# THE CONSEQUENCES OF NOT KNOWING LOW- AND HIGH-LATITUDE CLIMATE SENSITIVITY

BY D. RIND

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**REGIONAL EFFECTS OF TROPICAL/HIGH-LATITUDE CLIMATE SENSITIVITY.** In this section the impact of not knowing low- and high-latitude climate sensitivity is discussed on a region-by-region basis, starting in the tropics and working poleward. The potential regional influences are often associated with longitude-specific SST gradients. Understanding tropical and high-latitude climate sensitivity involves more than just knowing how these regions will react as a whole, it also involves being able to forecast how different locations will warm relative to those at the same latitude but different longitudes. This regional differentiation can then affect the climate change in other regions, some far removed. The conclusions given in Rind (2008)—that atmospheric dynamics are more a function of temperature gradients than absolute temperature—applies to regional temperature gradients as well, so what we have learned for current climate relationships should be applicable, at least to some extent, for the future warmer world (Chamberlin et al. 2004; Joly et al. 2007).

*Tropics*. Not knowing tropical climate sensitivity obviously has very large consequences for the tropics itself. Houghton et al. (2001) estimate the impact that regional temperature changes will likely have on various societal activities, including human health, agriculture, and ecosystems. Not surprisingly, in all cases the effects worsen with greater warming, and the words used to describe the effects become more dire above about 2°C. This is especially true in the tropics, where each of these aspects will be endangered to some extent with significant warming [e.g., with respect to agriculture, see Rosenzweig and Parry (1994)]. The melting of tropical glaciers, and its effect on water availability, is also obviously related to how warm it becomes.

The direct thermal impacts are not the only problem. In an assessment we made of the likelihood of future drought, we compared precipitation with potential evapotranspiration in the Goddard Institute for Space Studies (GISS) model using a Supply–Demand Drought Index (SDDI), analogous to the Palmer Severity Drought Index (Rind et al. 1990). An update of that assessment is shown in Fig. S1, this time from a range of models. Results are fairly similar to that shown with the older GISS model and, as noted in Rind et al., the effect is strongly temperature driven (via its impact on potential evapotranspiration). It is also highly variable among the models in the specific region. For example, over the Amazon in the last 30 yr of this century, droughts that today occur once every 50 yr become the *average* condition using data from the Third Hadley Centre Coupled

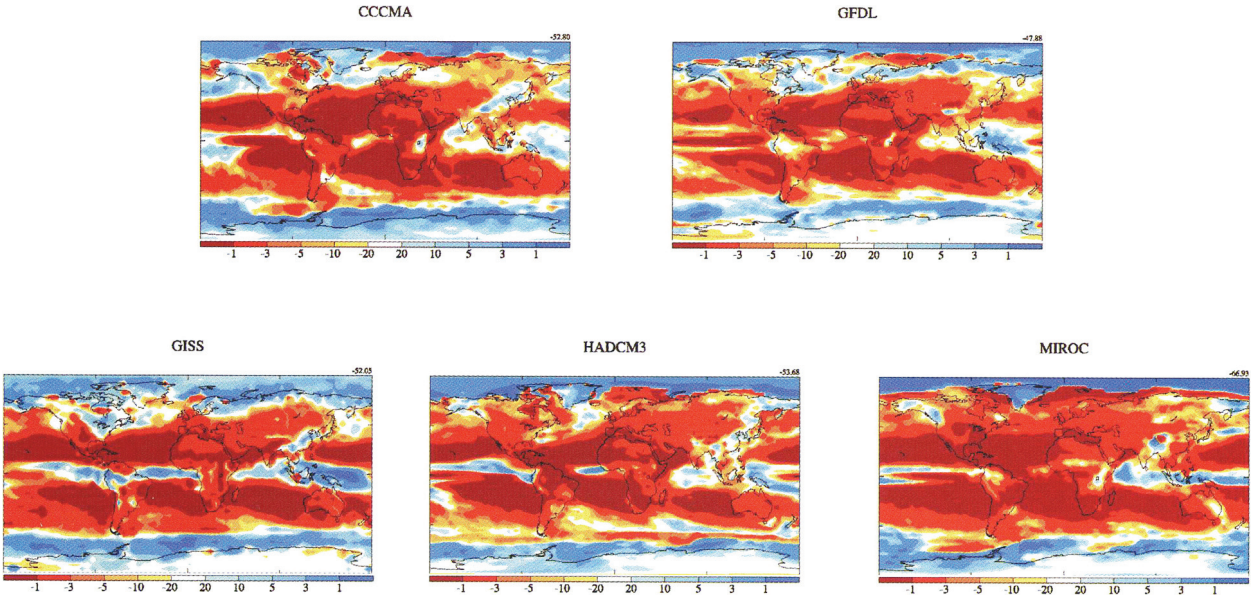


**Fig. S1. Supply–Demand Drought Index, indicating frequency of droughts (negative values) and floods (positive values) for summers from 2071 to 2100 in five different GCMs using the A2 scenario. The units refer to the frequency of drought/flood relative to that in the control run (e.g., dark red between 0 and –1 indicates average drought conditions that for the current climate occur less than once every 100 yr). [Courtesy J. Alltop; see Rind et al. (1990) for details.]**

Ocean–Atmosphere GCM (HadCM3), or they occur once every 6 yr [with the Canadian Center for Climate Modeling and Analysis (CCMA) model], or there is no drought increase at all (GISS model). The differences relate to the degree of warming (larger in HadCM3) and precipitation (greater in the GISS model).

