31022.

Hansen et al (2005) Efficacy of climate forcings. JGR, 110, D18104.

Hansen J, Sato M, Kharecha P, avon Schuckmann K (2011) Earth's energy imbalance and implications. Atmos Chem Phys 11:1–29

Henning, C. and Vallis, G. K. (2005). The Effects of Mesoscale Eddies on the Stratification and Transport of an Ocean with a Circumpolar Channel. J. Phys. Oceanogr., 35, 880-896.

Hogg, A., Meredith, M.P., Blundell, J.R. and Wilson, C, (2008): Eddy heat flux in the Southern Ocean: response to variable wind forcing. *J. Clim*, 21, 4, 608-620

Holland, M.M., C.M. Bitz, (2003): "Polar amplification of climate change in coupled models," Climate Dynamics (2003) 21: 221–232

Holland, P.R., A. Jenkins, and D.M. Holland. (2010): Ice and Ocean Processes in the Bellingshausen Sea, Antarctica. *J. Geophys. Res.*, 115, C5, C05020

Hwang, Y., D. Frierson, J. Kay, "Coupling between Arctic feedbacks and changes in poleward energy transport." (2011), Geophysical Research Letters

Kushner, P. J., I. M. Held, and T. L. Delworth, (2001): Southern Hemisphere atmospheric circulation response to global warming. J. Climate, 14, 2238-2249.

Lefebvre, W., H. Goose, R. Timmermann and T. Fichefet, (2004): Influence of the Southern Annular Mode on the sea ice–ocean system, J. Geo. Res, VOL. 109, C09005

Lefebvre W. and H. Goosse, (2008). Analysis of the projected regional sea-ice changes in the Southern Ocean during the 21st century. Climate Dynamics 30, 59-76

Little, C.M., D.N. Goldberg, A. Gnanadesikan and M. Oppenheimer, (2012): On the coupled response to ice shelf basal melting, J. Glaciology, 208, 203-215.

Marshall, J and Adcroft, A and Hill, C and Perelman, L and Heisey, C. (1997a): A finite-volume, incompressible Navier Stokes model for studies of the ocean on parallel computers, JOURNAL OF GEOPHYSICAL RESEARCH-OCEANS, vol. 102, no. C3, pp. 5753-5766.

Marshall, J and Hill, C and Perelman, L and Adcroft, A (1997b): Hydrostatic, quasi-hydrostatic, and nonhydrostatic ocean modeling, JOURNAL OF GEOPHYSICAL RESEARCH-OCEANS, vol. 102, no. C3, pp. 5733-5752.

Marshall, J. and K. Speer (2012): Closure of the meridional overturning circulation through Southern Ocean upwelling, Nature Geoscience, vol. 5, no. 3, pp. 171-180, 2012

Marshall, J., J. Scott, K. Armour, J-M, Campin, M. Kelley and A. Romanou (2013): The ocean's role in the transient response of climate to abrupt Greenhouse Gas forcing. Submitted to Climate Dynamics

Martinson, D. G., S. E. Stammerjohn, R. A. Ianuzzi, R. C. Smith, and M. Vernet, (2008): Western Antarctic Peninsula physical oceanography and spatio-temporal variability, *Deep Sea Res.* II, 55, 1964–1987.

McLandress, C., Shepherd, T.G., Scinocca, J.F., Plummer, D.A., Sigmond, M., Jonsson, A.I. and Reader, M.C., (2011): Separating the dynamical effects of climate change and ozone depletion: Part 2. Southern Hemisphere Troposphere. J. Clim., 24, 1850–1868.

Meredith, M.P and A.M Hogg (2006) Circumpolar response of Southern Ocean eddy activity to a change in the Southern Annular Mode. Geo. Res. Lett, Vol. 33, L16608

Perovich, D.K., J. A. Richter-Menge, K. F. Jones, and B. Light, (2008): Sunlight, water, and ice: Extreme Arctic sea ice melt during the summer of 2007. Geo. Res. Lett, Vol. 35, L11501

Polvani, Lorenzo M., Darryn W. Waugh, Gustavo J. P. Correa, Seok-Woo Son, 2011: Stratospheric Ozone Depletion: The Main Driver of Twentieth-Century Atmospheric Circulation Changes in the Southern Hemisphere. J. Climate, 24, 795–812.

Polvani, L.M., and K. L. Smith (2013) Can natural variability explain observed Antarctic sea ice trends? New modeling evidence from CMIP5. GRL, 40, 3195-3199.

