NOTES AND CORRESPONDENCE

A Possible Relationship of Volcanic Aerosol Variations with El Niño Southern Oscillations—Lidar Observations of Volcanic Aerosols in the Atmosphere¹

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1. Introduction

Dunkerton (1983) showed that the temperature rise in the tropical (20°S–20°N) stratosphere after Agung volcanic eruption 1963 was consistent with those inferred from the thermal wind structure of the abnormal QBO (Quasi Biennial Oscillations) calculated by his one-dimensional model. The situation would be similar to those after El Chichon eruption 1982 and the recent observed data seem to support this presumption (Labitzke and Naujokat 1984). A necessary condition of his theory is volcanic aerosols confined near the equator for about 5 years. He showed that aerosol atmospheric heating can modify QBO. Also he suggested a second possibility that the modification of QBO was due to the excitation by abnormal equatorial waves in the troposphere which might be related to the ENSO (El Niño Southern Oscillation). This event started almost at the same time as the eruption of El Chichon (Philander, 1983).

In his theoretical treatment the effect of the aerosol was limited within the altitude range from 17 to 30 km and judging from the observed results for the El Chichon event (Hirono *et al.* 1984), the assumed residence time of aerosols near the equator would be

too long.

In this paper we shall make estimates of the atmospheric heating by the aerosols falling through the upper troposphere for one season after the eruption of El Chichon 1982 April. The results suggest that the heating is strong enough to modify the Hadley circulation to initiate the variation of the trade wind leading to the ENSO.

2. Variations of column concentration of aerosols and equatorial Pacific surface temperature

Yag lidar system was used at Fukuoka (33°N, 131°E), since 1979 at the wavelength $F=1.06\ \mu\text{m}$, and $S=0.53\ \mu\text{m}$. The integrated aerosol backscattering $B_c=\int\beta_A(F)dz$ is shown in Fig. 1 for the period from 1979 through Feb. 1983, (Hirono and Shibata, 1983; Hirono *et al.*, 1984).

El Chichon eruption in April 1982 is clearly a tropical one, located at 17.33°N, 93.20°W, according to the classification by Lamb (1970).

In May 1982, $B_c=(1.08\pm0.48)\times10^{-3}\ \text{sr}^{-1}$ and the optical depth at the wavelength 550 nm $\tau(550)$ is about 0.108 ± 0.05 and attained a maximum value 0.2 in January 1983, decreasing to 0.02 in May 1984 about one tenth of the maximum value (not shown in Fig. 1).

Recently Handler (1984) showed that the ENSO took place after very large volcanic eruptions at low latitudes ($<20^\circ$ tropical);

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Fig. 1 Variation of integrated backscattering $B_c(F)$ for the period 1980–1982 and sea surface temperature anomaly averaged over eastern tropical Pacific: 5°S – 5°N , 90°W – 140°W .

Krakatoa 1883, Santa Maria 1902, Agung 1963 and El Chichon 1982. Schatten *et al.* (1984) made numerical calculations by a zonally symmetric model and showed that low latitude volcanic eruptions might trigger ENSO. They suggested that the eruption of El Chichon may provide a trigger for inducing the 1982–83 anomalous ENSO. But the latter authors advocated urgent need for more sophisticated theories on the subject.

The averaged sea surface temperature anomaly (SSTA) over the eastern equatorial Pacific can be a good measure of the ENSO. In Fig. 1 SSTA values over 90°W – 140°W , 5°S – 5°N (Quiroz, 1983) are also shown. It should be noted that the SSTA started to increase dramatically at the same time as the eruption of El Chichon. This fact suggests a much shorter response time of SSTA to volcanic eruptions, than those of Robock (1985) and Hansen *et al.* (1978). It seems to emphasize the importance of the dynamical effect on the atmosphere and ocean, in addition to the radiative and thermodynamical equilibrium models of these authors.

3. Estimation of the volcanic aerosol concentration in the upper troposphere at northern low latitudes and its atmospheric heating

According to Hofmann and Rosen (1983), initial loading of stratospheric El Chichon cloud in April 1982 was about 20 Tg ($T=10^{12}$), probably one half of which settled through tropopause for early several months. The center of the zonal stream of El Chichon cloud was about 20°N (Thomas *et al.*, 1983) until August 1982. The active settling of giant particles at south Texas for early few months can be estimated from the value of accumulated aerosol concentration $N_{.95}$ with radii larger than $.95 \mu\text{m}$ averaged from tropopause to 30 km being about 0.67 and 0.038 in May and August '82, respectively (Hofmann and Rosen, 1983). But only one observation in May is available at south Texas for three months after El Chichon eruption.