Of course, the intensity of the hydrologic cycle is driven by the energy available at the surface, which, because of the predominance of low cloud differences among models, really relates to solar absorption changes (in models). Potential evapotranspiration is just a surrogate for this energy, and perhaps not a very good one. However, it does highlight that warming in the tropics has the potential to increase evaporative demand, and over land in this region, even the sign of the zonal mean precipitation change is uncertain in coupled atmosphere–ocean models (Douville et al. 2006). The greater the warming, the more frequent the severe and extreme droughts in this region, and the more poleward the drought impact spreads (Rind et al. 1990). The drought index changes look very similar to the agricultural response in impact models (e.g., Rosenzweig and Parry 1994) when one assumes no direct $CO_2$ fertilization effect.

*Tropical oceans.* As shown in Fig. S1, the effect described above influences the tropics as a whole; for more specific locations, the pattern of SST warming becomes important. Cloud feedback will obviously impact that as well, aided by ocean dynamical changes driven by atmospheric forcing. Considering the cloud feedback first, the low cloud response in models seemed to dominate in the eastern and (to some extent) western Pacific in many of the models, while the high cloud response was more dominant in the central Pacific and Indian Oceans [generalizing the results of Webb et al. (2006)]. Cloud responses in some models were nonsymmetric around the equator in the Atlantic. Given the uncertainty in both types of cloud changes, the particular pattern of warming would be expected to differ among the models because of this effect alone.

With respect to changes in ocean dynamics, the prime mode of sea surface temperature variability and a dominant influence on hydrological variability throughout the tropics is associated with ENSO cycles. Hence, knowing what the mean state in the tropical Pacific will be, and how the phase or amplitude of ENSO cycles may vary both become key questions. With respect to the mean state, there are several conflicting effects. For example, it is easier to warm cold water than warm water because latent heat flux is less important when the temperature is colder; and water with shallow mixed layer depths, and thus less heat capacity, will also warm faster. Both aspects suggest that the eastern Pacific will warm more than the west. On the other hand, the eastern Pacific surface ocean energy balance includes the upwelling

of water that may not have experienced warming, potentially acting as a thermostat to keep the eastern Pacific cold (Clement et al. 1996). There is a tendency in coupled atmosphere–ocean models for the first effect to win out, producing a quasi-permanent El Niño; 12 of the 15 models assessed in the Fourth Assessment Report (AR4) had this response, while the other 3 were more La Niña–like (Soloman et al. 2007). Consistent with a quasi-permanent El Niño is the general reduction in the Walker circulation (Vecchi and Soden 2007). As far as ENSO variability itself is concerned, Merryfield (2006) surveyed 15 coupled atmosphere–ocean models and found that for future projections, almost half exhibited no change, 5 showed reduced variability, and 3 showed increased variability. Solomon et al. (2007) did not place great credence in results for either the mean or variability changes, essentially concluding that the future of ENSOs and eastern Pacific temperatures were basically unknown.

El Niño conditions incite longitudinal circulation cells that influence other tropical ocean basins. They are associated with warmer Indian Ocean temperatures, most likely via this "atmospheric bridge" effect, impacting surface winds and thus surface heat fluxes, and perhaps also cloud cover (Klein et al. 1999; Yu and Rienecker 1999; Alexander et al. 2002; Lau and Nath 2003). El Niño conditions also may be related to a particular sea surface temperature pattern within the Indian Ocean, known as the Indian Ocean dipole or Indian Ocean zonal mode, perhaps at least partly through an oceanic bridge between the Pacific and Indian Oceans (Bracco et al. 2007). El Niño also influences temperatures in the tropical Atlantic, again via the atmospheric circulation effect on surface fluxes, with reduced northeast trades leading to warmer water north of the equator. Without knowing the future ENSO state, these various influences on the other tropical ocean basins cannot be forecast. Barsugli et al. (2006) compared the SST trends in 13 model projections for the first 50 yr of this century; the standard deviation among the models was about 40% of the projected trends themselves in the eastern Pacific and South Atlantic, and about 25% in the Indian Ocean and North Pacific. The simulated trends also differ from the observed SST and variability over the past half-century, which suggests that the real uncertainty may be higher.

*Northern Australia/Indonesia/western South America.* The precipitation response over northern Australia and Indonesia, as well as western South America, depends upon what will really happen to ENSOs and the tropical east Pacific background state. Drier conditions in Australia–Indonesia and wetter conditions in western tropical South America would be expected with warmer eastern tropical Pacific SSTs. Given the degree of uncertainty that exists, the multimodel standard deviation in precipitation change forecast for these regions exceeds the mean change itself, implying no confidence.

