GEOPHYSICAL RESEARCH LETTERS, VOL. 35, L03705, doi:10.1029/2007GL032489, 2008

Evolution of stratospheric temperature in the 20th century

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Received 26 October 2007; revised 24 December 2007; accepted 8 January 2008; published 14 February 2008.

[1] We employ a coupled atmosphere-ocean climate model to investigate the evolution of stratospheric temperatures over the twentieth century, forced by the known anthropogenic and natural forcing agents. In the global, annual-mean lower-to-middle stratosphere (∼20–30 km.), simulations produce a sustained, significant cooling by ∼1920, earlier than in any lower atmospheric region, largely resulting from carbon dioxide increases. After 1979, stratospheric ozone decreases reinforce the cooling. Arctic summer cooling attains significance almost as early as the global, annual-mean response. Antarctic responses become significant in summer after ∼1940 and in spring after ∼1990 (below ∼21 km.). The correspondence of simulated and observed stratospheric temperature trends after ∼1960 suggests that the model's stratospheric response is reasonably similar to that of the actual climate. We conclude that these model simulations are useful in explaining stratospheric temperature change over the entire 20th century, and potentially provide early indications of the effects of future atmospheric species changes. **Citation:** Schwarzkopf, M. D., and V. Ramaswamy (2008), Evolution of stratospheric temperature in the 20th century, *Geophys. Res. Lett.*, *35*, L03705, doi:10.1029/2007GL032489.

1. Introduction

[2] Recent studies [*Santer et al.*, 1996; *Ramaswamy and Schwarzkopf*, 2002, hereafter RS02; *Schwarzkopf and Ramaswamy*, 2002, hereafter SR02; *Santer et al.*, 2003; *Shine et al.*, 2003; *Ramaswamy et al.*, 2006; *Santer et al.*, 2006; *Intergovernmental Panel on Climate Change* (*IPCC*), 2007] have shown that changes in the temperatures of the atmosphere and surface during the last few decades of the twentieth century are very likely related to anthropogenic emissions of trace gases, ozone and aerosols. These attribution studies have generally focused on explaining the evolution of the vertical temperature profile in the troposphere and lower stratosphere, below ∼20 km., especially in the period since satellite observations began in the late 1970s. An unexplored question is whether identifiable human influences on the stratospheric temperature occurred prior to the satellite era. A related issue is the quantitative roles of the major natural and anthropogenic forcing agents in producing these changes, both on the global, annual scale and on seasonal and regional scales (e.g., polar winter and spring). Of key importance is the timing of the stratospheric temperature response to these forcings: i.e., does the stratospheric response becomes significant (exceeding the variability of the unforced system) earlier than the response at lower levels, including the surface.

[3] Since stratospheric observations exist only post-1960 for sondes, and after the late 1970s for satellites [see *Ramaswamy et al.*, 2006], we investigate these questions by using simulations from a state-of-the-art climate model. By comparing the model results with the available observations, we assess the model's ability to reproduce observed stratospheric temperature trends. This, in turn, enables confidence in the use of the model in simulating the evolution of past (pre-observation) stratospheric temperatures as well as future stratospheric temperature change due to projected constituent changes.

2. Model Experiments

[4] We employ the GFDL coupled atmosphere-ocean-land climate model [*Knutson et al.*, 2006] (hereafter CM; the "CM2.1" version is used) to perform simulations of the atmospheric response to the evolution of anthropogenic and natural forcing agents over the 1861 to 2003 period. The principal simulation ('AllForc') gives the response to the known changes in all anthropogenic (well-mixed greenhouse gases (WMGGs), ozone, tropospheric aerosols, land use) and natural (solar, volcanic aerosol) factors in this period. Our emphasis is on the changes in the global, annual-mean temperatures in the lower-to-middle stratosphere (∼20–30 km.). We also investigate the seasonal temperature responses, especially in polar regions.

