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PERTURBATIONS OF THE ATMOSPHERE AND OF THE CLIMATE SYSTEM INDUCED BY VOLCANIC ERUPTIONS: RELEVANCE OF SOME SPECIFIC PROCESSES

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Abstract

Progress in the understanding of those physical and chemical processes which preside over the perturbations induced by volcanic eruptions on the atmosphere, but whose effects may be felt by the entire climatic system, is reviewed in the light of current research related to the aftermath of the El Chichon eruption of 1982.

Introduction: Early Attempts to Characterize the System and Successive Developments in Identifying and Parameterizing the Effects of the Main Variables

Large eruptions provide rare occasions to observe the transient response of the atmosphere to a quantifiable perturbation, whose amplitude and duration may be such that significant effects can be expected also on the other elements of the climate system.

The tenet that volcanic eruptions would affect the climate is neither new (Benjamin Franklin is credited with the original suggestion; for an early statement of the problem see Humphreys, 1940) nor basically disputed, but the state of knowledge was, and perhaps still is, such that the dynamic response of the system could not be reliably predicted.

The 1982 eruptions of the volcano El Chichon have occurred at a time when the stage of development of climate models and the capability of carrying out measurements of a large number of variables on a global scale suggest that consistent checks of the theory are possible. Also, a great deal will be learned in detail from this concerted effort about interactive, atmospheric processes involving aerosols and their precursors, radiation, dynamics.

It may be of interest to outline relatively recent developments. On the basis of a comprehensive analysis of volcanic chronology since 1500 AD, Lamb (1970) had introduced a criterion, the Dust Veil Index (DVI), for relating empirically the climatic effects to the characteristics of the eruption. The DVI was expressed with three alternative formulae, claimed to be equivalent, each expression being the product of three quantities:

$$DVI = 0.97 \times R_{Dmax} \times E_{max} \times t_{mo}\,, \tag{1}$$

$$DVI = 52.5 \times T_{Dmax} \times E_{max} \times t_{mo}\,, \tag{2}$$

$$DVI = 4.4 \times q \times E_{max} \times t_{mo}\,, \tag{3}$$

where R_{Dmax} is the greatest percentage of depletion of direct radiation following the eruption, E_{max} is a coefficient related to the geographical extent of the veil (whole Earth = 1), t_{mo} is the duration in months, T_{Dmax} is the estimated lowering of avg. temperature in degrees Celsius over the middle latitude zone of the hemisphere, q is the estimated value in cubic kilometers of solid matter dispersed as dust in the atmosphere. The coefficients are set up so that for Krakatoa DVI = 1000. For El Chichon, assuming $E_{max} = 1$, $t_{mo} = 24$, $q = 3$, we obtain DVI = 320, $T_{Dmax} = 0.25$ C, and $R_{Dmax} = 13\%$.

The DVI does not take into account the detailed nature of tbe process and is based on a fair amount of estimation: it can be argued, for instance, that it is not the total amount of dust that matters, but rather the fraction that reaches the stratosphere, particularly as a sulfur compound, and how that content evolves in time; also, it is not just the change in direct radiation that has an effect, but the redistribution, in the entire spectrum, of the direct and diffuse radiation, etc. The DVI is an approach to identify and relate the relevant and measurable parameters; it gives a useful insight by associating volcanic and climatic chronologies. In the words of the author this is "about as far as one can go towards objectivity in the assessment of past eruptions". In this respect it reflects the state of knowledge on the problem in 1970.

Substantial climatic anomalies have been attributed to the large eruptions of the past, Krakatoa, Tambora, Katmai, etc. On the basis of a statistical analysis, Schneider and Mass (1975) have shown that there is an effect of volcanic dust on long term temperature records. Bryson and Goodman (1980), using available data and a simple model of dispersion and fallout, have made an historical analysis of the role of volcanic eruptions in modulating the direct beam radiation and have found that the resulting time series would explain about 77% of the radiation variance (see also Yamamoto, 1977; Taylor, 1980; Robock, 1981; Rampino *et al.*, 1979). In order to improve the value of estimates of historical volcanism, useful to assess its role on past climatic variations, Newhall and Self (1982) have proposed a composite indicator of the magnitude of past explosive eruptions termed the Volcanic Explosivity Index (VEI). The measured concentration of sulfur in ice cores collected in Greenland and Antarctica is a useful marker of past volcanic activity (Dansgaard and Hammer, 1978; Stothers and Rampino, 1983).

