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Stratospheric ozone during the Last Glacial Maximum

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[1] Stratospheric ozone during the Last Glacial Maximum (LGM) is investigated in on-line simulations with the GISS Global Climate/Middle Atmosphere Model 3. LGM boundary conditions and atmospheric concentrations are employed in three simulations: without interactive ozone, with ozone photochemistry appropriate for that time period and with the LGM climate but current atmospheric composition for chemistry. Results show stratospheric ozone increased during the LGM due to reduced NOy and chlorine, while warmer stratospheric temperatures (from reduced stratospheric CO_2) decrease ozone with current photochemistry. The stratospheric residual circulation intensified in the lowermost stratosphere, increasing stratosphere/troposphere exchange at higher latitudes, although for most of the middle atmosphere the circulation decreased; the age of air in the Middle Atmosphere increased by up to one year. Compared with the vastly different LGM conditions, increase in stratospheric ozone of 2% by mass had little effect on atmospheric dynamics, and increased the global radiation balance by <0.1 Wm^{-2}. **Citation:** Rind, D., J. Lerner, C. McLinden, and J. Perlwitz (2009), Stratospheric ozone during the Last Glacial Maximum, *Geophys. Res. Lett.*, *36*, L09712, doi:10.1029/2009GL037617.

1. Introduction

[2] The stratosphere during the LGM (~20,000 yrs ago) was likely to have been considerably different than it is today. Not only were the radiative constituents altered (CO_2, H_2O, CH_4) but so was the forcing from the troposphere, due to the presence of large ice-sheets in the Northern Hemisphere. Our previous modeling studies [*Rind et al.*, 2001] suggested the following characteristics: a warmer stratosphere, due primarily to reduced CO_2; and an increase in the residual circulation in the lower stratosphere with a decrease above. The latter effect was due to increased tropospheric eddy kinetic energy associated with stronger latitudinal temperature gradients and greater (ice sheet) topography, which in turn resulted in greater planetary wave forcing of the lower stratospheric circulation. However, less favorable propagation conditions led to less wave forcing above for much of the middle atmosphere. The topic addressed in this article is the likely response of stratospheric ozone to the ice age conditions, and its associated radiative forcing.

2. Model Experiments

[3] The model used is the GISS Global Climate Middle Atmosphere Model (GCMAM) 3 [*Rind et al.*, 2007], the version with 4° × 5° resolution and 53 layers, model top at the mesopause. The relevant boundary conditions and atmospheric composition were changed to be consistent with our best understanding for that time period; the boundary conditions used were those adopted for the Paleoclimate Modeling Intercomparison Project 2 (PMIP 2) simulations (http://pmip2.lsce.ipsl.fr). In particular, more extensive land ice and sea ice occurs in the Northern Hemisphere, with reduced sea surface temperatures in most places, although the tropical cooling used is relatively minimal (see the discussion of this issue by *Rind et al.* [2001]). The atmospheric composition is changed to be consistent with that found in ice core data (in particular, the CO_2 concentration is set to 200 ppm, CH_4 to 400 ppb, N_2O to 200 ppb). The annual average global surface air temperature cooling in the model is 4.6°C. The reduced methane produces less water vapor in the upper stratosphere, roughly 1/4 of that generated in the current climate (based on monthly varying latitude-height source functions derived from a 2D chemical transport model [*Fleming et al.*, 1999; *Schmidt et al.*, 2006]). Note that both the model and the boundary conditions are different from those used in the previous simulations [*Rind et al.*, 2001].

[4] Stratospheric ozone is calculated using the LINOZ scheme [*McLinden et al.*, 2000], with coefficients altered to be consistent with the changed atmospheric composition. (The version of LINOZ used here does not parameterize the ozone hole, so all comparisons with the control run are independent of that feature.) In addition to the values indicated above, the calculations used a slightly reduced water vapor value of 3.2 ppm based on the slightly cooler tropical upper troposphere temperatures from the simulations of *Rind et al.* [2001]. NO_y uses the current profile scaled by LGM/current N_2O ratio. Cly equals 0.45 ppb (preindustrial) (current value is ~3 ppb) based on decomposition of CH_3Cl; and Br_y = 5.0 ppt (preindustrial), with a scaled current profile. The linearization in the LINOZ scheme is centered around stratospheric temperature 3°C warmer than today [from *Rind et al.*, 2001].