*India.* The Indian monsoon has a complex relationship with tropical SSTs. Warmer waters in the Indian Ocean lead to greater monsoon rainfall. In addition, the positive phase of the Indian Ocean dipole, characterized by cooler SSTs in the southeast and warmer anomalies in the central and western tropical Indian Ocean, lead to a circulation pattern that helps advect moisture onto the Indian subcontinent. Both of these are positively related to El Niño. However, El Niño also has an opposite effect—with reduced rainfall in the tropical western Pacific, a Rossby wave pattern is set up, resulting in subsidence over India (Lau and Nath 2000), in conjunction with cooler waters in the northern Australia–Indonesia region, another feature related to reduced monsoon rainfall (Nicholls 1995). It is hypothesized that the two ENSO effects work against one another, and that overall Indian Ocean warming over the last decade has reduced the El Niño influence on Indian monsoon rains (Krishna Kumar et al. 1999).

Associated with the ENSO influence, as well as the high cloud cover response, in about 60% of the models used in AR4 the Indian Ocean warmed by more than 2°C by the end of the twenty-first century, while in the other 40% the warming was less than that (Solomon et al. 2007, A1B scenario). While overall an increase in precipitation is forecast for the Asian monsoon, the intermodel standard deviation was higher than the average change, again showing no confidence (Solomon et al. 2007), presumably resulting from both ENSO and Indian Ocean warming variations, among other things.

*Brazil.* The Nordeste region is directly influenced by tropical Atlantic SSTs, with drought favored when the North Atlantic warms more than the South Atlantic, driving the ITCZ northward. Hence, this too is associated with El Niño (e.g., Moura 1994). Of course, increased greenhouse gases might also affect this north–south SST gradient.

In the AR4 model projections, temperature changes both north and south of the equator in the Atlantic show 50% probability of being greater than or less than 2°C, respectively, often associated with

reduced low- and high-level cloud cover (Webb et al. 2006). In some models the warming is greater north of the equator, while in others it is greater to the south (Barsugli et al. 2006). The Nordeste region also shows a mean precipitation change that is less than the standard deviation of the model projections (Solomon et al. 2007)

*Africa.* An interhemispheric gradient also affects western Africa, including the Sahel region, in the opposite sense from Nordeste—greater warming to the south (in the Atlantic and Indian Oceans) with cooling to the north (in the Atlantic and Pacific) produces drier conditions, as occurred from the 1960s into the 1990s (e.g., Joly et al. 2007). While El Niño promotes just the opposite latitudinal SST effect in the Atlantic, and hence should be encouraging rainfall, warming in the tropical east Pacific and Indian Oceans incites longitudinal cells that promote subsidence and drought in this region (e.g., Fontaine and Janicot 1996). Drought conditions are also more common in Zimbabwe when eastern Pacific SSTs are warm (Cane et al. 1994). The model disagreements in the tropical ocean response make all such forecasts unreliable. While the multimodel mean predicts an increase in the south portion of the west African monsoon and some drying over the Sahel (Cook and Vizy 2006), these regions too have mean precipitation changes less than the multimodel projection standard deviation (Solomon et al. 2007).

*Tropical storm impact regions, particularly in the Atlantic.* The influence of climate warming on tropical storm development, both currently and in the future, is a contentious topic. As the climate warms, the moist-adiabatic lapse rate is reduced because of added moisture, so the atmosphere becomes more stable. This has the general effect of favoring reduced cyclonic development and rainfall (Yang et al. 2003), and the particular effect of helping to reduce tropical storm frequency in some models (Yoshimura et al. 2006). On the other hand, increased sea surface temperatures favor stronger storms (e.g., Knutson and Tuleya 2004). The two effects work against one another, and which will win may depend on the magnitude of warming or on a third factor—changes in vertical shear, particularly in the Atlantic.

The same features that influence drought in western Africa (ENSO and the temperature gradient between northern and southern oceans) help establish circulation patterns that influence vertical shear. The El Niño phase (or a semipermanent El Niño) promotes an equatorially confined zonal circulation with upper-level west winds and lower-level easterlies, and hence strong vertical shear that helps suppress Atlantic basin hurricanes. To the extent that this same phase also reduces the intensity of the west African monsoon, it can affect easterly waves, and the occurrence of strong hurricanes (Goldenberg and Shapiro 1996). Similar effects arise with greater warming in the southern Atlantic (and southern oceans in general) than in the north (Fontaine and Janicot 1996). Knutson and Tuleya (2004) argue that many models do not show large changes in the vertical wind shear, but given the uncertainty in the patterns and magnitude of tropical SST change, the issue is far from settled.