If aerosol unit mixing ratio $m_I=10^{-9} \text{ g/g}$ air is used, in May near the altitude 10 km at south Texas the aerosol mixing ratio $m(a)$ is estimated at $m(a)\sim 5.6m_I$. At Fukuoka there are many lidar observation data and the re-normalization was made on May 17, 19, 21 and 22 (Hirono *et al.*, in this issue) giving $m(a)=f(70\pm 70)m_I$ ($0.26 < f < 1$, probably $f\sim 0.5$), for the altitude 9–11 km in this month and $m(a)\sim 10m_I$ on August 19 at the same altitude. The Yag lidar does not emit fluorescence and hence a mechanical shutter is unnecessary. Therefore the profile near 10 km is reliable if the normalization is correct. The Mauna Loa lidar observation data suggest $m(a)\sim 100m_I$ in April and May 1982 near altitude 10 km as shown in the next section. These values give some reliable bases to discuss concentrations of settling aerosols in the upper troposphere.

According to Czeplak and Junge (1974) and Kida (1983), at the tropics from latitudes 0° to 25° in the troposphere the horizontal eddy diffusion is smallest so that the dispersion of the pollutant is mainly due to mean atmospheric motion, i.e., Hadley circulation. In higher latitudes than 25° the horizontal eddy diffusion gradually increases with latitude and is the main cause of dispersion. This point

is an important basis of the present analysis. The meridional mean velocity $\langle v \rangle$ for Mar.-May at 10° - 20° N is about 0.6 m/s and -0.3 m/s (northward positive) at altitudes 10 and 2 km, respectively (Newell *et al.* 1972).

The conventional treatment of settling aerosols through the troposphere is to neglect their radiative impact upon the atmosphere because of rapid dispersion in the troposphere. In the following, we shall show that if the concentration of the settling aerosols is sufficiently large, the atmospheric heating by them may be strong enough to modify the Hadley circulation and the residence time near the injected region would be much longer than the conventional estimates.

Suppose that the mean radius of the settling aerosols through altitude 15 km be $2\ \mu\text{m}$ zonally distributed with peak at 20° N and meridional half width 10° , *i.e.*, from 15° N to 25° N and the vertical profile of the aerosol

Fig. 2 Schematic illustration of zonal El Chichon volcanic cloud. Densest part is shown by shading and dotted region is discussed on radiative and thermodynamical theory in the text.

Fig. 3 Atmospheric diabatic heating rate Q_D for March-May, which includes condensational, radiative, and boundary layer heating (Newell *et al.* 1974).

over 15 km is so disposed that total amount $10\ Tg$ settles down through 15 km in 3 months (Fig. 2). It can be calculated from these figures that at 15 km, 20° N, the aerosol concentration μ_s is $20\ \mu\text{g}/\text{m}^3$ and the number concentration is $0.3/\text{cm}^3$. This may be compared with the observed value at south Texas (28° N) $N_{0.95} \sim 0.2/\text{cm}^3$ in May 1982, which would give less μ_s than the above value.

Then the mixing ratio $m(a)$ at 15 km in the center of the zonal belt is estimated at $10^{-7}\ \text{g/g air} \sim 100m_I$. When the settling continues downward, at the lower altitude 10 km the mixing ratio will be reduced to $60m_I$ due to the increase of air density, although the settling velocity decreases as shown in section 4. The latter value may be several times

I

II

III

Fig. 4 Aerosol profiles and heating rate:

I Profile of aerosol mixing ratio in units $m_I = 10^{-9}\ \text{g/g air}$, which is estimated to be present for one season after eruption of El Chichon April 1982.

II Profiles of heating rate by aerosols Q_A induced by absorption of radiation at 20° N (in units K/day): Curves (A), (B), and (C) denote different aerosol refractive indices (see text).

III Profiles of Q_D at 20° N and 0° N (Newell *et al.* 1974) and $Q_D + Q_A$ at 20° N.

those estimated by lidar at Fukuoka (33°N) $f(70 \pm 70)m_I$ ($f \sim 0.5$). In May the meandering of El Chichon zonal flow was appreciable, and hence at times they might be of the same order of magnitude. At the present stage the influence of the mean Hadley circulation to the distribution of aerosols is neglected, as is that of horizontal diffusion.