Measurements carried out in connection with the 1963 Mt. Agung eruption provided a data base useful for detailed analyses. Lidars, then in the first stages of development, showed the height-resolved structure and evolution of the stratospheric aerosol layer (Fiocco and Grams, 1964; Grams and Fiocco, 1967). Upper level sounding indicated that, as a consequence of the eruption, the temperature of the tropical lower stratosphere increased by several degrees (Newell, 1970, 1971; McInturff *et al.*, 1971). Tropical tropospheric temperatures decreased by about one-half a degree (Newell and Weare, 1976; Angell and Korshover, 1977); there was a depletion of the solar radiation received at the ground (Burdecki, 1964; Moreno *et al.*, 1965; Flowers and Viebrock, 1965; Volp, 1969) and the surface also appeared to have cooled (Angell and Korshover, 1977), although with some uncertainty due to the large fluctuations of the climatic signal. It is possible that the break in the quasi-biennial oscillation of the atmosphere observed in 1963 was related to the eruption (Sparrow, 1965, 1971; Berson, 1966; Ebdon, 1967; Newell, 1970, a, h). A long-term trend of the stratospheric water vapor concentration in the years following the eruption has also heen tentatively associated witb it (Mastenbrook, 1971; Newell *et al.*, 1974; Crescentini and Fiocco, 1983).

Hansen *et al.* (1978), with a one-dimensional radiative-convective model, simulated the thermal response of the atmosphere to the increased amount of aerosol at low latitudes. The resulting stratospheric temperature increase, which matches reasonably well the average heating observed, is due mainly to the assumed stratospheric aerosol properties, but also to the tropospheric temperature, because of the absorption of upwelling infrared radiation; the simulation gave also an indication of tropospheric cooling. It was pointed out that the main limitation of the computation was the lack of an adequate treatment of atmospheric dynamics.

The heating pattern induced in the lower equatorial stratosphere by the Mt. Agung eruption was reproduced in a model calculation and associated to the changes in the albedo of the underlying system, thus providing also a measure of the absorption properties of the aerosols in the visible part of the spectrum (Fiocco *et al.*, 1976, 1977; Mugnai *et al.*, 1978). These authors have also investigated the changes in the photodissociation rates of ozone and other species resulting from the increased levels of the stratospheric diffuse radiation (Fiocco *et al.*, 1978; Petroncelli *et al.*, 1980).

Pollack, Toon, Turco and collaborators have extensively studied the characteristics of aerosols in the context of climate research (Pollack *et al.*, 1976; Toon and Pollack, 1975; Turco *et al.*, 1980, 1982). Relevant contributions in this field have been given by Reck (1974) and by Harshvardan and Cess (1976). Studies related to the nucleation and growth of sulfur stratospheric aerosols have been performed by Friend *et al.* (1980), by Hamill *et al.* (1977), and by Steele and Hamill (1981). Problems related to transport and dynamics of the volcanic clouds have been considered by Danielsen (1968, 1974), Dyer and Hicks (1968) and Cadle *et al.* (1976).

The 1980 eruption of Mount St. Helens was well documented: although the explosive force was large, the mass of sulfur injected into the stratosphere was small and did not cause a significant increase of its optical depth (Newell and Deepak, 1982; Deepak, 1982).

On the Microphysics of the Aerosols and Their Radiative and Thermodynamic Role

A volcanic eruption releases in the atmosphere large quantities of solid and gaseous materials: depending on the explosive nature of the eruption and on the diffusive properties of the medium, a fraction of these materials arrives at stratospheric heights where the lack of water and precipitation as well as the high vertical stability favor substantially longer residence times than in the troposphere. The radiative field is modified by the presence of these constituents; because of the long times involved, the resulting perturbations may become climatologically effective. The constituents that are of primary concern are the ash and the sulfur-bearing gases.

The ash has a comparatively short lifetime, depending on the relatively large size of the grains, and its radiative effects are dominant in the first few months following the eruption. The sulfur-bearing gases, through a series of imperfectly known reactions, undergo a conversion process which produces a population of aerosol particles consisting mainly of a mixture of sulfuric acid and water. Fine ash and impurities are also included in the mixture and may play a role in determining the nucleating and absorption properties of the particles.