*Midlatitudes, particularly North America.* By this latitude, the tropical forcing has less impact on variability, and total SST forcing accounts for no more than 15%–20% of the variance, even in regions that are strongly affected (e.g., Seager et al. 2005). For example, the AR4 models show a general poleward movement of the winter storm track in the Pacific–North American sector despite the increased El Niño–like background state that prevails in many of them, which should shift the storm track equatorward. Clearly, extratropical influences will affect much of what goes on at these latitudes (one example of this is given at the end of this section). However, tropical forcing plays a role today, and will likely continue to do so.

As noted in Rind (2008; "Large-scale responses to low- and high-latitude climate sensitivity" section), models run for the Intergovernmental Panel on Climate Change (IPCC) AR4 project show, in general, a poleward expansion of the Hadley circulation, with an intensification and northward movement of some of the subtropical dry regions (e.g., Seager et al. 2007). As noted earlier, this effect is directly related to tropical warming and latitudinal gradient changes. The influence of the tropical warming on hydrologic change at midlatitudes is also strongly pattern-dependent. Cool conditions in the eastern tropical Pacific have been related to annual U.S. droughts in various studies (e.g., Barlow et al. 2001; Schubert et al. 2004; Seager et al. 2005), although they are more capable of influencing the U.S. climate in late winter when the tropical winds are more conducive to allowing the tropical–extratropical connection (Newman and Sardeshmukh 1998; Lau et al. 2006). Conversely, warm conditions associated with El Niño are known to amplify rainfall values along the southern border of the United States in association with an amplified subtropical jet stream, and El Niño has also been related to specific, very wet seasons (e.g., summer 1993; Bates et al. 2001). Warm conditions in the western Pacific–Indian Ocean region are capable of instigating drought in the United States year-round (Hoerling and Kumar 2003; Lau et al. 2006), but especially in spring (Chen and Newman 1998), by helping to generate amplified ridges at upper midlatitudes.

The exact mechanism(s) that connect tropical SST changes and midlatitude circulation features is still

under debate. Warmer sea surface temperatures result in increased convection and diabatic heating, which can help produce Rossby waves (either through the tropical heating or via subsidence in the subtropics). Such waves generated in the Indian Ocean–west Pacific domain can propagate far distances, and thus influence the position of ridges and troughs over the United States (Chen and Newman 1998; Newman and Sardeshmukh 1998). Alternatively, tropical sea surface temperature changes will alter the strength of the subtropical jet and North Pacific high, affecting both the generation and propagation of transient waves; several studies suggest that it is these waves, rather than a stationary wave train, that are really generating the drought-producing ridges over the United States (Seager et al. 2003, 2005; Lau et al. 2005; L'Heureux and Thompson 2006). Because no particular sea surface temperature perturbation is associated with a large proportion of the variance of drought conditions over the United States, it is conceivable that various locations and mechanisms can influence the midlatitude response. When severe drought occurs, it may be because several mechanisms/regions are acting in tandem. Hence, to understand the likelihood of midlatitude droughts from these dynamical effects would require the various tropical SST pattern changes to be known. In this regard, differing model results likely contribute to the lack of consensus concerning droughts or soil moisture changes at midlatitudes (Solomon et al. 2007).

At these latitudes, the effect of the uncertain high-latitude sensitivity begins to become noticeable. In one example, Sewall and Sloan (2004) found that warmer temperatures associated with reduced sea ice, in particular over the Barents and Kara Seas, increased 500-mb heights south of those areas, and as part of the planetary wave response resulted in increased ridging and decreased precipitation over western North America during winter (when most of the rainfall for this region occurs).

*Japan, Alaska, and Canada.* The North Pacific index (NPI), the area-averaged sea level pressure for the region encompassing the Aleutian low (Trenberth and Hurrell 1994) is strongly related to temperature and precipitation in Alaska and Japan; it is a proxy record for the midtroposphere Pacific–North American teleconnection pattern, which in one phase features a stronger Aleutian low and amplified ridging over the northwestern United States and Canada. Decadal variations in the NPI have been related to sea surface temperature variations in the tropical Indo–Pacific region (Branstator 2002; Deser et al. 2004a) and, especially on interannual time scales, to ENSO perturbations in the tropical eastern Pacific (e.g., Trenberth et al. 1998; Deser et al. 2004a). Clearly, potential changes in this index depend upon the future state of the background and/or ENSO variability in the eastern and central Pacific as well as tropical warming in general.

*Europe, northern Asia.* The North Atlantic Oscillation (NAO) is defined as the sea level pressure difference between Portugal and Iceland, indicative of an oscillation in atmospheric mass between midlatitudes and higher latitudes over this ocean basin; a positive phase has been related to warm and wet winters over northern Europe and Siberia. It, along with the NPI, make up the Cold Ocean/Warm Land (COWL) index (Wallace et al. 1995), which is somewhat indicative of recent temperature trends. Over the past few decades (though not particularly in the last decade), this index has tended to be positive, featuring lower pressure at higher latitudes. Decadal variations in the NAO have been related to a change in the SST gradient in the tropical west Pacific (Kucharski et al. 2006), or to Indian Ocean temperatures (Hoerling et al. 2004), possibly in opposing ways (Kucharski et al. 2006 op. cit.). The tropical east Pacific may also be able to affect the NAO state (Lu et al. 2004), although historically the correlation with the Niño-3.4 index explains less than 15% of the variance (Kurcharski et al. 2006). Not to be left out, tropical Atlantic SST perturbations may also impact the NAO, via convection over the Amazon and Caribbean Sea (Terray and Cassou 2002; Cassou et al. 2004).

