Changes in Variability Associated with Climate Change

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Abstract In this paper, we briefly discuss changes in large-scale oscillations such as the El Nino/Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), and the northern and southern annular modes (NAM and SAM), changes in the polar and tropical troposphere, and interactions between the stratosphere and troposphere in a changing climate. We consider both changes in variability as well as trends in the

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mean state. We conclude, that to fully understand how modes of variability will change in a changing climate, we need additional analysis of observations, both paleo and present day, and a solid fundamental understanding of mechanisms. Understanding of mechanisms necessarily requires use of models, ranging from simple to complex. Such models need to be fully coupled, between atmosphere and ocean, and need to include a fully resolved middle atmosphere as well.

Keywords Climate variability • Climate change • Annular modes • El Niño Southern Oscillation • Sea ice • Greenhouse gases • Ozone • Stratosphere

1 Introduction

Climate change involves changing statistical properties in the climate system over an extended period of time. Such changes may be induced through long-term changes in solar or orbital parameters, long periods of enhanced volcanic activity or through long-term changes in radiatively significant gases. Whatever the actual forcing is, the end result will likely be long-term changes in the mean state or in the variability of the system or both. There are multiple ways to assess whether such changes may be occurring. Extended model simulations, where appropriate forcing parameters are varied are one means of assessing changes in the mean state or variability of the climate system. These simulations can then be used to provide estimates of the climate response to changing forcings as well as assess the internal variability, both of which are needed for detection and attribution studies. Past changes in climate variability can be addressed via analysis of historical data, using both recent measurements as well and geologic records or ice core records. As we are currently in the midst of a large scale climate change experiment, with changes in radiative gases and surface conditions induced by anthropogenic activity, analysis of existing climate data over the industrial era is another means of assessing impacts on variability due to changes in forcings of the climate system. We are interested in changes in the mean circulation and variability of that circulation ultimately because it impacts surface temperature and precipitation.

In this paper, we briefly cover an extremely broad topic: "How climate change impacts climate variability" with focus on the identification and mechanisms for modes and regimes of large-scale variability in different climates. We are basing the content of this paper largely on work discussed at the WCRP Open Science Conference held in Denver in 2011. Although there are many modes of variability that can be addressed in a review paper such as this, we will concentrate our efforts on just a few topics. In particular, we will briefly discuss changes in large-scale oscillations such as the El Nino/Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), and the northern and southern annular modes (NAM and SAM), changes in the polar and tropical troposphere, and interactions between the stratosphere and troposphere in a changing climate. We consider both changes in variability as well as trends in the mean state in our discussion. We will then make recommendations as to key issues that require additional research.

2 How Do Changes in Greenhouse Gases and Solar and Orbital Parameters Impact the Tropics?

2.1 ENSO

Decades of observational, theoretical, and numerical modeling research has shown that El Nino/Southern Oscillation (ENSO) is the result of dynamical coupling between the ocean and atmosphere which results in growth of perturbations to the tropical Pacific climate on seasonal to interannual timescales, generally referred to as the 'Bjerknes feedback.' However, while the fundamental mechanism of ENSO is fairly well understood, there are still some important open questions, particularly with regards to how ENSO will change in the future in response to anthropogenic forcing.

One way to address this is to look for detectable trends in the behavior of ENSO over the twentieth century that might be attributable to external forcing. For example, in the 1990s, it was argued that the persistent, weak El Nino that occurred from 1990 to 1995 was highly unlikely given the character of the record prior to that time (Trenberth and Hoar 1996, 1997), and that this may be an indication of 'El Nino and Climate Change' (though no formal attribution statement was made in those studies). What we have learned since then is that the details of ENSO continue to present new puzzles with practically every realization of the phenomenon (Stevenson et al. 2012). For example, there seems to be an increasing number of 'central Pacific' events, which, in contrast to the classic eastern Pacific event, have their maximum temperature response confined to the central basin (Yeh et al. 2009; Newman et al. 2011). There also appears to be variations in the predictability of ENSO, which depend on the mean state (e.g. Kirtman and Schopf 1998). In short, the 'natural' behavior of ENSO is so varied that detecting anthropogenic trends in likely to take an extremely long record (Wittenberg 2009).

One might look back further using paleoclimate records, and then the basic story is actually fairly straightforward: A simple, first-order answer which is supported both by paleoclimate records and by climate models is that the Pacific is characterized by large seasonal and interannual variability with global impacts no matter what the state of the mean climate is. Seasonally-resolved tropical Pacific paleoclimate records from periods in the Earth's history that were both warmer and colder than today show that ENSO-like interannual variability was present. Available Pliocene records, for example, when the Earth was several degrees warmer than present and ice sheets were minimal in extent, show that ENSO frequency and amplitude were not significantly different from today (Watanabe et al. 2011; Scroxton et al. 2011). The same goes for the glacial climates: Koutavas and Joanidis (2009) have shown using isotope measurements on individual forams that there is large variability at the Last Glacial Maximum, and coral records from prior glacial stages also suggest considerable interannual variability (Tudhope et al. 2001). Climate models have thus far not been able to rid the tropical Pacific of ENSO variability by either warming (Huber and Caballero 2003; Galeotti et al. 2010; von der Heydt et al. 2011) or cooling the climate (Zheng et al. 2008). Nor does there appear to be an obvious relationship between radiative forcing and ENSO behavior over the last millennium, when solar and volcanic forcing as well as the mean climate state all have varied (Cobb et al. 2003; Emile-Geay et al. 2012).

The only external climate forcing that, thus far, has been shown to influence ENSO in a systematic way that is consistent in observations and models is precessional forcing. It appears that when perihelion occurs in Northern Hemisphere summer (every 21 kyr), ENSO variance is reduced. The mechanism varies from model to model, but fundamentally it is the altered annual cycle of the large-scale atmosphere–ocean circulation that appears to influence ENSO. Models underestimate the influence of this effect compared with observations (c.f. Brown et al. 2008).