[5] To isolate the effects due to the anthropogenic and natural factors, we perform additional simulations comprising changes in: anthropogenic forcing agents only ('Anth'); natural forcing agents only ('Nat'); WMGGs and ozone only ('WmGhgO3'); and carbon dioxide only ('CO2'). In these simulations, WMGG values over 1861–2003 are adopted from the work of *Ramaswamy et al.* [2006] and *Knutson et al.* [2006]. The WMGGs include CO_2, CH_4, N_2O and 4 halocarbons (CFC11, CFC12, CFC113 and HFC22). Solar monthly spectral irradiances for 1882–2003 are taken from the work of *Lean et al.* [2005]; values for 1861–1882 are constant (at 1882 values). Stratospheric volcanic aerosol opacities and radiative parameters are taken from the work of *Stenchikov et al.* [2006]. Tropospheric ozone and aerosol concentrations are interpolated from calculations made at 10-year intervals using the MOZART chemical transport model [*Horowitz*, 2006; *Ginoux et al.*, 2006]. The monthly stratospheric ozone profile between 1861 and 1975 is invariant, employing the 1979 values from the work of *Randel and Wu* [1999]; values from 1979 onward are adapted from the work of *Randel and Wu* [2007]. Ozone monthly values between 1975 and 1979 are interpolated between the 1975 monthly values and the 1979 values from *Randel and Wu* [2007]. Stratospheric and tropospheric values are merged at the tropopause. Annual land-use changes follow *Knutson et al.* [2006].

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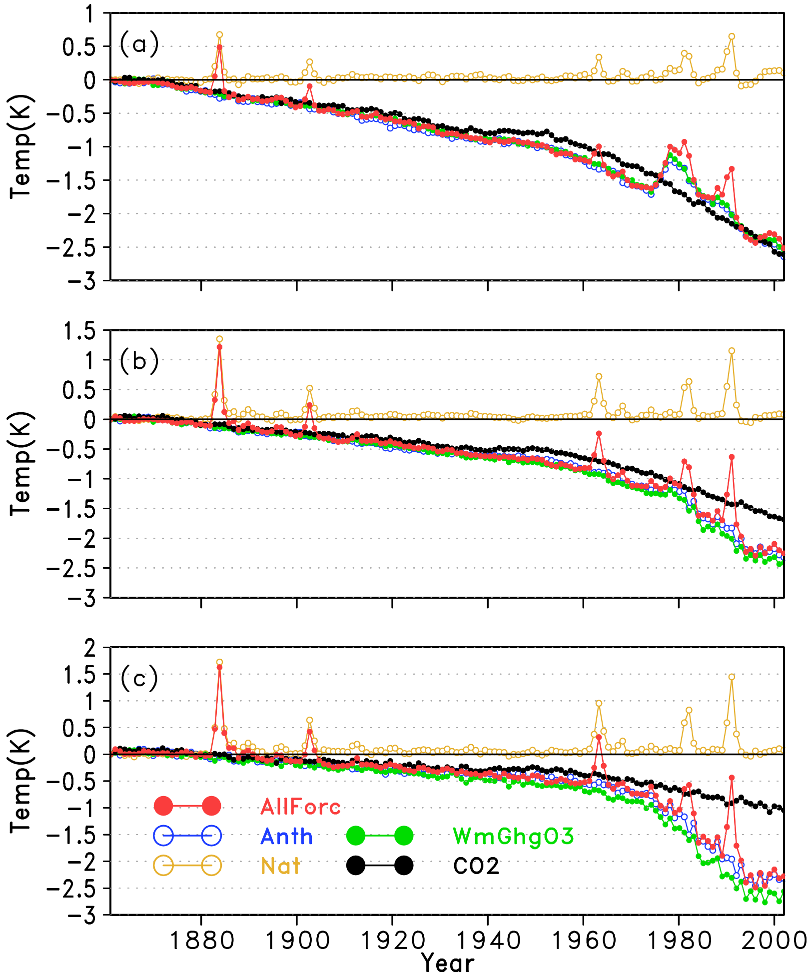

Figure 1. Evolution of the global, annual-mean temperature change relative to 1861 due to changes in radiative forcing agents at (a) 10 hPa; (b) 30 hPa; and (c) 50 hPa. The forcing agents are 'AllForc'; anthropogenic only ('Anth'); natural only ('Nat'); trace gases only ('WmGhgO3'); carbon dioxide only ('CO2').

[6] An ensemble of 5 members is used for the 'AllForc' simulations; other simulations use three-member ensembles. Ensemble members for each experiment start from different years of an 'unforced' integration with 1860 values of forcing agents [*Delworth et al.*, 2006; *Knutson et al.*, 2006]. The results are based on ensemble means. To evaluate the statistical significance of the simulated changes, the model's internal variability is estimated using a 1000-year section of the 'unforced' integration. This is taken as a surrogate for the variability of the actual climate system, in the absence of long-term global atmospheric measurements. At the surface, where such measurements exist, the CM's internal variability is at least as large as observations [*Knutson et al.*, 2006]. If the temperature change for any year relative to 1861 is larger than twice the standard deviation of the 'unforced' simulation, the response due to the imposed forcing (termed 'signal') is deemed significant for that year. For robustness, an identifiable impact in the stratosphere requires the presence of a significant response sustained over time.

