Temporal Variation of the Stratospheric Aerosol Layer after the Fuego Eruption observed by Lidar in Fukuoka

By Toshikazu Itabe*, Motowo Fujiwara and Motokazu Hirono

Department of Physics, Kyushu University, Fukuoka 812 Japan
(Manuscript received 21 April 1977, in revised form 13 September 1977)

Abstract

The eruption of the Fuego in Guatemala injected a large amount of volcanic substances to the lower stratosphere in mid-Oct. '74. About a month later, an extraordinary intense scattering layer was observed by ruby lidar in the lower stratosphere over Fukuoka. The peak value of the non-molecular radar cross section showed a maximum in Dec. '74. It decreased gradually after spring '75 and attained a secondary maximum in early autumn of the year. The height of the layer peak was located a few kilometers below the normal Junge layer and the half width of the layer which was initially a few kilometers increased as the peak value of the non-molecular radar cross section decreased.

The gross feature of the temporal variation of the peak value observed by lidar is in contradiction to the results expected from the 2-dimensional transport models proposed up to this time and the contradiction can not be explained if we consider only the case where the observed radar cross section is assumed to be a good conservative quantity. It is suggested that some kinds of *in situ* formation process should occur in the lower stratosphere.

1. Introduction

Following the eruption of the Fuego (14.5°N, 90.0°W), a large increase of the stratospheric aerosols was observed in the world by various methods *i.e.* measurements with twilight color (Volz, 1975), lidar (Fegley and Ellis, 1975), *in situ* optical particle counter (Hofmann and Rosen, 1975), impactor (Ferry and Lem, 1975), and so on. Lidar observations of the stratospheric aerosols in Fukuoka had begun in Oct. '72 and revealed the appearance of an extraordinary intense scattering layer in mid-Nov. '74 associated with the Fuego eruption. The peak values of non-molecular radar cross section β_M which stands for 4π times differential back-scattering coefficient at the lidar wavelength reached some ten times those of the pre-volcanic layer and varied in a rather complex manner from mid-Nov. '74 to Feb. '75. After spring '75 it decreased gradually and had a secondary maximum in early autumn of the year. The provisional results of lidar observations from Oct. '74 to Feb. '75 were reported in a previous paper (Fujiwara *et al.*, 1975).

In this paper, results of the lidar observations

after Mar. '75 are presented and are compared with those obtained from 2-dimensional models proposed by Gudiksen *et al.* (1968) and Cadle *et al.* (1976). We discuss only the case where the observed value β_M is assumed to be a good conservative property. The results from the 2-dimensional models mentioned above can not explain the temporal variations of β_M observed both in Fukuoka and Menlo Park (S. R. I.) consistently, in this case. If we adopt the Gudiksen's model, another case has to be considered, where *in situ* particle formation process in the lower stratosphere is taken into account. This case will be discussed in more detail in a later paper (Hirono *et al.*, to be published).

2. Method of observation and analysis

The instrumentation of our lidar was described in the previous paper in detail including the results of the early stage observation (Hirono *et al.*, 1974). The equipment consists of a ruby laser which produces 1 to 1.5 joule giant pulse for the optimum input energy, a 10 cm ϕ transmitting telescope and a 30 cm ϕ receiving telescope. For every night observation the number of total firing is about 300 shots with the repetition of 6 pulses per min. The return photon

* Present affiliation: Radio Research Laboratories.

signals from the atmospheric constituents are converted to electronic pulses by a photomultiplier. The pulses are counted in each 0.75 km height interval and the counts are accumulated for the total shots of a night. Then the height profiles of the photon counts are drawn for a height range between 12 and 17 km for the period from Mar. '75 to Jan. '76 and between 9 and 30 km after Feb. '76.