After the initial nucleation phase, the physical and chemical evolution of the aerosol cloud, leading eventually to its disappearance, is dictated, on the long term, by the mechanisms of coagulation and sedimentation and, on the short term, by processes related to its radiative and thermal balance which may determine a variable H_2O/H_2SO_4 ratio.

Coagulation leads slowly and irreversibly to bigger and heavier particles; gain or loss of water through evaporation and condensation may rapidly change their density and radius. In either case the sedimentation velocity is affected. Of particular interest are these adaptive processes in those regions (e.g. the Hadley cell) where the existence of updraughts may essentially sustain the particles for a very long time.

Aerosols are not a passively transported species and cannot be regarded strictly as tracers of motion. If present in large quantities their radiative properties may contribute to the heating and cooling of air to the extent of having a significant effect on atmospheric dynamics. In the 15-20 km region the heating rates attributed to the atmospheric gases are very small and the aerosol contribution may prevail.

The modifications of the radiative field induced by the presence of aerosols, if sufficiently large, may have a detectable effect on the climate system. Aerosols absorb and scatter solar radiation: as a result there is an attenuation of the direct solar field, on whose basis it is generally assumed that the energetic input to the underlying system is reduced.

Since however a large fraction of the scattering is in the forward direction, the depletion of solar radiation reaching the ground is not as large as expected merely on the basis of extinction. In addition the presence of the absorbing aerosol modifies the infrared characteristics of the atmosphere. The net effect depends on the size and refractive index of the particles.

Specific aspects of the problem, related to the individual as well as to the collective properties of the aerosols, their spatial variation and temporal evolution, can be identified as follows:

a) the size distribution,

b) the composition and the complex refractive index,

c) the radiative field,

d) the heating rates,

e) the thermodynamic properties, stability and lifetime of the aerosol layers,

f) diffusion and transport by large scale atmospheric motions.

An Overview of Observational Results Related to the El Chichon Eruptions of 1982

In 1982 the stratosphere was characterized by a very large increase of the aerosol content, consequence of large injections of volcanic material. At the beginning of the year a volcanic explosion at an as yet uncertain location manifested itself in the so-called mystery cloud (De Luisi, 1982). Subsequently, the eruptions of El Chichon, Mexico (17 N, 93 W), which occurred on 28 March and 3-4 April, have determined what is possibly the most intense stratospheric perturbation of the century, likely to have its effects felt on climatic conditions for a few years.

Several lidar stations detected the initial spreading of the cloud and followed its subsequent evolution. In the northern hemisphere (and in order of both increasing latitude and delay in the time of first arrival of the cloud overhead) these lidars are located in Hawaii (19.5 N: De Luisi *et al.*, 1982), in Japan (33.7 N: Hirono and Shibata, 1982. 35 N: Iwasaka *et al.*, 1983), in Italy (41.8 N: Adriani *et al.*, 1983. 42.3 N: D'Altorio and Visconti, 1983), in Germany (47.5 N: Reiter *et al.*, 1983); in the Southern hemisphere the only reporting station is located in Brazil (23 S: Clemesha and Simonich, 1983). Fig. 1 gives a sequence of the lidar observations in Frascati, Italy.

Airborne lidar observations have secured meridional cross sections of the cloud (McCormick, 1982) Balloon sightings of the El Chichon cloud were obtained (Ackerman and Lippens, 1983) one month after the eruption.

Data from the environmental satellites NOAA-7 and GOES showed that the eruptions had penetrated the tropopause and injected material into the stratosphere. The cloud produced by the April 4 eruption was tracked by visible band imagery in its westward march, filling the latitude band 10 N to 30 N in about three weeks (Robock and Matson, 1983).

Measurements of the gaseous sulfur dioxide released were obtained with the Total Ozone Mapping Spectrometer (TOMS; Krueger, 1983) and with the Solar Backscatter Ultraviolet Spectrometer (SBUV; Heath *et al.*, 1983), both carried on the Nimbus 7 satellite. Three instruments on board the Solar Mesosphere Explorer (SME) also revealed features of the cloud: the Infrared Radiometer measured the thermal emission from the aerosols, while the Visible and Near Infrared Spectrometers measured the backscattered solar radiation. The three instruments are limbscanning and view the atmosphere along the track of the sunsynchronous polar orbit (Barth *et al.*, 1983; Thomas *et al.*, 1983). Ground based and airborne spectrophotometric measurements of sulfur dioxide have also been carried out (Evans and Kerr, 1983).