Purkey, S. G., G. C. Johnson, (2010): Warming of Global Abyssal and Deep Southern Ocean Waters between the 1990s and 2000s: Contributions to Global Heat and Sea Level Rise Budgets*. J. Climate, 23, 6336–6351

Rigor, I.G., J. M. Wallace, R. L. Colony, (2002): Response of Sea Ice to the Arctic Oscillation. J. Climate, 15, 2648–2663.

Russell, J.L., K.W. Dixon, A. Gnanadesikan, R.J. Stouffer, and J.R. Toggweiler (2006): The Southern Hemisphere Westerlies in a Warming World: Propping Open the Door to the Deep Ocean. *J. Clim.* 19, 24, 6382-6390

Screen, J., N. P. Gillett, D. P. Stevens, G. J. Marshall and N. K. Roscoe, (2009): The Role of Eddies in the Southern Ocean Temperature Response to the Southern Annular Mode. J. Climate, Vol.22, 806-818.

Serreze, M.C., and R.G. Barry, (2011): Processes and impacts of Arctic amplification: A research sythesis. Global and Planetary Change, 77, 85-96.

Shindell, D. T., and G. A. Schmidt, (2004): Southern Hemisphere climate response to ozone changes and greenhouse gas increases, Geophysical Research Letters Volume 31, Issue 18

Sigmond, M., J. C. Fyfe, (2010): Has the ozone hole contributed to increased Antarctic sea ice extent? Geophysical Research Letters, Volume 37, Issue 18

Smith, K.M., L.M. Polvani and D.R. Marsh (2012): Mitigation of 21st century Antarctic sea ice loss by stratospheric ozone recovery, Geophys. Res. Lett., 39, L20701

Son, S.-W., E. P. Gerber, J. Perlwitz, L. M. Polvani, N. P. Gillett, K.-H. Seo, V. Eyring, T. G. Shepherd, D. Waugh, H. Akiyoshi, J. Austin, A. Baumgaertner, S. Bekki, P. Braesicke, C. Brühl, N. Butchart, M. P. Chipperfield, D. Cugnet, M. Dameris, S. Dhomse, S. Frith, H. Garny, R. Garcia, S. C. Hardiman, P. Jöckel, J. F. Lamarque, E. Mancini, M. Marchand, M. Michou, T. Nakamura, O. Morgenstern, G. Pitari, D. A. Plummer, J. Pyle, E. Rozanov, J. F. Scinocca, K. Shibata, D. Smale, H. Teyssèdre, W. Tian, Y. Yamashita, (2010): "Impact of stratospheric ozone on Southern Hemisphere circulation change: A multimodel assessment," Journal of Geophysical Research: Atmospheres (1984–2012), Volume 115, Issue D3

Stroeve, J., M. M. Holland, W. Meier, T. Scambos, and M. Serreze, (2007): Arctic sea ice decline: Faster than forecast. Geophysical Research Letters 34, L09501

Stroeve, J. C., V. Kattsov, A. Barrett, M. Serreze, T. Pavlova, M. Holland, and W. N. Meier, (2012): Trends in Arctic sea ice extent from CMIP5, CMIP3 and observations. Geophys. Res. Lett. 39, L16502.

Taylor et al, (2012): An overview of CMIP5 and the experimental design. Bull. Amer. Meteor. Soc. 93, 485-498

Thompson, D.W.J. and Solomon, S., (2002): Interpretation of recent Southern Hemisphere climate change. Science 296: 895-899.

Thorne, P.W., (2008): Arctic tropospheric warming amplification? Nature 455, E1-E2, doi:10.1038/nature07256

Turner, J. et al., (2009): Non-annular atmospheric circulation change induced by stratospheric ozone depletion and its role in recent increase of Antarctic sea ice extent. G. Res. Lett., 36, L08502.

Wang, M.Y., and J.E. Overland, (2009): A sea ice free summer Arctic within 30 years? Geophys. Res. Lett. 36, L07502, 10.1029/2009GL037820 (2009)

Winton, M. (2008), Sea ice-albedo feedback and nonlinear Arctic climate change, in Arctic Sea Ice Decline: Observations, Projections, Mechanisms, and Implications, Geophys. Monogr. Ser., vol. 180, edited by E. T. DeWeaver, C. M. Bitz, and L.-B. Tremblay, pp. 111–131, AGU, Washington D. C.