*Middle and high northern latitudes.* Higher-latitude regions in both hemispheres are characterized by zonally symmetric leading modes of variability that are thought to be affected by tropical SST variations. Large-scale influences on the northern annular mode (NAM) were discussed in Rind (2008; "Large-scale responses to low/high-latitude climate sensitivity" section). The NAM is highly correlated with the NAO, because the Atlantic sector variability dominates the latitudinal average pressure differences at these latitudes. In that respect, it too can be influenced by Indian Ocean–west Pacific temperature variations, although opposing effects on the Pacific and Atlantic sectors may limit the response on the zonal average (Kucharski et al. 2006). The influence of the eastern-central tropical Pacific may be more subtle; Quadrelli and Wallace (2002) found that El Niño produced a more extensive low pressure area over the pole, extending into Siberia, which is probably due to the alteration in the longitudinal extent of the subtropical jet across the Pacific. As with many of the extratropical relationships claimed for SST variations in this region, the issue is far from settled. L'Heureux and Thompson (2006) found no relationship between ENSOs and the NAM, primarily because the ENSO influence on the extratropics does not occur sufficiently far poleward to alter wave refraction and eddy momentum transports at higher latitudes. Schneider et al.

(2003), using the Center for Ocean–Land–Atmosphere Studies (COLA) GCM found that the influences from the Indian Ocean and west Pacific were out of phase, as was the influence of the Indian–west Pacific region and the eastern-central Pacific region.

As discussed in some detail in Solomon et al. (2007), more models than not project a more positive phase of the NAO/NAM. The magnitude of the effect is not particularly large, a result that is probably affected by the models' different tropical SST warming patterns and magnitudes. As noted in Rind (2008), in the GISS model, both strong tropical warming in the upper troposphere and large high-latitude warming at low levels can generate a more positive NAO/AO phase, so the differing high-latitude responses may contribute to the uncertainty.

*Southern Australia, New Zealand, Antarctica.* The equivalent leading mode of variability in the Southern Hemisphere, the southern annular mode (SAM), also is representative of mass oscillation between higher and midlatitudes. SAM has been more positive the last few decades, driven at least in part by stratospheric ozone reductions in spring and summer (Thompson and Solomon 2002), but perhaps also because of global warming (Marshall et al. 2004; Shindell and Schmidt 2004). Most of the models used for AR4 project an increasingly positive SAM phase, with greater significance than was true for the NAM (Solomon et al. 2007), even independent of stratospheric ozone reductions. Associated with this is a southward displacement of storm tracks, which results in reduced rainfall in the southern Australia–New Zealand region. In addition, if a more positive SAM does continue, it will presumably continue to be associated with amplification of warming in the Antarctic peninsula region, with a tendency toward cooling or reduced warming over Antarctica itself.

Considering other influences, the association of ENSOs and SAM is equally as contentious as the Northern Hemisphere relationship. One study found the preferred relationships in summer of El Niño with negative SAM and La Niña with positive SAM, which accounted for 25% of SAM variability (L'Heureux and Thompson 2006); others found no such relationship either in winter (Marshall and Connolley 2006) or throughout the year (Gallelgo et al. 2005). One reason for the different results may be the dominance of natural variability in the Southern Ocean region, which is capable of swamping the signal (Lachlan-Cope and Connolley 2006). In our own simulations, the association of La Niña and positive SAM during late spring/early summer (months of maximum correlation) occurred 37% of the time, as compared with the 21% for La Niña and negative SAM.

Another reason for the conflicting results may be that the primary influence of ENSO on high-latitude circulation is not zonally symmetric. El Niño is accompanied by higher pressure in the eastern Pacific (the Amundsen–Bellingshausen Sea region, e.g., Lachlan-Cope and Connolley 2006) and lower pressure in the Weddell Sea, constituting a Southern Ocean dipole, which has been observed in sea ice variations (Yuan and Martinson 2000). We explained this phenomenon as being the result of more equatorward storm tracks in the Pacific during El Niño, depriving the storms of the strong available potential energy associated with the high-latitude temperature contrast produced by Antarctica, while at the same time storm tracks (and the subtropical jet) shift poleward over the Atlantic and the storms are conversely strengthened (Rind et al. 2001a). Alternatively, the effect may be related to stationary wave propagation from the tropics (Lachlan-Cope and Connolley 2006). The latter study notes that for warmer SSTs to initiate the deep convection responsible for Rossby wave generation, it is necessary for them to occur in regions of background ascent; however, if warming exceeded 4°C (with the current background stability) it could initiate convection regardless of the background vertical motion. This implies that strong tropical global warming may well initiate tropical–extratropical interactions that have heretofore not occurred.