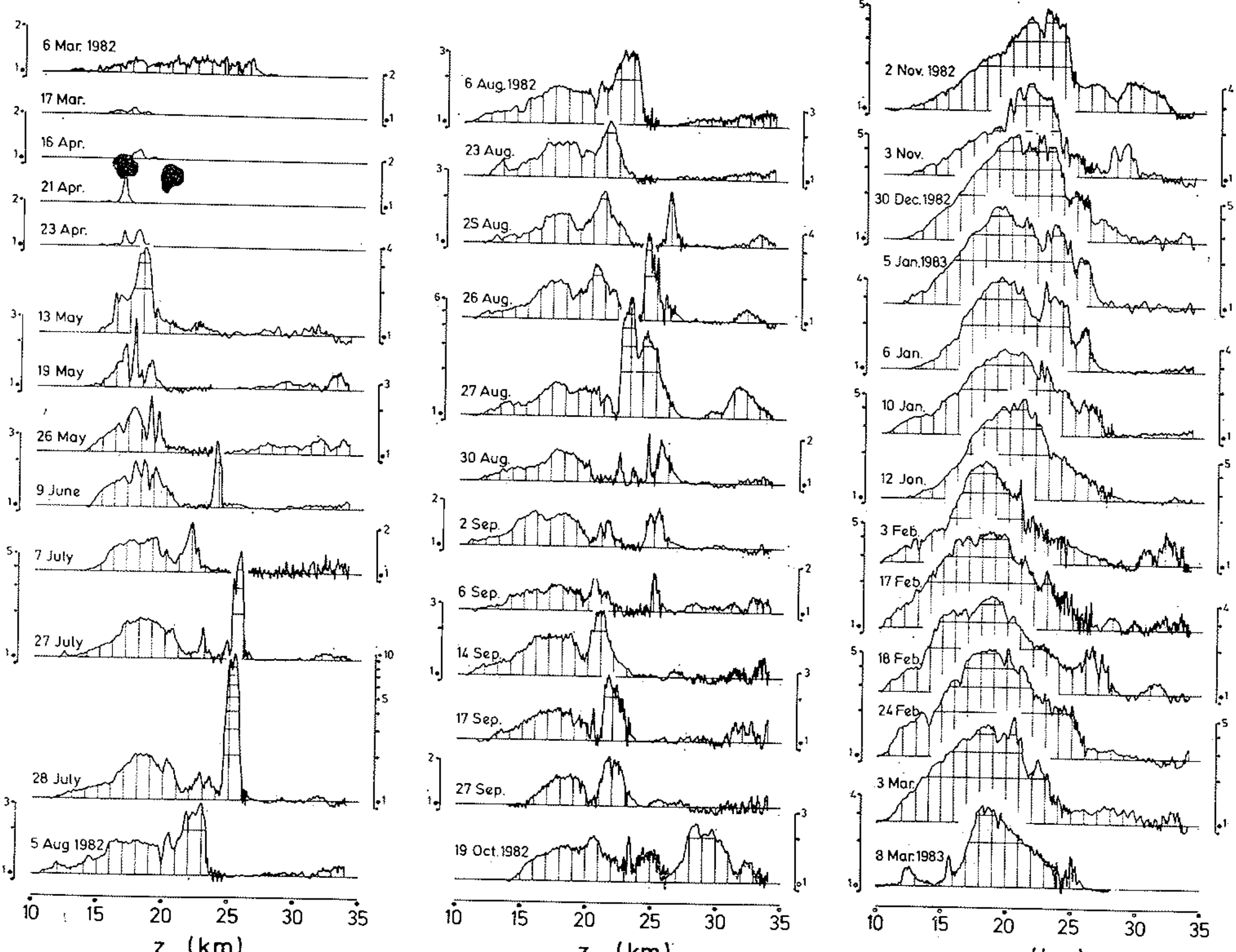
High latitude aerosol observations have been secured by the Stratospheric Aerosol Measurement (SAM II) satellite system: these have shown that the aerosol extinction profiles measured within the northern polar vortex differ significantly above 18 km from those measured outside the vortex (Mc Cormick *et al.*, 1983).

