Fingerprint of ozone depletion in the spatial and temporal pattern of recent lower-stratospheric cooling

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OBSERVATIONS of air temperatures in the lower stratosphere from 1979 to 1990 reveal a cooling trend that varies both spatially and seasonally[1]. The possible causes of this cooling include changes in concentrations of ozone or of other greenhouse gases[2,3], and entirely natural variability, but the relative contributions of such causes are poorly constrained. Here we incorporate the observed decreases in stratospheric ozone concentrations[4] over the same period into a general circulation model of the atmosphere, to investigate the role of the ozone losses in affecting patterns of temperature change. We find that the simulated latitudinal pattern of lower-stratospheric cooling for a given month through the decade corresponds well with the pattern of the observed decadal temperature changes. This result confirms the expectation, from simpler model studies[2,3,5], that the observed ozone depletion exerts a spatially and seasonally varying fingerprint in the decadal cooling of the lower stratosphere, with the influence of increases in concentrations of other greenhouse gases being relatively small. As anthropogenic halocarbon chemicals are important causes of stratospheric ozone depletion[2,3], our study suggests a human influence on the patterns of temperature change in the lower stratosphere over this 11-year period.

The ozone molecule, through its absorption of ultraviolet and visible solar radiation and absorption and emission of longwave radiation, is an important component in the stratospheric energy balance. Thus, a loss of ozone is expected to perturb the climate of the stratosphere[6–12]. The version of the GFDL 'SKYHI' general circulation model (GCM) used here to assess such perturbations, has a latitude–longitude resolution of $\sim 3° \times 3.6°$ and has 40 layers extending from the surface to 80 km (ref. 13). The model has fixed cloud distributions in the troposphere and climatologically varying sea surface temperatures. The total air-column values of the zonal, monthly-mean ozone losses over the past decade are taken from satellite measurements[4]. The observed ozone losses occur principally in the lower stratosphere (up to ~23 km, depending on latitude) of the middle (20°–60°) to high (60°–90°) latitude zones almost throughout the year[2]. There exists some uncertainty concerning the exact vertical profile of the loss, particularly around the tropopause region, which is important for the radiative perturbations[2,3,14,15]. Following other satellite[16] and ground-based observations[2], we assume a simple vertical profile of the loss, with the depletion extending from the model's climatological 'tropopause' level to ~7 km above it and with a constant percentage loss within this altitude range[12]. The model's climatological 'tropopause' pressure level is specified to vary linearly with latitude, from 92 hPa (~17 km) at the Equator, to 180 hPa (~12 km) at 45° and 280 hPa (~9 km) at the poles. This definition, together with the assumed vertical loss profile, is aimed at describing the influence of ozone losses that occur at altitudes just above the troposphere–stratosphere boundary. The GCM's 'control' and perturbation experiments are each run for 10 years, and the time-averaged response over this period is analysed.

The GCM's zonally and annually averaged profile of the temperature response due to the radiative perturbation (Fig. 1) shows a cooling in almost the entire lower-stratospheric region. The cooling is greater in the northern than in southern middle latitudes owing to larger ozone losses (hence, a greater radiative perturbation) occurring more towards the Equator in the Northern Hemisphere[4]. There is a cooling even in those regions where there are no ozone losses imposed in the model, for example the lower stratosphere between the Equator and 15° latitude. Above the domains with both ozone loss and lower-stratospheric cooling, there is a warming, particularly evident at the Southern Hemisphere high latitudes. These features are a manifestation of the dynamical changes in which the ozone-induced radiative perturbations alter the circulation in the global lower stratosphere, increasing the net adiabatic cooling in the tropics and the subsidence-related heating at higher latitudes. The dynamical changes are qualitatively similar to those found in a GCM study of the Antarctic ozone hole[10]. There is also a cooling below the ozone-loss region in the global upper troposphere, in part due to reduced infrared emission from the stratosphere. Figure 1 indicates that the annual-mean response is highly significant (that is, it has a high signal-to-noise ratio) between ~13 and 21 km in the ~20°–50°

