

Future climate change in the Southern Hemisphere: Competing effects of ozone and greenhouse gases

J. M. Arblaster,^{1,2,3} G. A. Meehl,¹ and D. J. Karoly³

Received 7 September 2010; revised 14 October 2010; accepted 19 October 2010; published 19 January 2011.

[1] Future anthropogenic climate change in the Southern Hemisphere is likely to be driven by two opposing effects, stratospheric ozone recovery and increasing greenhouse gases. We examine simulations from two coupled climate models in which the details of these two forcings are known. While both models suggest that recent positive summertime trends in the Southern Annular Mode (SAM) will reverse sign over the coming decades as the ozone hole recovers, climate sensitivity appears to play a large role in modifying the strength of their SAM response. Similar relationships are found between climate sensitivity and SAM trends when the analysis is extended to transient CO₂ simulations from other coupled models. Tropical upper tropospheric warming is found to be more relevant than polar stratospheric cooling to the intermodel variation in the SAM trends in CO₂-only simulations. **Citation:** Arblaster, J. M., G. A. Meehl, and D. J. Karoly (2011), Future climate change in the Southern Hemisphere: Competing effects of ozone and greenhouse gases, *Geophys. Res. Lett.*, 38, L02701, doi:10.1029/2010GL045384.

1. Introduction

[2] A consistent response to increasing greenhouse gases (GHGs) in climate model experiments is the change in the Southern Hemisphere (SH) extratropical atmospheric circulation [e.g., *Kushner et al.*, 2001]. The response includes a poleward shift in the SH storm tracks and a strengthening of the polar vortex and has been documented extensively in coupled climate model experiments, including those that participated in the Coupled Model Intercomparison Project Version 3 (CMIP3) [*Yin*, 2005; *Miller et al.*, 2006; *Arblaster and Meehl*, 2006]. These future changes project onto the Southern Annular Mode (SAM), the leading mode of variability in the SH extratropical circulation [e.g., *Rogers and van Loon*, 1982]. A trend in the SAM towards its positive polarity, which indicates a poleward shift in the westerly jet and corresponding increases in mean sea level pressure over SH mid-latitudes and decreases over Antarctica, is found under all future emission scenarios to the end of the 21st Century [*Miller et al.*, 2006; *Arblaster and Meehl*, 2006].

[3] Stratospheric ozone depletion impacts the SH extratropical circulation in a similar way to GHGs, however the tropospheric response is mostly restricted to austral summer [*Thompson and Solomon*, 2002]. With the ozone hole projected to recover over this century [*Stratospheric Processes and their Role in Climate (SPARC) CCMVal*, 2010], ozone and GHG forcing will no longer combine to produce strong positive summertime SAM trends, but instead will oppose each other with ozone recovery leading to negative SAM trends and GHGs continuing to shift the SAM towards positive values. Projections of the SAM in summer in CMIP3 class models suggest that GHG forcing is dominant, with the positive multi-model trends that began in the 20th Century continuing into the future [*Miller et al.*, 2006]. However, *Son et al.* [2008] and *Perlwitz et al.* [2008] examined SAM trends in coupled chemistry-climate models (CCMVal) and found the opposite response, with the dominance of ozone recovery leading to a reversal of the recent positive summertime SAM trends in projections to 2050.

[4] As both CMIP3 and CCMVal models can reproduce past trends in the SAM [*Miller et al.*, 2006; *Son et al.*, 2008], it is difficult to ascertain which modeling framework provides more plausible projections of SAM in the future. While chemistry-climate models include a more sophisticated treatment of stratospheric processes and extend higher in the atmosphere, they are typically not coupled to an ocean model but instead forced with sea surface temperatures from a CMIP3 simulation. CMIP3 models, on the other hand, are fully coupled but do not extend as high in the atmosphere and have coarse resolution in the stratosphere, both of which may impact the way the troposphere responds to external forcing [*Son et al.*, 2008; *Shaw et al.*, 2009].