Particle sampling devices have been flown on balloons (Hofmann and Rosen, 1983) and on high-altitude aircraft (Oberbeck *et al.*, 1983; Knollenberg and Huffman, 1983; Wilson *et al.*, 1983; Gooding *et al.*, 1983; Woods and Chuan, 1983).

Stratospheric sulfate and other gases were collected by high altitude aircraft and balloons (Gandrud *et al.*, 1983; Mroz *et al.*, 1983; Vedder *et al.*, 1983). These measurements indicated an injection of about 7.6 Tg of sulfate into the global stratosphere. Ground and airborne extinction measurements were also carried out (Clarke *et al.*, 1983; Coulson, 1983; Dutton and De Luisi, 1983; Spinhirne, 1983; Witteborn *et al.*, 1983) as well as comparisons with Lidar measurements (Swissler *et al.*, 1983). Other reports on the initial phases of this phenomenon and attempts at drawing preliminary conclusions have by now appeared: some will be mentioned in the following section.

Some Specific Aspects

Determining the size distribution and composition of the stratospheric aerosols and their evolution have been among the main questions posed since the discovery of the layer (Junge *et al.*, 1960). The steady state



6 Mar. 1982
17 Mar.
16 Apr.
21 Apr.
23 Apr.
13 May
19 May
26 May
9 June
7 July
27 July
28 July
5 Aug 1982
6 Aug. 1982
23 Aug.
25 Aug.
26 Aug.
27 Aug.
30 Aug.
2 Sep.
6 Sep.
14 Sep.
17 Sep.
27 Sep.
19 Oct. 1982
2 Nov. 1982
3 Nov.
30 Dec. 1982
5 Jan. 1983
6 Jan.
10 Jan.
12 Jan.
3 Feb.
17 Feb.
18 Feb.
24 Feb.
3 Mar.
8 Mar. 1983
10
15
20
25
30
35
z (km)

background is considerably altered in the aftermath of an eruption, and a large change in the aerosol characteristics, relatively fast in the initial phase, occurs over a time interval of a few years.

The inversion of the results for the zonally-averaged infrared extinction, obtained with the SME Infrared Radiometer, has given an indication of the rapidity of the gas-to-particle conversion: the aerosol increased in mass to a maximum of 8 Tg about 15 weeks after the eruption (Thomas *et al.*, 1983).

Different functional expressions have been adopted to fit the available size distribution measurements and provide a base for analytic modeling. Mainly used at present are the log normal and the equivalent zeroth order logarithmic distribution (ZOLD), bell-shaped, monomodal curves characterised by a maximum and a width parameter.

Hofmann and Rosen have investigated the question of size distributions through the use of balloon-borne dust sondes. Data obtained shortly after the eruption showed the difficulty of fitting standard distributions (Hofmann and Rosen, 1983). In a recent study of the temporal evolution of the cloud they obtained log-normal fits, after approx. 200 days with mean radii around 0.15 μm in the 17-20 km region and around 0.27 μm in the 20-25 km region, and around 0.22 μm after approx. 400 days (Hofmann and Rosen, 1984).

Size distributions obtained in the height region from 15 to 20 km by different airborne techniques, 8 to 10 months after the El Chichon eruption, show the tendency of secondary modes with larger particle radii to develop and eventually take over, indicative of the importance of sedimentation at this stage in the evolution of the cloud.

Knollenberg and Huffman (1983), using laser-based size spectrometers, find a large-particle sedimentation mode with modal radius of about 1.8 μm at 15 km and about 1.3 μm at 20 km. This fact is in agreement with the assumption that the amount of H_2SO_4 in the particles remains relatively constant and that particles at the two heights are characterized by a different dilution. In that case, if particles at 20 km had a 75% H_2SO_4 concentration, the concentration at 15 km would be around 30%.

Oberbeck *et al.* (1983), using a wire impactor, also find evidence for bimodality, their large-particle sedimentation mode having a modal radius around 0.6 μm with a geometric standard deviation of 1.5 at 20.7 km.

In fact it may appear that the recently observed monomodal distributions lack those properties of symmetry implied by the use of a log-normal distribution function and that on the larger-radii side the slope of the function is sharper: the recognition of this fact has important consequences in calculations of the optical cross section where the large particles weigh a lot more than the smaller ones.