Looking forward using climate models, there is also considerable uncertainty. First, there are large biases in climate model simulations of the mean tropical Pacific climate, which may impact their ability to simulate ENSO (Roberts and Battisti 2011). That said, if models are run into the future with greenhouse gas forcing, they robustly simulate an enhanced equatorial warming (IPCC 2007; Liu et al. 2005), but this should not be thought of as a change in ENSO. Rather, models do not simulate a consistent response in ENSO, and the changes are generally small even with the large $4\times CO_2$ forcing (Fig. 1, from Guilyardi et al. 2012). An analysis of CMIP3 models by DiNezio et al. (2012) suggests an explanation: They argue that ENSO is fairly insensitive to greenhouse gas forcing because the winds and thermocline actually have opposing effects on ENSO. As the climate warms, the Walker circulation weakens, which on its own would weaken ENSO variability. However, the ocean response to the weaker trade winds is a less tilted, but *shallower* thermocline, and this effect would strengthen ENSO variability. DiNezio et al. (2012) argue that these competing mechanisms can explain why climate models simulate overall little change in ENSO in response to greenhouse gas forcing, and the same arguments can be made for interpreting past climate changes as well.

These prior studies point to some important, outstanding questions about the response of ENSO to external forcing:

1. Can we develop a more complete characterization of the ENSO system with paleoclimate proxies including the internal variability as well as the radiatively-forced changes?
2. What are the mechanisms by which changes in the mean state can influence interannual variability of ENSO? Can models represent these mechanisms, and can observational networks support their characterization in the real system?
3. What are the mechanisms that contribute to the different ‘flavors’ of ENSO, and how may these be altered by a changing mean state?
4. What are the prospects for improving the predictability of ENSO on seasonal, interannual and even decadal timescales? Is predictability altered by an externally forced change in mean state?

2.2 Width of the Tropics

A geographic definition of the “tropical belt” is the region between the Tropics of Cancer and Capricorn, which are currently at 23º27′S and 23º27′N, respectively. This geographic definition of the tropical edge latitudes is a consequence of the

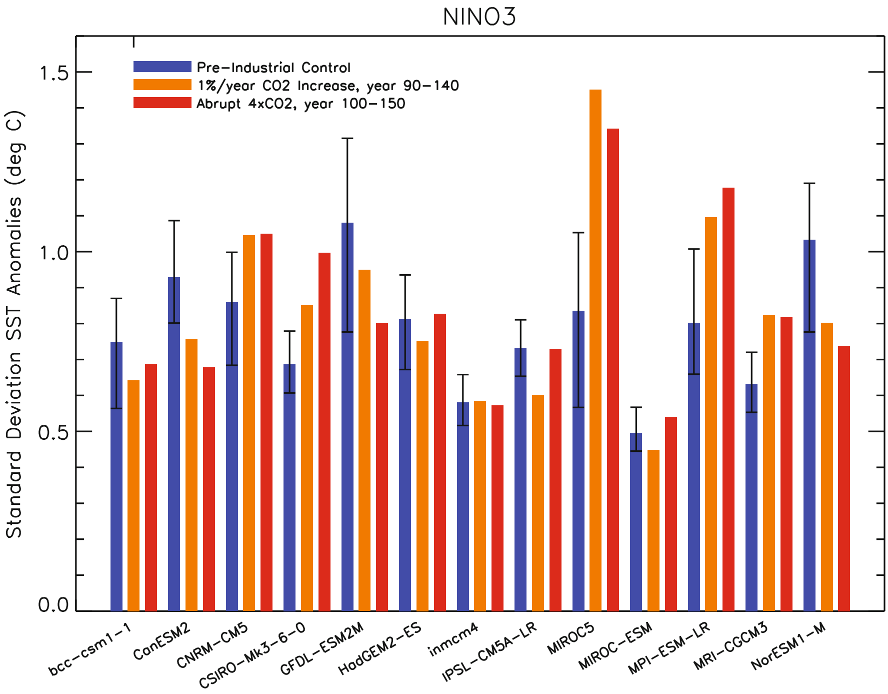

Fig. 1 Standard deviation of Niño3 SST anomalies for CMIP5 model experiments. *Blue bars*, preindustrial control experiments, *orange bars*, years 90–140 from the 1 %/year CO_2 increase experiments, *red bars* years 50–150 from the abrupt $4\times CO_2$. Calculations are performed for the models indicated on the x axis. The *black* '*error bar*' indicates the minimum and maximum of 50 year windowed standard deviation of Niño3 anomalies computed from the multi-century control experiments. Thus, when the Niño3 standard deviation in one of the CO_2 runs falls *below* or *above* the error bar, the changes are deemed to be significant. If significant changes are seen in both experiments that indicates a more robust response in that model (After Guilyardi et al. 2012)

axial tilt of the earth, which varies with a periodicity of approximately 41,000 years. From an atmospheric perspective, there is no similarly simple definition of the latitudinal extent of the tropical belt, but most definitions refer approximately to the region equatorward of the Hadley cell edges. These atmospheric tropical edge latitudes have been quantitatively diagnosed from observations and models by identifying arbitrary thresholds or local extremes in meteorological properties (e.g., winds, tropopause height) as they change with latitude from their tropical to extratropical values. Figure 2 shows examples of tropical edge diagnostics considered in the literature. From a surface climate perspective, the tropical edges are significant because they are essentially the poleward boundaries of the subtropical deserts. Potential changes in the latitudinal extent of the tropics are thus related to changes in precipitation patterns, and could lead to significant regional impacts in the ecosystem health, water resources, and agriculture.