[5] Understanding the differences between these two classes of models is complicated not only by their different modeling frameworks, but also by differences in the forcing that drives the experiments. While both use the Special Report for Emissions Scenarios (SRES) A1B scenario for GHG and aerosol forcings, the ozone forcing in CCMVal models is determined by chlorofluorocarbon emissions and by the dynamics and chemistry of the models while in CMIP3 the ozone is prescribed in a zonally symmetric manner [*Crook et al.*, 2008; *Waugh et al.*, 2009]. Even amongst the CMIP3 models a number of prescribed ozone datasets and sometimes fixed ozone were used [*Miller et al.*, 2006]. As the ozone forcing datasets are not readily available for comparison, isolating the cause of each model's radiative and dynamical responses to ozone depletion and recovery is difficult.

[6] Here we look at two CMIP3 class models for which the forcing for each simulation can be quantified and seek

¹NCAR Earth System Laboratory, National Center for Atmospheric Research, Boulder, Colorado, USA.

²Centre for Australian Weather and Climate Research, Bureau of Meteorology, Melbourne, Victoria, Australia.

³School of Earth Sciences, University of Melbourne, Melbourne, Victoria, Australia.

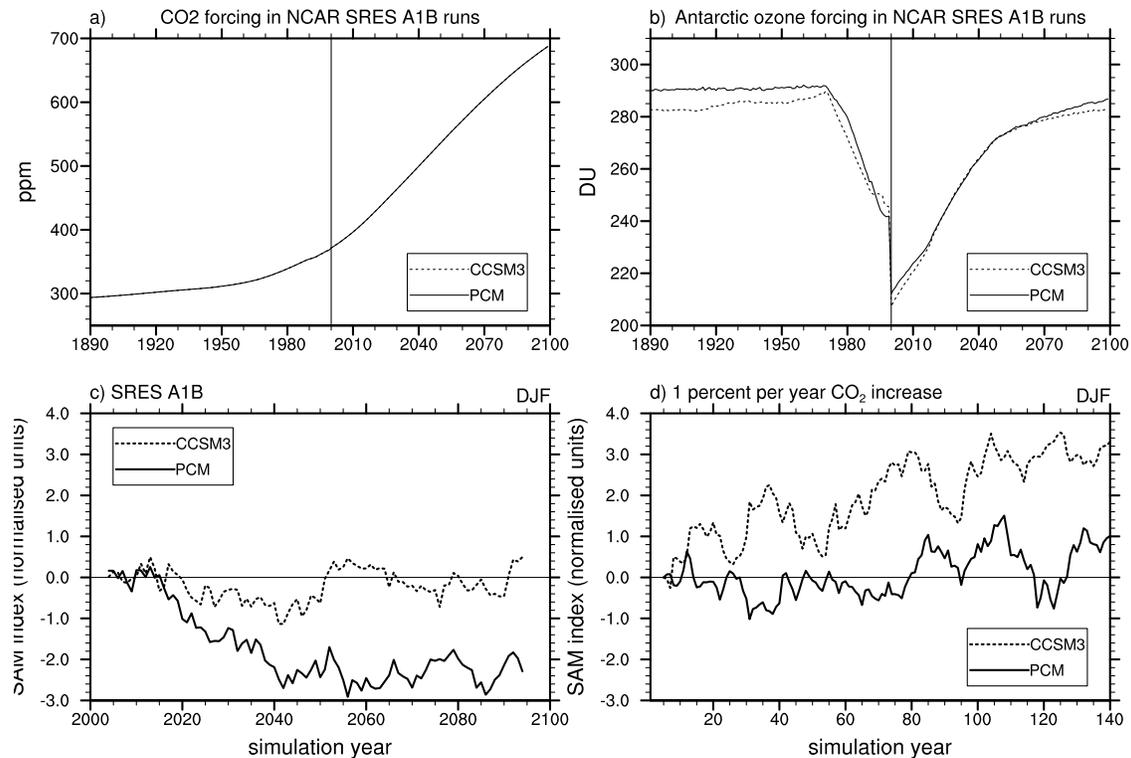


Figure 1. Time series of (a) CO₂ concentrations (ppm) and (b) annual mean total column ozone (Dobson Units) over the Antarctic polar cap (all gridpoints south of 70°S) from 1870–2100 and DJF SAM index from the (c) SRES A1B scenario (ensemble mean) and (d) 1% per year CO₂ increase (to quadrupling) simulations in PCM (solid) and CCSM3 (dashed). The SAM time series are smoothed with a 10 year running mean. Note the discontinuity at 2000 in the ozone series is the point at which the observed forcing ends and the scenario forcing begins.

