GEOPHYSICAL RESEARCH LETTERS, VOL. 19, NO. 19, PAGES 1927-1930, OCTOBER 2, 1992

BUFFERING OF STRATOSPHERIC CIRCULATION BY CHANGING AMOUNTS OF TROPICAL OZONE A PINATUBO CASE STUDY

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ABSTRACT. Stratospheric aerosol from Mount Pinatubo heated the tropical lower stratosphere by about 0.3 K/day mainly due to absorption of terrestrial infrared radiation. This heating was dissipated by (1) an observed increase in stratospheric temperatures, which enhanced the radiative cooling, (2) additional mean upward motion, observed for the aerosol cloud, which led to adiabatic cooling and (3) reductions in ozone concentrations resulting from enhanced upward motions. Each of these processes operated on a different time scale: maximum temperatures were observed after about 90 days; maximum ozone losses of about -1.5 ppm occurred after 140 days when the enhanced vertical velocities effectively lifted the ozone profile by about 2 km. We believe this shows that ozone plays an important role in buffering vertical motion in the tropical lower stratosphere, and hence the residual Brewer Dobson circulation of the whole stratosphere.

Introduction

Solar and infrared heating of stratospheric aerosols from the eruption of Mt. Pinatubo in June 1991 raised stratospheric temperatures in the tropics by as much as 3°C, as had aerosols from earlier eruptions of Mt. Agung in 1963 and El Chichon in 1982. We show that aerosol heating in the Pinatubo cloud was sufficiently strong to account not only for warmer temperatures, but also for an additional upward motion. Increased upward motion led to an ozone decrease, which provided a strong negative feedback on the heating rates.

Below, stratospheric heating rates for aerosols resembling those from Mt. Pinatubo are derived. Then, potentials for warming and lifting are analyzed, including buffering by associated feedback processes. Comparisons to observations are made. Next, the relation between vertical motion and ozone in the tropics is discussed. Finally, simple feedback simulations are performed and derived ozone changes are compared to observed ozone variations.

Aerosol Heating Rates

A two-stream radiative transfer model is used to determine the heating and cooling rates of Mt.Pinatubo aerosol in a tropical atmosphere. The presence of tropospheric clouds is included, as cold clouds in particular reduce the potential of infrared heating for the aerosol layer. Based on a 3D-NEPH cloud climatology for the tropics, we assume 55% low clouds (below 2 km, τ=25), 35% mid-level clouds (warmer than 250 K, τ=18) and 15% high clouds (colder than 250 K, τ=2). Our model calculations assume random cloud overlap.

The radiative effects of atmospheric particles depend on refractive indices, shape, size and abundance. Volcanic stratospheric particles are primarily composed of supercooled sulfuric acid solution droplets for which refractive indices are known. Single scattering properties are determined, using Mie theory, for two bimodal aerosol particle size distributions. Defining log-normal parameters are given in Table 1 where N, r_m, σ and r_e are the particle density, mode radius, standard deviation and effective radius (volume/surface).

Table 1. Size distribution parameters

aerosol type	r_e μm	'small mode' r_m μm	σ	N cm^{-3}	'large mode' r_m μm	σ	N cm^{-3}
type(1)	0.4	0.08	1.39	33	0.25	1.61	26
type(2)	0.6	0.09	1.45	17	0.33	1.68	11

Type(1) aerosol represents average values, type(2) maximum values, for r_m and σ from midlatitude ER-2 aircraft measurements at altitudes of 18 km several months after the Mt.Pinatubo eruption (R. Pueschel, private communication). Values for N reflect visible extinction coefficients of 0.025 /km for type(1) and 0.019 /km for type(2).

Sun-photometer measurements at Hawaii (E. Dutton, private communication) show a steady increase in aerosol effective radius r_e from about 0.4 at two months to 0.6 μm at six months after the eruption. Monthly averaged visible aerosol optical depths (extinction), reached a maximum of about 0.21 three months after the eruption, followed by a steady decline, to about 0.15 at month six. 8 km thick layers of the two aerosol types (Table 1) appear to bound the 'Pinatubo' properties for two to six months after the eruption. The aerosol placement between 18 and 26 km altitude is based on lidar observations in the tropics (e.g. DeFoor et al., 1992).

Background heating rates and heating rate changes in the Mt. Pinatubo aerosol layers are given in Table 2. Infrared heating, due to the large thermal difference between the aerosol

Paper number 92GL01937
0094-8534/92/92GL-01937$03.00

layer and the tropical troposphere, is much stronger than (daily averaged) solar heating. A similar aerosol heating of 0.3 K/day is computed for both Pinatubo cases, as, over time, radiative effects of increasing particle size are matched by effects of decreasing optical depth. Thus, for the first half-year after the eruption, heating in the aerosol layer remained near 0.3 K/day.