[5] For the troposphere, we do not attempt to model the LGM values, we use a simple relaxation approach to 20 ppb ozone in the boundary layer, with an e-folding time of 2 days (a slight alteration of the 25 ppb ozone from *McLinden et al.* [2000], to better fit observations); any tropospheric changes shown here are simply the result of altered exchange with the stratosphere. This limitation is discussed in section 5.

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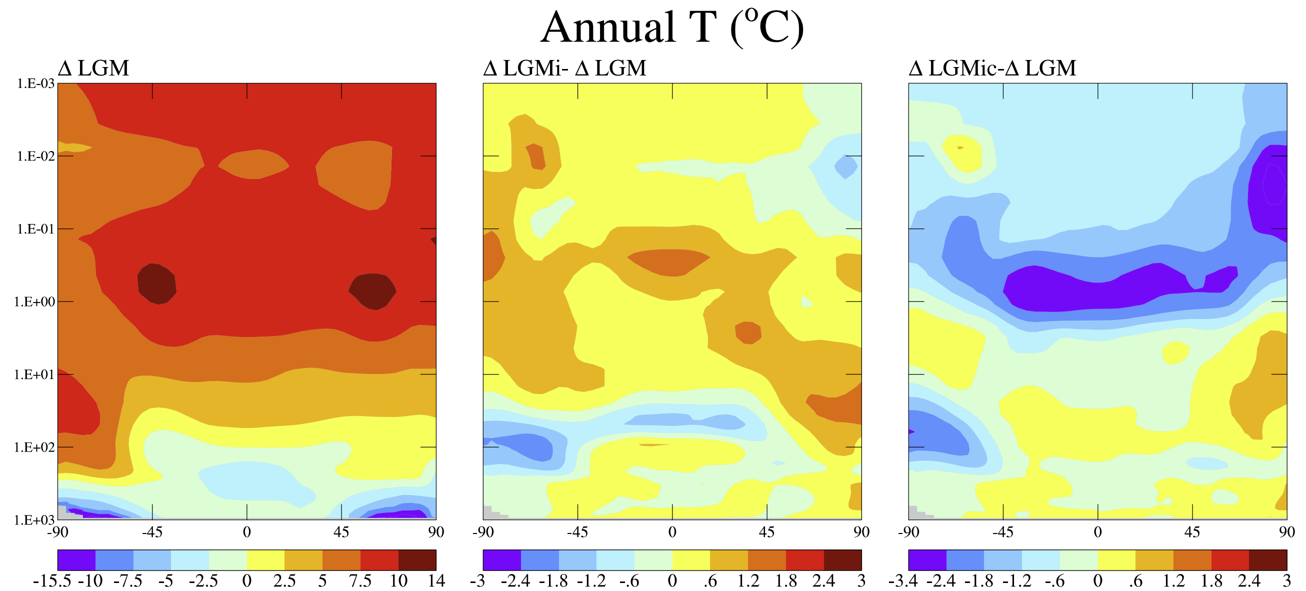

Figure 1. Annual average temperature changes for (left) LGM-CONT, (middle) (LGMi-CONTi) minus (LGM-CONT) and (right) (LGMic-CONTi) minus (LGM-CONT). See text for details of the simulations.

[6] Five simulations were performed: (1) Control run with current day SSTs and prescribed current day ozone ("CONT"); (2) Control run with current SSTs and calculated ozone (LINOZ scheme with modern atmosphere coefficients ("CONTi" (for interactive ozone)); (3) LGM boundary and atmospheric conditions with current day ozone ("LGM"); (4) LGM boundary and atmospheric conditions with calculated ozone (LINOZ scheme with LGM atmospheric composition) ("LGMi"); and (5) LGM boundary and atmospheric conditions with calculated ozone via the LINOZ scheme but with current day atmospheric composition ("LGMic")). LGM minus CONTROL indicate the changes induced by the ice age conditions; LGMi minus CONTROLi indicate the changes induced by ice age conditions plus the change in ozone; and LGMic minus CONTROLi gives the change induced by changes in ozone associated only with the altered ice age circulation and temperatures, as the composition for chemistry calculations was returned to current values.

