1. Classification *INPE-COM.10-PE* *CDU: 550.3A*	2. Period	4. Distribution Criterion internal ☐ external ☒
3. Key Words (selected by the author) *AEROSOLS, STRATOSPHERE, ATMOSPHERIC CIRCULATION, VOLCANIC EMISSIONS, LIDAR*		
5. Report No. *INPE-1138-PE/097*	6. Date *October, 1977*	7. Revised by
8. Title and Sub-title *STRATOSPHERIC DUST MEASUREMENTS, 1970-1977*		9. Authorized by *Nelson de Jesus Parada* *Director*
10. Sector *DCE/FISAT*	Code *412*	11. No. of Copies *13*
12. Authorship *B. R. Clemesha* *D. M. Simonich*		14. No. of Pages *25*
13. Signature of first author		15. Price

16. Summary/Notes

Stratospheric aerosol measurements have been made by laser radar at São José dos Campos (23ºS, 46ºW) since July 1970. From 1970 to 1974 the measured scattering ratios for the 20 km layer were similar to those reported for other locations. A large increase was observed in April 1975, six months after the eruption of Volcán de Fuego in Guatemala. This delay in the transport of the Fuego dust to our latitude is interpreted as being due to the inhibition of southward eddy transport, during the southern summer, by the mean meridional circulation. A maximum monthly average scattering ratio of 1.28, measured at a wavelength of 5890 Å, was observed at 20 km in August 1975. Since early 1976 the 20 km scattering ratio has oscillated around a value of about 1.15, but the integrated columnar back--scattering coefficient above 17 km, indicative of the total stratospheric dust loading, has increased by about 40%. Scattering from the upper stratosphere, at heights above 30 km, was observed in the southern springs of 1971, 1972 and 1973. There appears to be some evidence that the dust responsible for this scattering was of extraterrestrial origin.

17. Remarks

STRATOSPHERIC DUST MEASUREMENTS, 1970 - 1977

B. R. Clemesha and D. M. Simonich

Instituto de Pesquisas Espaciais (INPE)
Conselho Nacional de Desenvolvimento Científico e Tecnológico (CNPq)
São José dos Campos, São Paulo, Brazil

ABSTRACT

Stratospheric aerosol measurements have been made by laser radar at São José dos Campos (23°S, 46°W) since July 1970. From 1970 to 1974 the measured scattering ratios for the 20 km layer were similar to those reported for other locations. A large increase was observed in April 1975, six months after the eruption of Volcán de Fuego in Guatemala. This delay in the transport of the Fuego dust to our latitude is interpreted as being due to the inhibition of southward eddy transport, during the southern summer, by the mean meridional circulation. A maximum monthly average scattering ratio of 1.28 measured at a wavelength of 5890 Å, was observed at 20 km in August 1975. Since early 1976 the 20 km scattering ratio has oscillated around a value of about 1.15, but the integrated columnar backscattering coefficient above 17 km, indicative of the total stratospheric dust loading, has increased by about 40%. Scattering from the upper stratosphere, at heights above 30 km, was observed in the southern springs of 1971, 1972 and 1973. There appears to be some evidence that the dust responsible for this scattering was of extraterrestrial origin.

1. OBSERVATIONS

The stratospheric aerosol profile has been monitored by laser radar at São José dos Campos ($23^{o}S$, $46^{o}W$) since 1970. Some preliminary results were published by Clemesha and Rodrigues (1971), and the observation of an unusually high altitude dust layer in October 1971 was reported by Clemesha and Nakamura (1972). In this paper we describe the results of observations made up till July 1977.