Additionally, there are more subtle implications regarding transport of mass and species into the stratosphere. The tropical upper troposphere is the primary location

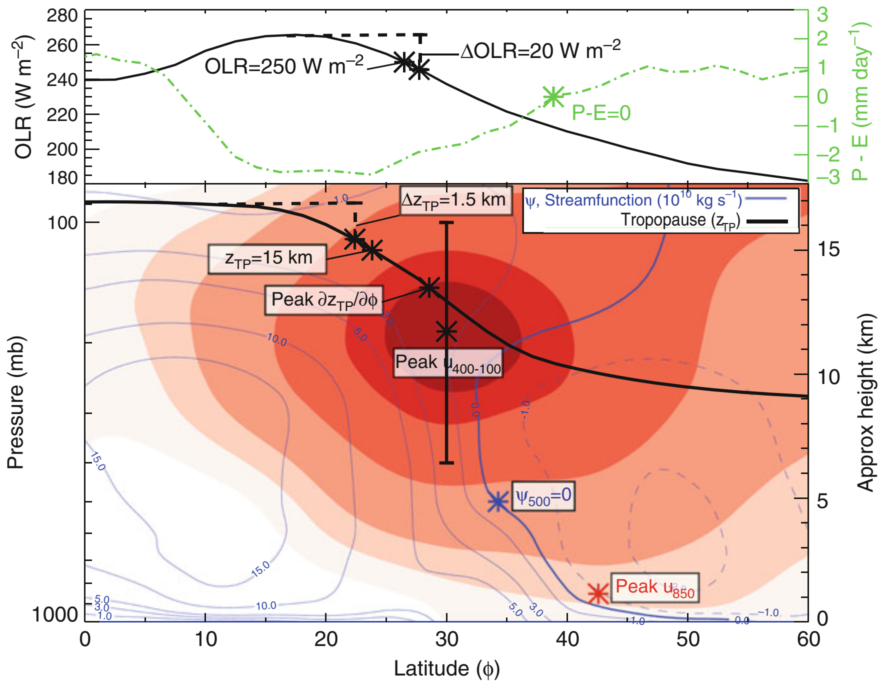

Fig. 2 An example monthly-mean zonal-mean cross section of atmospheric properties from the NCEP/NCAR reanalysis and NCEP OLR dataset for January 2008. The zonal-mean zonal wind is *shaded* (5 m s^{-1} contours), and the mean meridional streamfunction (ψ) based on the zonal-mean meridional wind is contoured in *blue*. Tropical edge diagnostics are denoted by *asterisks*. *Dashed vertical bars* illustrate the relative threshold metrics, and the solid vertical bar illustrates the vertical range of averaging for the peak wind metric (From Davis and Rosenlof 2012)

through which air enters the stratosphere; it is the gateway for stratospheric entry of tropospheric trace gasses, some of which are potentially ozone depleting or radiatively active. If the tropical upwelling region widens, it can conceivably alter the amount of such species entering the stratosphere, especially if the upwelling region encompasses a greater concentration of populated regions emitting anthropogenic pollutants. Hence, how the tropical belt responds to natural and anthropogenic forcings is not only significant for regional surface climate in the subtropics, it also has a potentially global-scale impact via changes in stratospheric composition and radiative forcing.

Multiple independent analyses using chemical constituent measurements, meteorological observations, and meteorological fields from reanalyses have identified changes in the latitudinal extent and character of the tropical belt during the past 40–50 years. Specifically, studies have noted an intensification and poleward expansion of the Hadley cell as defined by OLR and the meridional mass streamfunction (Hu and Fu 2007; Johanson and Fu 2009; Mitas and Clement 2005), the region of high-altitude tropical tropopause (Lu et al. 2009; Seidel and Randel 2007; Seidel et al. 2008), and the region of “tropical”-like low column ozone (Hudson et al. 2006).

Other studies based on reanalyses have suggested changes in both the strength and position of the subtropical and polar jet streams (Archer and Caldeira 2008; Strong and Davis 2007), and a poleward shift in storm tracks (Fyfe 2003; McCabe et al. 2001). A lack of trend has also been noted (Swart and Fyfe 2012). Although changes in the eddy-driven jets are discussed in Sect. 3.1, and are not strictly related to the other tropical edge diagnostics discussed here, it is worth noting that the jet latitudes and Hadley cell edge latitudes are correlated during Austral summer in the SH (Kang and Polvani 2011).

Tropical widening and poleward expansion of the jets has been detected in climate model simulations with anthropogenic forcings, and pre-industrial control runs indicate that the magnitude of the late-twentieth century widening cannot be explained by natural variability alone (Johanson and Fu 2009; Lu et al. 2009; Yin 2005). However, the rate of widening is greater in observations than in models for the few diagnostics that have been tested (Johanson and Fu 2009). For example, the late-twentieth century poleward expansion rates from several Hadley cell diagnostics span a range of ~ 0.6–1.8° decade^{-1}, whereas comparable model estimates are 0.1–0.2° decade^{-1} (Hu and Fu 2007; Johanson and Fu 2009).

A better understanding of the dynamical mechanisms for tropical belt expansion is very important for assessing the relative importance of ozone depletion and anthropogenic greenhouse gas (GHG) forcing of tropical widening, and may help in reconciling the discrepancy between observations and models, thus allowing for better predictions of future widening. Several dynamical mechanisms have been proposed for explaining the poleward expansion of the tropics and jets, and in general these mechanisms involve interactions between the atmospheric thermal structure/gradients, winds, and wave breaking. More specifically, tropical belt changes have been proposed to occur due to polar stratospheric cooling (Polvani and Kushner 2002; Polvani et al. 2011a, b; Tandon et al. 2011), increases in upper tropospheric static stability associated with global warming (Frierson et al. 2007; Lu et al. 2007), warming in the tropical Indo-Pacific ocean (Johanson and Fu 2009; Lau et al. 2008), and changes in wave propagation and breaking associated with changes in the near-tropopause meridional temperature gradient (Chen and Held 2007; Lorenz and Deweaver 2007; Simpson et al. 2009).