We find that an enhanced cirrus cloud cover (from 15% to 24%), potentially due to the El Nino coinciding with Pinatubo, is unable to offset Pinatubo heating, as aerosol heating would be reduced by only 0.08 K/day above 25 km and less than 0.05 K/day below 22 km altitude (Table 2).

Sources of Cooling

Radiative heating of the stratosphere by the Pinatubo aerosols must be balanced by warmer temperatures, enhanced upward motions (adiabatic cooling), increased eddy transports of heat, or a combination of all. We examine the first two as the primary response of a tropical stratosphere to Pinatubo.

Warmer Temperatures

A warmer stratosphere cools more effectively. The effects of 3, 6 and 9 K warmer temperatures on aerosol type(1) heating rates are given in Table 2. To offset a heating of 0.3 K/day below 25 km altitude a temperature increase of 5-9 K is required. These results agree with similar calculations of Pollack and Ackerman (1983), who show that the time needed to achieve these temperature changes is about two months. However, observed temperature increases in the tropical lower stratosphere since the eruption in June 1991 (Labitzke and McCormick 1992) are much smaller, as shown in Table 3.

The warmest temperatures occur after about three months, but compensate then for only 0.15 and 0.12 K/day aerosol heating at 24 and 21 km. This is less than half the aerosol heating, thus, other methods of cooling can be expected.

Adiabatic Cooling

A scale analysis of the thermodynamic equation for the lower stratosphere in the tropics (Dunkerton, 1978) indicates that the heating term is balanced by the vertical motion term, so that

$$w^* = (1/T \cdot Q/c_p) \; / \; (d\ln\theta/dz)$$

Q/c_p is the aerosol induced differential heating, T the atmospheric temperature, θ the potential temperature and $d\theta/dz$ the lapse rate for the potential temperature. If the added heating of 0.3 K/day were countered by adiabatic ascent, the matching vertical motion would be 1/40 km/day or 0.75 km/month (superimposed on any preexisting vertical motion). This lift is reduced, not only by direct warming, but also due to perturbations to the ozone profile discussed below.

Evidence of upward motion is given by a time-series of lidar data (DeFoor et al., 1992). Derived lifting rates are summarized in Table 4.

The lower values for the 'mean lift' during the initial months, take into account that the aerosol plume had not spread over all longitudes. The lifting rates refer to the monthly movement at cloud base, as lifting rates at the cloud top are distorted by sulfate chemistry. The observed rise of the aerosol plume slowed down over time and ceased after a total lift of 1.8 km.

Chemical Response: O3 Reductions

Changes in temperature and vertical motion may indirectly affect the heating rates by perturbing ozone. Small ozone reductions are expected from warmer temperatures and altered photolysis rates (Michelangeli et al. 1989). Heterogenous reactions on volcanic aerosol (Hofmann and Solomon 1989, Prather 1992) are not very effective in the tropics, because ozone production greatly exceeds loss.

We have simulated ozone concentrations with a mechanistic model for upwelling stratospheric air based on the photochemical model of Prather and Jaffe (1990). A parcel of air with clean upper tropospheric composition ascends in a tropical atmosphere. A key assumption of this simplistic model is that chemical evolution of the parcel is driven by local photochemistry, not by horizontal mixing with the extratropics. This approximation may hold at best only at low latitudes ($\pm 15^o$) and within a scale height of the tropopause.

The calculated O3 profiles for different ascent rates are compared in Table 5 to the ozone profile of a standard tropical atmosphere and to observed ozone profiles near the equator prior to the Mt. Pinatubo eruption (Grant et al., 1992).

The standard tropical ozone profile can be approximately reproduced with the photochemical model assuming an ascent rate of about 1/60 km/day. Such an ascent rate is in agreement with average heating rates based on NMC data prior to the Pinatubo eruption (Holton, 1990). If all of the Mt. Pinatubo aerosol heating of about 0.3 K/day led to an upward motion, the ascent rate would increase to about 1/25 km/day, and ozone reduction would be significant.

Table 2. Differential (to background) heating rates for aerosol type(1) and type(2) (K/d)