The profile of photon count $C(h)$ is given by

$$C(h) = K \frac{\beta_R(h) + \beta_M(h)}{h^2} \quad (1)$$

where K is a constant for the particular lidar including a transmission coefficient of the atmosphere and a height interval Δh , $\beta_R(h)$ and $\beta_M(h)$ are molecular and non-molecular radar cross sections, respectively. The non-molecular radar cross section $\beta_M(h)$ can be determined using the observed values of $C(h)$, if informations on the parameters $\beta_R(h)$ and K are given. The molecular radar cross section $\beta_R(h)$ is calculated by a molecular density profile (U.S. Standard Atmosphere 1962) neglecting its temporal variations at the present stage. It is, however, very troublesome and difficult to determine the value of K only by the usual one color lidar. In the present, therefore, we adopt the matching method *i.e.* we assume that the atmosphere at a certain height h_0 within the height region spanned by lidar involves negligible aerosol content. At the height h_0 , eq. (1) is given as follows,

$$C(h_0) = K \frac{\beta_R(h_0)}{h_0^2} \quad (2)$$

and from eqs. (1) and (2), $\beta_M(h)$ is given by

$$\beta_M(h) = \left\{ \frac{C(h)}{C(h_0)} \cdot \frac{h^2}{h_0^2} \cdot \frac{\beta_R(h_0)}{\beta_R(h)} - 1 \right\} \beta_R(h) \quad (3)$$

In the analysis of lidar data by the matching method, it is very important how to choose the height of normalization h_0 . We choose the height h_0 according to the following rule: when the resulting non-molecular radar cross section $\beta_M(h)$ for an arbitrary h_0 is positive or zero within the limits of the probable error of the photon counting over the observed height region, then the height h_0 is chosen to be the height of normalization. The height h_0 is located typically below nearly 15 km or above nearly 25 km. At present, the photon counts from the upper region are so few that the photon counting error is fairly large. This error gives bad influence on the accuracy of $\beta_M(h)$ at all heights as can be seen readily in

eq. (3), if we choose the height of normalization to be around 25 km. So we adopted the data which resulted from the choice of the height h_0 below 15 km and omitted the other ones of which the height h_0 was located around or above 25 km. It is necessary to know the matching method can lead to underestimate the stratospheric aerosol content in any case.

The scattering ratio $R(h)$ which is usually used in the discussion of lidar observations is given as follows,

$$R(h) = 1 + \frac{\beta_M(h)}{\beta_R(h)} \quad (4)$$

In this paper, however, we use the radar cross section $\beta_M(h)$.

3. Observational results and comparison with 2-dimensional models

Examples of the vertical profiles of β_M and R analyzed by the matching method mentioned in the previous section are shown in Fig. 1. According to the local rawinsonde soundings carried out by the Fukuoka Meteorological Observatory the mean height of a tropopause which extends from the equatorial region is about 16 km in all seasons and in winter another tropopause appears at height lower than the former. The peak height of the post Fuego stratospheric aerosol layer is about 2 km higher than the upper tropopause, as shown in Fig. 1. Taking account of the lidar observations in the other stations, *i.e.*, Hawaii, Virginia and California, the volcanic substances injected by the Fuego would have been transported horizontally along the height of the tropopause extended from equator in the lower latitude region. The thickness of this layer was about 3 km in the first stage after the Fuego eruption and became broader gradually. Then secondary small peak began to appear occasionally above 20 km after spring '75.

Peak values of β_M observed in Fukuoka are plotted in Fig. 2 for the period from Oct. '74 through Sep. '76. The results of Stanford Research Institute observed in Menlo Park, California (Russell *et al.*, 1976) are also plotted in the figure. The provisional results during five months after the Fuego eruption were published previously (Fujiwara *et al.*, 1975). The peak values decrease up to Aug. '75 and then have a secondary maximum in early autumn, as shown in Fig. 2. According to Dyer and Hicks (1968), it is indicated by the analysis of the decrement of the direct solar radiation that the winter maxima of

Fig. 1 Examples of height profiles of (a) scattering ratio $R(h)$ and (b) non-molecular radar cross section $\beta_M(h)$.