A comparison of a variety of estimates of changes in the width of the tropics over the past 30 years is given in Davis and Rosenlof (2012). This study demonstrated that there is a large spread among tropical width trends calculated from different edge definitions, as well as from different reanalyses. The study also shows that the use of objective definitions gives trends that are smaller than previous subjective estimates (i.e., $z_{TP} = 15$ km and OLR = 250 W m^{-2} in Fig. 2). For one metric (the Hadley cell streamfunction, ψ_{500}), the reanalysis trends are large (> 1° decade^{-1}), statistically significant, and significantly larger than those derived from climate models. Other than the Hadley cell metric, reanalysis trends over the past 30 years from objective definitions are mostly positive but not statistically significant.

To date, much work has focused on trends, and relatively little work has been done comparing the co-variability of various metrics on seasonal, interannual, hemispheric, and longitudinally-resolved scales. Such comparisons, for models, observations, and between models and observations, would help give a clearer picture of what aspects of tropical widening are robust. Clearly, additional studies are needed to ascertain the reasons for the differences in model- and observation-based trends, and mechanisms for the changes in tropical width need to be further explored. Key questions that need answering include:

1. Are historical (i.e., satellite- and reanalysis-based) trends in tropical width accurate? How well can the observational- and model-based trends of the past several decades be reconciled?
2. What are the predominant drivers of historical and future tropical width trends in models (e.g., natural variability, greenhouse gasses, stratospheric ozone depletion)?
3. What are the dynamical mechanisms by which these drivers affect the tropical width? To what extent can trend variations (e.g., as a function of season, hemisphere, definition) be used to test these proposed mechanisms?
4. How do tropical width trends relate to changes in other modes of climate variability?
5. Are there feedback processes operating whereby tropical width changes impact stratospheric composition, leading to a radiative impact on surface temperatures or further tropical width changes?

3 How Does Climate Change Impact Middle and High Latitudes?

3.1 The Northern Annular Mode and Related Latitudinal Shifts of the Eddy-Driven Jet

The Northern Annular Mode (NAM), also called the Arctic Oscillation, is the main atmospheric mode of variability in the northern extratropics (Thompson and Wallace 2000). It is usually defined as the first empirical orthogonal function (EOF) of Northern Hemisphere (20°–90°N) winter sea level pressure but other definitions exist. While the NAM structure is very similar to the North Atlantic Oscillation (NAO) pattern in the Atlantic, it exhibits stronger anomalies over the North Pacific, leading to a more zonally symmetric structure. It has been argued that the two patterns, NAM and NAO, may in fact represent two different paradigms of the Northern Hemisphere variability: the sectoral paradigm (NAO) and the annular one (NAM) (Deser 2000; Ambaum and Hoskins 2002). While the debate of which of them is more appropriate remains unresolved, here we take a simple and pragmatic viewpoint: it likely depends on the asked scientific question and context. Consequently, we will be alternatively using both paradigms. In the Atlantic, the NAM/NAO is also strongly related to latitudinal displacements of the eddy-driven jet although other modes of variability (such as the East Atlantic pattern) are also needed to fully account for the jet variability (Woollings and Blackburn 2012). NAM/NAO positive phases are characterized by a strong subpolar jet (the eddy-driven jet) that is well separated from the subtropical jet. During negative phase periods, in contrast, the two jets merge and lead to a more zonal circulation across the Atlantic. The NAM/NAO is an intrinsic mode of atmospheric variability as it always appears in long atmospheric simulations with climatological SST forcing. NAM and related jet stream variations are due to mean flow forcing associated with the breaking of transient, synoptic-scale Rossby wave (Benedict et al. 2004; Franzke et al. 2004; Riviere and Orlanski 2007). The NAM/NAO can also be viewed as a stochastic process with a characteristic e-folding time around 10 days (Feldstein 2000). On longer interannual time scales, it exhibits long-range dependence with more power than a simple red-noise process (Stephenson et al. 2000).

The observed interannual persistence of positive NAO phase winter events in the 1990s following the mostly negative phases in the 1960s has led to a strong NAM/NAO trend and many related climate impacts in the Northern Hemisphere (Hurrell et al. 2003; Hurrell and Deser 2009). This remarkable phenomenon has spurred a strong interest in the research community on the possible influence of low-frequency external forcings, such as interannual-to-decadal SST variability or the increasing GHG concentrations, onto the NAM/NAO. A couple of studies then suggested detection of an anthropogenic influence on sea level pressure (SLP) with a response pattern projecting strongly on the NAM in the northern extratropics (Gillett et al. 2003, 2005). However, they also pointed out that the climate models used in these studies strongly underestimated the amplitude of the response. A more recent study (Gillett and Stott 2009) carried out a global seasonal SLP detection and attribution analysis suggesting detection of an anthropogenic influence for the tropics but not for the southern and northern extratropics. This indicates that the NAM pattern did not dominate the anthropogenic fingerprint identified in previous SLP detection results. Note however that this last study uses a single climate model and needs to be extended using a multimodel approach to assess the robustness of the findings. Furthermore, the recent winter NAO/NAM trend has considerably weakened when updated to 2011 due to the dominance of negative phase years since 2000, to the extent that it is no longer significant at the 5 % level.