Observations were made with three different lidars. The first system was of low sensitivity, and gave useful results only up to 35 km. This system, equipped with a ruby laser, was used to obtain scattering profiles in July and August of 1970. The second system, also using a ruby laser, was used from July 1971 till December 1972; it was of much greater sensitivity, and gave profiles up to 80 km. The third system, which uses a dye laser tuned to 5890 Å, has been in use since March 1972. The specifications of the three lidars are given in Table 1. All three systems used the same photon counting receiving system, the photomultiplier of which is protected against the laser pulse scattered in the lower atmosphere by a rotating shutter. A second rotating shutter was used to cut off fluorescence from the ruby laser. This shutter is not used in the case of the dye laser because the much shorter fluorescence decay time renders it unnecessary. The signal back-scattered from the atmosphere is recorded in 2 km range bins in a multichannel scaler. The scaler channels are periodically recorded on magnetic tape and processed for display purposes on an on-line programmable calculator.

2. DATA ANALYSIS

The precision of measurements made with a lidar using photon counting in the receiver depends on the number of photon pulses counted. The reasonable assumption is made that the pulses follow a Poisson distribution in time, so that the standard deviation of a measurement corresponds to the square root of the number of pulses involved, appropriate allowance being made for background noise. In our case a typical night's measurements using the dye laser is based on several thousand laser shots, resulting in a precision of better than 1% at 20 km, to about 10% at 45 km for the more recent measurements. In the case of the ruby laser a smaller number of shots was normally employed, but the much greater energy of this laser led to higher counting rates, giving typical precisions of 1% at 40 km to 5% at 60 km. The least accurate measurements were those made in 1970, with precisions of about 1% at 10 km to 20% at 30 km.

TABLE 1

LIDAR SPECIFICATIONS

SYSTEM	1	2	3
Transmitted Energy	0.5	5	0.03 Joules
Pulse Duration	2	2	1.5 μs
Pulse Rate	0.16	0.16	0.3 s^{-1}
Wavelength	6943	6943	5890 Å
Beamwidth	0.6	0.6	0.8 mr.
Receiver Area	0.01	0.39	0.39 m^2
Receiver Bandwidth	20	20	10 Å
Height Resolution	2	2	2 km

In order to derive aerosol scattering profiles the expected Rayleigh scattering from the molecular atmosphere must be subtracted from the experimental profiles. For heights up to 30 km we have calculated the Rayleigh scattering on the basis of an annual mean density profile obtained from 2 years of data from meteorological balloon ascents made at São José dos Campos. Above 30 km we have used a profile which is the mean of the 15^{o}N and 30^{o}N curves from the U.S. Standard Atmosphere Supplements (1966). We have compared this to a series of 8 rocket profiles taken between July and October 1976 at Marambaia, 200 km east of São José dos Campos, and find good agreement.

We estimate that in no case would the use of our standard atmosphere model, instead of the true atmospheric density variation, lead to an error of more than 0.05 in the derived scattering ratio (ratio of total to Rayleigh scattering).

Comparison between the lidar profiles and the calculated molecular atmosphere was made by visual curve fitting. It was sometimes possible to fit the curves both above and below the height at which any aerosol scattering present was observed, but in cases where aerosol scattering was present at the lowest data point, the fitting was done only at heights of 30 km or above. Pettifer (1974) has suggested that signal induced noise can cause serious errors in aerosol measurements obtained by laser radar, and that the nature of these errors depends on the manner in which the fitting of the lidar profile to the expected molecular scattering curve is done. The photomultiplier used in our lidar has been carefully tested, and signal induced noise effects were found to be negligible (Clemesha, 1977).

3. RESULTS AND DISCUSSION

In Figure 1 we show the time-height variation of the aerosol scattering ratio from mid-1970 until mid-1977. The 1970 measurements were made at a wavelength of 6943 Å, as were the observations at 24 km and above from July 1971 to December 1972. All other measurements were made at 5890 Å. Except in the case of October 1972, no adjustments have been made to the scattering ratios shown in Figure 1 to allow for the different wavelengths used. During October 1972 the dye laser measurements were limited to a maximum height of 26 km, and the ruby laser measurements, extending down to a minumum height of 24 km, showed a scattering ratio of 1.06 at this height. For reasons which will be explained later, it was assumed that his should correspond to a scattering ratio of 1.04 at 5890 Å, and the dye laser measurements were normalised accordingly.