Wong, A. P. S., N. L. Bindoff, J. A. Church, (1999): Large-scale freshening of intermediate waters in the Pacific and Indian oceans, *Nature*, 400, 440-443.

Zunz, V., H. Goosse, and F. Massonnet, (2012): How does internal variability influence the ability of CMIP5 models to reproduce the recent trend in Southern Ocean sea ice extent? The Cryosphere Discussions, 6 (5), 3539–3573

8 Figures

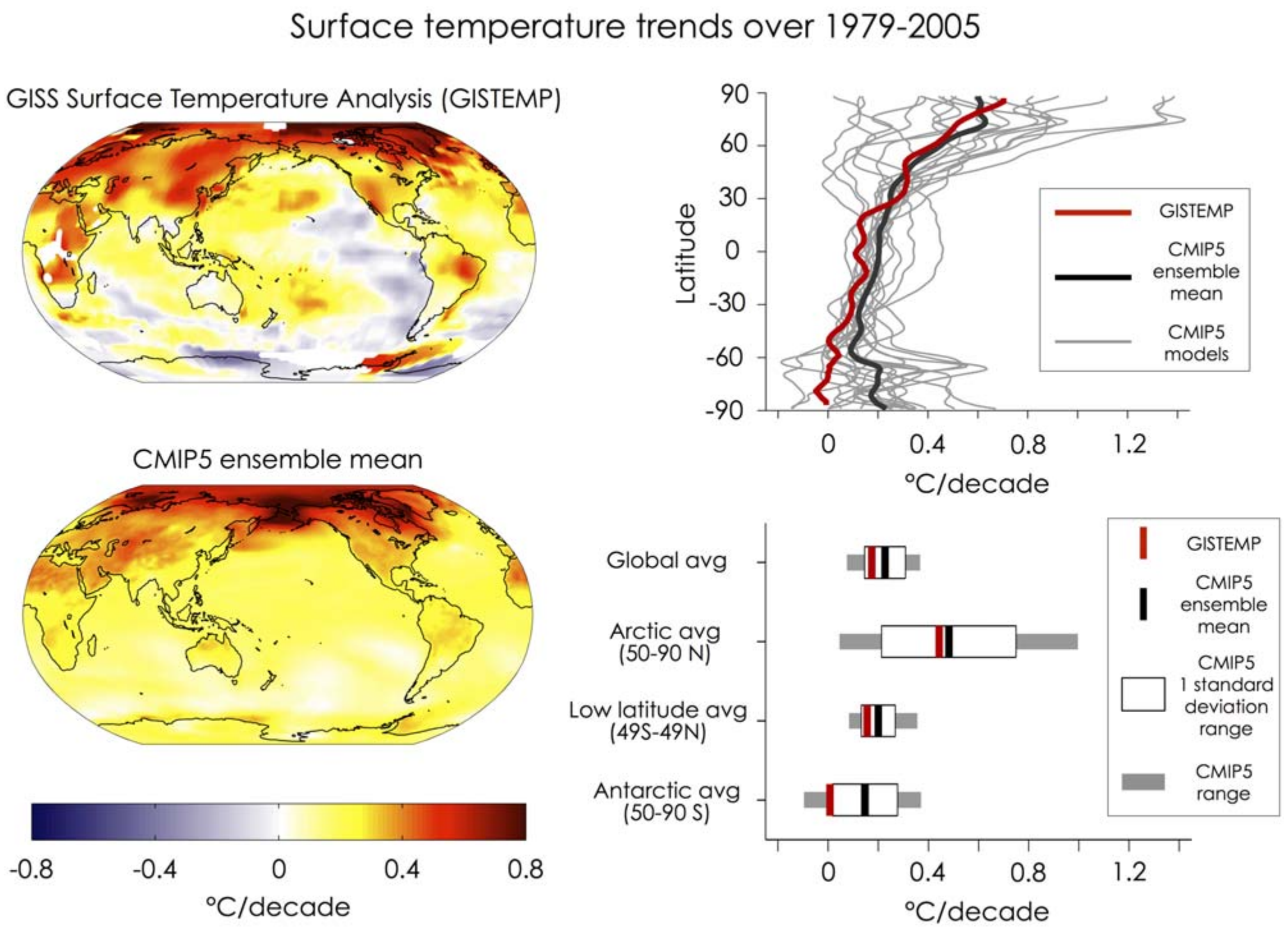

Figure 1: Surface temperature trends over 1979 to 2005 from (left, top) the GISS Surface Temperature Analysis (GISTEMP; Hansen et al 1999) (left, bottom) an ensemble of CMIP5 models. (right, top) Zonal-mean surface temperature trend from GISTEMP (red line), CMIP5 ensemble mean (black line), individual CMIP5 models (grey