The study of the influence of external forcing upon the extratropical atmospheric circulation has often been based on the following paradigm (often termed the nonlinear paradigm): the forced response and the leading mode(s) of the unperturbed climate have similar structure implying also that the dominant patterns of intrinsic variability remain unchanged. Among the various arguments which have been proposed to sustain this paradigm (Branstator and Selten 2009), the following explanation is the most often invoked. The atmospheric variability exhibits preferred flow states or regimes such as blocked and zonal flows (Vautard 1990; Cheng and Wallace 1993; Kimoto and Ghil 1993; Hannachi 2007). In this framework, the response to anthropogenic forcing may be a change in the residence frequency of the most dominant regimes such as the phases of the NAM/NAO (Palmer 1999; Corti et al. 1999; Terray et al. 2004). Recent work suggests that this paradigm might not be adapted

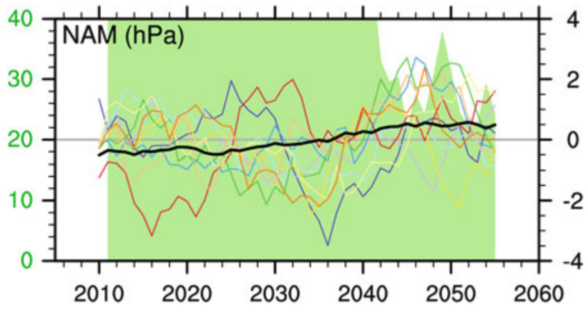

Fig. 3 Projected changes in the 10 year running mean wintertime Northern Annular Mode (NAM) from CCSM3 forced by the SRES A1B GHG scenario. The *thick black curve* shows the forced response, estimated from the average of the 40 realizations. The *thin colored curves* show results for ten individual realizations in which internal variability is superimposed upon the forced response. The *green shaded curve* shows the minimum number of model realizations needed to detect a 95 % significant change relative to the decade centered on 2010 (After Deser et al. 2012)

to fully capture the atmospheric response to anthropogenic forcing. First, while the structure of the NAM is mainly barotropic, it has been suggested from CMIP3 model studies that the response to anthropogenic forcing has a strong baroclinic component in the Arctic due to strong surface warming induced by ice melting (Woollings 2008). Second, the horizontal pattern of the mean response is never exactly the NAM (and even less so the NAO) but rather projects onto the NAM/NAO with different amplitudes depending on the models and periods used, the size of the ensembles, and other parameters. Third, in the context of the anthropogenic influence on the Northern Hemisphere extratropical circulation, the signal-to-noise ratio is likely to be low as shown by a study of a very large ensemble (40 members) of twenty-first century climate scenarios performed with one climate model (Deser et al. 2012). For example, more than 25 members are needed to detect the forced response in the NAM as estimated by the ensemble mean (Fig. 3).

Many studies have documented the possible impacts of increased amounts of greenhouse gases (GHG) upon the mid-to-high latitude atmospheric circulation changes in the Northern hemisphere. Among them, one can cite the rise in the height of the tropopause (Lorenz and DeWeaver 2007), the increase in dry static stability (Frierson 2006) and a NAM-related poleward shift of the tropospheric jet streams and storm tracks (Yin 2005). The latter is well marked in the Southern Hemisphere and less so in the Northern one, where it can actually be missing in some models. When present, this change is related to changes in baroclinicity with different effects between low and upper-level baroclinicity in the context of the twenty-first century GHG increase. Stronger warming in the tropical upper troposphere leads to an increase in upper-level horizontal temperature gradients in mid-latitudes while the increased warming of the polar regions due to sea ice melting leads to a decrease of low-level baroclinicity. Several mechanisms have been proposed to support the dynamical interpretation of the jet stream poleward shift due to enhanced upper-level baroclinicity. They usually invoke the increase and poleward shift of eddy kinetic energy as well as an increase in eddy length scale and its effects on the nature of wave breaking (Kidston et al. 2010; Riviere 2011). Other changes such as the tropopause height increase (Lorenz and DeWeaver 2007) or changes in subtropical stability (Lu et al. 2010) could also lead to similar effects.

While much has been learned about the impact of radiative forcing associated with changes in GHG and other constituents upon the NAM/NAO and related changes in the jet streams and other characteristics of the extratropical circulation, some key questions remain, including:

1. What are the relative impacts of the known mechanisms of the NAM and jet streams response? To what extent do competing changes in low and upper-level baroclinicity explain the model spread in the poleward shift of the Jet streams? If yes, what are the relative roles and spread of the surface and upper-level temperature response?
2. Does the NAM affect subtropical and tropical atmospheric circulations, and if so, by what processes? Is the potential interaction between tropical and extratropical modes going to change with increasing GHGs?
3. Do stratospheric dynamics play a role in the tropospheric NAM response? Is there a two-way interaction in the response to GHGs?
4. Is the non-linear paradigm still useful? Should we think instead in terms of two-way interaction between the response and variability?

To answer these, both observational and modeling approaches (including coupled ocean–atmosphere-land-sea ice models, those with a well resolved stratosphere and with and without interactive atmospheric chemistry) are needed. Simpler models such as the three-level quasi-geostrophic (QG) model or dry GCMs with simple setups must be widely used to provide dynamically coherent mechanisms and support complex GCM analysis.