3. Results

3.1. Annual Global-Mean Response

[7] Figures 1a, 1b, and 1c show the global, annual-mean temperature evolution from 1861 at 10 hPa (Figure 1a), 30 hPa (Figure 1b) and 50 hPa (Figure 1c) for 'AllForc', 'CO2', 'WmGhgO3', 'Anth' and 'Nat'. The evolution of 'WmGhgO3' accounts for most of the evolution of 'Anth' and 'AllForc', except in years with major volcanic eruptions, when the 'Nat' evolution greatly affects the stratospheric temperatures. Prior to ~1975, the evolution of 'CO2' accounts for most of the evolution of 'WmGhgO3' and thus 'AllForc'. After ~1979, ozone decreases and CO_2 increase, together with the effects of the El Chichon and Pinatubo volcanic eruptions combine to produce a step-like temperature evolution of 'WmGhgO3' and 'AllForc' throughout the stratosphere (as documented for the lower stratosphere [*Ramaswamy et al.*, 2006]). The effects of ozone decrease are most noticeable at 50 hPa, where after ~1990 the ozone response is larger than the CO_2 response; this result is similar to the conclusions by *Santer et al.* [2003] for the T4 level in the global, annual-mean, and to results of RS02 and *Langematz et al.* [2003].

[8] Two periods of interest are the flattening of the 'CO2' evolution between ~1935 and ~1955 at all levels, and the changes between 1975 and 1979 seen in 'WmGhgO3', 'Anth' and 'AllForc'. The flattening of 'CO2' results from

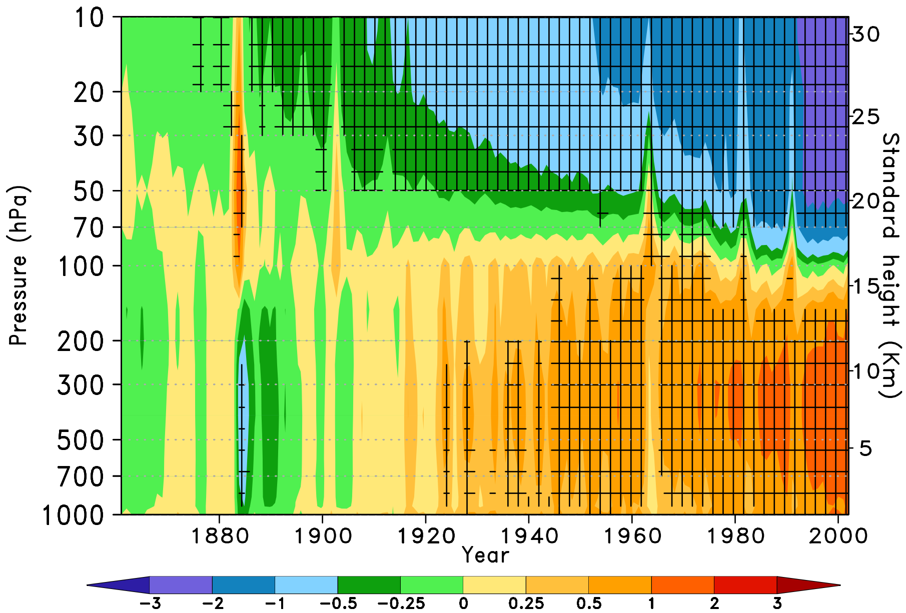

Figure 2. Evolution of the global, annual-mean temperature change relative to 1861 due to changes in all known anthropogenic plus natural forcing agents ('AllForc'). The hatched area denotes the time-height region in which the temperature change from 1861 exceeds twice the standard deviation of annually averaged temperatures for years "1001–2000" of the 'unforced' simulation.

an almost constant CO_2 concentration in that period in the prescribed input (http://www.c4mip.cnrs-gif.fr/C4MIP/ATMCO2/co2force_spline.gr). Since CH_4 and N_2O concentrations increase in this period, a noticeable difference emerges between 'CO2' and 'WmGhgO3'.