Fig. 2 Temporal variation of peak values of radar cross section $\beta_M(\text{peak})$, their approximated curve and mixing ratios of W^{185} calculated according to Gudiksen *et al.* For the approximated curve, the constants, a , b and c , are $1.07 \times 10^3 \text{ sec cm}^{-1}$, 1.32 cm^{-1} and $3.93 \times 10^{-4} \text{ sec}^{-1} \text{ cm}^{-1}$ respectively.

the volcanic dust progressed polewards within a few years following the Agung eruption in 1963. In the case of post Fuego, however, other stations which have been observing the stratospheric aerosol layer have not reported the secondary maximum. The secondary maximum in early autumn observed in Fukuoka may, therefore, be rather local phenomenon. In this paper, we consider only a gross feature of temporal variation of the peak values and compare it with the 2-dimensional models proposed by Gudiksen *et al.*

(1968) and Cadle *et al.* (1976).

In order to consider the gross temporal variation of the peak values of β_M , we draw a curve of approximation in Fig. 2. The curve is in the form of a function which involves an asymptotic solution of a diffusion equation in the first term as follows,

$$\beta_M(\text{peak}) = \frac{a}{t} + b + ct \quad (5)$$

where t is time in unit of day reckoned from the Fuego eruption, a , b , and c are constants which are determined by the least square method. The approximated function fits fairly well the observed values of β_M , as shown in Fig. 2. Assuming that the stratospheric aerosols and gases injected by the Fuego eruption have been homogeneous along the latitudinal circle since Mar. '75, we compare the approximated curve with the calculated values of 2-dimensional models.

When we compare the temporal variation of the stratospheric aerosols by lidar with the 2-dimensional models, it is necessary to examine whether or not the observed value of β_M is a conservative quantity. The radar cross section β_M is written as follows,

$$\beta_M(h, t) = 4\pi \int \sigma(r, m, \lambda) n(r, h, t) dr \quad (6)$$

where σ is backscattering cross section of an aerosol particle with radius r , m the refractive index for a wavelength λ , and $n(r, h, t)$ the number density of aerosols with radii between r and $r+dr$. As shown in eq. (6), the observed value β_M depends on the two parameters, refractive index m and size distribution $n(r, t)$. The former depends on the composition and phase of the aerosol particles. The recent *in situ* measurement indicated that the aerosol particles injected by the Fuego eruption were sulfuric acid droplets after Mar. '75 (Ferry and Lem, 1976; Hofmann and Rosen, 1975). We adopt the refractive index of the post Fuego aerosols to be 1.41 which is the same one as the pre Fuego aerosols composed of a 75% solution of sulfuric acid (Toon and Pollack, 1973). Then σ can be computed for an aerosol particle with any radius by Mie theory.

The size distribution of the aerosol particles is controlled by many complex processes, coagulation, nucleation, sedimentation and so on. In the present, it is impossible to solve a time dependent equation involving all processes and to describe the evolution of the size distribution. So we confine our attention only to the coagulation process. More detailed discussion including other processes, especially, nucleation, will be described in a later paper. The equation including only a coagulation process is the simplest one, which can be solved numerically. The results of calculation shows that the size distribution within the optically effective size range between nearly 0.1 and 1.0 μm in the lower stratosphere does not change significantly if the initial distribution is assumed to be Haze H proposed by Deirmendjian (1969; see also Russel *et al.*, 1974) and power law proposed by Friedlander (1960; see also Hirono *et al.*, 1974), and that the evolution of the size distribution has little effect on the temporal variation of the observed values of β_M . In this case where the size distribution depends only on the coagulation, the observed value β_M is a good conservative quantity and the function of eq. (5) should be fitted fairly well to the calculated ones by the 2-dimensional transport model.