We assumed the distribution of settling aerosols vs. altitude as shown in Fig. 4 I for the latitude 20°N , which would be expected for 3 months after the injection. The initial injection of some volcanic aerosols in the same altitude range might be present. The distribution is also assumed to have Lorentzian distribution with half width 10° , *i.e.*, the aerosol mixing ratio at the latitude ϕ is given by $m(\phi) = m(\phi_0)\Delta^2 / \{(\phi - \phi_0)^2 + \Delta^2\}$, where $\phi_0 = 20^\circ\text{N}$ and $\Delta = 5^\circ$.

We calculated atmospheric heating rate Q_A due to aerosols induced by the absorption of solar short wave and terrestrial long wave radiations, according to the theoretical study by Mugnai *et al.* (1978) as shown in Fig. 4 II. The heating rate is compared with diabatic

atmospheric heating rate (radiative, condensational and boundary layer heatings all included) for Mar.-May by Newell *et al.* (1974) which is thought to be the motive power of the atmospheric general circulation (Fig. 3). The curves (A), (B) and (C) in Fig. 4 II are obtained by using aerosol models with refractive index $m = n - ik$:

(A) $\kappa(vis.) = 0$: sulphuric acid aerosol with no impurities. *vis.* denoting the value for the visible wave length range.

(B) $\kappa(vis.) = 0.005$: impure sulphuric acid,

(C) volcanic ash particles.

The case (B) would be most probable for the present event. In Fig. 4 III, diabatic heating rate Q_D at the equator, 20°N and combined heating rate $Q_A + Q_D$ at 20°N are shown. In the calculation of Mugnai *et al.* (1978) the heating of the atmosphere does not contain atmospheric absorption of the aerosol infrared emission and hence it may be underestimated to some extent. In Fig. 5 the diabatic heating rate Q_D at the altitudes 12, 10, 8, and 6 km are shown in I, II, III and IV, respectively by dashed curves. The combined heating rate

Fig. 5 Diabatic heating rate Q_D at altitudes I : 12 km, II : 10 km, III : 8 km, IV : 6 km. denoted by dashed lines (Newell *et al.* 1974) and A : $Q_D + Q_A$, B : $Q_D + 2Q_A$ at every altitudes.

$Q_D + Q_A$, and $Q_D + 2Q_A$ are shown by the solid curves A and B, respectively in the same figures at every altitude. It may be seen that Q_A or $2Q_A$ are of the same order of magnitude as Q_D near 20°N and have opposite sign at some altitude range.

Holton (1972a) shows by diagnostic equation that tropical Hadley circulation is generated by meridional gradient of heating rate in the upper troposphere. The heating Q_A over $15^\circ\text{--}25^\circ\text{N}$ may greatly reduce the meridional gradient due to Q_D alone and hence may slow down the Hadley circulation in northern tropical region or even reverse the sense of the mean motion, if the value of $m(a)$ is more than the present estimates for the early period of one season of concern.

According to Quiroz (1983) the westerly wind at the altitude 200 mb (11.5 km) over tropical Pacific ($165^\circ\text{W--}115^\circ\text{W}$) and easterly wind at the altitude 850 mb (1.5 km; over $135^\circ\text{E--}170^\circ\text{W}$) began to change to opposite directions, respectively in April 1982. The triggering of El Niño Southern Oscillation represented by the variation of SST might be accomplished by the mechanism mentioned above, followed by coupled oscillations of atmosphere and ocean as shown by a preliminary model by Philander *et al.* (1984).

The neglect of mean atmospheric motion to determine the distribution of settling aerosols is reasonable under the considerations, but the horizontal diffusion in higher latitudes than 25°N may have some effect. The northern boundary of the zonal flow may be more obscure than that explained here.

4. Discussion of the results

Stratospheric warming following El Chichon eruption was discussed by Labitzke *et al.* (1983), Angell and Korshover (1983) and Fujita (1984). According to the first law of thermodynamics, the heating rate Q is related to the variation of the atmospheric temperature as follows:

$$-\frac{d}{dt}C_pT = \alpha \frac{dp}{dt} + Q \quad (8)$$

where conventional notation as in Holton (1982a) is used.