3.2 The Southern Annular Mode

The Southern Annular Mode (SAM) refers to a seesaw of atmospheric mass between the middle and high latitudes of the Southern Hemisphere (SH; e.g. Thompson et al. 2000; Thompson and Wallace 2000; Marshall 2003; Fogt and Bromwich 2006). It is the leading pattern of tropospheric circulation variability over the extra-tropical SH, accounting for the largest fraction of variance on time scales longer than a few weeks (e.g. Thompson et al. 2000). The positive phase of the SAM is associated with reduced Sea Level Pressure (SLP) at polar latitudes and increased SLP at middle latitudes, evident as a strengthening of the westerly winds along their poleward flank; the negative phase shows opposite-sign changes (Thompson et al. 2000). The SAM is an intrinsic mode of atmospheric variability resulting from unstable dynamical feedbacks between the climatological zonal flow and high-frequency transient eddies along the storm track (e.g. Limpasuvan and Hartmann 2000). Although it is an intrinsic property of the atmosphere, it is also sensitive to a variety of external forcing factors including changes in radiative forcing associated with the build-up of greenhouse gas (GHG), depletion of stratospheric ozone concentrations, and alterations in earth's orbital parameters (e.g., Arblaster and Meehl 2006; Arblaster et al. 2011; Son et al. 2009, 2010; Polvani et al. 2011a, b; Hall and Visbeck 2002; McLandress et al. 2011). The SAM is also sensitive to changes in Sea Surface Temperatures (SSTs) both in the extra-tropics (e.g., Sen Gupta and England 2007) and in the tropics in association with the El Nino/Southern Oscillation (ENSO) phenomenon (e.g., L'Heureux and Thompson 2006; Seager et al. 2005, 2010; Fogt et al. 2011; Schneider et al. 2011). While present year-round, the SAM is most prominent in austral summer (December-February) and autumn (March-May).

Assessing the response of the SAM to each of the forcing agents listed above is a complex task due to: (1) the limited duration of the observational record; (2) the high level of unforced (internal) variability in the SAM; and (3) the covariability of the forcings (e.g., the build-up of GHGs and the depletion of stratospheric ozone have occurred in tandem over the past 50 years or so). Reliable instrumental records of barometric pressure suitable for documenting the SAM extend back to approximately 1957 (Marshall 2003). Attempts have been made to extend this record further back in time, but the degree of reliability of such efforts is not clear. The recent positive trend in the SAM since the late 1950s has been argued to be in part a response to both the increase in GHG concentrations and the decrease in polar stratospheric ozone amounts, based on a variety of atmospheric general circulation model experiments (e.g., Kushner et al. 2001; Arblaster and Meehl 2006; Deser and Phillips 2009; Son et al. 2009; Gillett and Thompson 2003). Arblaster and Meehl (2006) show further that the impact of ozone depletion is mainly limited to austral summer while the effect of increased GHGs is evident year-round. The case for the impact of ozone depletion upon the SAM has also been made in observational analyses, relying on the time-lagged response of the lower troposphere to radiative changes in the stratosphere to argue cause-and-effect (e.g., Thompson and Solomon 2002). A positive trend in tropical Pacific SSTs associated with ENSO since the late 1950s has also been shown to contribute to the upward trend in the SAM in austral summer (e.g., Schneider et al. 2011).

Lower-stratospheric ozone levels are expected to return to near-normal conditions in the next 30–50 years as a result of the Montreal Protocol to reduce ozone-depleting chemicals (Waugh et al. 2009). The increase in ozone levels is expected to drive a negative trend in the SAM which will counteract the tendency associated with increased GHG forcing (Perlwitz et al. 2008; Arblaster et al. 2011; Son et al. 2010). The net result may be a near-cancellation of radiative forcing and a lack of trend in the SAM, at least in austral summer (e.g., Polvani et al. 2011b).

It should also be noted, that while many of the same issues pertain to the NAM and SAM responses to anthropogenic forcing, there are some differences in the factors affecting the two annular modes. A primary consideration is that polar stratospheric ozone depletion has been stronger in the Southern Hemisphere (SH) than in the Northern Hemisphere (NH) over the past few decades. Given that ozone depletion and GHG increases both act to strengthen the SAM and to shift it poleward, one may expect the annular mode response to anthropogenic forcing to be stronger in the

SH compared to the NH in the late twentieth century and weaker in the twenty-first century due to SH ozone hole recovery.

While much has been learned about the impact of radiative forcing associated with changes in GHG and stratospheric ozone concentrations upon the SAM, some key questions remain, including:

1. What are the mechanisms of the SAM response, and to what extent are changes in SSTs (in both the tropics and extra-tropics) and sea ice involved?
2. To what extent does the SAM affect subtropical and tropical atmospheric circulations, and by what processes?
3. To what extent do changes in the stratospheric Brewer-Dobson circulation impact the SAM, and by what mechanisms?
4. To what extent do known modes of multi-decadal climate variability such as the "Pacific Decadal Oscillation" and the "Atlantic Multi-decadal Oscillation" affect the SAM?

To answer these, both observational and modeling approaches (including coupled ocean–atmosphere-land-sea ice models with and without interactive atmospheric chemistry) are needed. In particular, paleo-climate proxy records with demonstrated sensitivity to the SAM are needed to provide a longer-term perspective on past variations in the SAM and associations with fluctuations in CO_2 and SSTs, both tropical and extra-tropical. It is important that these proxy records provide information on austral summer and autumn conditions when the SAM is most prominent and distinguishable from another important mode of atmospheric circulation variability, the "Pacific-South American" pattern, which also affects middle and high latitude SH climate. While progress has been made towards answering these questions, for example Kang et al. (2011) address #2 and Li et al. (2010) address #3, additional studies are still needed.