*Arctic sea ice.* One of the impacts of changes in the high-latitude modes of variability, either in terms of phase or location, relates to its effect on sea ice. As has been well documented, Arctic sea ice at the end of the September melt season has declined rapidly, especially over the last several decades. While GCMs have forecast Arctic sea ice decrease, their rate of change is noticeably slower, some 1/2–1/3 of what has been observed, depending on the averaging period (Stroeve et al. 2007). One potential explanation offered was that the rapid change was due to a more positive NAO signal from the mid-1980s through the mid-1990s (e.g., Rigor et al. 2002), helping to advect sea ice out through the Fram Strait. The more positive NAO has also been associated with increased heat advection through both the Barrents Sea and the eastern Fram Strait into the Arctic, which has acted to diminish the cold halocline layer separating sea ice from the warm Atlantic layer below (Grotenfendt et al. 1998; Steele and Boyd 1998; Dickson et al. 2000). [However, it has since recovered somewhat (Boyd et al. 2002), consistent with a less positive NAO.] GCMs in general have not produced as positive an NAO trend as was observed and thus might be expected to miss some of these effects (either because the models are deficient in this response to global warming, or because part of this effect is just natural variability). The GCM sea ice loss is then primarily associated with the warming and thinning of sea ice, leading to increased ocean heat fluxes, further warming the atmosphere as a

positive feedback; another positive feedback arises because of melting and reduced sea ice albedo. Indeed, warming is thought to play a significant role in these ways in real world trends as well, primarily affecting sea ice production (Rothrock and Zhang 2005).

However, the sea ice decrease has continued while the NAO state has been relatively neutral; this was explained as a delayed response resulting from the thinner ice that had been left in the Arctic by the previous positive NAO (Rigor and Wallace 2004). Closer examination has revealed that the atmospheric flow field has continued to be favorable for ice advection, just not in an NAO mode. Maslanik et al. (2007) found that two other patterns—one featuring low pressure over the central Arctic, and the other featuring a dipole pattern of high pressure over the Canadian Arctic in conjunction with low pressure over the Siberian Arctic—also act to remove ice, and both of these have picked up since the mid-1990s. As well, warm ocean water continues to be advected into the Arctic, both from the Atlantic (Shimada et al. 2006) and the Pacific (Polyakov et al. 2005).

The AR4 models predict continued Arctic sea ice reductions during this century, with some producing ice-free Arctic conditions during late summer (Solomon et al. 2007). Given the weak positive NAO/AO phase being projected, presumably the model response is again primarily thermodynamic, and could be under/overestimated if different circulation trends arise because of the advective effect alone. Obviously, the (interactive) high-latitude sensitivity also bears on the question of future sea ice conditions.

Does the sea ice disappearance act as a positive feedback to potential NAO (or NAM) changes? If so, then high-latitude sensitivity directly affects both this mode and all of the midlatitude regions it influences. Various studies have examined this issue, and the results seem to depend on exactly which sea ice changes are considered. The local response, however, is similar in different models: reduced sea ice leads to greater surface heat fluxes, destabilizing the atmosphere and producing lower sea level pressure and warmer temperatures—a positive feedback. The warmer temperatures also cause the atmosphere to expand, producing ridging above the boundary layer in the vicinity of the sea ice change. In addition, in the National Center for Atmospheric Research (NCAR) Community Climate Model (CCM) the effect of sea ice reductions east of Greenland projects onto the leading mode of internal variability in such a way as to produce a small negative NAO tendency, which is opposite to the trend that had produced the real world sea ice change (Magnusdottir et al. 2004; Alexander et al. 2004; Deser et al. 2004b). In the Hadley Centre GCM, reduced sea ice in the Arctic had no impact on the NAO per se, although it did result in reduced sea level pressure in the Arctic, similar to what has been observed (Singrayer et al. 2006) and similar to one of Maslanik et al.'s (2007) patterns that produced more sea ice advection. In that sense, the sea ice loss did act as a positive feedback, just not to the NAO. Sea ice reductions in the North Pacific also produced a small positive feedback on the circulation in the Pacific–North American region (Alexander et al. 2004). If the positive feedbacks prove to be real and substantial, they could encourage rapid Arctic sea ice loss; this effect may already be happening.

*Southern Ocean sea ice.* In the Southern Ocean, the prominent sea ice changes have consisted of strong reductions in the southern Bellingshausen/Antarctic peninsula (B/AP) region, with the sea ice season duration about 85 days shorter, and strong growth in the Ross Sea, and an ice season duration about 60 days longer (Stammerjohn et al. 2008). Thus, while the overall Southern Ocean trend is small (growth of about 0.8% decade$^{-1}$, Cavalieri et al. 2003), it masks a much greater regional variability that is already having an impact on local biology (e.g., Massom et al. 2006). These changes appear to be related to intensified low pressure in the South Pacific sector of the Southern Ocean, with increased north winds over the B/AP and the reverse flow over the Ross Sea. The effects have been associated with the more positive SAM possibly interacting with the ENSO phase, as discussed above. To the extent that the more positive SAM (in summer) is related to the ozone hole (e.g., Gillett and Thompson 2003), this effect is likely to diminish with time, although as noted above, models project that greenhouse warming by itself will produce a more positive SAM. The sea ice edge helps position the baroclinic temperature gradient, encouraging low pressure areas to the northeast of the ice extent maximum (Yuan et al. 1999); were sea ice to diminish substantially in the Southern Ocean (a somewhat doubtful proposition due to strong negative feedback from ocean heat fluxes), that would encourage a more negative SAM, or perhaps new modes of variability.