height z(km)	bgrd. heat.	type(1)-heating total (SOL,IR)	...with warmer T T+3K	T+6K	T+9K	...with lifted O3 1km^	2km^	3km^	cloud Ci+9%	type(2) heating total (SOL,IR)
25-26	0.06	0.30 (.08 .22)	-.02	-.28	-.55	0.12	-.05	-.25	0.22	0.35 (.09 .26)
24-25	-.07	0.32 (.07 .25)	0.13	-.08	-.31	0.10	-.12	-.30	0.24	0.35 (.07 .28)
23-24	-.09	0.30 (.05 .25)	0.14	-.03	-.22	0.09	-.09	-.20	0.24	0.35 (.06 .29)
22-23	-.08	0.29 (.04 .25)	0.16	0.02	-.13	0.13	0.02	-.07	0.24	0.34 (.05 .29)
21-22	0.02	0.28 (.03 .25)	0.16	0.04	-.08	0.18	0.10	0.02	0.23	0.33 (.04 .29)
20-21	0.23	0.27 (.03 .24)	0.17	0.06	-.05	0.18	0.11	0.07	0.23	0.32 (.03 .29)
19-20	0.31	0.25 (.03 .22)	0.18	0.08	-.01	0.18	0.13	0.12	0.22	0.30 (.03 .27)
18-19	0.22	0.23 (.02 .21)	0.16	0.08	-.01	0.21	0.21	0.20	0.20	0.28 (.03 .25)

Table 3. Temperature changes since June 1991 (K)

	30 mb (24.0 km) 10°N	30 mb (24.0 km) 20°N	50 mb (20.8 km) 10°N	50 mb (20.8 km) 20°N
July 1991	+ 1.3	+ 1.0	+ 1.2	+ 0.9
Aug. 1991	+ 2.6	+ 1.8	+ 3.1	+ 2.5
Sep. 1991	+ 3.2	+ 2.3	+ 3.5	+ 2.8
Oct. 1991	+ 2.7	+ 2.0	+ 2.6	+ 2.2

Some caution is due with such a simple model for atmospheric transport. It can accurately reproduce the O3 profiles with ascent rates believed to be typical of the lower tropical stratosphere (e.g. Holton, 1990), and it also matches the correlation of CO and O3 above the tropopause (Murphy et al. 1992). However, the fall-off of N2O above the tropopause (i.e. the correlation between O3 and N2O) is predicted to be much smaller than observed (WMO, 1981; Murphy et al., 1992). One obvious implication is that the horizontal mixing of air from the extratropics (with low N2O) must be occurring simultaneously with the rapid ascent or at least when averaged over the annual cycle. Such processes should not significantly affect the scale analysis of the impact of vertical motions on the ozone profile presented here.

The net result of our calculations is an effective 'lifting' of the ozone profile (e.g. by 2 km from 1/60 to 1/25 km/day). The effects of lifted ozone profiles by 1, 2 and 3 km on aerosol type(1) heating rates due to lift associated reductions in solar ozone heating are given in Table 2. Strongest impacts occur above 24 km altitude, where ozone reductions of about 1.5 ppm due to a 1.5 km lift are sufficient to balance a heating of 0.3 K/day. Below 21 km, the ozone feedback mechanism cannot work as ozone heating rates are inconsequential. At these altitudes the aerosol concentrations will decrease, as the lower edge of the layer lifts, and thus, the excess heating will be automatically reduced.

Table 4. Monthly upward movement (km)

	Jun/ Jul	Jul/ Aug	Aug/ Sep	Sep/ Oct	Oct/ Nov		Jun/ Nov
plume lift	+1.5	+0.8	+0.4	+0.1	+0.0		
mean lift	+0.7	+0.6	+0.4	+0.1	+0.0	=	+1.8

Table 5. Ozone vertical profile (ppm, per volume)

height (km)	std. trop.	obs. trop.	ascent rate (km/day) 1/25	1/40	1/60	1/80
24.5	4.85	4.90	2.92	4.05	4.96	5.49
23.5	3.85	3.75	2.04	2.93	3.75	4.28
22.5	2.90	2.70	1.40	2.05	2.73	3.21
21.5	2.10	1.85	.96	1.41	1.92	2.31
20.5	1.60	1.20	.65	.94	1.30	1.60
19.5	1.15	.68	.44	.62	.85	1.06
18.5	.70	.37	.29	.40	.54	.66
17.5	.35	.15	.20	.26	.33	.40
16.5	.19	.10	.14	.16	.19	.22

Table 6. Aerosol properties. Extinction profile

	June/ July	July/ Aug.	Aug./ Sept.	Sept./ Oct.	Oct./ Nov.
τ (.55μm)	0.15	0.20	0.21	0.19	0.17
r_e (μm)	0.25	0.40	0.45	0.50	0.55
27-28 km				*	*
26-27 km		*	**	**	**
25-26 km	*	**	***	***	***
24-25 km	**	***	****	****	***
23-24 km	***	***	****	***	***
22-23 km	***	***	***	***	**
21-22 km	**	***	**	**	**
20-21 km	**	**	**	*	*
19-20 km	*	**	*		
18-19 km	*	*			

where '*','**','***' and '****' represent extinctions of .01, .02, .03 and .04/km.