[9] The interpolation of stratospheric ozone amounts between 1975 and 1979 introduces an artifact in the temperature evolution of 'WmGhgO3', 'Anth' and 'AllForc' over this period. Separate tests show that this artifact produces only a very small change to the post-1979 temperature evolutions at 30 and 50 hPa; at 10 hPa, the 'WmGhgO3', 'Anth' and 'AllForc' evolutions would show a contribution from ozone depletion, but remain dominated by the effects of CO_2 increase.

[10] We next display the 'AllForc' response at all atmospheric levels (Figure 2), with particular emphasis on the onset of the statistically significant response and its retention over the rest of the period. The stratospheric temperature decline becomes significant (hatched area) by ~1905 at ~30 km (10 hPa) and at ~24 km (30 hPa) and by ~1920 at ~21 km (50 hPa). By contrast, the signal reaches significance only after ~1970 at 70 hPa, and never, in a sustained manner, at 100 hPa. The warming signal in the troposphere reaches significance, at most altitudes, near ~1950, ~30–60 years after the attainment of significance in the lower-to-middle stratosphere. The cooling signal (warming in the troposphere) then increases with time at all altitudes. Volcanic aerosol-induced warming produces a large, statistically significant stratospheric temperature increase which briefly arrests the long-term cooling trend [RS02; SR02; *Santer et al.*, 2006; *Ramaswamy et al.*, 2006].

Table 1. Simulated and Observed Global and Northern Hemispheric Temperature Trends Over the 1964–1979 period at 30 hPa and 50 hPa and Over the 1979–2003 Period in the Lower Stratosphere[a]

			30 hPa			50 hPa		
Period	Region	Season	CM	RATPAC	FUB	CM	RATPAC	FUB
1964–1979	Global	annual	−0.34 (.20)	−0.40 (.29)		−0.52 (.22)	−0.54 (.32)	
	N. Hem.	annual	−0.30 (.18)	−0.25 (.29)	−0.37 (.35)	−0.45 (.18)	−0.48 (.34)	−0.59 (.32)

			Lower stratosphere	
Period	Region	Season	CM	MSU
1979–2003	Global	annual	−0.35 (.14)	−0.35 (.17)
	60N–90N	JJA	−0.47 (.11)	−0.37 (.20)
	60S–90S	DJF	−0.92 (.15)	−0.67 (.35)
	60S–90S	SON	−1.66 (.44)	−0.52 (1.80)

[a]Temperature trends measured by K/decade; 30 hPa, ~24 km, 50 hPa, ~21 km; lower stratosphere is T4 level. "CM" denotes the model simulation. "FUB" (K. Labitzke et al., The Berlin stratospheric data series, 2002, available at http://strat-www.met.fu-berlin.de/products/cdrom/html/main.html) and "RATPAC" [*Free et al.*, 2005] are datasets constructed from rawinsonde observations. "MSU" is the MSU (T4) dataset from RSS (ftp://ftp.ssmi.com/msu/data, version 3.0). CM values for this level use the T4 altitude weighting function (ftp://ftp.ssmi.com/msu/weighting_functions/std_atmosphere_wt_function_chan_4.txt). "FUB" is available for Northern Hemisphere only. Trend uncertainties (standard error) are given in parentheses.