an understanding of the mechanisms surrounding their responses to future stratospheric ozone and GHG forcing. The NCAR PCM [Washington *et al.*, 2000] and NCAR CCSM3 [Collins *et al.*, 2006] are both fully coupled models of the atmosphere–ocean–land–sea–ice system and contributed to the CMIP3 archive. The CCSM3 is a later generation NCAR model with higher horizontal resolution compared to the PCM and improvements in all components. Arblaster and Meehl [2006] documented recent changes in the SH tropospheric circulation in the PCM model, finding that observed changes in the SAM were reproduced when all forcings were combined, with single forcing runs indicating that ozone depletion contributed the most to recent summertime SAM trends. Similar 20th Century SAM trends to the PCM and observations are found in the CCSM3 all-forcing runs. However, in the CCSM3 single forcing experiments, GHG increases and stratospheric ozone depletion have approximately equal contributions to recent summertime SAM trends (based on an ensemble of only 2 members; not shown).

[7] The PCM and CCSM3 both exhibit negative SAM trends in their SRES A1B simulations from 2001–2050 for the DJF season, consistent with the majority of CCMVal models and unlike most of the CMIP3 models (even those with ozone recovery) which have weakly positive trends [Son *et al.*, 2008, Figure 3c]. However, there is a large difference in the magnitude of the trends between the two NCAR coupled models, with the negative trend in the PCM much greater in magnitude than the weakly negative trend in CCSM3. What leads to this large difference between two

similar models? Is it a difference in forcing or a difference in their response to the same forcing?

2. Results

[8] Figure 1a shows time series of the CO₂ and Figure 1b shows SH high latitude ozone forcing used in the 20C3M and SRES A1B experiments of the PCM and CCSM3. It is clear that identical CO₂ concentrations were used in both models. For both models ozone forcing is prescribed using the Kiehl *et al.* [1999] dataset for the 20th Century and an idealized profile of stratospheric ozone recovery by ~2050 for the 21st Century [Meehl *et al.*, 2006]. Some discontinuities in ozone forcing are apparent at the year 2000, the year of transition between observed historical forcing and the beginning of the scenario experiments. However, the spatial patterns of ozone trends over the Antarctic polar cap are very similar between the two models (not shown).

[9] Time series of the SAM index (Figure 1c; defined as the difference in standardized sea level pressure between 45°S and 60°S after Gong and Wang [1999], a unitless index) show that the difference in A1B DJF trends between the PCM and CCSM3 are robust and not limited to a choice in trend period. The PCM exhibits a strong decline in the summertime SAM until ~2050, while the CCSM3 SAM declines only slightly.

[10] Since the forcing appears to be similar in both models, could climate sensitivity be playing a role in their different SAM responses? One way to isolate the impact of GHG forcing is to look at the SAM index in 1% per year

Table 1. DJF SAM Trends From NCAR Coupled Model Simulations^a

	SRES A1B	1%/year CO ₂	SRES A1B (GHG)	SRES A1B (Ozone) ^{estimated}
Based on years	2005–2050	1–140	2005–2050	2005–2050
PCM	–0.56 ^b (±0.40)	0.07 ^b (±0.07)		–0.66
CCSM3	–0.24 (±0.42)	0.22 ^b (±0.07)	0.34 ^c (±0.39)	–0.57

^aTrends (unit/decade) are based on linear regression of single member ensembles except for the SRES A1B which use ensemble means (PCM: 4 members, CCSM3: 7 members). 95% confidence intervals are given in brackets.

^bSignificance at the 95% level.

^cSignificance at the 90% level.

CO₂ increase experiments shown in Figure 1d. Since CO₂ is the only forcing that varies in these simulations, a clean signature of its role in SAM changes can be found. It is clear that the CCSM3 is responding more rapidly to CO₂ forcing than the PCM in DJF. This suggests that the weaker negative trend in the CCSM3 when all forcings are combined (such as in the A1B scenario) could be due mostly to a larger sensitivity to CO₂ in CCSM3, which offsets its response to ozone recovery.