[7] The results shown are for 12 years following a 15 month spin-up (with the LINOZ scheme). In addition, the initial atmospheric conditions came from a multi-year spinup (without LINOZ). Nevertheless, the rapid approach to equilibrium in the meteorology is really the result of the use of specified (unchanging) SSTs (the standard deviation of the global annual surface air temperature change for LGMi over the 12 years is 0.6°C with no apparent trend). The stratospheric ozone response also shows little trend, and the standard deviation of the annual total ozone in each of these runs after the first year is $\leq$0.3% (while the difference in total ozone between the simulations is 10$\times$ larger). As an additional check, we extended each run five years, with again very similar results. To better understand the transport changes, we also included in each simulation other on-line tracers [*Rind et al.*, 2007], of which SF_6 is utilized below.

3. Simulation Results

[8] The annual average temperature change for the LGM run with current day ozone is shown in Figure 1 (left). The lower stratosphere warms by ~1°C, the middle stratosphere by ~4°C, and the upper stratosphere by some 9°C. The differences in the LGMi and LGMic experiments relative to their control run are compared with this temperature change in Figures 1 (middle) and 1 (right), respectively. In LGMi, the upper stratosphere temperature has warmed about 1°C more associated with its ozone change, while in LGMic, it has warmed 2–3°C less. The opposite effects arise at the tropical tropopause – less warming in LGMi, and slightly more in LGMic.

[9] The annual average ozone changes are shown in Figure 2. In the LGMi experiment (Figure 2, left), ozone decreased by about 7% in the lower stratosphere (~3 ppbv), increased by 10–15% (~10 ppbv) in the mid-stratosphere, and by ~3% (~1 ppbv) in the upper stratosphere. These changes were the result of several conflicting influences. The chemistry change itself is due to two factors: the warmer temperatures, leading to less net ozone production due to the temperature dependence of the ozone loss rate coefficients; and the altered chemical composition, most importantly the reduced NOy and chlorine, leading to more net ozone production. This is most clearly shown by comparison with the LGMic results (with modern day composition for chemistry). Here the warmer temperatures are combined with the less favorable atmospheric composition, and it has the lowest ozone of any of the simulations, with a 10–15% deficit in the upper stratosphere. The altered ozone in the upper and lower stratosphere in the two runs change the short and longwave radiation absorption, and help explain most of the temperature differences shown in Figures 1 (middle) and 1 (right).

[10] In addition to the photochemical changes, ozone is also affected by changes in the dynamics associated with the LGM climate. The change in age of air in these experiments is shown in Figure 3, here derived from an SF_6-like tracer [see *Rind et al.*, 2007]. The increased age of air is the result of a decrease in the middle atmosphere residual circulation. The lower stratosphere increased circulation found in the previous simulations [*Rind et al.*, 2001] is confined in these runs to the lowest polar stratospheric levels, especially in