One of the 2-dimensional models was proposed by Gudiksen *et al.* (1968) in order to explain the global spread of W^{185} injected into the lower stratosphere by nuclear weapon test. (the Hardtack test, 1958). For the numerical calculation the trace substance W^{185} was assumed to have sources at 11°N in the heights of 18 and 19.5 km

on June 15, '58. The latitude of the volcanic injection is 14.5°N which is not so far from 11°N and interval of latitude between 11°N and 30°N is equivalent to that between 14.5°N and 33.5°N. Because the principal mechanism of the transport in their model is the horizontal eddy diffusion of which the coefficients are nearly constant in any seasons, the initial time of their model can be shifted to the date of the volcanic eruption (Oct. 16, '74). Peaks of the vertical mixing ratio profiles of W^{185} calculated by their model are located near the height of 18 km at 30°N during 18 months. The variation of peak mixing ratio is shown by rectangles in Fig. 2, where it is indicated that the overall decay of the approximated curve mentioned above is much slower than the calculated one. The 2-dimensional model proposed by Gudiksen *et al.* can not, therefore, explain the results of lidar observations.

Fig. 3 Relative concentrations of the trace substance at altitude 18 km on different latitudes according to the 2-dimensional model proposed by Cadle *et al.*

Another 2-dimensional model was proposed by Cadle *et al.* (1976; see also Louis, 1974) and is able to explain our approximated curve better. Fig. 3 shows the relative concentrations of the trace substance at different latitudes according to the model of Cadle *et al.* The decay feature of this model has, however, large variability against the latitude near 30–40°N, in contradiction to the results observed at two different stations, Fukuoka (33°N) and Menlo Park (38°N), which are not so different except the appearance of the secondary maximum in early autumn in Fukuoka. The model proposed by Cadle *et al.* also can not explain the decay feature observed at two different stations, consistently.

Taking account of the variability of the decay feature against latitude between 30°N and 40°N, the 2-dimensional model proposed by Gudiksen *et al.* is more suitable than the one proposed by Cadle *et al.* in order to analyze the gross temporal variation of the post-Fuego stratospheric aerosols. Thus we adopt the model proposed by Gudiksen *et al.* for our purpose. It is important to note that the gross decay feature observed by lidar is not faster but slower than predicted one calculated by Gudiksen *et al.* It is suggested that some kinds of *in situ* formation process should be necessary to explain the results of lidar observation according to the 2-dimensional model proposed by Gudiksen *et al.* On the contrary the sedimentation has remarkable influence on the removal of the stratospheric large particles of aerosol ($r > 0.3 \mu\text{m}$) and if the return of the lidar emission comes mainly from these particles, the decay feature would be faster than the calculated one in the absence of any growth or formation of particles. Thus the influence of increased sedimentation will enhance the discrepancy between the observed and calculated results. The *in situ* formation in the lower stratosphere would occur by mean of the homogeneous or heterogeneous heteromolecular nucleation in which a solution droplet is formed from two gaseous species, water and sulfuric acid vapor. In a later paper we will discuss the case where the *in situ* formation in the lower stratosphere is taken into account.

Temporal variation of the height integrated β_M between 15.75 and 24.75 km is shown in Fig. 4, where the secondary maximum can also be seen in early autumn '75. In spring '75 the integrated β_M reached about six times as great as that of pre-volcanic layer which was 0.7×10^{-3} on average and then it gradually decreased. The mean value of the integrated β_M is 2.1×10^{-3} between Mar. '75 and Feb. '76 and 1.1×10^{-3} between Mar. '76 and Sep. '76. Assuming the

Fig. 4 Temporal variation of the height integrated β_M .

size distribution of the aerosol particles at each height is represented by the same normalized size distribution function, the height integrated β_M is proportional to aerosol optical thickness for a ruby wavelength $0.69 \mu\text{m}$, $\tau(0.69)$, which is defined as follows,

$$\tau(0.69) = \frac{1}{P(\pi)} \int_{15.75 \text{ km}}^{24.75 \text{ km}} \beta_M(h) dh \quad (7)$$

where

$$P(\pi) = \frac{\beta_M}{\beta_{\text{ext}}} \quad (8)$$

is a phase function and β_{ext} is a non-molecular extinction coefficient. Based on the Mie theory we computed the values of β_{ext} for three different kinds of the size distribution; Power Law $n_p(r)$ (Junge *et al.*, 1961), Lognormal $n_l(r)$ (Friend, 1966; Pinnick *et al.*, 1976) and Deirmendjian Haze H $n_h(r)$ (Russell *et al.*, 1974). Their specific form are:

$$n_p(r) = A_p r^{(\nu+1)}, \quad r_{\min} < r < r_{\max} \quad (9)$$

where $\nu=3$, $r_{\min}$ and $r_{\max}$ are minimum and maximum radius respectively,

$$n_l(r) = A_l \frac{1}{\sqrt{2\pi}s} \exp \left[-\{1n(r/r_0)\}^2 / 2s^2 \right] \quad (10)$$

where $r_0 = 0.35 \mu\text{m}$, $s = 0.30$ proposed by Friend and $r_0 = 0.0725 \mu\text{m}$, $s = 0.621$ proposed by Pinnick *et al.*,

$$n_h(r) = A_h r^{-2} \exp(-20r) \quad (11)$$

where A_p , A_l and A_h are constants for normalization. Calculated values of $P(\pi)$ for these size distributions are shown in Table 1.

Table 1. Phase functions for various size distributions.

Size Distribution	$r_{\min}$ (μm)	$r_{\max}$ (μm)	$P(\pi)$
Junge Power Law $\nu=3$	0.05	0.5	0.2344
	0.08	0.5	0.2172
	0.1	0.5	0.1993
	0.05	1.0	0.2334
	0.08	1.0	0.2212
	0.1	1.0	0.2086
Lognormal (Friend)			0.1692
Haze H (Russell et al.)			0.1644
Lognormal (Pinnick et al.)			0.1938

Usually the optical thickness for a wavelength $0.55 \mu\text{m}$, $\tau(0.55)$, is used and the one for $0.69 \mu\text{m}$, $\tau(0.69)$, is converted to $\tau(0.55)$ assuming that the non-molecular extinction varies as λ^{-1} . The

values of $\tau(0.55)$ of the post-Fuego aerosols are between 3.2 and 0.59 ($\times 10^{-2}$) for Haze H size distribution and between 2.3 and 0.45 ($\times 10^{-2}$) for Junge power law size distribution in which the radii of aerosols are in the size range $0.08 < r < 1 \mu\text{m}$.

4. Concluding remarks

The temporal variation of the post-Fuego stratospheric aerosol layer from Mar. '75 to Sep. '76 observed by lidar in Fukuoka is described and is compared with the 2-dimensional transport models proposed by Gudiksen *et al.* and Cadle *et al.* In this paper, we confine our attention to the case in which the production of the particles in the embryonic size range is neglected and so the observed value β_M is well conservative. The decay of the observed values is much slower than the one expected from the result calculated by Gudiksen *et al.* Another model proposed by Cadle *et al.* also can not explain consistently the decay features observed at two different stations, Fukuoka and Menlo Park, which show nearly parallel decrements at these stations. Consequently, none of the two 2-dimensional models can explain the results of lidar observations in the case of no *in situ* formation process in the stratosphere. If the model by Gudiksen *et al.* is adopted at least to give appropriate rate of decay in low latitudes, some kinds of *in situ* formation process should be considered to explain the observed results.

The temporal variation observed in Fukuoka shows a secondary maximum in early autumn '75, which has not been discernible at other stations.

The variation of integrated β_M from 15.75 km to 24.75 km is also described in order to show the optical thickness for several size distributions.

