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

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We still can't predict future climate responses at low and high latitudes, which constrains our ability to forecast changes in atmospheric dynamics and regional climate.

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The first work summarizing research on prospective climate change resulting from increasing greenhouse gases was known colloquially as the Charney report (Charney et al. 1979). All of 22 pages long, it compared results from several different general circulation models (GCMs), whose range for doubled-CO_2 ($2 \times CO_2$) warming went from 1.3° to 4.2°C; thus, it concluded that climate sensitivity to $2 \times CO_2$ was between 1.5° and 4.5°C. In the coming decades, we have not moved very far from that range, although recently, the Intergovernmental Panel on Climate Change (IPCC) has indicated a low probability for $2 \times CO_2$ warming less than 2°C (Solomon et al. 2007).

In neither the Charney report nor the follow-up assessment (Smagorinski et al. 1982) was it noted that the global mean temperature differences between models were indicative of other important discrepancies, in particular, the tropical and high-latitude responses. The Geophysical Fluid Dynamics Laboratory (GFDL) model tended to produce a tropical warming resulting from doubled CO_2 that was less than 1/2 of that in the Goddard Institute for Space Studies (GISS) model; on the other hand, its high-latitude sensitivity was about doubled (see Fig. 1). From the standpoint of atmospheric dynamics, and possibly hydrology, this was a worse discrepancy than the global sensitivity difference (which between those two models was only about 33%).

We have not managed to trim the uncertainty in our global sensitivity number very much; but, have

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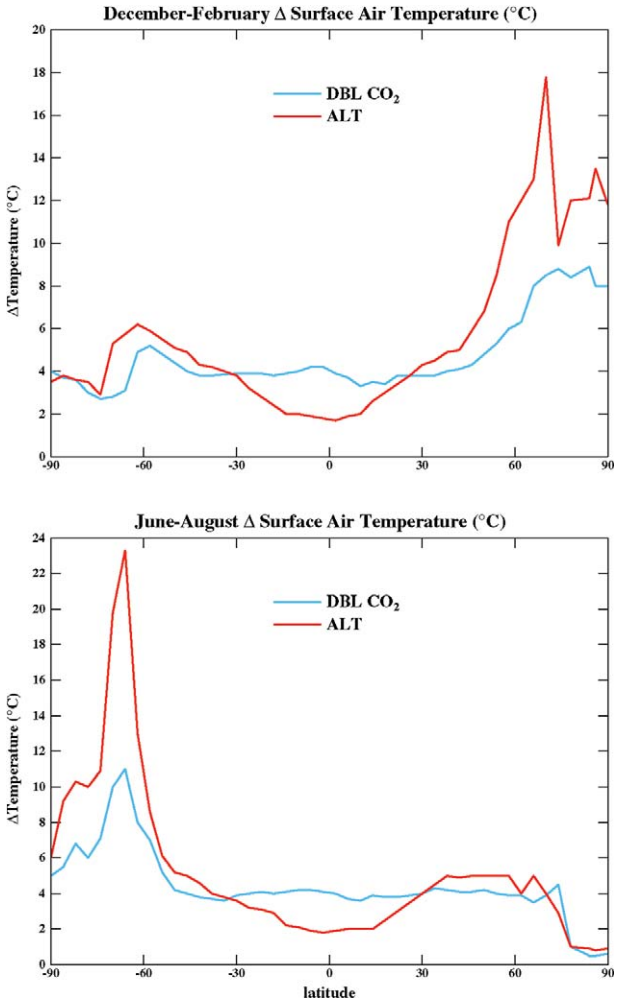

Fig. 1. Surface air temperature change in the two solstice seasons when using doubled-CO_2 sea surface temperatures as calculated in the GISS (DBL CO2) and GFDL (ALT) models circa early–mid-1980s. [From Rind (1987).]

we at least improved our assessment of high- and low-latitude sensitivity, and hence the change in the latitudinal temperature gradient? Doubled-CO_2 simulations of one sort or another were performed by many of the Fourth Assessment (AR4) climate models, and the results show that this uncertainty is also still as great as ever, with a factor of 2 at both high and low latitudes. The same is true for transient climate change simulations.

Of course, a lot *has* changed between the early models and our current ones: now coupled dynamical oceans without flux corrections are being used; cloud liquid water routines are incorporated to calculate changes, not only in cloud cover but in cloud optical depth; the horizontal and vertical resolutions are noticeably finer; and the number of different models has increased by a factor of 5. However, the same net uncertainties still exist; their causes are somewhat better known, perhaps, but a solution is not yet in sight.

Do climate changes that have been happening recently give us any indication of the proper sensitivity? Since 1900, the estimated temperature change in the tropics (24°N–24°S) has been about 2/3 that of the extratropics (the rest of the globe, averaging the two hemispheres together), and the same is true for the last 25 yr, with tropical warming of 0.3°C (see the GISS temperature reconstruction, information available online at http://data.giss.nasa.gov/gistemp/graphs/Fig.B.lrg.gif). This would seem to imply a relatively large tropical sensitivity. However, the tropical data are sparse and of uncertain quality (especially going back 100 yr); the tropics are more likely to respond faster, given the shallow mixed layer depths; and the tropical warming is affected by tenths of degrees due to variations in ENSO phase frequency, with a recent tendency for more El Niño events, which may not be related to global warming.