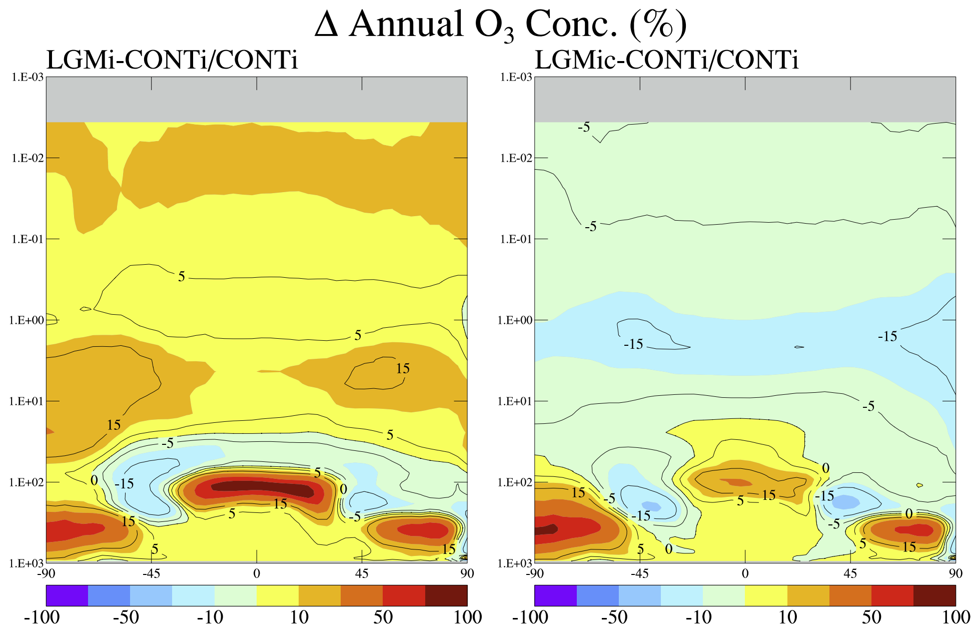

Figure 2. Annual average percentage ozone change for (left) LGMi-CONTi and (right) LGMic-CONTi.

the Northern Hemisphere; it does, however, imply more rapid stratosphere/troposphere exchange in the extratropics, and is responsible for the increase in tropospheric ozone due to transport seen at those latitudes (Figure 2). Transport of ozone from higher altitudes downward through the 200 mb level increases by about 11%, producing a similar increase in tropospheric ozone (from transport alone). This increased high-latitude exchange is associated with greater tropospheric eddy kinetic energy (by 17%, similar with and without the stratospheric ozone response). By the mid-stratosphere, however, eddy energy is reduced (by some 20%) due to less favorable propagation conditions associated with stronger extratropical west winds. Global poleward eddy sensible heat flux (proportional to the vertical flux of wave energy) is larger at 100 mb by 2–3% in the two LGM runs, equal to the control run values at 50 mb, and 10–14% less by 15 mb. This in turn leads to the reduced residual circulation in general and greater age of air evident for most of the middle atmosphere (Figure 3). There is then less loss of ozone from the upper stratosphere (~4–9 mb) by dynamical transport in each of the LGM runs. The ozone change shown in Figure 2 is the net effect of the altered chemical and dynamical influences.

[11] Shown in Figure 4 are the changes in total O_3, dominated by the changes in the lower stratosphere. Increased relative stratospheric subsidence over Antarctica (with a more positive Southern Annular Mode, the leading

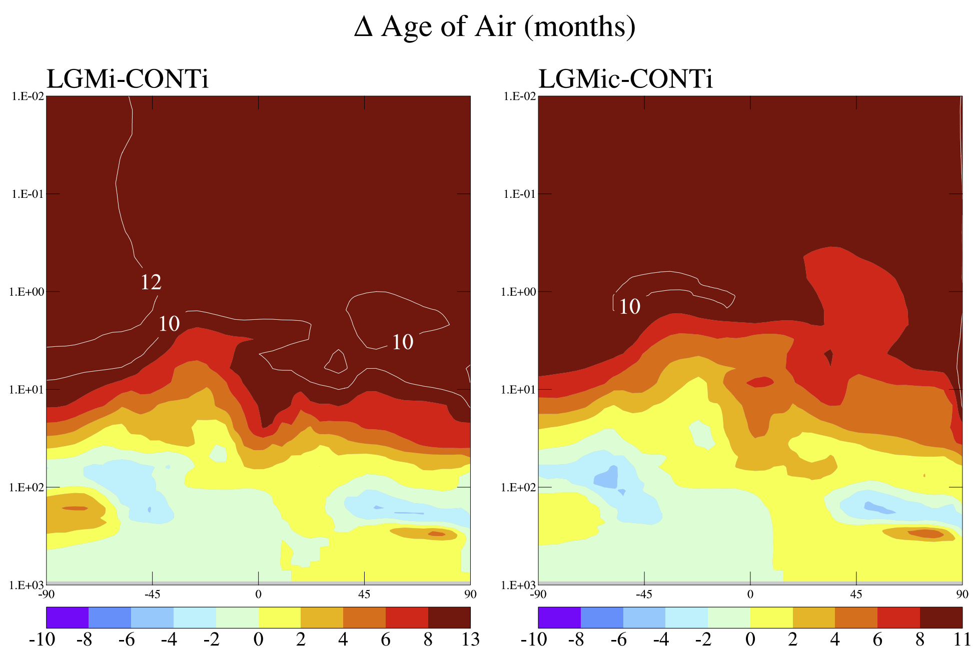

Figure 3. Change in the age of air for (left) LGMi-CONTi and (right) LGMic-CONTi.