[11] An examination of a CCSM3 A1B simulation in which all forcings except GHGs were kept fixed [Shindell *et al.*, 2008] enables a quantitative test of this hypothesis. Table 1 gives DJF trends in the SAM from the various experiments with PCM and CCSM3. Assuming no other 21st Century forcings (e.g. aerosols) have a large impact on the SAM, one can estimate the SAM trend in the CCSM3 due to ozone recovery by SAM_{A1B} minus SAM_{A1B(GHG-only)}. Since the CCSM3 A1B GHG-only simulation gives a SAM trend per decade of approximately 1.5 times the 1% CO₂ increase experiments (presumably due to stronger GHG forcing per decade in the A1B scenario), one can estimate the

SAM_{A1B(ozone-recovery)} trend in PCM by SAM_{A1B} – 1.5 * SAM_{1pct}. This method results in similar estimations of SAM_{A1B(ozone-recovery)} of –0.6/decade for both PCM and CCSM3.

[12] We can extend this analysis to other models submitted to the CMIP3 archive. Figure 2 shows a scatter diagram between trends in global temperature (a measure of climate sensitivity) and trends in the SAM index from the 1% per year CO₂ increase experiments to quadrupling, where each symbol is the trend from a different CMIP3 model. A robust relationship exists between climate sensitivity and the trend in the SAM in all seasons, with the larger the climate sensitivity the larger the trend in the SAM. The most consistent relationship is found in DJF and a regression analysis indicates that a 1 degree increase in global temperatures leads to positive shift in the SAM of 0.5.

[13] Figure 3 explores the physical processes that may contribute to the strong link between SAM and climate sensitivity under increasing CO₂. The left column shows scatter diagrams between the trend in globally averaged annual surface air temperatures and annual mean trends in

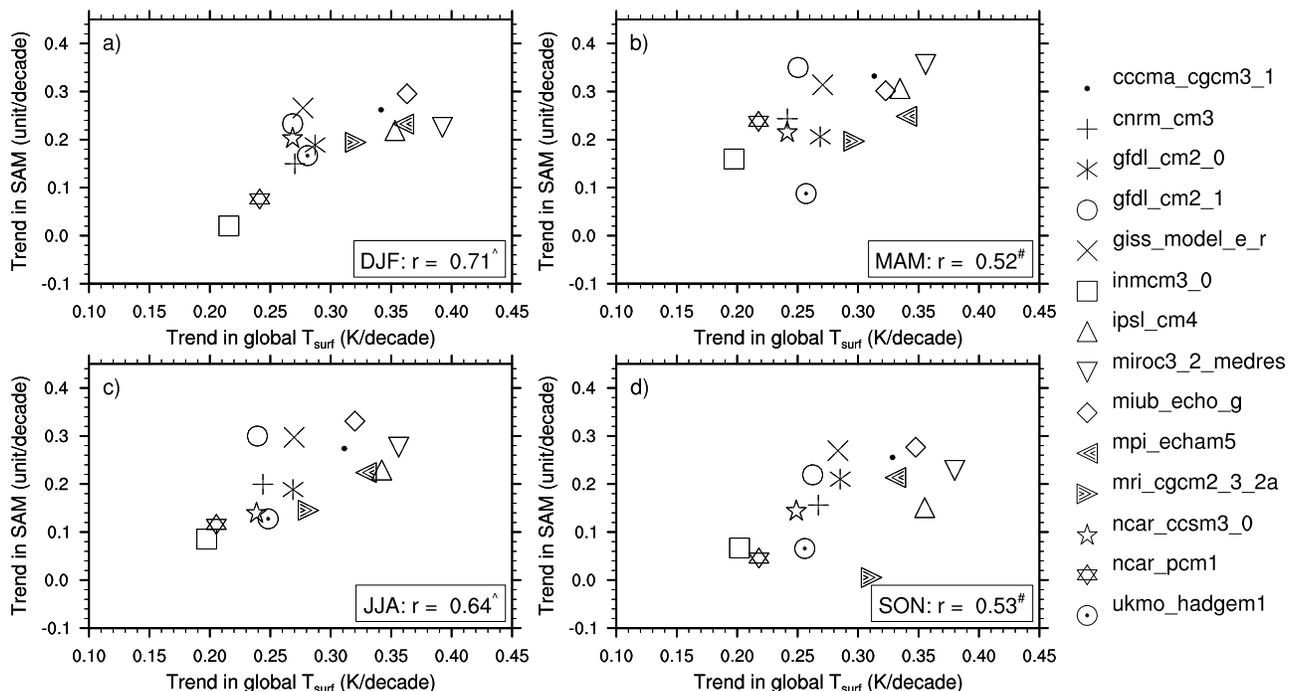


Figure 2. Scatter diagram between the trend in global temperatures (K/decade) and the trend in the SAM index (unit/decade) from the 1% per year CO₂ increase (to quadrupling) simulations of CMIP3 models for (a) DJF, (b) MAM, (c) JJA, and (d) SON seasons. All trends are based on timeseries from the beginning of each simulation to quadrupling of CO₂. Each symbol represents a different CMIP3 model, with the same symbol used for each model. The circumflex and pound symbols indicate the correlations are significant at the 95% and 90% level, respectively.