Over the past 25 yr, the region north of 65°N has warmed about 1°C, 3 times more than the tropics; this would seem to imply a large high-latitude sensitivity. The recent warming is not necessarily indicative, however; high latitudes are more variable, and high southern latitudes have shown little zonal-average temperature change (though with large regional variations). Russian data collected during the past century show that the high-latitude response (poleward of 62°N) is dominated by a low-frequency oscillation (two cycles, of about 60-yr length), with very little overall trend in either temperature or sea ice when the full record is considered (Polyakov et al. 2002). Taken literally, this would imply no high-latitude amplification of climate warming. Consistent with this "oscillation," the Greenland warming of 1920–30 was apparently as large as that for the last decade, and the rate was actually 50% higher (Chylek et al. 2006). Furthermore, a significant portion of the observed warming is associated with the more positive phase of the Arctic Oscillation (AO)/North Atlantic Oscillation (NAO; Shindell et al. 1999), which may or may not be related to global warming. On the other hand, the recent record may actually underplay the ultimate response: high latitudes may well have critical points (nonlinearities) associated with cryospheric melting, after which their sensitivity could increase strongly, preventing past from being prologue.

Can we use paleoclimate data from equilibrium climate changes of greater magnitude, which presumably contain the nonlinearities, to determine the proper latitudinal response? Unfortunately, the

answer is a resounding no. Tropical climate sensitivity has been a big uncertainty from the paleoperspective ever since the CLIMAP (1981) production of the Last Glacial Maximum dataset. The Climate: Long-Range Investigation, Mapping, and Prediction's (CLIMAP's) significant finding was that the tropics were relatively invariant, with a cooling of some 1.5°C, compared with the high-latitude change of greater than 10°C (and in the vicinity of the continental ice sheets, much larger than that). Their data were driven by sea surface temperature reconstructions based on the location of forams, diatoms, and cocoliths; no change in the location relative to today was viewed as no change in sea surface temperature [the modern analogue technique (MAT)]. At the same time, they largely overlooked land evidence of tropical glacial and tree-line descent that implied strong cooling. In the GISS GCM either the two datasets were found to be incompatible, or the model was seriously flawed (Rind and Peteet 1985). Subsequent work by other modelers reached the same conclusion. Different ocean datasets (isotopes, alkenones, Mg/Ca ratios) now give somewhat different answers, but each has its own set of assumptions; no clear quantitative picture has emerged, there is just a general tendency to think that the land is somewhat less ambiguous, and so, the tropics probably had more significant cooling than was originally thought.

High latitudes were, of course, much different at the Last Glacial Maximum, but here, as well, the models produce quite different responses, in part due to different ice sheet elevation changes. Simulations for the Paleoclimate Modeling Intercomparison Project 2 (PMIP2) strongly underestimate the reconstructed glacial-to-interglacial Greenland temperature change, even when corrected for elevation changes (Masson-Delmotte et al. 2006). However, some of the models do perform better for Antarctica. Models also underestimate the Northern Hemisphere mid-Holocene high-latitude warming. Dynamical processes associated with storm tracks are thought to be a possible cause (Masson-Delmotte et al. 2006, op. cit.), although the observed temperature changes are derived from isotope–temperature relationships that have their own questions (Charles et al. 1994).

A similar uncertainty for both the tropics and high latitudes arises with the pre-Pleistocene climates. Here, when CO_2 was assumed to be higher, data reconstruction for most of the time and most of the tropics has generally shown little warming. On the other hand, paleodata invariably show large high-latitude warming, on the order of 10 times the tropical change (e.g., Dowsett et al. 1996). However, much of the tropical data are determined by the same method (MAT) as was used for the Last Glacial Maximum, so how good is it? We do not know, but some more recent data, suggesting a possible permanent El Niño–like state in the east Pacific (Wara et al. 2005) and the use of other data types (alkenone reconstructions; Haywood et al. 2005), are positing that tropical warming may have been underestimated. Complicating the issue is that for the best-explored time period, the mid-Pliocene (some 3 million years ago), the atmospheric CO_2 level is not estimated to have been much larger than today's (e.g., Solomon et al. 2007, their Fig. 6.1). What we do know is that no GCM can produce the archetypal pre-Pleistocene climate response of little tropical warming and large high-latitude warming, so if it really occurred, and was due to increased CO_2, then our model simulations are highly flawed. [A way out has been suggested by positing that the climate change is associated with increased ocean circulation (Rind and Chandler 1991; Raymo et al. 1996); this would work because the increased poleward heat transport melts sea ice, resulting in strong high-latitude warming, and at the same time removes heat from the tropics, limiting any temperature change there. However, there is no firm evidence that the ocean circulation change occurred.]

Thus, the current situation is one where we cannot constrain tropical or high-latitude sensitivity to within a factor of 2; this has some very large consequences for our ability to predict future climate change impacts. In the following sections, the causes of different model sensitivities are discussed, followed by some of the consequences of our uncertainty, starting with large-scale responses, and then turning to regional effects from the tropics to high latitudes.