Coupled Simulation

A month-by month simulation with simple temperature, aerosol and ozone feedbacks has been performed. Monthly averages of observed temperatures (Labitzke and McCormick, 1992), observed particle radii r_E and visible optical depth τ (Dutton et al., 1992), and aerosol extinction vertical profiles τ_e [km^{-1}] (DeFoor et al., 1992) are used, as outlined in Table 6.

After accounting for the observed monthly temperature changes, the remaining heating rates are translated into vertical motion, which subsequently modifies ozone concentrations in all layers. The new ozone profile provides initial conditions for next month's calculations.

After November the predicted heating due to Pinatubo is less than .05 K/day. Such small rates signal a new equilibrium, as no further ozone reductions are needed. The simulated vertical lift, with a maximum of 1.75 km in the center of the aerosol layer, compares well to observations presented in Table 4. Ozone changes, associated with the calculated vertical lift, are given in Table 7 and in general agreement to observations.

Table 7. Cumulative O_3 changes (ppm, per volume)

height (km)	O3 June	δO3 Jul	δO3 Aug	δO3 Sep	δO3 Oct	δO3 Nov	obs.* Aug/Sep
28	8.1	.0	.1	-.0	-.2	-.3	-.8
27	7.3	.1	.1	-.2	-.4	-.5	-1.1
26	6.4	-.1	-.3	-.9	-1.3	-1.5	-1.1
25	5.4	-.4	-1.1	-1.5	-1.7	-1.7	-1.1
24	4.3	-.5	-1.0	-1.2	-1.4	-1.4	-.9
23	3.4	-.3	-.9	-1.1	-1.2	-1.2	-.7
22	2.4	-.2	-.4	-.5	-.6	-.6	-.5
21	1.8	-.1	-.2	-.3	-.3	-.3	-.3
20	1.4	-.1	-.2	-.1	-.1	-.1	-.1
19	1.0	-.1	-.1	.0	.1	.1	.1
18	0.5	-.0	-.1	.0	.1	.1	.1

* observed stratospheric ozone changes between April and August/September 1991 at Brazzaville (4°S, 15°E) by Grant et al.(1992).

Our simulated column ozone amounts suggest the largest ozone reductions at four to five months after the eruption. This is in agreement with satellite data (Schoeberl et al., 1992) and other ozonesonde data (J. Fishman, private communication), as shown in Table 8.

Table 8. Ozone column and changes (DU)

	O3 Jun	δO3 Jul	δO3 Aug	δO3 Sep	δO3 Oct	δO3 Nov
simulation	286	-7	-17	-22	-26	-26
TOMS-satellite	256	+5	-2	-8	-14	-18
ozonesondes			-14	-18	-21	-22

Our simulated ozone losses are larger than observed. This overestimate could be expected with this simple model, because we have ignored other mechanisms that cool the tropical stratosphere and could reduce the lift, most notably horizontal mixing. However, horizontal mixing does not seem significant above 23 km, because Pinatubo aerosol above that altitude remained confined to tropical regions, despite warmer temperatures and vertical advection. Three-dimensional models with interactive aerosol ozone chemistry are needed to assess the importance of horizontal mixing at all altitudes.

Conclusion

Heating by volcanic debris from Mt. Pinatubo has not only to warmed the lower tropical stratosphere, but also induced upward motion. Both effects, warming and lifting, provide their own negative feedbacks on aerosol heating. Simulations of these feedback processes show (in agreement to observations), that after about five months a new 'equilibrium' is reached with 1.5 K warmer temperatures and reduced ozone amounts by 10 to 30%. The temperature feedback appears to respond more rapidly than the 'motion induced' ozone feedback; however, the ozone feedback eventually dominates.

This Pinatubo study suggests an automatic buffering link between heating rates, temperature and ozone concentration in the tropical lower stratosphere. We propose that a forced change in the heating rates in the tropical stratosphere will be buffered in part by the temperature change itself, but mainly by changes in ozone concentrations, forced by changes in the vertical motion. Thus, the overall stratospheric-residual circulation appears to be stabilized by this ozone feedback. For example, if greenhouse gases were to perturb the net heating (and hence diabatic circulation) of the tropics to give a faster mean rate of ascent, the resulting ozone decrease would counter the initial radiative forcing. The reverse cycle also works. Perhaps this mechanism - the production of ozone and subsequent ozone heating in the lower tropical stratosphere - stabilizes the Brewer-Dobson circulation over a large range of natural fluctuations.

Further insights into relations between ozone concentrations, heating due to volcanic aerosols or greenhouse gases, and stratospheric circulation require the use of multidimensional models, in which radiation, aerosols, stratospheric chemistry and dynamics are coupled more realistic than in our model.

Acknowledgements. We thank E. Dutton, J. Fishman and R. Pueschel for the use of their data. M. Prather acknowledges the support from the atmospheric chemistry programs of NSF and NASA.

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Received: July 20, 1992;
accepted August 12,1992