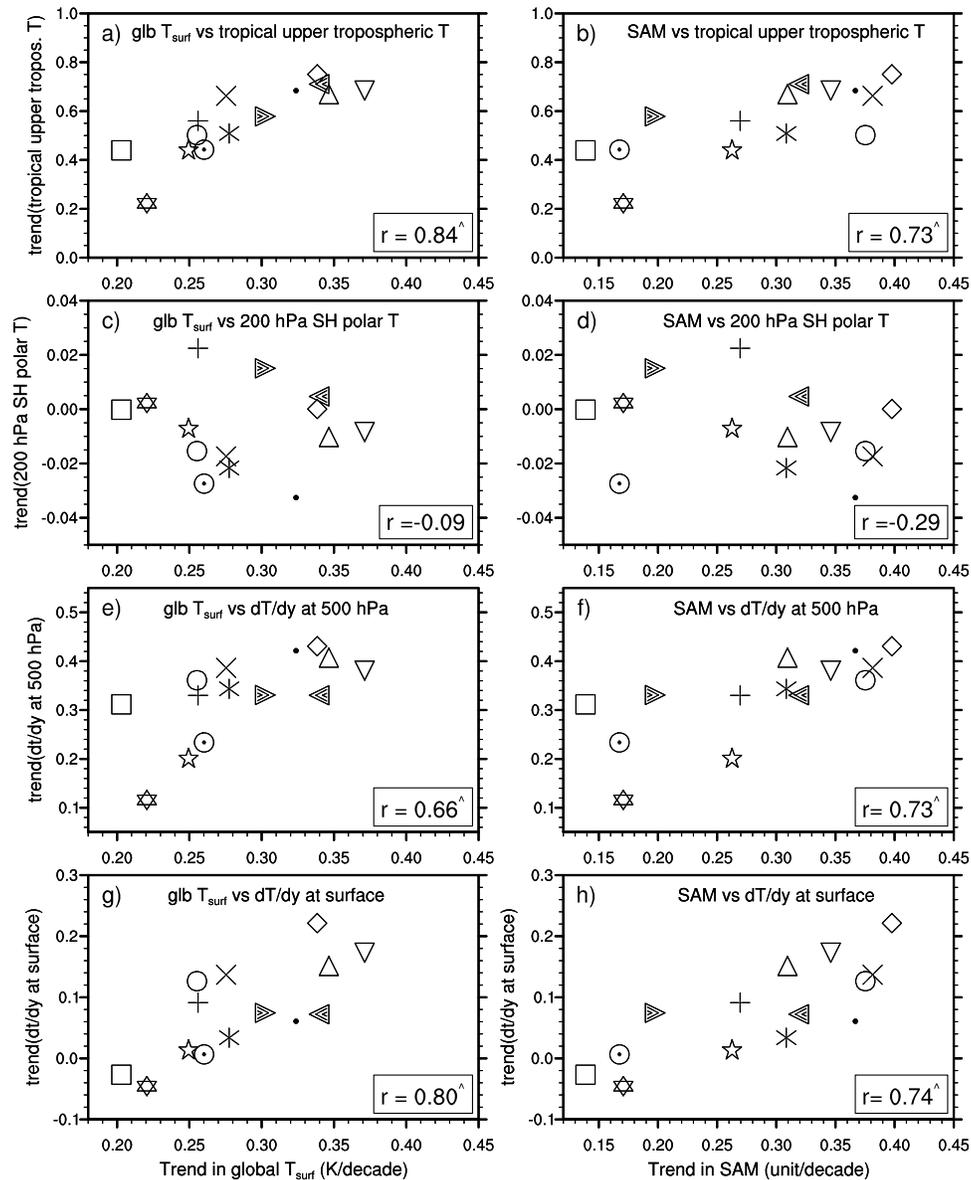


Figure 3. Scatter diagram between the trend in (left) global temperatures and (right) the trend in the SAM index with (a and b) change in temperature in the tropical mid-troposphere (c and d) change in temperature at 200 hPa over the polar cap (the use of 100 hPa gives similar results) and (e and f) meridional temperature gradient at 500 hPa and (g and h) meridional temperature gradient at the surface from the 1% per year CO_2 increase (to quadrupling) simulations of CMIP3 models. All trends are based on time series from the beginning of each simulation quadrupling of CO_2 using annual mean values. The circumflex indicates the correlations are significant at the 95% level.

the a) tropical upper tropospheric temperature, c) lower polar stratosphere temperatures, e) the equator-to-pole temperature gradient at 500 hPa (the difference between zonal temperatures at 500 hPa averaged over the area extending from the equator to 20°S compared with $70\text{--}90^\circ\text{S}$), and g) the equator to pole temperature gradient at the surface (the difference between zonal surface temperatures averaged from the equator– 20°S compared to $50\text{--}70^\circ\text{S}$). Similar analysis is shown with respect to trends in the SAM in the right column. All regions have been discussed previously [Kushner et al., 2001; Lorenz and DeWeaver, 2007; Lim and Simmonds, 2009; Son et al., 2008; Butler et al., 2010] as potentially contributing to shifts of the westerly jet in future projections.

[14] Tropical upper tropospheric warming is a known feature of anthropogenic climate change (Figure 3a) and is clearly related to the strength of the SAM trend (Figure 3b) in the CMIP3 models. Lower stratospheric polar cooling, a signature of both global warming and SH stratospheric ozone depletion, appears to have less bearing on the variation in CO_2 induced SAM trends (Figure 3d). This contrasts with the results of Lorenz and DeWeaver [2007], who found a significant intermodel correlation between this region and changes in the 850 hPa zonal wind (a measure of the SAM). However, their study analyzed the spread of CMIP3 models under the A1B scenario compared to 20th Century simulations. While GHGs are the dominant forcing by the end of the 21st Century, ozone forcing in the simulations likely