The relation between the variations of Q and T is not straightforward and can have opposite sign. In order to discuss the dynamical effect, the heating rate Q should be estimated as in the present note.

It follows from equation (8) that

$$\left(\frac{\partial}{\partial t} + \mathbf{V} \cdot \nabla\right) \ln \theta + w \frac{\partial \ln \theta}{\partial z} = \frac{Q}{C_p T} \quad (9)$$

where w is vertical velocity and θ is potential temperature.

In the absence of the first term of the left hand side of (9), the value of w is given by

$$w = (H/\kappa)(Q/C_p T) \quad (10)$$

where $\kappa = (\partial \ln \theta / \partial z)H$ is proportional to the static stability. Essentially similar equation is a basic process of QBO modification theory by Dunkerton (1983). The aerosol heating would produce upward advection, but in reality the value of Q should include other kind of radiative cooling superposed on the aerosol heating as shown in Fig. 5. Therefore unless the net value of Q becomes positive, cancellation of the falling velocity of aerosols would be small. Moreover the first term of the left hand side of (9) will be important.

As shown in the previous section, the aerosol heating would reduce the meridional gradient of the diabatic heating rate, and hence the meridional velocity of the Hadley cell. If atmospheric heating by the infrared aerosol radiation is taken into account, the values of Q_A may be increased by a factor being probably not greater than 2.

Thus the dispersion in the upper troposphere would be much reduced at subtropical latitudes $15^\circ\text{--}25^\circ\text{N}$, and hence the residence time of settling aerosols near the original injection point would become much longer than those suggested by conventional theory.

For simplicity, we may use the following equations to show the total amount of falling aerosols:

$$v_s \cdot \mu_s \cdot ST = M_{st} \quad (11)$$

$$m(a) = \mu_s / \rho_{air} \quad (12)$$

where v_s = falling velocity of aerosols, μ_s = aerosol mass concentration per unit volume at the center of the zonal flow, ρ_{air} = air density,

S =bottom area of zonal flow of El Chichon cloud from 15° to 25°N , $T \sim 7.8 \times 10^6 \text{ s}$ (3 mo.) and $M_{st} \sim 10 Tg$. According to U.S. Standard Atmosphere Supplements 1966 at 15°N the air density ρ_{air} increases downward more slowly in the upper troposphere than it does at 30°N . The value of $v_s = (2/9)(\rho_p r^2 g / \eta)(1 + A\lambda/r)$ decreases downward, where η =dynamic viscosity coefficient, λ =mean free path, ρ_p =density of material of particles, A =a constant $=0.864$, g =acceleration by gravity. And hence the value of μ_s increases downward from the equation of continuity. Thus we obtained the vertical distribution of $m(a)$ shown in Fig. 4 I as the intermediate values between 15°N and 30°N . We have $v_s (\sim 0.15 \text{ cm/s})$ at 15 km and it can be readily seen that μ_s is inversely proportional to $(v_s T)$ for a given value of M_{st} .

We utilized the calculated results by Mugnai *et al.* (1978) which are made based on the U.S. Standard Atmosphere Supplements 1966 at 15°N , and the infrared flux due to the theoretical method of Rodgers and Walshaw (1966).

The latter showed that if the tropical troposphere is reasonably cloudless, an absorber near the tropopause can cause considerable heating due to absorption of radiation from ground at high temperature. Near 20°N it will fit the situation and now the absorber is the aerosols consisting mainly of sulfuric acid and of few impurities like silicate and carbon denoted by (B) in the previous section. The imaginary part of the refractive index for the visible wavelength $\kappa(vis.)$ is 0.005 for the aerosols and 0.05 for the silicate. The solar heating is about twice the above mentioned infrared heating at the altitude near 10 km and the net heating is shown in Fig. 4 II. The solar heating will increase with possible increase of impurities in the aerosols.

The amount of heating rate 1 K/day for the altitude range 9–11 km is due to the absorption of radiative flux 9 W/m^2 . This value is larger than 2 W/m^2 : the mean rate of dissipation of atmospheric kinetic energy by the friction at the surface of the globe. The absorption 6 W/m^2 of the above 9 W/m^2 is about 2% of the solar radiative flux 350

W/m^2 and remaining 3 W/m^2 is about 1% of the infrared upward flux from the globe. Therefore the perturbation of the radiation field by the presence of the above mentioned aerosols will be very small.