lines). (right, bottom) Surface temperature trends averaged over latitude bands (global, Arctic, low to mid latitudes and Antarctic), for GISTEMP (red line) and CMIP5 ensemble mean (black line); the white boxes show the CMIP5 ensemble $\pm$ one standard deviation range, and the gray boxes show the full CMIP5 ensemble range. All trends are expressed in $^{\circ}$C/decade.

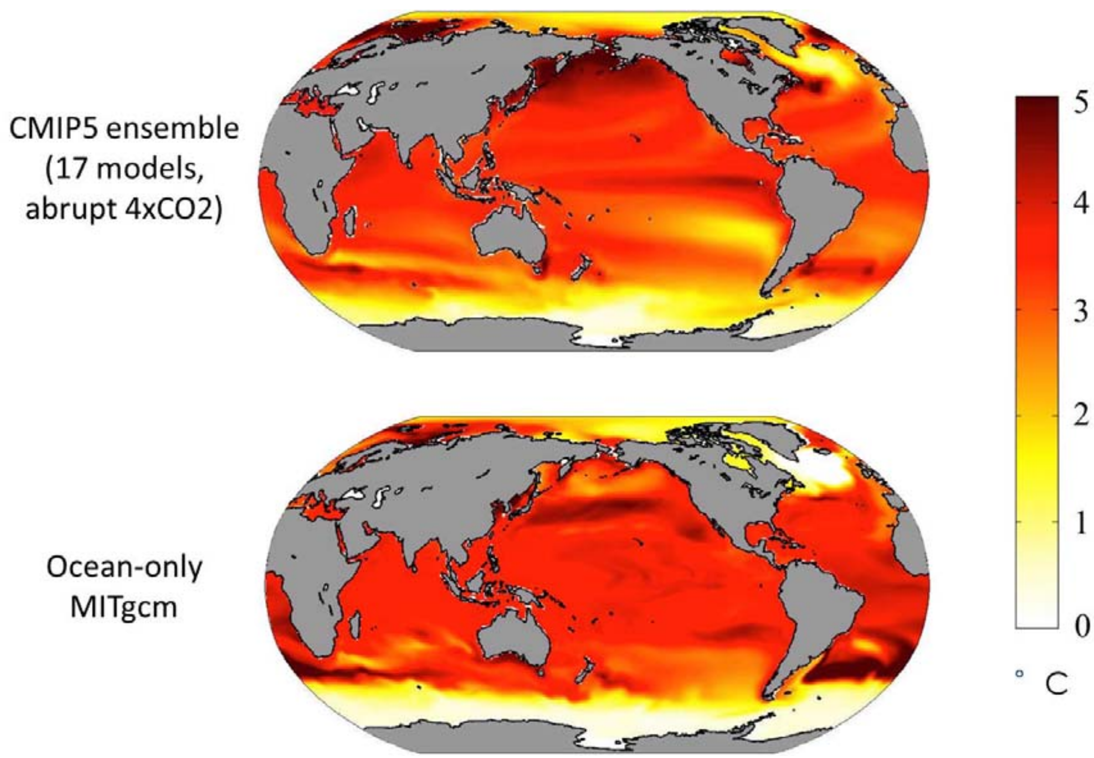

Figure 2: (top) Ensemble-average SST anomalies 100 years after abruptly quadrupling of GHGs in 17 CMIP5 models. (bottom) SST anomalies after 100 years of an ocean only configuration of the MITgcm induced by a uniform downwelling longwave flux of $4\mathrm{W\,m^{-2}}$ and damped by climate feedbacks at a rate of $1\mathrm{W\,m^{-2}\,K^{-1}}$, as described in Section 2.