The lowest height at which the scattering can be measured is limited by saturation of the photon counting receiver system.

Variations in the performance of the dye laser, and in the atmospheric transmission, result in fairly large variations in this height. The scattering ratios shown in Figure 1 are monthly averages, and the minimum height shown for each month represents the lowest height at which data were obtained during that month. For this reason the scattering ratio at this height is not normally the average of as many days observations as that for greater heights. Similar considerations apply to the greatest height shown for each month. For any given profile the greatest height was taken as that height were the standard deviation in the signal was about 5%. For the most recent measurements this height was usually greater than 40 km. During late 1971 and part of 1972, some observations were made using the ruby laser with a neutral filter attenuating the received signal in such a way as to reduce the lowest observable height from 30 km to about 24 km. The profiles obtained in this way overlapped the dye laser profiles at low altitudes and the normal ruby laser profiles at high altitudes. In this way it was possible to construct scattering profiles from 10 km to greater than 60 km.

3.1 - THE LOWER STRATOSPHERE

The scattering ratio at the peak of the 20 km aerosol layer in July and August 1970 was about 1.4. Measurements made at a similar epoch in Boulder, Colorado (Schuster, 1970) and Adelaide, Australia (Gambling et al, 1970) showed very similar ratios for the same wavelength. From August 1970 till June 1971 no observations were made, and observations of the 20 km layer were only resumed in March 1972, when the dye laser was put into operation. The observations from March 1972 till April 1975 frequently showed no measurable scattering from the 20 km layer. It is difficult to estimate the minimum detectable scattering ratio, because this depends on the height distribution of the aerosols. A constant scattering ratio at all heights would be undetectable even at quite high aerosol concentrations, whereas a narrow layer would be obvious even with a scattering ratio as low as 1.05. In general, we believe that our observations would have reliably detected an aerosol layer similar to that observed between 1964 and 1970 in many locations, i.e., one which peaks at about 20 km and has a half-width of 5 to 10 km, if the maximum scattering ratio were 1.08 or more.

Fernald and Schuster (1977), on the basis of an extensive series of observations in the northern hemisphere, made with an airborne lidar system, find that 1973 was in a period of low aerosol activity. They find peak scattering ratios between 1.04 and 1.2, and, since these workers used a dye laser tuned to 5850 Å, their results should be directly comparable to ours. The average 20 km scattering ratios found by Fernald and Schuster were about 1.12, except at latitudes less than 10^0. We made observations on a total of 33 occasions during 1973, on only 7 of which scattering was detectable from 20 km, with scattering ratios between 1.06 and 1.12. Comparison of our results with those of Fernald and Schuster suggests that the concentration of dust was slightly higher in the northern hemisphere during this period of low aerosol activity. It should be remembered, however, that the scattering ratios which we observed were close to the lower limit of detectability of our lidar, and thus are not very accurate. It should also be noted that our measurements would tend to underestimate the monthly average scattering ratios, because individuals days on which the scattering ratio was very small would be taken as showing zero scattering, whereas the errors in this parameter measured when the aerosol was clearly observable should be normally distributed, and thus tend to cancel out.

Dustsonde measurements made during 1973 by Rosen et al (1976), show little difference between northern and southern hemispheres. Mie theory calculations and actual comparison between lidar and dustsonde experiments (Pinnick et al, 1976) indicate that the average peak aerosol concentrations measured by Rosen et al should correspond to scattering ratios of 1.09 to 1.17 at 6943 Å. The same workers find that the scattering cross section for a typical aerosol distribution should vary as λ^{-1}, and that this wavelength dependance is relatively insensitive to the parameters of the size distribution. As the Rayleigh cross-section varies as λ^{-4} the ratio of aerosol to Rayleigh scattering (scattering ratio - 1) should vary as λ^3. This means that the ratio measured at 5890 Å should be approximately 60% of that measured at 6943 Å, i.e. 5890 Å ratios corresponding to Rosen et al's observations would be in the range 1.05 to 1.1, in good agreement with our results. It seems, then, that our results are in slightly better agreement with the dustsonde measurements than with other lidar observations.