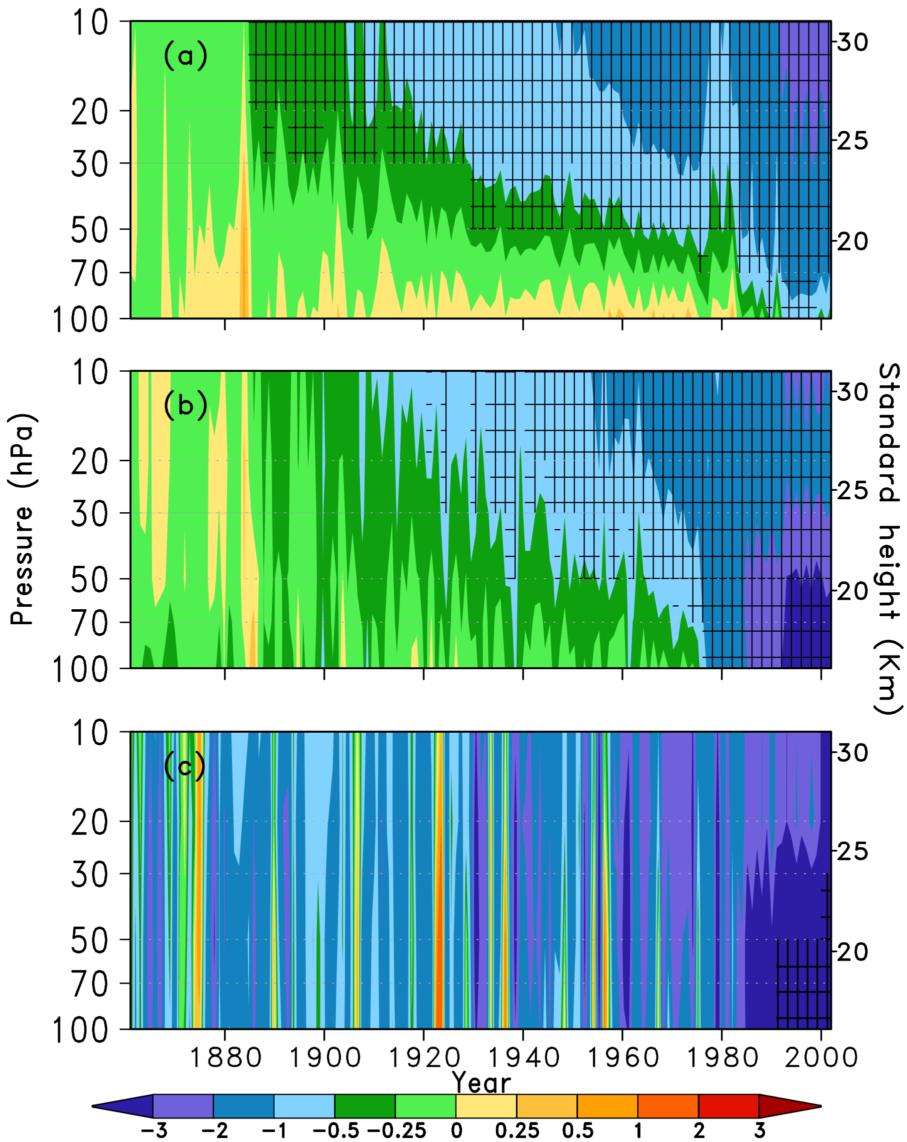

Figure 3. Evolution of the seasonal temperature change relative to 1861 for 'AllForc' for: (a) Arctic (60N–90N) summer (JJA); (b) Antarctic (60S–90S) summer (DJF); (c) Antarctic spring (SON). Hatched areas denote the time-height region with statistically significant temperature change, as in Figure 2.

[11] Together, these results show: (1) that the global, annual-mean signal above ~21 km is statistically significant by the early 20th century, mainly due to increases in CO_2; (2) this signal is detectable prior to the onset of decreases in stratospheric ozone in the late 20th century, which then contributes to enhanced cooling [RS02; SR02; *Shine et al.*, 2003; *Ramaswamy et al.*, 2006]; (3) the signal attains significance earlier in the middle stratosphere than in any lower region.

[12] Comparison of the global, annual-mean CM signal with annual observational trends is possible only from the early 1960s, when reliable, widespread rawinsonde observations became available; global satellite observations exist only from ~1979. We thus compare the CM results to two rawinsonde-based datasets, FUB (K. Labitzke et al., The Berlin stratospheric data series, 2002, available at http://strat-www.met.fu-berlin.de/products/cdrom/html/main.html) and RATPAC [*Free et al.*, 2005], over the 1964–1979 period, and to satellite observations thereafter. Both the model and the rawinsonde observations show a cooling trend (Table 1). The trends at 30 and 50 hPa generally agree to within ~25 percent, and exhibit statistical significance, except for the Northern Hemisphere RATPAC trend at 30 hPa. The lower stratospheric CM temperature evolution post-1980 has been shown to be in substantial agreement with MSU observations [*Ramaswamy et al.*, 2006] (Table 1). We have not compared the middle stratospheric model trends to SSU satellite measurements, owing to continuing uncertainties in satellite measurements and trends.

3.2. Regional and Seasonal Response

[13] The global-mean evolution of 'AllForc' in the four seasons (DJF, MAM, JJA, SON) at 10, 30 and 50 hPa (not shown) is very similar to the evolution of the global, annual-mean at these levels. The annual and seasonal tropical

evolution (not shown) is also similar to the global-mean pattern, except that significance is attained somewhat later.