Two-wavelength lidar observations (Gobbi *et al.*, 1984) indicate a change in the ratio of the aerosol cross sections respectively at 1.064 μm and 0.532 μm which can be attributed to the existence of a narrow mode in the size distribution, or, possibly, to marked differences in the imaginary part of the refractive index at the two wavelengths.

Patterson *et al.* (1982) have studied the optical properties of the ash from the El Chichon in the wavelength region 300-700 nm. The samples were collected on the ground at various distances from the volcano. In the absence of in-situ data the refractive index for the stratospheric silicate was given as $\tilde{n} = 1.53 - 0.001\,j$.

Less is known at this time about the complex refractive index of the sulfur aerosol, and its variation with height, in view of the role tbat dilution and impurities may have, particularly in determining the imaginary part $\tilde{n}''$. Knowledge of the refractive index is essential in determining heating effects, and would help in assessing the aerosol composition.

As regards the worldwide spreading of the cloud, in a first approximation, models start by considering the effect of an undisturbed circulation on the motion of a conservative tracer. More intriguing is to ascertain to what extent aerosols can be considered conservative tracers and the possible effects of the heating induced by the cloud on the circulation.

Satellite observations, although lacking the resolution and sensitivity of other methods, give a global view of the phenomenon. Partial conclusions to be drawn from the satellites' observations are: the high altitude cloud took twenty one days to first circle the Earth moving from East to West; its maximum density occurred about eight weeks after the eruption. After the first week in June the aerosol completely encircled the Earth in the latitude band from the Equator to 30 N.

The subsequent spreading of the cloud towards higher latitudes in the Northern hemisphere can be followed through the sequence of ground-based lidar observations. On April 18 Hirono and Shibata (1983) reported intense layers at 16, 25.5 and 26 km. Fig. 1 shows the sequence of observations at Frascati: the profiles give the ratio of the total backscattering cross section of air (aerosols and molecules) to the molecular backscattering cross section as a function of altitude.

On April 21 a relatively weak layer appears around 16 km, possibly the first sign of arrival of the cloud. It is not until June that upper stratifications can be seen at that site. The intensity of the echoes reported is by an order-of-magnitude weaker than those observed at Fukuoka by Hirono and Shibata (1983). The latitude of the two stations differs by about 8°. The May-June data show, in the 16-20 km region, complex sets of stratifications, a few hundred meters thick separated by less than 1 km, which disappear by July. From now on, the 16-20 km region appears characterised by a relative homogeneity. After June 9th conspicuous stratifications are visible above 20 km, with particularly large R values obtained in July around 25 km and in August at various heights in the 21-27 km interval. At the end of summer the aerosol load shows a decrease and is down to early summer values. A consistent increase begins at the end of October: by now the layers have lost most of the small scale features and the lower stratosphere appears continuously filled from 10 to 25 km. After the beginning of the year a decreasing trend in the optical thickness is evident from these and subsequent unpublished data.

The thin stratifications appearing on the Frascati lidar data for several months after the eruption cannot be accounted for on the basis of simple models of turbulent vertical diffusion acting on inert particles. As already mentioned, their long lifetime is related to a balance between the different and variable mechanisms which affect the vertical transport of sulfur and water substance, in particulate and gaseous form. Particulates are acted upon by sedimentation, by vertical advection where that exists, e.g. in the Hadley cell, and by turbulent diffusion; they may gain or lose water. Gases move under the action of vertical advection and molecular or turbulent diffusion. Conditions existing in the upgoing branch of the Hadley cell may favor the formation of thin layers.

There is evidence that polar air masses, one and half years after the eruption, carry a good deal less aerosols indicating that meridional mixing is incomplete. In an analysis of the SAM II data satellite extinction data for the Northern hemisphere winter of 1982, supplemented by airborne lidar observations, McCormick *et al.* (1983) found the polar vortex to be an area of substantially low aerosol content where the El Chichon cloud does not seem to have penetrated and that either an aerosol sink or a supply of clean air exists in the polar winter vortex.

This observation is an additional indication that large horizontal and vertical inhomogeneities may arise as a result of the circulation, long after the eruption. These effects were noticed also after the Mt. Agung eruption and were a basis to explain, in part, correlations found between ozone and aerosol content (Grams and Fiocco, 1967).