*Greenland ice sheet.* In addition to disappearing sea ice, another cryospheric response presumably affected by high-latitude sensitivity relates to changes in the Greenland mass balance. Widespread glacier acceleration has been documented, with double the ice sheet mass deficit in the last decade (up through 2005) (Rignot and Kanagaratnam 2007). However, there was a strong decrease in ice discharge in 2006, bringing rates back down to those near previous loss values (Howat et al. 2007). To some extent, this might be compensation for the strong previous losses; Truffer and Fahnestock (2007) point out that this rapid variability has an analog in large tidewater glaciers, such as the Columbia Glacier. Although retreat rates for this gla-

cier are highly variable, the overall effect has been a retreat and thinning in the past 20 yr. GCMs cannot comment directly on ice sheet loss because of the lack of embedded ice sheet dynamics models and the lack of information on conditions at the glacial surface (i.e., infiltration) and base (e.g., frozen to the ground or free water providing an hydraulic lift). Global positioning measurement data suggest that summer melting may be reaching the glacial bed and accelerating the flow (Zwally et al. 2002). The lack of definitive modeling studies prevented Solomon et al. (2007) from determining what the likely ice sheet contribution to sea level rise will be. Considering what GCMs do include, the models predict increased mass in Northern Greenland due to higher precipitation, with mass loss around the southern edges and a relatively modest effect on sea level (Solomon et al. 2007).

*West Antarctic ice sheet.* A similar uncertainty affects predictions of changes in the west Antarctic ice sheet. Glaciers draining into the Amundson Sea are losing mass because of accelerated flow, possibly already affecting the salinity of the Ross Sea (Jacobs et al. 2002). Ocean warming has been identified as a possible trigger for this trend (Payne et al. 2004). With respect to the Antarctic peninsula, there has also been the notable collapse of portions of the Larsen Ice Shelf, with accelerated ice discharge from various ice streams (Rignot et al. 2004). Nevertheless, in both regions some of this loss has been cancelled by additional snowfall (see the discussion in Shepherd and Wingham 2007). Models predict a net mass gain for Antarctica in the future, despite the fact that it (as well as Greenland) is estimated to be currently experiencing a net loss (Shepherd and Wingham 2007). At this point it is impossible to determine whether, if the high-latitude sensitivity is underestimated, this would result in massive glacial melting raising sea level, or more accumulation of snow, lowering it.

*North Atlantic Deep Water.* An additional component of the high-latitude sensitivity relates to model predictions of North Atlantic Deep Water (NADW) changes. Most (but not all) of the models run for AR4 predict a decrease in the overturning circulation (Solomon et al. 2007); no model predicts an increase. The effect of the slowing is to reduce the high-latitude warming especially over the North Atlantic. Both freshening and warming of waters in this region are responsible for the slowdown, and there are numerous freshening possibilities: increased rainfall over both the Arctic and North Atlantic associated with warmer temperatures; increased river runoff into the Arctic in the Eurasian sector, again due to greater precipitation, followed by increased freshwater flux from the Arctic into the North Atlantic. Opposing this tendency in the AR4 models is decreased future sea ice melt in the Greenland–Iceland–Norwegian Seas, resulting from reduced transport through the Fram Strait (Holland et al. 2006). Currently a freshening does appear to be occurring as a result of many of these same processes (Peterson et al. 2007), although at this time there is increased sea ice melt to help as well, and greater sea ice export resulting from the current wind forcing. This latter effect furthermore appears to be helping to freshen the Labrador Sea (e.g., Koenigk et al. 2006). In the future there will be less sea ice to be transported, and that influence will diminish; a continued advective loss associated with the atmospheric circulation is somewhat uncertain as well. Because the models do not include a substantial future Greenland ice melt component (which is providing a small freshening now), underestimation of that effect would imply even larger potential NADW reductions. Additionally, changes in NADW production will alter the ocean everywhere and will affect ocean $CO_2$ uptake.

*Southern Ocean.* If, as forecast, climate change brings with it a more positive SAM phase, the Southern Hemisphere ocean circulation should be affected, because the stronger and southward-shifted west winds produce greater Antarctic divergence with a consequent acceleration of the Deacon cell. The downwelling side of this cell acts to sequester anthropogenic $CO_2$ (currently 40% of the total; Sabine et al. 2004) and that effect should then grow, perhaps counteracting the increased stability associated with the warming waters (Russell et al. 2006). The upwelling side of the cell exposes deeper water with higher $CO_2$ levels (from dissolved inorganic carbon) to the atmosphere, and thus acts as a natural source; it is this effect that dominates today (Lovenduski et al. 2007). A more frequently positive SAM would then be associated with less Southern Ocean $CO_2$ uptake as a whole, a positive feedback (Lovenduski et al. 2007, op. cit.).