Scattering from the 20 km layer continued to be very weak until April 1975, with only occasional short-lived layers being seen at other heights. The scattering from 10 to 16 km in March 1972 was observed on 5 days out of a total of 6 on which observations were made. A layer at 10 km in April 1972, with a scattering ratio of 1.04, was observed on only one day out of 7, and a 12 to 14 km peak in May 1972, with a peak scattering ratio of 1.22, was seen on one day only. Again, the 16 km peaks observed in October 1972 and February 1973, and the 14 km scattering seen in October 1973 all refer to single observations.

Monthly averages of the 20 km scattering ratio and the integrated columnar non-Rayleigh backscattering coefficient are shown in Figure 2. In order to prevent variations in the lowest height of measurement from influencing the integrated backscattering coefficent, this parameter was calculated from 17 km up, 17 km being the lowest minimum height for any month during which measurements were made. It is clear from Figure 2 that a major increase in the stratospheric aerosol occurred in April 1975. We attribute this increase to particles injected into the stratosphere by the eruption of the Volcān de Fuego in Guatemala, which took place in mid-October 1974.

3.2 - DUST FROM THE FUEGO ERUPTION

Enhanced scattering from the stratosphere, caused by particles injected by the eruption of Volcān de Fuego in Guatemala, was first reported by Meinel and Meinel (1975), who observed unusual sunset effects in California. Quantitative observations were made by lidar

(Fegley and Ellis, 1975, Fujiwara et al, 1975, Fernald and Frusch, 1975, Remsberg and Northam, 1975, Russel and Hake, 1975, Russel and Hake, 1977) and by balloon and aircraft sampling techniques (Gras, 1976, Hofman and Rosen, 1977, Ferry and Lem, 1975). In the northern hemisphere, maximum particle concentrations appear to have been observed one to two months after the eruption, but the situation is complicated by the fact that increased stratospheric aerosol was observed a few days before the main eruption (Hofman and Rosen, 1976, Fegley and Ellis, 1975), suggesting that a smaller additional injection took place at an earlier date.

In the southern hemisphere, Gras (1976) observed an aerosol layer at Mildura (34^{O}S) 6 weeks after the eruption, and Rosen et al (1975) found an enhancement of the 15 km to 20 km particle concentrations by a factor of 3 in January 1975, as compared with one year previously, at McMurdo Sound (78^{O}S) and South Pole Station. Both of these groups consider that the initial enhancements were unconnected with the Fuego eruption. A sounding at Mildura in February 1975 showed no enhancement (Rosen et al, 1975), and Gras's observations showed a second increase, in May, which he attributes to the Fuego event.

Gras (1976) suggests that the dust layer which he observed at 34^{O}S on November 29, 1974, and the layers observed in the northern hemisphere on October 8, (19^{O}N - Fegley and Ellis, 1975 and 41^{O}N - Hoffmann and Rosen, 1975), might have originated from the Manan Island Volcano (4^{O}S, 14^{O} E), active between May and late September 1974. We made observations on the 7th, 9th, 10th and 14th of October and on the 4th, 14th and 18th of November. On the 10th and 14th of October we observed higher than average scattering ratios, with a maximum of 1.17 at 25 km on October 10. This value is only slightly greater than those observed on a number of other occasions during 1974, and cannot be considered as necessarily indicating a specific volcanic injection. Furthermore, the layer was vertically diffuse, lacking the sharp peak just above the tropopause which appears to be characteristic of recent volcanic injections. It should also be pointed out that stratospheric scattering layers were observed in October in the three previous years, and it is possible that they were not volcanic in origin. These layers are discussed in section 3.4.