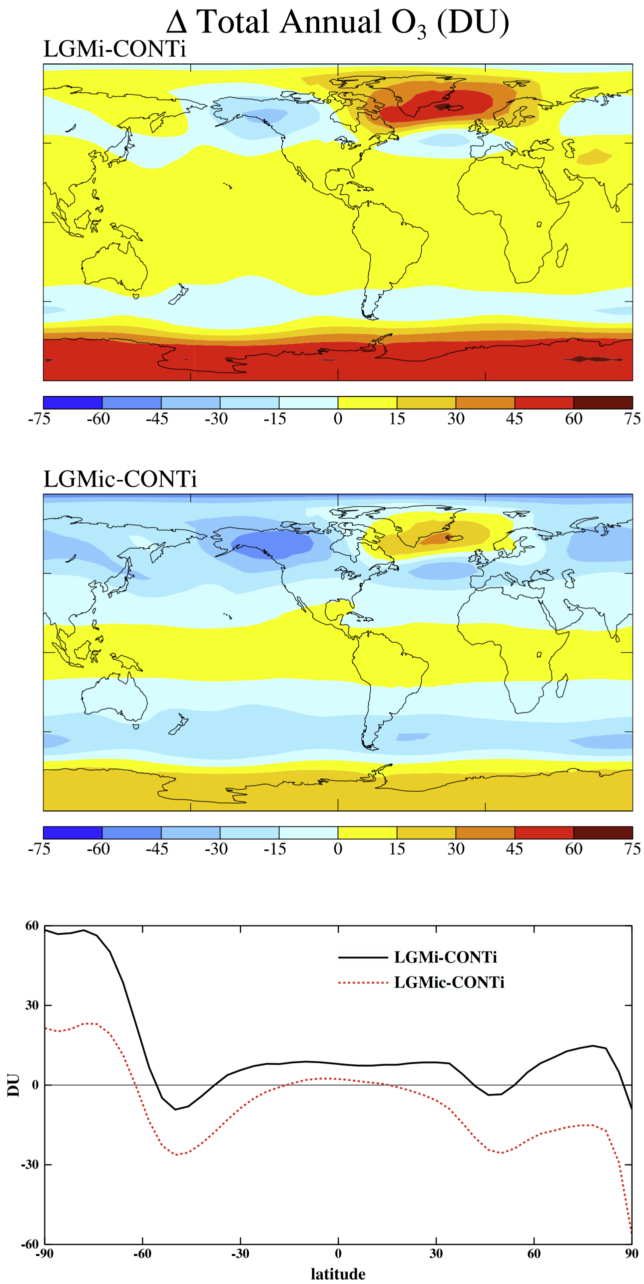

Figure 4. Annual average change in total ozone (in Dobson units) for (top) LGMi-CONTi and (middle) LGMic-CONTi. (bottom) Zonal average differences are also shown.

mode of variability with lower pressure at high latitudes compared with mid-latitudes) results in greater total ozone (~10% increase), with larger values in LGMi (than LGMic) due to its greater mid-and upper stratospheric ozone. Increased values in the tropical lower stratosphere result primarily from the residual circulation change in LGMic, and both chemistry and circulation in LGMi. An amplified Aleutian High/polar vortex in the lower stratosphere is responsible for the pattern seen at high northern latitudes (peaking at ~15% increase in the polar vortex). Overall, LGMi had a 2.4% increase in total ozone mass relative to the control run, while LGMic had a 2.3% decrease. The latitudinal average differences are given in Figure 4 (bottom), and emphasize the large differences between the polar regions. The highest southern latitudes have a more direct middle atmosphere circulation driven by increased wave energy flux throughout the stratosphere, while the reverse is true at high northern latitudes, as noted earlier.