3.3 Sea Ice and Associated Atmospheric and Oceanic Circulations

Sea ice responds directly to the changes in wind stress and heat fluxes associated to standard modes of atmospheric variability. For instance, when the NAO is in its positive phase, the enhanced south-westerly atmospheric flow in the Barents Sea induces a reduction of the sea ice cover while the more northerly winds in the Labrador Sea favors a higher sea ice extent there (Deser et al. 2000; Rigor et al. 2002). Anomalous circulation over the North Pacific also has a potential impact on sea ice in the Bering Sea up to the central Arctic (Overland and Wang 2005). In the Southern Ocean, SAM is associated with a decreased ice extent in the Bellingshausen Sea and an increase in the Ross Sea (Lefebvre et al. 2004). ENSO also has a clear impact in the Bellingshausen Sea, adding or subtracting its effect to the one of SAM there, depending on the polarity of the two modes (Stammerjohn et al. 2008; Pezza et al. 2011). Any change in atmospheric variability, as discussed in the other sections of this paper, thus have a clear impact on the sea ice cover. In a similar way, changes in the state of the ocean, bringing more or less heat to the sea ice, have an imprint on the sea ice in both hemispheres (e.g. Polyakov et al. 2010). However, the role of the ocean in explaining sea ice variability has been much less studied than the one of the atmosphere.

In turn, variations in the ice concentration or thickness modify the surface albedo as well as the heat and freshwater transfers between the ocean and the atmosphere, inducing temperature and circulation changes in those two media (e.g., Deser et al. 2007; Raphael et al. 2011) Those changes are present both locally, close to the anomalies, and at the large-scale. In this framework, the effect on the atmospheric circulation of the reduced ice cover during the recent summers has received particular attention because of the expected further reduction in the coming decades. In particular, it has been suggested that a low summer ice extent could be associated with stronger easterly winds (or reduced westerlies) in the following seasons, leading to colder conditions in some regions of Eurasia in Autumn and winter (e.g., Honda et al. 2009; Overland and Wang 2010; Petoukhov and Semenov 2010). However, additional work is still required to confirm and refine this hypothesis.

When analyzing changes in sea ice variability as a function of the mean conditions, we must take into account that sea ice displays a fundamental difference compared to the ocean and atmosphere as the surface it covers depends directly on the state of the system. By comparing various model results in different set up, it has been shown that the standard deviation of the summer ice extent in the Southern Ocean is roughly proportional to the root square of the mean extent (Goosse et al. 2009). The proposed explanation simply states that the largest fraction of the variability occurs nears the mean ice edge. A larger ice extent corresponds thus to a longer ice edge and thus to a larger variability. The variability of the ice extent is also strongly dependent on the mean state in the Arctic. However, its geometry, with a central basin surrounded by continents compared to the roughly annular Southern Ocean, induces a maximum standard deviation of the ice extent in summer for a sea ice cover of about 3×10^6 km^2, i.e. when enough ice remains to sustain large variability but the mean limit of the ice edge is still far away from the continent to allow strong variability of its position both southward and northward during cold and warm years (Goosse et al. 2009; Eisenman 2010).

Consequently, sea ice obviously plays a larger role in the setting up the mean conditions and the variability of the climate during cold periods such as the Last Glacial Maximum than in much warmer ones where it was absent of the surface of the Earth (Polyak et al. 2010). The few paleorecords of sea ice concentration are generally related to the presence or absence of sea ice during some periods covering centuries to millennia, with not much information on interannual to decadal variability. However, information from the early Holocene suggest low frequency variations of the sea ice transport, likely related to changes in atmospheric circulation and possibly to the forcing changes, as well as periods with larger multi-decadal variability of the ice cover compared to more quiet ones (e.g. Funder et al. 2011). The large-scale structure of those suggested changes and the mechanisms potentially responsible for them are still largely unknown and thus deserve attention.

In the future, models suggest an increase of the variability of the summer ice extent in the Arctic as sea ice extent is reduced (Holland et al. 2008; Goosse et al. 2009). This is consistent with the geographic arguments mentioned above but could also be related to a thinning of the ice pack (Notz 2009). Such a higher variability, combined with the reduction caused by anthropogenic forcings, can explain the very low ice extent observed during some recent years and the abrupt reductions of the sea ice extent simulated in response to the warming (e.g., Holland et al. 2006, 2008). An alternative explanation is that those large reductions would occur when the system is crossing a threshold (or tipping point) but this hypothesis appears unlikely on the basis of our present knowledge (Holland et al. 2008; Notz 2009; Eisenman and Wettlaufer 2009).

This brief overview illustrates that, although we have learned much about sea ice variability over the past decades, many questions remain. Some keys ones include

1. What is the response of atmospheric and oceanic circulation to anomalies in the sea ice cover?
2. Will the knowledge of the sea ice concentration and thickness bring predictability at the seasonal to decadal time scale?
3. Are warm states in polar regions (as for instance in the Arctic during the mid-twentieth century, the so-called Medieval Climate Anomaly roughly 1,000 year ago and the early Holocene) characterized by different amplitude/modes of variability than colder periods?
4. What is the role of ocean in setting up sea-ice variability at multidecadal time-scales?

Answering the first two questions will help us to understand and predict the impact of changes of the sea ice cover (mean and variability). Answering the third one will provide essential information on the behavior of the system that will help us to refine our projections of future changes. However, this will require additional high-resolution proxy records and simulations devoted to the target periods. Additional long time series will also be required in order to address question 4 but this will allow better estimates of the mechanisms responsible for the multi-decadal variability of the ice extent. A striking example is the occurrence of the big Weddell Polynya, covering about 250 10^3 km^2 during the entire austral winters of 1974, 1975, and 1976 (Carsey 1980). We still do not know if this major event is extremely rare one and may still occur in the future or if it was a recurring feature of the Southern Ocean that is not observed anymore because of some shifts in the Southern Ocean.