INFLUENCES ON LOW- AND HIGH-LATITUDE SENSITIVITY. We may not know tropical sensitivity any better than in the past, but we do know more about why models are producing their different sensitivities. It is not the water vapor response, which is relatively consistent from model to model. Whether the water vapor response is realistic is hard to assess, because we have not had recent climate changes of the magnitude forecast for the rest of this century, and our observations for the upper troposphere in the tropics are subject to both the sparseness of data and changes of the humidity sensors being used. Nevertheless, considering the past 20 yr, the response of models is similar to the clear-sky trend detected in the High Resolution Infrared Radiometer Sounder

(HIRS; Soden et al. 2005). As stated in Solomon et al. (2007), there is no evidence currently that the model water vapor feedback is at variance with that of the real world, with both acting to either (approximately) conserve relative humidity or to produce slight reductions in response to warmer sea surface temperatures (e.g., Minschwaner et al. 2006).

However, the water vapor feedback in absolute magnitude will vary with the temperature response from the Clausius–Clapeyron relationship, and thus is subject to the magnitude of the other feedbacks. In the tropics the prime difference between models relates to the disparate cloud cover changes. It would appear that the largest difference relates to low cloud cover changes (Webb et al. 2006). Boundary layer clouds today have a strong negative cloud radiative forcing and cover a large area of the tropics. We know these clouds are related to low-level stability, but they also are affected by shallow convection, precipitation processes, and changes in the capping inversion height. The latter two effects may be why ISCCP data indicate reduced cloud optical thickness as temperature warms (Tselioudis et al. 1998). Furthermore, boundary layer clouds may be subject to remote influences, like change in deep convection and the resulting humidity of subsidence regions (Larson and Klein 1999). All of these features, as well as the cloud cover amount in the control simulation, presumably come into play when determining a model's particular low-level cloud cover response.

High-level cloud cover in the tropics with significant optical thickness is predominantly associated with deep convection, and clouds of this nature play the second most important role in accounting for model differences (Webb et al. 2006). Today, these clouds have their long- and shortwave cloud radiative forcings nearly canceling out (Ramanathan et al. 1989), but it is not clear that this has to be the case, and variations occur in the different model climate change simulations. One suggestion is that a warmer climate would have increased precipitation efficiency, causing more moisture to rain out, with less detrainment and a smaller area of upper-level cloud cover, limiting the positive longwave forcing (Lindzen et al. 2001). The evidence for this effect is not definitive, and it certainly has not limited the tropical warming to this point in time. Deep convection and cloud microphysics, including cirrus anvil detrainments, are poorly parameterized processes in climate models. When cloud optical thickness is greater than about 5, shortwave cooling predominates, but the optical thickness will vary with distance from the convective plume, making it a difficult problem with subgrid-scale parameterizations of convection. Another issue affecting the future longwave response relates to the temperature of the tropical anvil clouds and whether it will remain unchanged as climate changes; for this to happen, the altitude would have to rise, perhaps associated with the location at which the clear-sky longwave radiative cooling rapidly declines with height (Hartmann and Larson 2002). Some component of this will be affected by the vertical resolution in different models, but more generally, anything as complicated as the amount of moisture carried aloft by convection will be suspect (and varying) in models.

Finally, there is the issue of how dynamical changes in the tropics will affect cloud cover. The models run for the IPCC AR4 assessment show a reduction in the Walker circulation (Vecchi and Soden 2007) and a poleward expansion of the Hadley circulation (Lu et al. 2007) in future climate change experiments. Both effects can be related to stability increases and/or decreased horizontal temperature gradients in a warmer atmosphere, and could alter cloud cover and its tropical radiative forcing. These predicted trends are in contrast to a strengthening of both circulation cells seen over the last decade (Chen et al. 2002; Mitas and Clement 2005). The strengthening itself did not seem to have much effect on the tropical mean radiation budget (Clement and Soden 2005), which was also true in idealized warming experiments (Bony et al. 2006). It is not clear that that will be true in all models, or in the real world with significant enough climate perturbations, especially if they result in a change in the ratio of moist convective and subsidence areas in the tropics (Pierrehumbert 1995; Larson and Klein 1999).

The physics/dynamics associated with both low and high cloud radiative forcing is thus still an active area of research. Because their radiative response to greenhouse gas increases strongly governs the tropical warming in models, it is not surprising that forecast warming varies. To put the model variation in perspective, in the coupled atmosphere–ocean models run for AR4, there is approximately a 50% probability that the tropical ocean warming will exceed 2°C in the last 20 yr of this century with the A1B trace-gas release scenario (Solomon et al. 2007) (i.e., a coin flip).

At high latitudes, several different factors come into play, none of which are well known. High latitudes, of course, could be warmer in one model projection than another because the tropics and midlatitudes are warmer, with this air advected into the polar regions. However, even when normalizing for

the global mean temperature change, models show up to a factor of 4 variation in the high-latitude Northern Hemisphere response for the end of this century compared to the end of the last one, using the A1B scenario (Bony et al. 2006). All told, with respect to the global mean temperature change the high-latitude amplification varied from 1.3 to a factor of 4 (Bony et al. 2006 op. cit.).

The most obvious difference with respect to the tropics is that here the cryospheric feedback needs to be understood, both with respect to sea ice and snow cover. Reduction in the surface cryospheric parameters, and the corresponding reduction in surface albedo, is the primary reason for the high-latitude amplification of surface air temperature change; in the GISS model when sea ice changes were not allowed, the high-latitude amplification went away entirely (Rind et al. 1995). Seasonally, when sea ice is absent in summer, more solar radiation is absorbed, and the stored heat comes back into the atmosphere with reduced sea ice in winter, producing maximum warming in that season.