biases their result, resulting in a stronger influence of lower polar stratospheric cooling than that indicated by Figure 3d (while slightly different methodologies between our studies could influence this result, we can reproduce a significant correlation when replacing our 1% simulations with A1B). The meridional temperature gradients at both 500 hPa and the surface are also strongly linked to the spread of the models response to global temperatures and the SAM.

3. Discussion and Conclusions

[15] We have suggested that climate sensitivity largely explains the difference between the strength of future summertime trends in the SH extratropical circulation in two NCAR coupled climate models. Ozone forcing under all SRES scenarios in these models reflects a recovery to 1980 levels by the mid-21st Century; in both the NCAR CCSM3 and PCM this recovery overwhelms the impact of increasing GHGs, shifting the westerly jet equatorward and leading to negative SAM trends from present day. By examining simulations where only CO₂ increases, the weaker SAM trends in CCSM3 compared to PCM can be explained by its larger sensitivity to CO₂ which offsets the ozone recovery-driven trends. We extended this analysis to all CMIP3 models, finding a strong link between climate sensitivity and the strength of SAM trends under transient CO₂ conditions in all seasons. Various thermal forcings were investigated for their role in driving these trends, with the strength of tropical upper tropospheric warming explaining more of the variation amongst the models than stratospheric cooling over the polar cap.

[16] These results complement recent findings that biases in the climatological SH jet position [Kidston and Gerber, 2010] and energy budget [Trenberth and Fasullo, 2010] are related to the magnitude of future shifts in the jet and global temperatures, respectively. This suggests that a more complete understanding of the associations between the mechanisms of the SAM and climate sensitivity may help in narrowing uncertainty in global projections.

[17] While the analysis in Table 1 suggests that sensitivity to CO₂ can largely explain the different PCM and CCSM3 A1B trends to 2050, it is possible that different parameterizations in the atmospheric components of the NCAR models, especially with respect to radiative and dynamical processes, could also result in different responses to similar stratospheric ozone forcing. Further investigation is required to explore this aspect. Deser *et al.* [2010] also caution that internal variability plays a dominant role in uncertainty of extratropical projections.