The observations reported here showed no excess scattering form above 15 km in November 1974, and only very weak scattering in February 1975. The February measurements were made on a single day only, the 24th, and showed a peak scattering ratio of 1.08 at 26 km to 32 km. Measurements made on the 3rd, 4th, 7th and 11th of March showed no aerosol scattering from above 17 km. On the 7th of April a maximum scattering ratio of 1.08 was measured at a height of 26 km. Strong scattering was first observed on April 8, with maximum scattering ratios of 1.4 and 1.27 being measured at 14 km (the lowest height of observation) and 18 km respectively. On the 10th and 11th of April peak scattering was observed at 18 km, but subsequent to the latter date the peak was almost invariably to be found at 20 km. During the following months an upward trend in the peak scattering ratio was observed, with a maximum monthly average of 1.28 at 20 km being reached in August. The highest ratio observed during the entire event was 1.43 at 20 km on July 10th. Figure 3 shows the height/time variation of the weekly average scattering ratio from March 1975 to July 1977, and Figure 4 shows the variation of the weekly average of the 20 km scattering ratio for the same period.

The integrated aerosol backscattering coefficient did not increase significantly after April 1975, indicating that the high 20 km scattering ratios observed in July and August were the result of changes in the height distribution of the dust rather than an increase in the total aerosol burden. The 20 km aerosol scattering decreased towards the southern winter of 1975/76, reaching its minimum value in March 1976, after which date it has shown a slight upward trend, with fairly large superimposed oscillations. The integrated aerosol backscattering coefficient, which is a measure of the total dust load, has increased more than the 20 km scattering ratio as a result of a widening of the layer clearly visible in Figures 1 and 3. In July 1977 the integrated scattering was actually greater than the maximum value reached in August 1975.

During 1975 the scattering ratio underwent large and rapid oscillations with a period of a few weeks. Both the amplitude and the frequency of the oscillations decreased in 1976. These oscillations presumably result from spatial inhomogeneities in conjunction with atmospheric circulation, in which case the decrease in amplitude and frequency would result from the destruction of the smaller scale inhomogeneities by horizontal mixing.

The fact that we did not observe dust from the Fuego eruption until April 1975 supports the opinions of Gras (1976) and Rosen et al (1976) that the increased stratospheric aerosol, which they observed in December and January in the southern hemisphere, was unconnected with the Fuego eruption. The 6 month delay in the transport of aerosols from 14^{o}N to 23^{o}S can be explained on the basis of the meridional circulation of the tropical atmosphere. During the northern summer a large tropical Hadley cell dominates the meridional flow as far as 30^{o}S, resulting in southward flow at heights above about 6 km (Newel et al, 1972). Our knowledge of the circulation of the tropical atmosphere is mainly based on meteorological balloon observations, which are inadequate for the purpose of determining the mean meridional circulation above a height of about 15 km. It seems probable, however, that the Hadley cell extends to greater heights; this is suggested by theoretical models (Manabee and Mahlman, 1976, Kasahara et al, 1973) and by the radioactive tracer measurements of Telegados (1971). Under northern summer conditions, the transport of particles to our latitude would be assisted by the upper part of the Hadley cell, in the region of 20 km. During the period immediately following the eruption, southward eddy transport in the stratosphere would be inhibited by the mean meridional circulation associated with the southern hemisphere summer cell penetrating to sub-tropical latitudes in the northern hemisphere.

We may compare our observations with the results of Cadle et al (1976), who have applied a dynamical atmospheric model to the spread of dust from the Fuego eruption to determine isopleths of particle mixing ratio as a function of height and latitude. They compare the results of their calculations with lidar observations from NCAR, Boulder, Colorado (Fernald and Frush, 1975) and Mona Lao, Hawai (Fegley and Ellis, 1975). In order to make this comparison Cadle et al use a conversion factor of 10 to convert from scattering ratio, R, to parts per billion, ppb, i.e. ppb = 10 (R - 1). They base this equivalence on various workers comparisons between lidar and dustsonde measurements and on Mie theory calculations. The lidars used at Boulder and Mona Lao operate at 6943 Å, so, in order to allow for the difference in wavelength, and thus make our results directly comparable, we have used a factor of 17 to give the parts per billion scale in Figure 4.