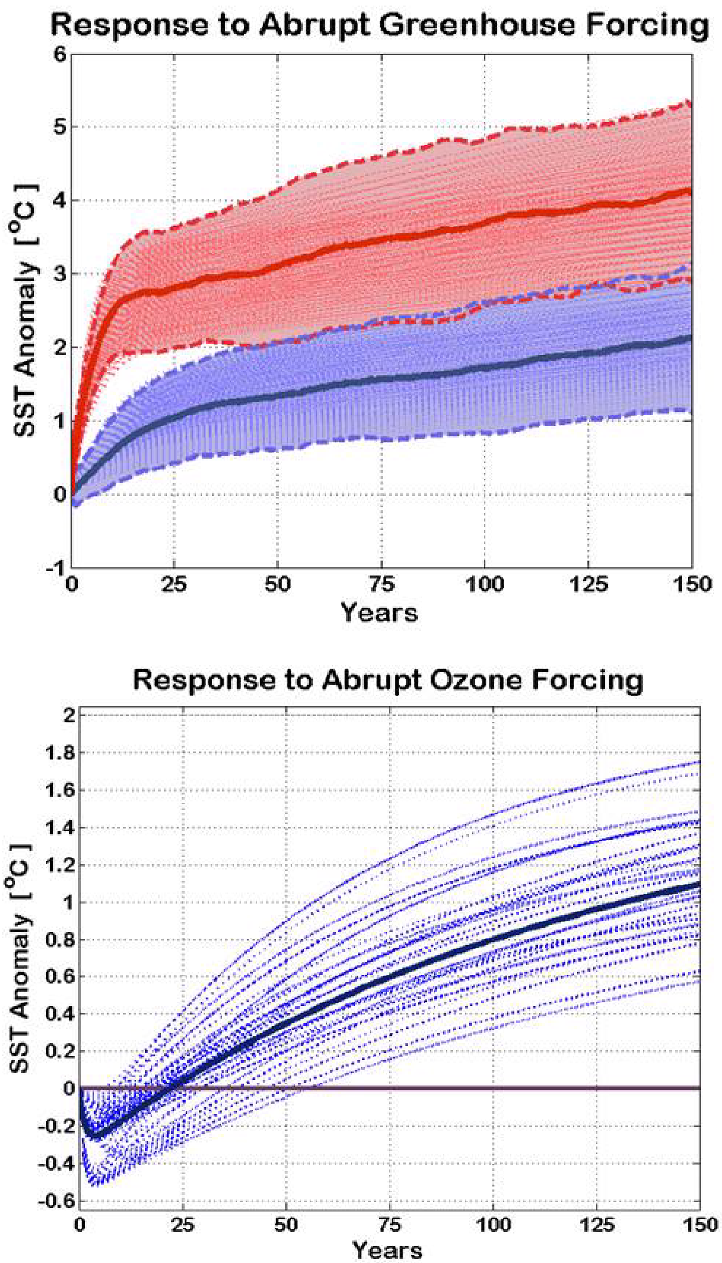

Figure 3: Sea surface temperature Climate Response Functions for (top) GHG forcing computed from an ensemble of CMIP5 models. CRF's are for SST anomalies averaged in the Arctic, between 50 and 70 °N (in Red; thick red line is the ensemble mean) and the Antarctic, between 50 and 70 °S (in Blue; thick blue line is the ensemble mean). (bottom) Ozone hole forcing based on the analytical expression Eq.(2). Thin lines in all figures are curves plotted from the analytical expression, Eq.(2) across the range of parameters given in Table 1.