*High-latitude peatlands.* Another greenhouse gas whose future release/uptake rate depends on the high-latitude response is methane. Some high-latitude regions, such as western Siberia, are covered with peat. Peatlands today take up $CO_2$ and release methane, and this has probably been true since the early Holocene (Smith et al. 2004). If the regions are warm and dry, they will degas $CO_2$ (adding about 4% to the current rise) but shut down methane outgassing, which could reduce the $CO_2$ outgassing effect by ~20% (Smith et al. 2004, op. cit.). On the other hand, if it becomes moderately warmer and wetter, peat would increase, and on the long term the draw down of $CO_2$ would be greater than the methane emission, reducing the greenhouse effect. However, if the warming were even stronger, there could be large

increases of methane release. Model projections show, on average, a small reduction in soil moisture in these regions (despite precipitation increases), but again with no associated confidence.

Considering the uncertainties in tropical and high-latitude sensitivity, it is conceivable that the latitudinal sea surface temperature gradient change might be very different in the Atlantic and Pacific Ocean basins. If North Atlantic Deep Water production decreases, the SST gradient in the tropical Atlantic could increase. If the central/eastern tropical Pacific really acts as a thermostat, the Pacific latitudinal temperature gradient could strongly decrease. Our experiments show that gradients of one sign in the Pacific produce a similar response to gradients of the opposing sign in the Atlantic (Fig. S2), due to their impact on longitudinal cells (Rind et al. 2001a), and together they could have a strong impact on atmospheric circulation features. Given the resistance of Southern Ocean sea ice changes to global warming, the latitudinal temperature gradient change in the Southern Hemisphere could be very different from that in the Northern Hemisphere, with the potential for interesting and new cross-hemispheric influences and preferential buildup of mass in one hemisphere, further affecting the circulation (Rind et al. 2001b).

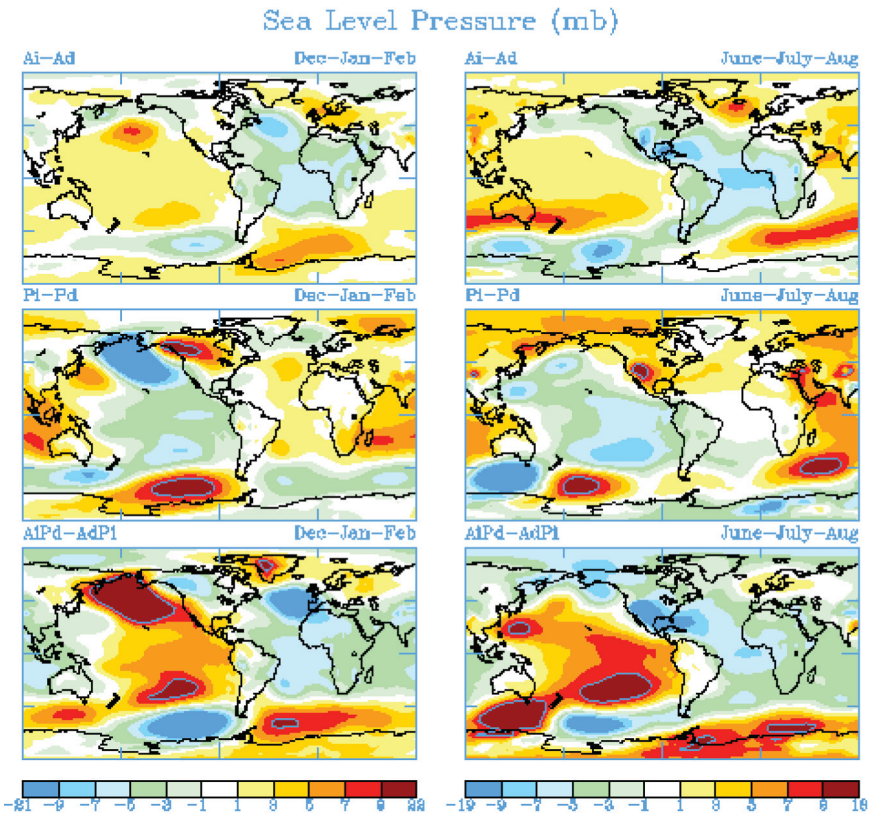


**Fig. S2. Sea level pressure changes in the two solstice seasons from a GCM simulation with an increased sea surface temperature gradient minus a simulation with a decreased gradient in the Atlantic (top), in the Pacific (middle), and (bottom) increased gradient in the Atlantic along with a decreased gradient in the Pacific minus the reverse. [Adapted from Rind et al. (2001b).]**

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