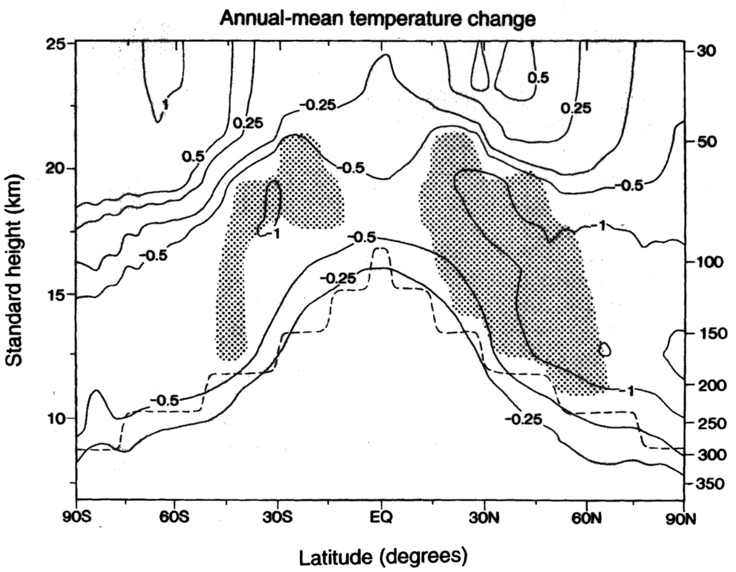

FIG. 1 Change in the zonally and annually averaged vertical profile of temperature, as simulated by a general circulation model (GCM) due to the ~1979–90 observed global ozone losses[4,16]. Shaded areas in the figure show statistical significance at the 99% confidence level, as determined using a Student's *t*-test. The assumed 'tropopause' level (thin, dashed line) in the model is also indicated.

latitude belt. The changes at high latitudes (>60°) fail the significance test because of large interannual variability in those regions, a feature corroborated by observations[17,18].

The latitude-month pattern of the decadal (~1979–90) temperature change simulated by the model, in the altitude range of the ozone changes (Fig. 2*a*), is compared with that derived from satellite observations[1] (Fig. 2*b*) of the lower stratosphere. The satellite-derived general cooling trend in the lower stratosphere is well supported by the radiosonde observations[19–21]. The observed decadal cooling in the middle-to-high latitudes exceeds 0.5 K for most of the year, reaching 2.5 K in the polar spring in both hemispheres; these values constitute substantial changes over a relatively short period. The simulated result also exhibits a marked cooling over a broad latitude–month domain, especially in the middle and high latitudes. The middle latitudes in Fig. 2*a* and *b* show a cooling from January to October in the Northern Hemisphere and from September to July in the Southern Hemisphere. The cooling in the middle latitudes of the Northern Hemisphere from about December to July, and in the Southern Hemisphere from about December to May, is statistically significant in both the model and in observations. Near the poles, both the simulations and the observations exhibit a cooling pattern almost throughout the year, with the occurrence of relatively large magnitudes during winter and spring. The simulated cooling in the Antarctic is highly significant during the austral spring (period of the 'ozone hole'), consistent with observations. The cooling in the Arctic during spring does not show a high significance in either Fig. 2*a* or *b* owing to a large dynamical variability[13,18,20]. The simulated cooling in the tropics is not significant for most of the year owing to the small temperature changes. There are some differences in patterns between the simulated and observed trends, especially during certain seasons in the polar regions. In particular, a larger domain of significance is seen in the simulation. This is due to a smaller variability in the model compared with the observations. Uncertainties arise in the simulation owing to incomplete observational knowledge of the vertical profile of global ozone loss near the tropopause, including that in the tropical areas[2,3,16]. While more comprehensive altitudinal measurements of ozone loss would lead to more precise simulations of temperature change, with cooling perhaps extending to even higher altitudes (for example springtime southern polar latitudes[10]), the lower stratosphere region, taken as a whole, can be expected to cool significantly given the magnitude of the observed[4,16] ozone losses. We conclude that the reasonable consistency of the simulated cooling pattern and magnitudes with those observed, including the regimes of statistically significant changes, coupled with the high correlations noted between