In current models, approximately half the albedo effect of the shrinking cryosphere as climate warms in the Northern Hemisphere is associated with snow retreat, and half is due to sea ice reduction (Winton 2006). The primary issue that causes divergence among the models concerning snow is how the surface albedo varies with temperature, with a factor-of-3 difference among models over both North America and Eurasia (Bony et al. 2006). Other factors than just snowmelt enter in, such as snow aging and the effect of impurities. The Coupled Model Intercomparison Project 2 (CMIP2) results showed that model simulations of sea ice were quite poor and that sea ice differences among models relate more to the atmosphere and ocean circulation than to sea ice physics (Flato 2004). In the Northern Hemisphere, the climate change response is strongly related to the sea ice extent, and especially thickness in the control run (thinner ice is easier to remove) (Rind et al. 1997; Holland and Bitz 2003).

Additional factors influence the magnitude of polar amplification in models. Most models predict an increase in low cloud cover as climate warms, but some produce high cloud cover increases as well (Holland and Bitz 2003; Webb et al. 2006). Changes in atmospheric transport also occur; models with greater tropical warming have an increased chance of producing greater poleward transport of latent heat, amplifying warming in the Arctic (Rind 1987; Alexeev et al. 2005). Francis and Hunter (2006), assessing sea ice edge changes in recent decades, found that the biggest driving force in the 1990s was an increased downward longwave energy flux, associated with greater water vapor (from increased moisture transport from lower latitudes, and evaporation from ice-free areas). Increased cloud cover and cloud liquid-water content can also contribute to greater downward longwave forcing. High-latitude amplification is furthermore significantly correlated to both the control climate ocean heat transport and the change in transport simulated by the model at high latitudes (Holland and Bitz 2003; Cai 2006). Considering all of these mechanisms, the range of model parameterizations potentially influencing the high-latitude response is quite large.

Given that the primary reasons for our uncertainty in both low- and high-latitude sensitivity is largely associated with model physics, it makes the problem that much harder to reduce by averaging. It has been suggested that we can have greater confidence in the multimodel mean changes than in that of any individual model for climate change assessments, which seems to be the case in weather forecasting experiments and simulations driven by specified SSTs. Considerable work has been done in deriving weights for the different models to provide the optimum forecast (e.g., Krishnamurti et al. 1999). However, in both of those situations the response is dominated by atmospheric dynamics, even when the physics are varied (e.g., with specified SSTs, the forcing is largely provided). The various models are all attempting to solve the same basic dynamical equations, albeit in different forms, and it is understandable that their errors can be minimized when different attempts are averaged together. For climate-change simulations, given their different physical parameterizations, models are not solving the same equations. The formulations for low and high clouds, or snow albedo change with temperature, vary. It is doubtful that averaging different formulations together will end up giving the "right" result, especially because we have no way of knowing whether the various choices that have been made even circumscribe the proper sensitivity. As noted by Wang (2005), an individual model may provide results that differ from the mean model response because it includes either an improved parameterization or a missing mechanism. Because the cloud cover and cryospheric feedbacks are providing a substantial part of the net forcing, in effect the models are being given a different forcing distribution. The model responses (e.g., tropical land precipitation) can often be of different signs, and there can be little confidence that averaging them together will produce a better result.

At this point we cannot determine the low- and high-latitude sensitivities, and we have no real way of obtaining them. What we can do is assess the consequences of not knowing, which is the subject of the next two sections.

LARGE-SCALE RESPONSES TO LOW- AND HIGH-LATITUDE CLIMATE SENSITIVITY. Our ignorance concerning the prospective low- and high-latitude response impacts our ability to forecast important aspects of climate change, on both the large and regional scales. Over the years, we have examined the large-scale response to this uncertainty in a number of different modeling studies, and a review of some of the main points is provided here. These large-scale effects provide the backdrop against which more regional changes may occur, as discussed in the next section.

The radiative and thermodynamic responses to tropical and high-latitude sensitivity are fairly obvious. The tropical response dominates the magnitude of the water vapor feedback and cloud feedback everywhere but at high latitudes (e.g., Rind 1987, 1998). High-latitude sensitivity is dominated by the cryospheric feedback, which affects cloud feedback locally, and, via its influence on global warming, the water vapor and cloud response at all latitudes (e.g., Rind et al. 1995). Given the interactive nature of these feedbacks, separate assessments for each of them are somewhat misleading, but still have some utility. For example, the cryospheric and water vapor feedbacks can have different relative sensitivities in different climate regimes. Increasing poleward ocean heat transports in nonglacial climate reduces sea ice and warms the globe, with extreme high-latitude amplification. Sea ice removal is associated with large high-latitude warming, while the increased ocean heat transport increase minimizes tropical warming [Rind and Chandler (1991); again a possible analogue for pre-Pleistocene climates]. In this case the cryospheric forcing is much stronger than the water vapor forcing (which is minimal with the lack of tropical temperature change). In contrast, increasing ocean heat transports in the ice-age climate results in colder global conditions; it has little effect on high-latitude sea ice (because the background climate is too cold for it to make a difference), but acts to substantially cool the tropics (Webb et al. 1997). Thus, here the water vapor feedback is stronger than the cryospheric response. The relative importance of these two feedbacks for the current global warming depends on, and is interactive with, low- and high-latitude climate sensitivity.