References

- Cadle, R. D., Kiang, C. S., and J. F. Louis, 1976: The global-scale dispersion of the eruption clouds from major volcanic eruptions. *J. Geophys. Res.*, **81**, 3125–3132.
- Deirmendjian, D., 1969: Electromagnetic scattering on spherical polydispersions. American Elsevier.
- Dyer, A. J., and B. B. Hicks, 1968: Global spread of volcanic dust from the Bali eruption of 1963. *Quart. J. Royal Meteor. Soc.*, **94**, 545–554.
- Fegley, R. W., and H. T. Ellis, 1975: Lidar observations of a stratospheric dust cloud layer in the tropics. *Geophys. Res. Lett.*, **2**, 139.
- Ferry, G. V., and H. Y. Lem, 1975: Changes in stratospheric aerosols collected after the eruption of volcan de Fuego. *EOS Trans. Amer. Geophys. Union*, **56**, 366.
- Friedlander, S. K., 1960: Similarity considerations for the particle size spectrum of a coagulating, sedimenting aerosol. *J. Meteorol.*, **17**, 479–483.
- Friend, J. P., 1966: Properties of the stratospheric aerosol. *Tellus*, **18**, 465–473.
- Fujiwara, M., Itabe, T., and M. Hirono, 1975: Sudden increase of stratospheric aerosol content after the eruption of Fuego volcano; lidar observations in Fukuoka. *Rep. Ionos. Space Res. Japan*, **29**, 74–78.
- Gudiksen, P. H., Fairhall, A. W., and R. J. Reed, 1968: Roles of mean meridional circulation and eddy diffusion in the transport of trace substances in the lower stratosphere. *J. Geophys. Res.*, **73**, 4461–4473.
- Hirono, M., Fujiwara, M., Uchino, O., and T. Itabe, 1974: Observations of stratospheric aerosol layer by optical radar. *Can. J. Chem.*, **52**, 1560–1568.
- Hirono, M., Fujiwara, M., and T. Itabe, 1976: Behavior of the stratospheric aerosols inferred from laser radar and small ion radiosonde observations. *J. Geophys. Res.*, **81**, 1593–1600.
- Hirono, M., Fujiwara, M., Itabe, T., and C. Nagasawa, 1977: On the long term variation of stratospheric aerosol content after the eruption of Fuego volcano observed by Lidar. (in preparation)
- Hofmann, D. J., and J. M. Rosen, 1975: Balloon observation of the time development of the stratospheric aerosol event of 1974–75. *EOS Trans. Amer. Geophys. Union*, **56**, 995.
- Junge, C. E., Chagnon, C. W., and J. E. Manson, 1961: Stratospheric aerosols. *J. Meteorol.*, **18**, 81–108.
- Louis, J. F., 1974: A two dimensional model of the atmosphere. *Ph.D. Thesis*, University of Colorado.
- Pinnick, R. G., Rosen, J. M., and D. J. Hofmann, 1976: Stratospheric aerosol measurements III: optical model calculations. *J. Atmos. Sci.*, **33**, 304–314.
- Russell, P. B., Vizee, W., and R. D. Hake, Jr., 1974: Lidar measurements of the stratospheric aerosol over Menlo Park, California; October 1972–March 1974. Final Report 2217, Stanford Research Institute.
- _____, _____, and _____, 1975: Lidar measurements of the stratospheric aerosol. Third Quarterly Report 4019, Stanford Research Institute.
- Toon, O. B., and J. B. Pollack, 1973: Physical properties of the stratospheric aerosols. *J. Geophys. Res.*, **78**, 7051–7056.
- Volz, F. E., 1975: Volcanic twilights from the Fuego eruption. *Science*, **189**, 48–50.

ライダーによって観測された福岡に於けるフェゴ噴火後の 成層圏エアロゾル層の時間的変動

板部 敏和*・藤原 玄夫・広野 求和

九州大学理学部物理学教室

1974年十月中旬にガテマラのフェゴ火山噴火は下部成層圏に大量の火山物質を投入した。約1ヶ月後ライダーによって福岡上空の下部成層圏に異常に強い散乱層が観測された。大気分子以外のレーダー断面積のピーク値は1974年12月に最大値を示した。1975年春以後は徐々に減少して行ったが、その年の秋に第2の極大値に達した。層のピークの高度は通常のエンゲ層の2~3 km下にあり、層の半値幅は最初の2~3 kmからピーク高度の下降と共に増大した。

ライダーによって観測されたピーク値の時間的変動の全体的な形は今日までに提唱された2次元輸送モデルから期待される結果と矛盾しているが、この矛盾は観測されたレーダー断面積が良い保存量であると仮定する限り説明できない。そこで下部成層圏に於て何らかの形でその場に於ける粒子生成が起っていることが示唆される。

* 現在所属：電波研究所