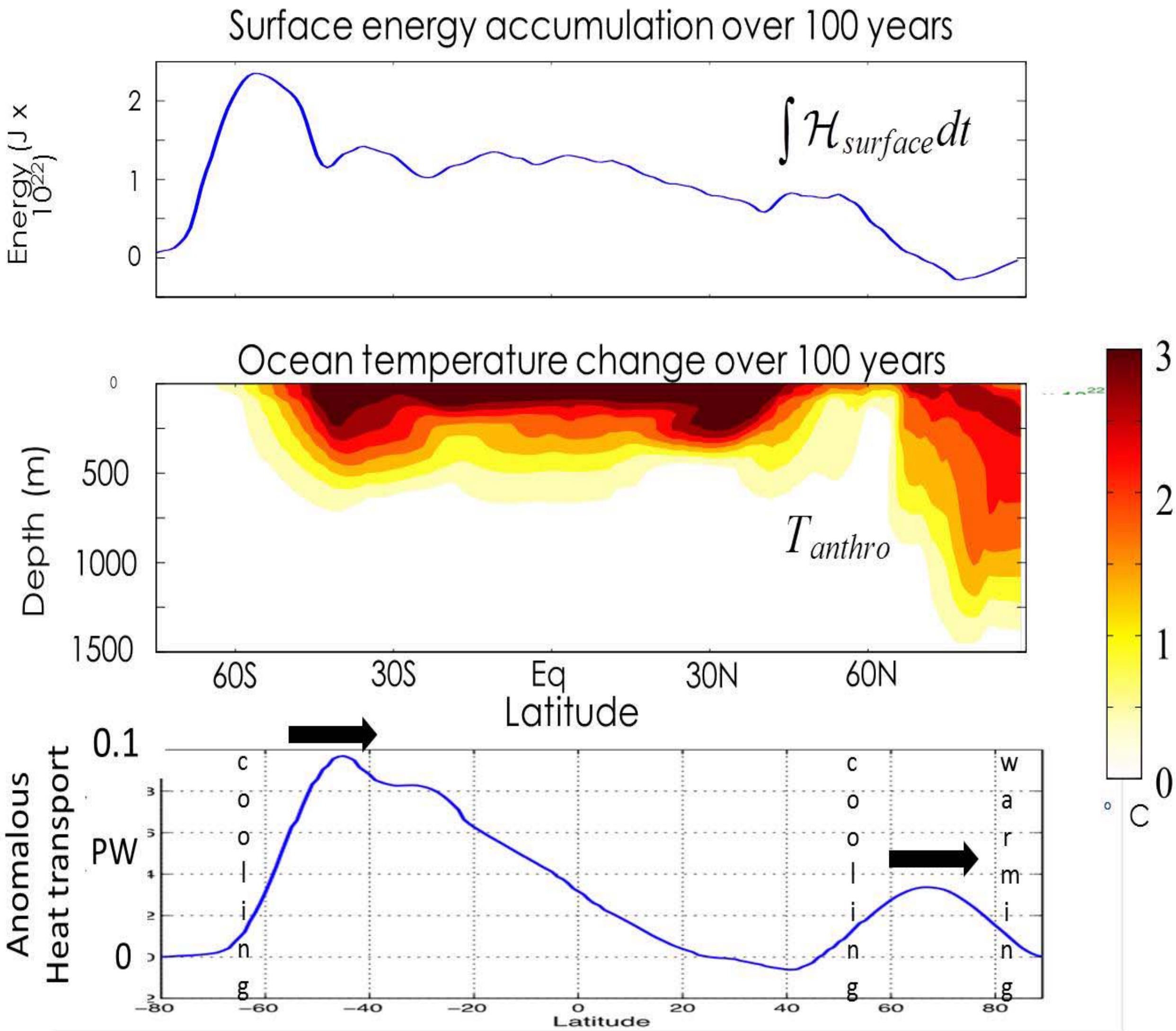

Figure 4: (top) Surface energy accumulation integrated over 100 years. (middle) Meridional section of zonal-average T_{anthro} after 100 years from the ocean-only configuration of MITgcm whose SST_{anthro} is shown in Fig.2 (bottom). (bottom) Anomaly in meridional ocean heat transport (in PW) after 100 years relative to a control integration. Latitudinal bands of implied warming and cooling are marked.