4. Impacts on Climate

[12] Both the dynamical and radiative conditions for the LGM are greatly different from those of the current climate due to the altered boundary conditions and atmospheric trace gas concentrations. Hence the differences in stratospheric ozone are a minor consideration in that regard. Indeed, the changes in most tropospheric parameters relative to the current day control runs are indistinguishable among the three different LGM experiments. Inclusion of specified SSTs further helps maintain similarity in the changes for the lower troposphere and various dynamical factors. If these temperatures had not been specified, would the changes in stratospheric ozone have made more of a difference?

[13] Considering the net radiation at the top of the atmosphere, for LGMi the difference relative to the control run was -2.7 Wm^{-2}, and for LGMic, -2.8 Wm^{-2} [note that the changes in tropospheric ozone, due here only to transport, were not allowed to affect the radiation]. In both cases, the model would have cooled relative to the values associated with the specified SSTs, due primarily to the specified relatively warm tropical and subtropical SSTs, as noted previously [e.g., *Rind and Peteet*, 1985]. The similarity of the magnitudes in the two runs shows that the stratospheric ozone differences had little effect. Similarly, with respect to the net radiation at the tropopause, for the LGMi and LGMic the numbers (relative to the control run) were -1.3 and -1.4 Wm^{-2} respectively; hence the stratosphere was contributing about 1/2 the net radiative loss at the top. Given that the ozone increased for LGMi and decreased for LGMic, the ozone change was not the dominating factor but rather it was the warmer stratospheric temperatures, associated primarily with reduced CO_2. The radiation associated with the ozone differences in the two experiments thus amounted to 0.1 Wm^{-2}.

5. Discussion

[14] The model results suggest that the combined effect of warmer stratospheric temperatures and the altered atmospheric composition and dynamics of the last ice age would have led to increased stratospheric ozone. Overall the changes of total ozone mass are only a few percent, and their effect on the radiation balance is small compared to other alterations in the climate system at that time. *Martinerie et al.* [1995] using a 2D transport model with coupled chemistry and radiation calculated a similar magnitude of tropical upper stratospheric ozone change driven by the altered atmospheric composition. They also found much higher percentage changes in polar regions, which were likely influenced by their parameterized transports.

[15] The stratospheric ozone changes do have an influence on the magnitude of warming in the stratosphere of a few °C, and on the global radiation balance, on the order of 0.1 Wm^2. These effects are small relative to the extreme nature of the climate changes driven by the altered ice age

boundary conditions and other components of the atmospheric composition.

[16] The model simulations suggest that an increase in stratosphere-troposphere exchange in the polar regions would lead to greater tropospheric ozone from that contribution; however, the magnitude of the effect likely depends upon the configuration of this circulation change, which has varied in the different GISS simulations. Other factors affected tropospheric ozone at this time. *Martinerie et al.* [1995] calculated a decrease of some 60% in tropospheric ozone due primarily to methane and NOx changes, with an associated 11% increase in tropospheric OH, whereas *Valdes et al.* [2005] found a 7% decrease in tropospheric OH due to reduced water vapor [*Pinto and Khalil*, 1991]. The changes in stratospheric ozone discussed in this paper would also lead to changes in the UV flux into the troposphere, further affecting ozone photochemistry [*Bekki et al.*, 1994]. These issues would likely be more important for tropospheric ozone than altered exchange with the stratosphere.

[17] With the stratospheric ozone burden being an order of magnitude larger than the tropospheric, and a large net flux of ozone into the troposphere, uncertainties in the tropospheric component should have a near negligible impact on the stratospheric columns. To verify this we ran Linoz in an off-line chemical transport model using tropospheric boundary conditions of 20 ppb and 40 ppb and examined the impact on stratospheric ozone columns. In the lowest stratospheric layers in the tropics, where any impact would be found, the difference was about $\sim$0.2% and thus not important.