With respect to atmospheric dynamics, the impact of the low- and high-latitude climate response is substantial. In general, atmospheric dynamics changes are related to changes in the latitudinal temperature gradient, much more so than changes of the mean temperature (Rind 1998). (The one main exception involves the amplifying effect of water vapor condensation, whose influence is obviously a function of the water vapor amount, and hence the temperature.) Of primary interest in the tropics and subtropics is what will happen to the strength of the Hadley cell (and the reference here is not to the absolute magnitude of the peak streamfunction, which depends on where the peak warming is relative to where the maximum was previously located, but to the circulation as a whole, with rising air in the tropics and sinking in the subtropics). With a stronger effective Hadley circulation, there is more rain in the tropics and less in the subtropics. As the sea surface temperature gradient between the tropics and lower midlatitudes increases, the Hadley cell intensifies (Rind and Rossow 1984; Rind 1998; Rind and Perlwitz 2004). The CLIMAP (1981) sea surface temperature reconstruction featured just such a latitudinal gradient increase, and it resulted in a stronger Hadley cell in our model despite the fact that there was less water vapor in the atmosphere (Rind 1986). When we simulated the ice-age climate using current ocean heat transports, the tropics cooled substantially, reducing this SST gradient, and the Hadley cell decreased in intensity, especially in June (Webb et al. 1997; Rind and Perlwitz 2004).

The latitudinal extent of the Hadley cell determines where the subsidence and drying occurs. It seems to be governed both by increased stability associated with warmer temperatures (see also Rind 1998) and by other midlatitude eddy processes (both heat and momentum transport convergences) that are affected by extratropical latitudinal temperature gradients. This latter influence is situation-specific, because it depends on the exact latitudes at which the change in eddy energy and momentum transport convergences occur (Rind and Rossow 1984; Rind 1987; Rind and Perlwitz 2004). Without knowing the high-latitude amplification, this aspect is impossible to predict.

In the extratropics, various eddy features (e.g., low- and high-pressure centers) can be discussed in terms of latitudinal gradients. Changes in eddy kinetic energy are related to changes in the vertically integrated temperature gradient (and, hence, zonal available potential energy). The ratio between transient and stationary eddy energy in our newer models is not associated with gradient changes

(Rind 1998) because both seem to respond in the same way; that is, greater latitudinal gradient results in greater baroclinic generation of transient eddies, and a greater tropical warming could be producing extratropical stationary eddies through vorticity generation of stationary Rossby waves (Inatsu et al. 2002) and increased zonal flow around mountains. Furthermore, there are interactions between standing and transient wave energy transports and their effect on the latitudinal temperature gradient (Rind et al. 2001a). This result is relevant for the discussion in the regional section, when both standing and transient waves are suggested as mediators of tropical–extratropical interactions. The biggest change in the eddy kinetic energy spectrum occurs with substantial tropical warming, which amplifies the longest planetary waves (Rind et al. 2002) and is important for driving the stratospheric circulation.

Stronger tropical warming enhances the subtropical jet stream, tending to divert winter storms to that latitude; a weakening of this tropical/lower midlatitude gradient is associated with a more poleward movement of the effective jet and storm activity (Rind et al. 2001b), which occurs in summer and La Niña situations. In the extratropics, an increase in stability, which occurs with warmer climates, is thought to shift the storm track poleward, but whether it happens depends on the latitudinal temperature gradient. With sufficient latitudinal gradient changes, the effect can be counteracted (Yin 2005), and may reverse what is already occurring (Fu et al. 2006). It is not yet clear whether this poleward shift is happening in the AR4 model future simulations, because different researchers are getting different results (Yin 2005; Bengtsson et al. 2006; Lambert and Fyfe 2006). The change in the storm track in any specific region will be affected by longitude-specific gradient changes, and thus could very well be model dependent. Effects associated with localized gradient changes are discussed in the regional section.

It might be thought that transports depend on low- and high-latitude sensitivity, but many studies have found a strong compensation between changes in atmospheric and ocean energy transports. Simulations with the GISS model largely verify this compensation, although there is a tendency for total energy transports to increase slightly when the tropics warm substantially, due to increased latent heat transport (Rind 1998). Large tropical warming does lead to increased interhemispheric tracer transport associated with the tropical mean circulation change (Rind et al. 2002). Furthermore, if large surface tropical warming is successfully translated into the upper troposphere, stratosphere–tropospheric exchange is greater; planetary wave refraction from the midlatitudes to tropics is enhanced, and stronger Eliasson–Palm flux convergences drive an accelerated residual circulation (Rind et al. 1998, 2002).

The influence of low- and high-latitude sensitivity on the phase of the prime mode of variability in the extratropics, the Northern Annular Mode (NAM; also called the Arctic Oscillation; Thompson and Wallace 1998, 2000), is multifaceted. The positive phase of this mode has lower pressure at higher latitudes and higher pressure in midlatitudes, which is emblematic of particular storm-track paths. Again, if substantial tropical warming occurs in the upper troposphere, the resulting equatorward planetary wave refraction is consistent with poleward momentum transport, producing a more positive NAM phase (e.g., Rind et al. 2005a). However, low-altitude latitudinal temperature gradient changes also play a role; with a strong reduction in this temperature gradient at mid- to high latitudes, the change with height of poleward heat transport is reduced, increasing the poleward quasigeostrophic potential vorticity flux, thus also producing a more positive NAM (Rind et al. 2005b.). Therefore, both stronger tropical warming (transmitted aloft) and stronger high-latitude warming at low altitudes can result in positive NAM phases (at least in the sea level pressure field). Finally, the high-latitude response by itself, for example, changes in snow (Cohen et al. 2001, 2002) or sea ice (Rind et al. 2005b), can influence the NAM phase, at least at low levels in the sea level pressure field, with more unstable conditions leading to a more positive phase. Whether the response is similar (quasibarotropic) aloft depends on which of these factors influences the wind field at higher levels. If strong tropical upper-tropospheric warming occurs, it leads to a more positive NAM in the lower stratosphere (Rind et al. 2005a), which can delay the onset of the Northern Hemisphere ozone hole recovery by a decade (Shindell et al. 1998).