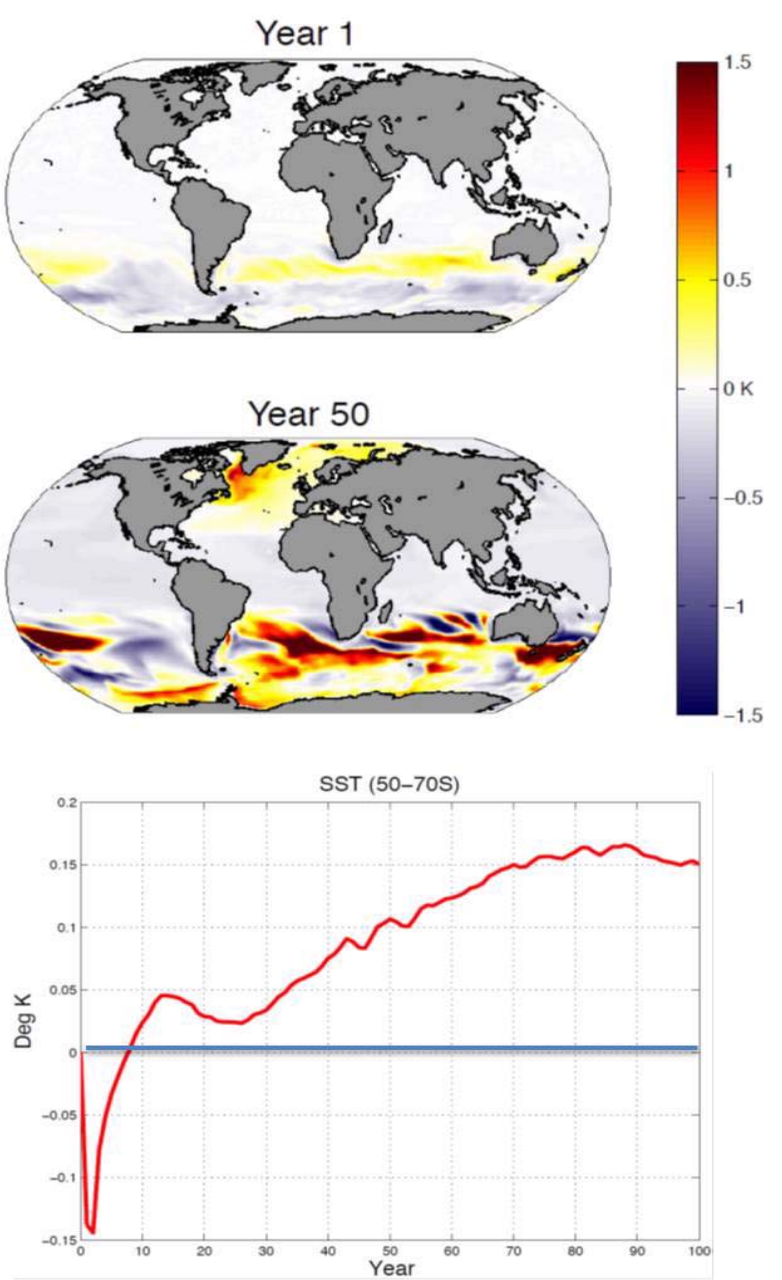

Figure 5: SST anomalies in °C after (top) 1 year and (middle) 50 years of an ocean only configuration of the MITgcm induced by anomalous SAM wind forcing around Antarctica, as described in Section 3. Red indicated warming and blue cooling. The temperature scale is on the right. (bottom). SST anomaly averaged between 50 and 70 °S from the SAM perturbation experiment, plotted as a function of time. This is analogous to the CRF for the Ozone Hole plotted in Fig.3 (bottom).

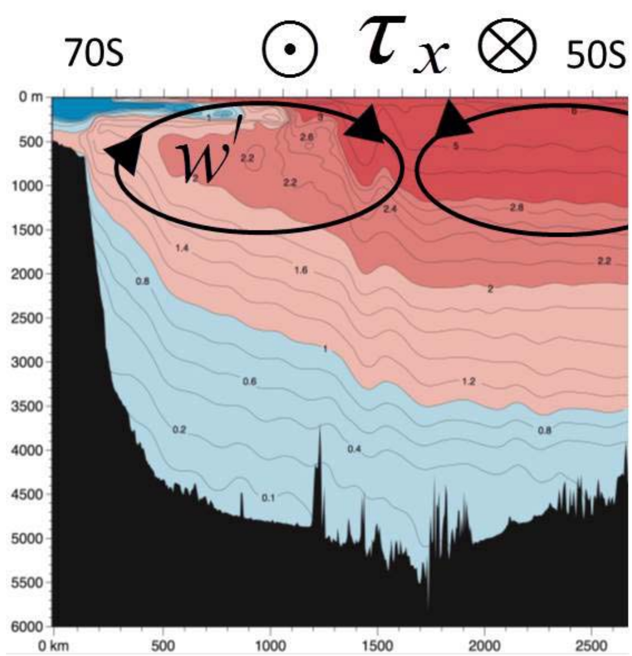

Figure 6: Meridional hydrographic section of temperature (WOCE section P19) stretching up to Antarctica on the left. The region of seasonal sea-ice is coincident with cold water (blue tongue) at the surface overlying warmer water (red) below. Superimposed is the sense of the meridional overturning circulation associated with a positive SAM anomaly, with upwelling around Antarctica and downwelling further equatorward. This acts to warm the ocean just beneath the surface layer.