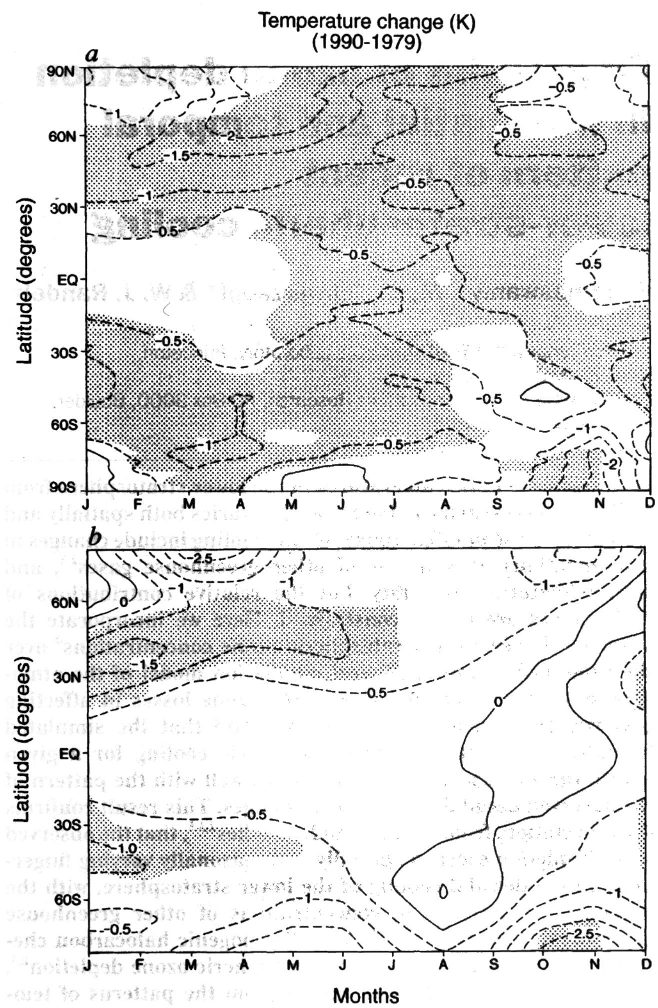

FIG. 2 Zonal, monthly-mean pattern of the lower-stratospheric temperature change over the past decade (~1979–90). *a*, As simulated by the GCM (90° S to 90° N) due to the observed global ozone depletion[4,16]; *b*, as inferred from satellite observations (82.5° S to 82.5° N)[1]. The satellite temperature trends are derived for an atmospheric layer essentially comprising the lower stratosphere[1]. The simulated results denote the mean temperature trends in the altitude region of the model where ozone concentration changes occur. Shaded areas denote the statistical significance at the 95% confidence level[1].

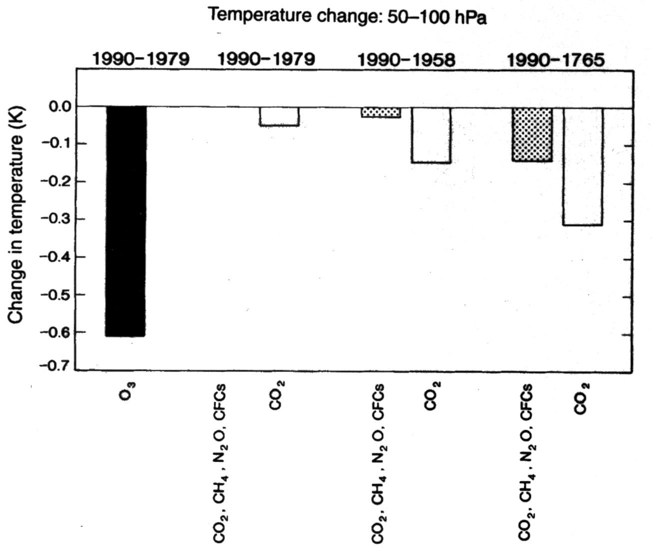

FIG. 3 Computed global, annual-mean temperature changes in the ~50–100 hPa (~16–21 km) lower-stratospheric region due to different trace-gas changes. The result for ozone corresponds to the ~1979–90 losses, and is computed using a GCM. The results for CO_2 alone, and for all the well mixed greenhouse gases (CO_2, CH_4, N_2O and CFCs) are calculated using a one-dimensional, radiative–convective model for three different periods: 1979 to 1990, 1958 to 1990 and 1765 to 1990. Note that the result for all the well mixed gases in the period 1979–90 is too small to be distinguishable from zero.

observed temperature changes and ozone losses[1], confirms the notion that ozone depletion has caused a substantial spatially and seasonally dependent effect in the lower stratosphere over the past decade.