An assessment of the impact of low- and high-latitude sensitivity on regional climates around the world is presented as an electronic supplement (online at DOI:10.1175/2007BAMS2520.2). It emphasizes the importance of not only the magnitude of the response, but also its longitudinal pattern, particularly the future changes in the tropical Pacific, about which there is currently no consensus (e.g., Barsugli et al. 2006; Merryfield 2006).

CONCLUDING REMARKS. As noted in the introduction, over the past 25 yr we have not been able to quantitatively improve our understanding

of low- and high-latitude (or even global) climate sensitivity. That does not mean we have not learned many things; we are more knowledgeable about why models are getting different responses in various locations, and, as the preceding discussion has shown, we are in a position to better understand the consequences of not knowing these sensitivities. However, at this point the uncertainties in latitudinal temperature gradient changes affect the confidence we can have in many of our projections of atmospheric dynamic and hydrologic responses to global warming.

The question then is as follows: When will we get a better handle on this issue, and therefore be able to make more confident regional projections?

Real progress will be the result of continued and newer observations along with modeling improvements based on these observations. Concerning clouds in the tropics, new satellite observations that can resolve cloud vertical structure are becoming available (e.g., CloudSat and Calipso; Stephens et al. 2002). Various groups are focused on understanding cloud parameterizations [e.g., Atmospheric Radiation Measurement (ARM); Global Energy and Water Cycle Experiment (GEWEX) Cloud System Study (Randall et al. 2003)]. This year represents the International Polar Year (IPY) organized through the International Council for Science and the World Meteorological Organization; it is focused on improving our understanding of polar processes in both the Arctic and Antarctic. Now there are many more modeling groups active, which increases the manpower focused on these issues. Computer power is increasing, allowing for finer-resolution simulations, with more focused on the scale at which some of these processes work (although they are still too coarse for others).

Nevertheless, forecasting even the large-scale response to climate change is not easy given the current uncertainties, and the regional discussion in the supplementary section highlights the fact that regional responses may be the end result of varying influences in part due to warming in different tropical and high-latitude regions. There is no guarantee that these issues will be resolved before a substantial global warming impact is upon us. How we proceed to act in an environment of uncertainty will, perhaps, become as great a challenge as dealing with global warming itself.

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REFERENCES

Alexeev, V. A., P. L. Langen, and J. R. Bates, 2005: Polar amplification of surface warming on an aquaplanet in "ghost forcing" experiments without sea ice feedbacks. *Climate Dyn.,* **24,** 655–666.

Barsugli, J. J., S.-I. Shin, and P. D. Sardeshmukh, 2006: Sensitivity of global warming to the pattern of torpical ocean warming. *Climate Dyn.,* **26,** 483–492.

Bengtsson, L., K. I. Hodges, and E. Roeckner, 2006: Storm tracks and climate change. *J. Climate,* **19,** 3518–3543.

Bony, S., J.-L. Dufresne, H. LeTreut, J.-J. Morcrette, and C. Senior, 2004: On dynamic and thermodynamic components of cloud changes. *Climate Dyn.,* **22,** 71–86.

——, and Coauthors, 2006: How well do we understand and evaluate climate change feedback processes? *J. Climate,* **19,** 3445–3482.

Cai, M., 2006: Dynamical greenhouse-plus feedback and polar warming amplification. Part I: A dry radiative-transportive climate model. *Climate Dyn.,* **26,** 661–675.

Charles, C. D., D. Rind, J. Jouzel, R. Koster, and R. G. Fairbanks, 1994: Glacial-interglacial changes in moisture sources for Greenland: Influences on the ice core record. *Science,* **263,** 508–510.

Charney, J. G., and Coauthors, 1979: *Carbon Dioxide and Climate: A Scientific Assessment.* National Academy of Sciences, 22 pp.

Chen, J., B. E. Carlson, and A. D. DelGenio, 2002: Evidence for strengthening of the tropical generation circulation in the 1990s. *Science,* **295,** 838–841.

Chylek, P., M. K. Dubey, and G. Lesins, 2006: Greenland warming of 1920–1930 and 1995–2005. *Geophys. Res. Lett.,* **33,** L11707, doi:10.1029/2006GL026510.

Clement, A. C., and B. Soden, 2005: The sensitivity of the tropical-mean radiation budget. *J. Climate,* **18,** 3189–3203.

CLIMAP, 1981: *Seasonal Reconstruction of the Earth's Surface at the Last Glacial Maximum.* Map and Chart Series Vol. 36, Geological Society of America.

Cohen, J., K. Saito, and D. Entekhabi, 2001: The role of the Siberian high in Northern Hemisphere climate variability. *Geophys. Res. Lett.,* **28,** 299–302.

——, D. Salstein, and K. Saito, 2002: A dynamical framework to understand and predict the major Northern Hemisphere mode. *Geophys. Res. Lett.,* **29,** 1412, doi:10.1029/2001GL014117.