[14] In the Arctic, the seasonal evolution shows substantial cooling in summer (JJA, Figure 3a), reaching significance by ~1890 at ~30 km and by ~1950 at ~21 km, almost as soon as the global, annual-mean response. The response below ~70 hPa never reaches sustained significance. In other seasons, the response never attains significance due to the large interannual variability.

[15] The Antarctic summer response (DJF, Figure 3b) is similar in magnitude but reaches significance considerably later (mostly after 1940), due to a greater interannual variability in the CM during this season than in the northern polar summer (At 50 hPa, the 2-sigma interannual variability of the 'unforced' integration is ~0.27 K in the Arctic summer and ~0.44 K in the Antarctic summer; the global, annual-mean value is ~0.12 K). During spring, the Antarctic response (SON, Figure 3c) is small prior to ~1980; afterwards, the large stratospheric ozone losses result in a large (>−3 K), temperature decrease, significant below ~21 km. after ~1990. Comparison with available MSU4 decadal trends between 1979 and 2003 for these three seasons (Table 1) indicate that the model trends generally correspond to the observations, although the model variability is considerably smaller than that of the MSU.

4. Conclusions and Implications

[16] This study shows that the CM simulation infers a signal of human influence upon the atmosphere in the global-mean lower-to-middle stratosphere by early in the 20th century, when carbon dioxide concentrations had increased by only about 10% over pre-industrial values (to ~300 ppmv). The early quantitative detection of the anthropogenic signal in this region is possible because: (1) the evolution of WMGGs and their radiative forcing is known accurately [*Ramaswamy et al.*, 2001b; *IPCC*, 2007]; (2) the stratosphere is strongly sensitive to radiative perturbations [*Fels*, 1984; *Kiehl and Solomon*, 1986]; (3) the resulting global-mean temperature evolution of this region between 1861 and 2003 is largely a radiative response explained by well-understood physics [*Manabe and Wetherald*, 1967; *Ramanathan and Dickinson*, 1979; *Fels et al.*, 1980; *Goody and Yung*, 1989; *Ramaswamy et al.*, 2001a]; (4) The CM representation of radiative processes is calibrated to high accuracy against benchmark computations [*Freidenreich and Ramaswamy*, 1999; *Schwarzkopf and Ramaswamy*, 1999]. The cooling signal is evident in this region earlier than in any lower region; in particular, it occurs before the tropospheric warming signal becomes significant after ~1950.

[17] The relatively small variability and large signal in the global-mean stratospheric temperature evolution indicates that the attribution of this trend mainly to increased CO_2 is unlikely to change if different choices of atmospheric model, forcing agent evolution and statistical methods are employed. Changes in stratospheric water vapor [*Ramaswamy et al.*, 2001a; *Shine et al.*, 2003] may affect the trend but there are considerable uncertainties in the estimates of global changes. Extension of the model into the upper stratosphere, use of interactive stratospheric chemistry or adoption of an improved ozone dataset (pre-1979) could improve quantitative aspects of the results, but are not very likely to alter the fundamental inferences of this study, since, as outlined above, the global-mean stratospheric temperature response to constituent changes is primarily radiative.

[18] The larger values for regional, seasonal interannual variability of stratospheric temperatures compared to the global-mean value affect the emergence of a significant signal. However, the early significance seen in the Arctic summer evolution (similar to SR02) suggests that monitoring that region may be fruitful for the detection of changes due to stratospheric forcing agents primarily affecting the Northern Hemisphere.

[19] The correspondence of CM and observed annual-mean trends after ~1960 suggests that the model response to changes in forcing agents may approximate the atmospheric response, even before ~1960. Thus, the early appearance of a statistically significant model signal of temperature change in the lower-to-middle stratosphere, especially as compared with the signal of surface temperature change, confirms that monitoring of this region is especially useful for detecting the early impacts of changes in WMGGs, especially CO_2 [*Kiehl*, 1983]. Similarly, future changes expected from ozone recovery, changes in aerosol concentrations, and changes in stratospheric water vapor might be detected rapidly using temporally continuous satellite temperature measurements of this region. In summary, these findings provide a strong basis for concluding that man-made emissions have exerted a significant influence upon the atmosphere virtually from the beginning of the industrial period, and that observations of temperature changes in the stratosphere may provide an early signal of the atmospheric response to changes in forcing agents, especially the long-lived greenhouse gases.

[20] **Acknowledgments.** We thank J. Austin and R. Stouffer for comments.

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