The calculation of heating rate by Mugnai *et al.* (1978) are made on the assumption: the amount of aerosols is so small that the variations of the radiation and the temperature distributions due to the aerosol heating in the atmosphere are negligible.

In the present work, it is further assumed that the variation of atmospheric temperature due to significant aerosol heating is very small due to the dissipation of increased internal energy by the mechanical work done to decrease or modify the Hadley cell mean motions and that the surface temperature is held almost constant for one season due to the large thermal inertia of the ocean. If there are significant temperature rise in the altitude range mentioned above, the infrared cooling would increase and hence the present calculations would show overestimated heating rate than the real value. Recent analysis of Angell and Korshover (1984) seems to support our treatment.

In the early period of one season after the Chichon injection, the height range 12–15 km near 20°N will be heated by aerosols. The equatorward gradient of heating would transport a small amount of aerosols towards ITCZ where very low temperatures ($< -83^\circ\text{C}$) exist at limited places (Newell and Gould-Stewart, 1981). The settling aerosols over ITCZ will form cirrus-like cloud, in such a very low temperature, according to the theory of Steele *et al.* (1983). The cirrus like cloud will act as a semi-transparent resonator with earth surface due to the size of its particles being not much less than the half wavelengths of the infrared earth radiation. The infrared earth radiation inside the resonator will be intensified and hence the absorption of the radiation by the atmosphere or aerosols will be increased. This zonally very inhomogeneous heating at the height range might modify the Walker circulation.

Thus far zonal homogeneity of the Chichon cloud was assumed. But according to Robock

and Matson (1983) appreciable zonal inhomogeneity and meandering of the cloud was present. It would be possible that the settling cloud concentration at the altitudes 8–12 km near Mexico within the longitudinal distance about 30° from El Chichon was ten times the zonal mean value assumed in Fig. 4 for several weeks. Then atmospheric heating by these aerosols will be about 5 K/day or more for about several weeks at the altitudes. If significant inclusion of impurity like carbon in addition to sulfuric acid is taken into account, heating would be appreciable. Meandering or broadening of the zonal wind would transport the cloud to lower latitudes temporarily for significant fraction of the period. According to the study of Holton (1972*b*) the heating of the atmosphere of this order of magnitude is possible to excite the mixed Rossby-gravity waves and Kelvin waves in the tropical region, of which the latter can significantly modulate the trade wind at the surface. In reality the weakening of the trade wind began in April 1983 (Philander, 1983; Quiroz, 1983).

The supposed modification of the circulation should be subject to numerical model simulations to test the validity. Much more work is still needed to confirm the situations.

When this note is written, Dr. J. J. DeLuise kindly sent us all lidar data at Mauna Loa (19°N) for one season after April 1982. A renormalization of the data in a similar way as in a companion paper (Hirono *et al.*, this issue) gave results supporting our model shown in Fig. 4 I. The details will be reported in a later paper.

5. Concluding remarks

In the tropical upper troposphere, the horizontal diffusion coefficient is thought to be lowest on the globe. The Chichon volcanic zonal cloud had its center at about 20°N for the first one season and the densest part was located at latitudes from 15°N to 25°N . The profiles of settling aerosols during this season is supposed to be within this subtropical region, where the amount of precipitation is also small, especially above altitude 5 km. The residence time of settling would be much

longer than about 10 days estimated by conventional analysis in the middle and higher latitudes. It is shown that the atmospheric heating rate around 15° – 25°N is sufficiently strong to modify Hadley circulation. For the season, the Hadley circulation in the northern tropics would be weakened and also the trade wind at the same time. This may be effective for the triggering of El Niño Southern Oscillation, which would be followed by the coupled atmosphere-ocean oscillations.

For several weeks after eruption, the Chichon cloud was zonally very inhomogeneous. In the region near Mexico, the concentration of falling aerosols including ash near the altitudes 8–12 km might be ten times those zonal mean values presented here (Hirono *et al.*, in this issue). Therefore the atmospheric heating by them might be strong enough to excite equatorial waves, of which Kelvin waves may modify the trade wind at the surface.

These mechanisms are qualitative at present. Much more work is essential to confirm the situation.

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火山変動とエル・ニーニョ南方振動の関連の可能性 一大気中の火山性エアロゾルのライダー観測

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