Dowsett, H., J. Barron, and R. Poore, 1996: Middle Pliocene sea surface temperatures: A global reconstruction. *Mar. Micropaleontol.,* **27,** 13–25.

Flato, G., 2004: Sea-ice and its response to CO_2 forcing as simulated by global climate models. *Climate Dyn.,* **23,** 229–241.

Francis, J. A., and E. Hunter, 2006: New insight into the disappearing Arctic sea ice. *Eos, Trans. Amer. Geophys. Union,* **87,** 509–511.

Fu, Q., C. M. Johanson, J. M. Wallace, and T. Reichler, 2006: Enhanced mid-latitude tropospheric warming in satellite measurements. *Science,* **312,** 1179.

Hartmann, D. L., and K. Larson, 2002: An important constraint on tropical cloud-climate feedback. *Geophys. Res. Lett.,* **29,** 1951, doi:10.1029/2002GL015835.

Haywood, A. M., P. Dekends, A. C. Ravelo, and M. Williams, 2005: Warmer tropics during the mid-Pliocene? Evidence from alkenone paleothermometry and a fully coupled ocean-atmosphere GCM. *Geochem. Geophys. Geosys.,* **6,** Q03010, doi:10.1029/2004GC000799.

Holland, M. M., and C. M. Bitz, 2003: Polar amplification of climate change in coupled models. *Climate Dyn.,* **21,** 221–232.

Inatsu, M., H. Mukougawa, and S-P. Xie, 2002: Tropical and extratropical SST effects on the midlatitude storm track. *J. Meteor. Soc. Japan,* **80,** 1069–1076.

Krishnamurti, T. N., C. M. Kishtawal, T. E. LaRow, D. R. Bachiochi, Z. Zhang, C. E. Williford, S. Gadgil, and S. Surendran, 1999: Improved weather and seasonal climate forecasts from multimodel superensemble. *Science,* **285,** 1548–1550.

Lambert, S. J. and J. C. Fyfe, 2006: Changes in winter cyclone frequencies and strengths simulated in enhanced greenhouse warming experiments: results from the models participating in the IPCC diagnostic exercise. *Climate Dyn.,* **26,** 713–728.

Larson, K., and S. A. Klein, 1999: The role of clouds, water vapor, circulation, and boundary layer structure in the sensitivity of the tropical climate. *J. Climate,* **12,** 2357–2374.

Lindzen, R. S., M.-D. Chou, and A. Y. Hou, 2001: Does the Earth have an adaptive infrared iris? *Bull. Amer. Meteor. Soc.,* **82,** 417–432.

Lu, J., G. A. Vecchi, and T. Reichler, 2007: Expansion of the Hadley cell under global warming. *Geophys. Res. Lett.,* **34,** L06805, doi:10.1029/2006GL028443.

Masson-Delmotte, V., and Coauthors, 2006: Past and future polar amplification of climate change: Climate model intercomparisons and ice-core constraints. *Climate Dyn.,* **26,** 513–529.

Merryfield, W., 2006: Changes to ENSO under CO_2 doubling in a multimodel ensemble. *J. Climate,* **19,** 4009–4027.

Minschwaner, K., A. E. Dessler, and P. Sawaengphokhai, 2006: Multimodel analysis of the water vapor feedback in the tropical upper troposphere. *J. Climate,* **19,** 5455–5464.

Mitas, C. M., and A. Clement, 2005: Has the Hadley cell been strengthening in recent decades? *Geophys. Res. Lett.,* **32,** L03809, doi:10.1029/2004GL021765.

Pierrehumbert, R. T., 1995: Thermostats, radiator fins, and the local runaway greenhouse. *J. Atmos. Sci.,* **52,** 4537–4540.

Polyakov, I. V., and Coauthors, 2002: Observationally based assessment of polar amplification of global warming. *Geophys. Res. Lett.,* **29,** 1878, doi:10.1029/2001GL011111.

Ramanathan, V., R. D. Cess, E. F. Harrison, P. Minnis, B. R. Barkstrom, and D. L. Hartmann, 1989: Cloud-radiative forcing and climate: Results from the Earth Radiation Budget Experiment. *Science,* **243,** 57–63.

Randall, D., and Coauthors, 2003: Confronting models with data: The GEWEX Cloud Systems Study. *Bull. Amer. Meteor. Soc.,* **84,** 455–469.

Raymo, M. E., B. Grant, M. Horowitz, and G. H. Rau, 1996: Mid-Pliocene warmth: Stronger greenhouse and stronger conveyor. *Mar. Micropaleontol.,* **27,** 313–326.

Rind, D., 1986: The dynamics of warm and cold climates. *J. Atmos. Sci.,* **43,** 3–24.

——, 1987: The doubled CO_2 climate: Impact of the sea surface temperature gradient. *J. Atmos. Sci.,* **44,** 3235–3268.

——, 1998: Latitudinal temperature gradient and climate change. *J. Geophys. Res.,* **103,** 5943–5971.

——, and W. Rossow, 1984: The effects of physical processes on the Hadley circulation. *J. Atmos. Sci.,* **41,** 479–507.

——, and D. Peteet, 1985: Terrestrial conditions at the last glacial maximum and CLIMAP sea-surface temperature estimates: Are they consistent? *Quat. Res.,* **24,** 1–22.