Modeling efforts aimed at simulating the worldwide diffusion of the cloud begin to appear: preliminary results gave rather faster transit times than have actually been observed (Capone *et al.*, 1983; Pitari *et al.*, 1984). While these results may be due to the scarce resolution of the circulation models, they may also point to the rather complex nature of the aerosol dynamics (see e.g. Crescentini and Fiocco, 1983). Related anomalies were found by us in utilising the meridional distribution model of Cadle *et al.* (1976), in efforts to simulate the effects of Mt. Agung (Fiocco *et al.*, 1977).

Essential in establishing the heating induced by El Chichon is the separation of other concurrent effects due to El Niño and to the quasi biennial oscillation.

Labitzke *et al.* (1983) have presented preliminary results, obtained almost in "real time", of a comparison between monthly mean stratospheric temperatures between 10° and 30° N for the years 1964-1982 with those of 1982: definite differences appear for the months of July and August with + 5 °C above the 18 year average at the 30 mbar level over 10° N.

Having examined the various periodic and quasi-periodic causes of fluctuation, Quiroz (1983) concludes that the amount of warming in the summer of 1962 due to the El Chichon at 30 mbar is 1 − 3 °C at latitudes between 35 N and the Equator, and that effects related to a volcanic modulation of the QBO would he of secondary importance.

Pollack and Ackerman (1983) have reported the results of calculation with a one-dimensional radiative-convective model which predict the El Chichon cloud to have caused an increase of planetary albedo of 10%, a decrease in total radiation at the ground of 2-3%, and an increase in temperature of 3.5 degrees at the 30 mhar level. The GCM of the European Center for Medium Range Forecast was utilised to model the perturbation introduced by a fixed layer with an optical thickness of 0.15 added to the background: a stratospheric warming of 3.5 °C in the stratosphere and a cooling of about 0.1 °C near the surface was obtained (Tanre and Geleyn, 1984).

In conclusion, the present intense research effort triggered by the El Chichon eruption should have a significant impact on aerosol science and on dynamic climatology. It is fair to expect that a detailed description of the spatial and temporal variation of the radiative field, of the fluxes reaching the surface, of the heating rates, should be obtained by the measurements and by the use of radiation codes. These values will be used as inputs to general circulation models, to check their ability to simulate the spreading of the aerosol clouds, and to recognize expected feedback effects.

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DISCUSSION

MARINI-BETTÒLO

I would like to know if there are figures about the average sulphur emissions by volcanoes per year.

FIOCCO

I can refer to two recent papers, one by Sedlacek *et al.*, the other by Berresheim and Jaeschke. The global volcanic sulphur release averaged between 1961-79, is estimated to be around $1 \cdot 5 \times 10^7$ t(SO_2)/a: of this the contribution due to volcanic eruptions appears to be around 7% of the non-eruptive emissions. At least 58% of the sulfate aerosol present in the lower stratosphere between 1971-78 was of volcanic origin.

ARNOLD

A brief comment concerning your mechanism which stabilizes the dense layer in your suggestion. Your mechanism would apply only to condensed matter, to particles. However, the sulphuric acid vapor which we have measured is also concentrated in a very thin cloud, and here it should be kept in mind that the vapor is formed *in situ* probably from SO_2 precursor gas and that the lifetime of the sulphuric acid molecule is much less than one day. In the undisturbed stratosphere it is about one day. Under these conditions, where there is more aerosol, the vapor has an even shorter lifetime. And this would tell us that also the gas, and that means the precursor gas, the SO_2 is concentrated in such a thin layer. So this does not necessarily apply only to the aerosols which are formed from the condensable gas formed in turn from the precursor.

CRUTZEN

On the question of confinement in thin layers, since in the initial cloud which gets into the stratosphere the SO_2 concentration is very high, SO_2 being a strong absorber of ultraviolet radiation, one can think of reaction chains in which SO_2 oxidation leads to ozone formation. I once did calculations on the heating rates one gets there, and it is on the order of 10 to 30 degrees per day, which is very large for the stratosphere. So I cannot see why these layers under such conditions can stay that confined; maybe what we are looking at is two phases really of the volcanic eruption: first, one has the initial eruption and then maybe a fast spread vertically, and then the SO_2 is oxidized and starts speading horizontally in sheets.