[18] We can compare these results with simulations of the $2 \times CO_2$ climate done with the GISS model. These have shown increased exchange between the troposphere and stratosphere, resulting in increases in tropospheric ozone (from this effect alone) of up to 30%, especially when there was strong tropical warming [*Rind et al.*, 2002]. While the LGM has the opposite sign of global temperature change, the use of CLIMAP SSTs keeps the tropics warm, and forces a similar UT/LS response.

[19] Reduction of the LGM residual circulation at higher levels in the middle atmosphere is plausible, as increased stratospheric west winds lead to less favorable wave propagation conditions. In this case, the results are opposite to that of the $2 \times CO_2$ climate with its increase in residual circulation [*Rind et al.*, 2002]. The details of the circulation changes are dependent on a number of uncertain factors for the last ice age, such as the magnitude of tropical SST changes, and the configuration of ice sheets, which would have affected both planetary and topographic gravity wave generation (see the discussion of *Rind et al.* [2001]). More refined simulations of the LGM middle atmosphere must await better determination of the relevant tropospheric boundary conditions.

[20] This paper represents a step toward understanding the stratospheric ozone distribution at the LGM. It is limited by linearization around decoupled perturbations, e.g., the ozone response to a change in chlorine loading depends on the NOy loading as well as temperature. In the current climate, non-linearities are (literally) second order effects, as the Linoz scheme (and others based on a linearization of the ozone tendency) is able to capture all key aspects of the ozone field in a climatological sense. The linearization behaves equally well for LGM levels of ozone, trace-gases and temperature, as determined in off-line testing. However, comparison with a GCM with full chemistry [e.g., *Eyring et al.*, 2006] will be a useful next step.

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References

Bekki, S., K. S. Law, and J. A. Pyle (1994), Effect of ozone depletion on atmospheric CH_4 and CO concentrations, *Nature*, *371*, 595–597.

Eyring, V., et al. (2006), Assessment of temperature, trace species, and ozone in chemistry-climate model simulations of the recent past, *J. Geophys. Res.*, *111*, D22308, doi:10.1029/2006JD007327.

Fleming, E. L., C. H. Jackman, R. S. Stolarski, and D. B. Considine (1999), Simulation of stratospheric tracers using an improved empirically based two-dimensional model transport formulation, *J. Geophys. Res.*, *104*, 23,911–23,934.

Martinerie, P., G. P. Brasseur, and C. Granier (1995), The chemical composition of ancient atmospheres: A model study constrained by ice core data, *J. Geophys. Res.*, *100*, 14,291–14,304.

McLinden, C. A., S. C. Olsen, B. Hannegan, O. Wild, M. J. Prather, and J. Sundet (2000), Stratospheric ozone in 3-D models: A simple chemistry and the cross-tropopause flux, *J. Geophys. Res.*, *105*, 14,653–14,665.

Pinto, J. P., and A. K. Khalil (1991), The stability of tropospheric OH during ice ages, inter-glacial epochs and modern times, *Tellus, Ser. B*, *43*, 347–352.

Rind, D., and D. Peteet (1985), LGM terrestrial evidence and CLIMAP SSTs: Are they consistent?, *Quat. Res.*, *24*, 1–22.

Rind, D., M. Chandler, P. Lonergan, and J. Lerner (2001), Climate change and the middle atmosphere: 5. Paleostratosphere in cold and warm climates, *J. Geophys. Res.*, *106*, 20,195–20,212.

Rind, D., 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*(D24), 4800, doi:10.1029/2002JD002483.

Rind, D., J. Lerner, J. Jonas, and C. McLinden (2007), Effects of resolution and model physics on tracer transports in the NASA Goddard Institute for Space Studies general circulation models, *J. Geophys. Res.*, *112*, D09315, doi:10.1029/2006JD007476.

Schmidt, G., et al. (2006), Present-day atmospheric simulations using GISS Model E: Comparison to in situ, satellite and reanalysis data, *J. Clim.*, *19*, 153–192.

Valdes, P. J., D. J. Beerling, and C. E. Johnson (2005), The ice age methane budget, *Geophys. Res. Lett.*, *32*, L02704, doi:10.1029/2004GL021004.

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