——, and M. Chandler, 1991: Increased ocean heat transports and warmer climate. *J. Geophys. Res.,* **96,** 7436–7461.

——, and J. Perlwitz, 2004: The response of the Hadley circulation to climate changes, past and future. *The Hadley Circulation: Present, Past and Future*, H. F. Diaz and R. S. Bradley, Eds., Advances in Global Change Research, Vol. 21, Springer, 511 pp.

——, R. Goldbert, J. Hansen, C. Rosenzweig, and R. Ruedy, 1990: Potential evapotranspiration and the

likelihood of future drought. *J. Geophys. Res.,* **95,** 9983–10 004.

——, R. Healy, C. Parkinson, and D. Martinson, 1995: The role of sea ice in 2 × CO_2 climate model sensitivity. Part I: The total influence of sea ice thickness and extent. *J. Climate,* **8,** 449–463.

——, ——, ——, and ——, 1997: The role of sea ice in 2 × CO_2 climate model sensitivity: Part II: Hemisphere dependence of sea ice thickness and extent. *Geophys. Res. Lett.,* **24,** 1491–1494.

——, D. Shindell, P. Lonergan, and N. K. Balachandran, 1998: Climate change and the middle atmosphere. Part III: The doubled CO_2 climate revisited. *J. Climate,* **11,** 876–894.

——, P. Lonergan, J. Lerner, and M. Chandler, 2001a: Climate change in the Middle Atmosphere: Part V: The paleostratosphere in warm and cold climates. *J. Geophys. Res.,* **106,** 20 195–20 212.

——, M. Chandler, J. Lerner, D. G. Martinson, and X. Yuan, 2001b: Climate response to basin-specific changes in latitudinal temperature gradients and implications for sea ice variability. *J. Geophys. Res.,* **106,** 20 161–20 173.

——, J. Lerner, J. Perlwitz, C. McLinden, and M. Prather, 2002: Sensitivity of tracer transports and stratospheric ozone to sea surface temperature patterns in the doubled CO_2 climate. *J. Geophys. Res.,* **107,** 4800, doi:10.1029/2002JD002483.

——, ——, and P. Lonergan, 2005a: AO/NAO response to climate change: 1. Respective influences of stratospheric and tropospheric climate changes. *J. Geophys. Res.,* **110,** D12107, doi:10.1029/2004JD005103.

——, ——, ——, and J. Lerner, 2005b: AO/NAO response to climate change: 2. Relative importance of low- and high-latitude temperature changes. *J. Geophys. Res.,* **110,** D12108, doi:10.1029/2004JD005686.

Shindell, D., D. Rind, and P. Lonergan, 1998: Increased polar stratospheric ozone losses and delayed eventual recovery owing to increasing greenhouse-gas concentrations. *Nature,* **392,** 589–592.

——, R. L. Miller, G. Schmidt, and G. Pandolfo, 1999: Simulations of recent northern winter climate trends by greenhouse-gas forcing. *Nature,* **399,** 452–455.

Smagorinsky, J., and Coauthors, 1982: *Carbon Dioxide and Climate: A Second Assessment.* National Academy of Sciences, 72 pp.

Soden, B. J., D. L. Jackson, V. Ramaswamy, M. D. Schwarzkopf, and X. Huang, 2005: The radiative signature of upper tropospheric moistening. *Science,* **310,** 841–844.

Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K. B. Avery, M. Tignor, and H. L. Miller, Eds., 2007: *Climate Change 2007: The Physical Science Basis.* Cambridge University Press, 996 pp.

Stephens, G. L., and Coauthors, 2002: The CloudSat mission and the A-Train: A new dimension of space-based observations of clouds and precipitation. *Bull. Amer. Meteor. Soc.,* **83,** 1771–1790.

Thompson, D. W. J., and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.,* **25,** 1297–1300.

——, and ——, 2000: Annular modes in the extratropical circulation. Part I: Month-to-month variability. *J. Climate,* **13,** 1000–1016.

Tselioudis, G., A. Del Genio, W. Kovari Jr., and M. S. Yao, 1998: Temperature dependence of low cloud optical thickness in the GISS GCM: Contributing mechanisms and climate implications. *J. Climate,* **11,** 3268–3281.

Vecchi, G. A., and B. J. Soden, 2007: Global warming and the weakening of the tropical circulation. *J. Climate,* **20,** 4316–4340.

Wang, G., 2005: Agricultural drought in a future climate: Results from 15 global climate models participating in the IPCC 4th assessment. *Climate Dyn.,* **25,** 739–753.

Wara, M., A. C. Ravelo, and M. L. Delaney, 2005: Permanent El Niño conditions during the Pliocene warm period. *Science,* **309,** 758–761.

Webb, R. S., S. J. Lehman, D. Rind, R. Healy, and D. Sigman, 1997: Influence of ocean heat transport on the climate of the Last Glacial Maximum. *Nature,* **385,** 695–699.

Webb, M. J., and Coauthors, 2006: On the contribution of local feedback mechanisms to the range of climate sensitivity in two GCM ensembles. *Climate Dyn.,* **27,** 17–38.

Winton, M., 2006: Surface albedo feedback estimates for the AR4 climate models. *J. Climate,* **19,** 359–365.

Yin, J. H., 2005: A consistent poleward shift of the storm tracks in the simulations of 21st century climate. *Geophys. Res. Lett.,* **32,** L18701, doi:10.1029/2005GL023684.