[18] Our results indicate that one factor leading to contrasts between CMIP3 and chemistry-climate models in future trends of the SAM could be climate sensitivity. The two NCAR models have smaller climate sensitivities than the majority of CMIP3 models [Meehl *et al.*, 2005], which likely contributes to their more negative SAM trends in DJF (when ozone recovery counters the opposing effect of GHGs) than the weakly positive trends in other CMIP3 models. Unfortunately, without estimates of climate sensitivity from chemistry-climate models (since most use prescribed SSTs) and detailed ozone forcing from CMIP3 models it is not possible to ascertain how large a role climate sensitivity plays in these differences. Future inter-comparisons such as CMIP5 will go some way to improving

our ability to address questions raised here in the next generation of models. In addition, a recent analysis by *Son et al.* [2010] finds smaller differences between a new set of chemistry-climate model simulations (CCMVal-2) and the CMIP3 models.

[19] **Acknowledgments.** We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) and the WCRP's Working Group on Coupled Modelling (WGCM) for their roles in making available the WCRP CMIP3 multi-model dataset. Support of this dataset is provided by the Office of Science, U.S. Department of Energy. Portions of this study were supported by the Office of Science (BER), U.S. Department of Energy, Cooperative Agreement DE-FC02-97ER62402, and the NSF. The National Center for Atmospheric Research is sponsored by the NSF. We thank Judith Perlwitz and Clara Deser for helpful discussions and Scott Power and two anonymous reviewers for their constructive comments.

References

- Arblaster, J. M., and G. A. Meehl (2006), Contributions of external forcings to southern annular mode trends, *J. Clim.*, *19*(12), 2896–2905, doi:10.1175/JCLI3774.1.
- Butler, A. H., D. W. J. Thompson, and R. Heikes (2010), The steady-state atmospheric circulation response to climate change-like thermal forcings in a simple general circulation model, *J. Clim.*, *23*(13), 3474–3496, doi:10.1175/2010JCLI3228.1.
- Collins, W. D., et al. (2006), The Community Climate System Model Version 3 (CCSM3), *J. Clim.*, *19*(11), 2122–2143, doi:10.1175/JCLI3761.1.
- Crook, J. A., N. P. Gillett, and S. P. E. Keeley (2008), Sensitivity of Southern Hemisphere climate to zonal asymmetry in ozone, *Geophys. Res. Lett.*, *35*, L07806, doi:10.1029/2007GL032698.
- Deser, C., A. S. Phillips, V. Bourdette, and H. Teng (2010), Uncertainty in climate change projections: The role of internal variability, *Clim. Dyn.*, in press.
- Gong, D., and S. Wang (1999), Definition of Antarctic Oscillation index, *Geophys. Res. Lett.*, *26*(4), 459–462, doi:10.1029/1999GL900003.
- Kidston, J., and E. P. Gerber (2010), Intermodel variability of the poleward shift of the austral jet stream in the CMIP3 integrations linked to biases in 20th century climatology, *Geophys. Res. Lett.*, *37*, L09708, doi:10.1029/2010GL042873.
- Kiehl, J. T., T. L. Schneider, R. W. Portmann, and S. Solomon (1999), Climate forcing due to tropospheric and stratospheric ozone, *J. Geophys. Res.*, *104*(D24), 31,239–31,254, doi:10.1029/1999JD900991.
- Kushner, P. J., I. M. Held, and T. L. Delworth (2001), Southern Hemisphere atmospheric circulation response to global warming, *J. Climate*, *14*(10), 2238–2249.
- Lim, E., and I. Simmonds (2009), Effect of tropospheric temperature change on the zonal mean circulation and SH winter extratropical cyclones, *Clim. Dyn.*, *33*(1), 19–32, doi:10.1007/s00382-008-0444-0.
- Lorenz, D. J., and E. T. DeWeaver (2007), Tropopause height and zonal wind response to global warming in the IPCC scenario integrations, *J. Geophys. Res.*, *112*, D10119, doi:10.1029/2006JD008087.
- Meehl, G. A., W. M. Washington, W. D. Collins, J. M. Arblaster, A. Hu, L. E. Buja, W. G. Strand, and H. Teng (2005), How much more global warming and sea level rise?, *Science*, *307*(5716), 1769–1772.
- Meehl, G., W. Washington, B. Santer, W. Collins, J. Arblaster, A. Hu, D. Lawrence, H. Teng, L. Buja, and W. Strand (2006), Climate change projections for the twenty-first century and climate change commitment in the CCSM3, *J. Clim.*, *19*(11), 2597–2616, doi:10.1175/JCLI3746.1.
- Miller, R. L., G. A. Schmidt, and D. T. Shindell (2006), Forced annular variations in the 20th century Intergovernmental Panel on Climate Change Fourth Assessment Report models, *J. Geophys. Res.*, *111*, D18101, doi:10.1029/2005JD006323.
- Perlwitz, J., S. Pawson, R. L. Fogt, J. E. Nielsen, and W. D. Neff (2008), Impact of stratospheric ozone hole recovery on Antarctic climate, *Geophys. Res. Lett.*, *35*, L08714, doi:10.1029/2008GL033317.
- Rogers, J. C., and H. van Loon (1982), Spatial variability of sea level pressure and 500 mb height anomalies over the Southern Hemisphere, *Mon. Weather Rev.*, *110*(10), 1375–1392, doi:10.1175/1520-0493(1982)110<1375:SVOSLP>2.0.CO;2.
- Shaw, T. A., M. Sigmund, T. G. Shepherd, and J. F. Scinocca (2009), Sensitivity of simulated climate to conservation of momentum in gravity wave drag parameterization, *J. Clim.*, *22*(10), 2726–2742, doi:10.1175/2009JCLI2688.1.
- Shindell, D. T., H. Levy, M. D. Schwarzkopf, L. W. Horowitz, J.-F. Lamarque, and G. Faluvegi (2008), Multimodel projections of climate change from