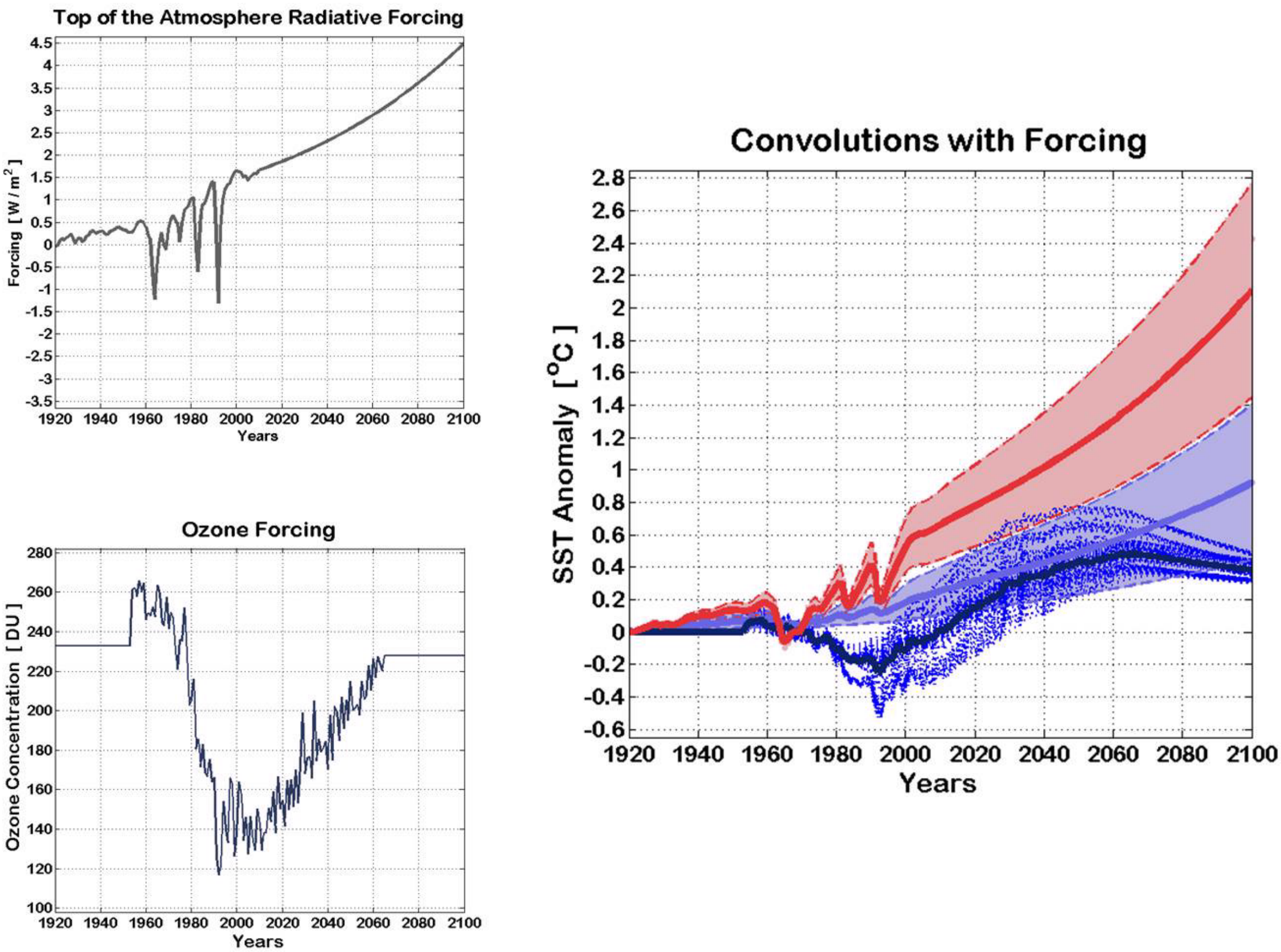

Figure 7: (left) Top: Historical net GHG forcing (in $\mathrm{W\,m^{-2}}$) from Hansen et al (2011) and projections in to the future assuming that the forcing increases smoothly to 4.5 $\mathrm{W\,m^{-2}}$ from 2010 to 2100, consistent with a standard RCP scenario. Bottom: Observed Ozone Concentration over the Antarctic and projections in to the future assuming the ozone hole heals at the same rate as it is observed to be doing now (courtesy of Diane Ivy, MIT). (right) Convolution of the GHG and Ozone Hole forcing plotted on the left, with the GHG and Ozone Hole Climate Response Functions plotted in Fig.3, to yield predictions and projections of SST anomalies between 50 and 70 °N (Arctic: red due to GHGs) and between 50 and 70 °S (Antarctic: light blue due to GHGs and dark blue due to Ozone Hole forcing).