Based on a postulated source function, Cadle et al computed sulphuric acid aerosol mixing ratios between 3 and 10 times the particulate mixing ratios measured by lidar in the northern hemisphere. Taking fine ash concentrations for our latitude from Figure 4 of Cadle et al (1976), and, based on the source function given in Table 1 of the same publication, multiplying by a factor of two to give approximate sulphuric acid aerosol mixing ratios, makes the maximum mixing ratio in the region on 20 km 30 ppb. This mixing ratio is reached by April 1975, and remains little changed at least until February 1976, which is as far as Cadle et al take their computations. The lidar measurements reported here show a peak mixing ratios in the region of 4 ppb during the summer of 1975, and thus the computed values are about 7.5 times too large. This factor falls within the range of values found for northern hemisphere measurements, i.e. 3 to 10. It may be seen, then, that Cadle et al's model is coherent with our observations, provided the source function is suitable modified.

3.3 - OTHER VOLCANIC ERUPTIONS

The total dust at heights above 17 km has increased by 40% since March 1976. This increase is presumably due to the in-situ formation of sulphuric acid by the oxidation of SO_2, and/or further volcanic injection of particulates. Major volcanic eruptions, possibly capable of injecting material to stratospheric heights, occurred in St. Augustine, Alaska in January 1976, Nyiragongo, Zaire in January 1977, Bazymianny, USSR in March 1977 and Alaskan Peninsula in March 1977. Of these only the St. Augustine and Nyiragongo eruptions have had stratospheric aerosol observations associated with them. Cadle et al, 1977, report a small increase in lidar returns measured at Boulder, Colorado, 20 days after the St. Augustine eruption, but in view of the fact that this increase does not even attain the level of the extrapolated decay curve for the Fuego dust, any attempt to relate the lidar observations with the St. Augustine volcanic eruption must be considered somewhat speculative. Fegley (1977) reports the observation of a dust layer over Hawaii 16 days after the Nyiragongo eruption. Confirmation of this injection must await the analysis of an adequate time series observations.

We are unable to positively identify an increase in stratospheric aerosol with any of the above mentioned events. It is possible, however, that high scattering ratios observed at 16 km on February 9 and 14, 1977 and at 14 km on March 11, 1977 might be associated with the Nyiragongo eruption. The average scattering ratios which we observed were 1.43 and 1.38 at 16 km on February 9 and 14, and 1.4 at 14 km on March 11. These were the largest observed since August 1975, and thus may well be indicative of significant injection of particulate material into the lower stratosphere. At least in the case of the March observation, equinox conditions should obtain, and the meridional circulation would be appropriate for the transport of dust from the latitude of the eruption (1^{o}S) to our location.

3.4 - SCATTERING FROM ABOVE 30 KM

Observations of the atmosphere above 30 km were made from June 1971 onwards. The high sensitivity ruby lidar was used until December 1972, giving useful data up to 80 km. After this date the dye laser produced results up to about 40 km. Up until October 1971 the scattering profiles were generally compatible with a clear atmosphere, although occasional measurements could be interpreted as showing a small amount of aerosol scattering. On October 19 a strong scattering layer was observed at 47 km, with a peak scattering ratio of 1.6 and a half-width of about 3 km. The appearance of this layer, and the subsequent variations in the scattering from the 30 km to 80 km region until January 1972 have been described by Clemesha and Nakamura (1972).

The authenticity of the 47 km layer observed on October 19, 1972 has been questioned in the literature (Volz, 1974). When the results showing this scattering layer were originally published it was thought unnecessary to show any additional evidence for its genuine nature, the laser radar technique being fairly well established by that time. In view of the criticism cited above, however, we present the following description of the checks which were made to ensure that the layer was a genuine atmospheric phenomenon, and not the result of a malfunction of the lidar.