To evaluate the importance of ozone depletion relative to changes in the concentrations of other greenhouse gases that are well mixed (CO_2, CH_4, N_2O, chlorofluorocarbons), we employ a radiative–convective model and determine the global, annual-mean temperature change in the lower stratosphere due to the known increases in their concentrations. The model's radiation and convection schemes follow those used in earlier studies[12,22]. The calculations are performed for three different periods[23]: ~1979 to 1990, 1958 (ref. 24) to 1990 (the period since routine CO_2 measurements began) and 1765 to 1990 (the period since the beginning of the industrial era). In Fig. 3 the global, annual-mean GCM temperature change due to the decadal ozone losses in the ~50–100 hPa (~16–21 km) lower-stratospheric region is compared with the corresponding effects due to increases in CO_2 only, and all well mixed greenhouse gases taken together. The global-mean decadal cooling due to ozone is ~0.6 K, with the middle latitudes (Fig. 1) making a substantial contribution; the value is comparable to the reported decadal trends[3,25]. Although the increase in CO_2 alone since 1765 yields a cooling of ~0.3 K, inclusion of the other well mixed gases, which together tend to warm the tropopause region[26] gives a cooling of ~0.15 K. The overall cooling effect in the lower stratosphere due to increases in the well mixed greenhouse gases contrasts with their warming effect on the surface[24,26]. It is thus clear that the computed 1979–90 ozone effect on lower-stratospheric temperature outweighs the effects of changes in other gases, not only over the past 10–30 years, but also over the past two centuries. This sharp contrast in the effects of ozone in relation to the other gases is qualitatively similar to other global-mean estimates[3,11,27].

Possible secular changes in other radiatively active species[22], including stratospheric volcanic aerosols and water vapour[28], are estimated to contribute considerably smaller decadal effects than the stratospheric ozone loss. Information on decadal changes in clouds is insufficient to estimate their influence. There is little evidence to suggest that forcings from the troposphere (for example, sea surface temperature changes[29]) or natural climate variability (gauged from the 'control' simulation here and in other analyses[30,31]) have significantly influenced the decadal global lower-stratospheric temperature change, although in the absence of rigorous long-term observations, a precise estimate of their contributions cannot be obtained. The knowledge available at present, and current model simulations strongly suggest that the 1979–90 ozone loss has played a dominant role in the latitude–month pattern of cooling observed in the global lower stratosphere. It is noted that ozone loss has also been reported[27] for the 1970s, although the observations then did not span the globe. The 1970s losses are also estimated[11,27] to have contributed significantly to the observed cooling[19–21] during that decade.

An ozone-induced cooling of the lower stratosphere implies a reduction in the longwave radiative emission from the stratosphere into the troposphere. This is a mechanism that leads to a negative radiative forcing of the surface–troposphere system[2,3,12]. The dynamically induced cooling of the tropical lower stratosphere, which we have found, suggests a negative surface–troposphere forcing at the low latitudes. This is a feature that could not be predicted by earlier calculations, and which would enhance the global-mean negative forcing caused by the observed stratospheric ozone loss[2,12].

Because of the connection with the emissions of man-made halocarbon chemicals[2,3], the depletion of ozone and its spatial and temporal fingerprint on global lower-stratospheric temperatures becomes a major anthropogenic component of stratospheric climate change—one that has occurred on a much shorter timescale (~one decade) than the effects due to the well mixed greenhouse gases (several decades). Further, the ozone-loss-induced spatial and seasonal cooling near the stratosphere–troposphere boundary becomes an important factor to be considered in the search for anthropogenic effects on the vertical temperature profile record[11,30]. □

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