- short-lived emissions due to human activities, *J. Geophys. Res.*, *113*, D11109, doi:10.1029/2007JD009152.
- Son, S., L. Polvani, D. Waugh, H. Akiyoshi, R. Garcia, D. Kinnison, S. Pawson, E. Rozanov, T. Shepherd, and K. Shibata (2008), The impact of stratospheric ozone recovery on the Southern Hemisphere westerly jet, *Science*, *320*(5882), 1486–1489, doi:10.1126/science.1155939.
- Son, S.-W., et al. (2010), Impact of stratospheric ozone on Southern Hemisphere circulation change: A multimodel assessment, *J. Geophys. Res.*, *115*, D00M07, doi:10.1029/2010JD014271.
- Stratospheric Processes and their Role in Climate (SPARC) CCMVal (2010), SPARC report on the evaluation of chemistry-climate models, edited by V. Eyring et al., *Rep. 5*. (Available at <http://www.atmos.physics.utoronto.ca/SPARC>).
- Thompson, D., and S. Solomon (2002), Interpretation of recent Southern Hemisphere climate change, *Science*, *296*(5569), 895–899, doi:10.1126/science.1069270.
- Trenberth, K. E., and J. T. Fasullo (2010), Simulation of present-day and twenty-first-century energy budgets of the southern oceans, *J. Clim.*, *23*(2), 440–454, doi:10.1175/2009JCLI3152.1.
- Washington, W. M., J. W. Weatherly, G. A. Meehl, A. J. Semtner Jr., T. W. Bettge, A. P. Craig, W. G. Strand Jr., J. Arblaster, V. B. Wayland, and R. James (2000), Parallel climate model (PCM) control and transient simulations, *Clim. Dyn.*, *16*(10–11), 755–774, doi:10.1007/s003820000079.
- Waugh, D. W., L. Oman, P. A. Newman, R. S. Stolarski, S. Pawson, J. E. Nielsen, and J. Perlwitz (2009), Effect of zonal asymmetries in stratospheric ozone on simulated Southern Hemisphere climate trends, *Geophys. Res. Lett.*, *36*, L18701, doi:10.1029/2009GL040419.
- Yin, J. H. (2005), A consistent poleward shift of the storm tracks in simulations of 21st century climate, *Geophys. Res. Lett.*, *32*, L18701, doi:10.1029/2005GL023684.

J. M. Arblaster and G. A. Meehl, NCAR Earth System Laboratory, National Center for Atmospheric Research, Boulder, CO 80307, USA. (jma@ucar.edu)

D. J. Karoly, School of Earth Sciences, University of Melbourne, Melbourne, Vic 3010, Australia.