4 How Do Greenhouse Gas Induced Climate Changes Interact with Ozone Depletion?

Just as the GHG induced climate changes impact stratospheric ozone, changes to the ozone layer will also affect climate (Forster et al. 2011; Eyring et al. 2010). Changes in both long-lived greenhouse gases and stratospheric ozone influence surface climate directly via radiative effects and indirectly by forcing circulation and temperature changes. There have been a number of studies looking at the impact of Antarctic stratospheric ozone depletion on climate. In particular, lower-stratospheric cooling associated with the Antarctic ozone hole during austral spring and early summer strengthens the SH polar stratospheric vortex compared with pre-ozone hole periods (see Thompson and Solomon 2002; Baldwin et al., 2007; Forster et al. 2011). There may also be impacts on rainfall patterns (Kang et al. 2011) and the latitudinal extent of the tropics (Lu et al. 2009; Polvani et al. 2011a). Additionally there has been work considering the climate interactions between greenhouse gas increases and stratospheric ozone recovery. Key issues under current research include assessing how climate change may impact ozone recovery, how the current levels of stratospheric ozone depletion have affected surface climate and how changes in ozone expected with the decreases in concentrations of ozone depleting substances will impact the troposphere.

It has been found that changes in stratospheric ozone, water vapor and aerosols all radiatively affect surface climate. Observations and model simulations show that the Antarctic ozone hole is the major contributor to SH circulation changes over the past 50 or so years (Polvani et al. 2011a), and these changes extend all the way to the SH tropics. Additionally, ozone increases expected with the reduction in ozone depleting substances in the stratosphere may act to counteract some SH circulation changes expected from CO_2 increases (Polvani et al. 2011b). Recent literature (Scaife et al. 2012) has shown that there is a significant impact of stratospheric changes on tropospheric climate. It is quite likely that stratospheric ozone changes will alter the temperature and wind structure of the stratosphere. This will ultimately impact surface climate regimes.

The horizontal structure, seasonality and amplitude of the observed trends in the SH tropospheric jet are only reproducible in climate models forced with Antarctic ozone depletion. The southward shift of the SH tropospheric jet due to the ozone hole has been linked to a range of observed climate trends over SH mid and high latitudes during summer. Because of this shift, the ozone hole has contributed to summertime changes in surface winds, warming over the Antarctic Peninsula, and cooling over the high plateau. Other impacts of the ozone hole on surface climate have been investigated but have yet to be fully quantified. These include a potential impact in sea ice area averaged around Antarctica (e.g., Sigmond and Fyfe 2010), a southward shift of the SH storm track and associated precipitation, warming of the sub-surface Southern Ocean at depths up to several hundred meters; and decreases of carbon uptake over the Southern Ocean (see Forster et al. 2011 and references therein). Robust connections between NH ozone depletion and surface climate have not yet been established with observational data, possibly due to the fact that NH ozone losses are much smaller than those observed in the SH.

In addition to ozone changes impacting surface climate as noted in the previous discussion here on the SAM, GHG changes also can alter stratospheric ozone chemistry. This is through the GHG contribution to stratospheric temperature change, which then impacts ozone concentrations through changes in reaction rates for ozone controlling chemical reactions that are temperature-dependent. GHG forced changes in the stratospheric circulation can in turn alter the ozone distribution in the stratosphere and the flux of ozone into the troposphere. As noted above, ozone depletion/NH climate connections are not robust, however it remains to be seen whether NH ozone losses will increase with expected increases in GHGs and associated stratospheric cooling, thereby potentially altering NH surface climate as well.

There may very well be coincident changes in ozone, water vapor (a key GHG) and circulation. Randel et al. (2006) demonstrated a strengthening in tropical upwelling led to decreases in stratospheric water vapor as well decreases in ozone in a narrow layer near the tropical tropopause. They note that part of the temperature changes may also be explained as a radiative response to the observed ozone decreases. The changes in water vapor were subsequently used in a model study that demonstrated that the water vapor change may have induced a surface temperature response (Solomon et al. 2010). There are clearly feedback processes operating here, but they are not fully understood and warrant additional study.

Important questions remain that require further research.

1. When will stratospheric ozone recover to values seen prior to the discovery of the Antarctic ozone hole?
2. How will stratospheric ozone recovery interact with changing greenhouse gas concentrations?
3. How will changes in ozone impact surface climate? (primarily see discussion on the SAM)
4. What are the feedbacks between ozone changes, other radiatively active gases, and circulation changes?

To answer these questions, observations of stratospheric ozone and ozone depleting constituents need to continue. Key to furthering our understanding of stratospheric ozone/climate relationships are development and analysis of climate models that fully represent stratospheric processes.

5 Summary

There are strong indications that some aspects of climate variability either will change or have already changed with variations in GHGs. In this paper, we have discussed climate changes related to ENSO, tropical width, the NAM and SAM, sea ice and variations in stratospheric ozone. There is solid fundamental knowledge in regards to control mechanisms. However, there are many open questions in regards to all of these as well.

In regards to ENSO, there is more work that can be done using paleoclimate proxies. There are also questions as to how mean state changes impact interannual variability and predictability. More work is needed both in regards to modeling and observations.

In analysis of the latitudinal extent of the tropics there remain many unknowns. First off, there are questions on the accuracy of historical trends and details on mechanisms. More observations are needed, and further analysis of models that include all potentially relevant processes, both tropospheric and stratospheric. Similar conclusions can be drawn in regards to the state of knowledge for the NAM and the SAM, stratospheric ozone and sea ice variability in a changing climate.

The bottom line is that to fully understand how modes of variability will change we need additional analysis of observations, both paleo and present day, and solid fundamental understanding of mechanisms. Understanding of mechanisms necessarily requires use of models, ranging from simple to complex. Because coupling, between ocean and atmosphere, and between different segments of the atmosphere is important for many of the phenomena discussed, there should be an emphasis on fully coupled general circulation models. Stratospheric processes are also likely to be important, so models also need to include ozone chemistry. These topics involve all of the core WCRP projects, and answering the key questions will involve cooperative research between all the WCRP communities.

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