On October 19 observations were made from 2040 LT till 2152 LT, when cloud cover stopped further measurements from being made. A total of 11 profiles of 25 lasers shots each was obtained. For each individual profile the signal from 47 km was typically 8 standard deviations above that expected from the standard atmosphere fitted at heights above and below this point. When observations were resumed at 1950 LT on October 20 the peak scattering ratio had dropped from 1.6 to 1.1. A series of 175 shots, with the lidar pointed towards the zenith, gave a signal in channel 17 of the lidar receiver 6 standard deviations above the interpolated return from adjacent higher and lower channels. In the configuration in use channel 17 corresponded to a height of 43 km. Immediately after the series of 175 shots, and starting at 2110 LT, the lidar was inclined at 30^0 to the zenith, and a series of 300 shots was taken at this angle. Peak excess signal 11 standard deviations above the interpolated return from adjacent higher and lower channels was found in channel 21, corresponding to a slant range of 51 km, or to a height of 44.2 km. The fact that the scattering peak was observed at an almost constant height, and not at a constant range, nor in a constant channel number of the multi-channel scaler, appears to us to conclusively prove the genuine existence of an atmospheric scattering layer.

The initially well defined layer was still visible, 3 km lower in height, 24 hours after its first observation, but by then excess scattering was visible at all heights from 35 km to 70 km. As may be seen from Figure 1, the measured scattering ratios decreased in subsequent months, and by April 1972 no excess scattering was observable. At heights above 50 km a small amount of excess scattering was again seen from September to October 1972, with maximum scattering in October. Aerosol scattering was also observed between 23 km and 37 km. No well-defined strongly scattering layers of the type observed on October 19, 1971 were observed during this period. The high altitude observations were discontinued in December 1972.

In October 1973 scattering layers were observed at a height of 34 km on the 4th and 21st, with peak scattering ratios of 1.4 and 2.2 respectively. On the 22nd of October 1973 the peak scattering ratio was 1.5 at a height of 36 km. Measurements made on the 3rd and 15th of October showed no excess scattering. From November 1973 onwards no scattering layers have been observed in the upper stratosphere, although the "tail" of the 20 km layer has occasionally been seen to extend as high as 34 km, and usually strong scattering from the region on 26 km was observed on October 10 and 14, 1974. It should be remembered, however, that the dye laser lidar in use since January 1973, would not reliably detect scattering layers at heights above 40 km.

The origin of the high altitude dust observed in October 1971 and in subsequent months is uncertain. An unusually large influx of extra-terrestrial material in the month of October has been suggested by Rosinski (1972), on the basis of the collection of magnetic spherules, and Rosinski et al (1975) report an extra large influx in October 1971. The fact that we gain observed scattering from the 50 km to 60 km region in the southern spring of 1972, and from a layer at 32 km to 34 km in October 1973, lends some weight to the hypothesis of an extra--terrestrial source, supposing the influx to result from the passage of the Earth through the dust cloud in the plane of a cometary orbit. Volz (1974) has reported the observation of unusual twilight color ratios in October 1971, and believes the dust responsible for this phenomenon to have originated in a volcanic eruption on the northern rim of the Pacific Ocean. If this was indeed the case any connection with the 47 km layer which we observed is unlikely, since it is improbable that the dust could have been transported so far south in a period of only one month. Furthermore, there appears to be no evidence for volcanic eruptions injecting dust to heights greater than 30 km.

A dust layer in the region of 50 km has been reported by Rossler (1968), who measured the diffuse sky brightness from an ascending rocket launched at Hamaguir, Sahara, in April 1963. Cunnold et al (1973) observed excess scattering from the same height range, using horizon inversion techniques on observations from an X-15-1 aircraft, and Giovane and Schuerman (1976) observed a layer at 48 km, using a solar occultation technique in observations made from Skylab during November 1973.

In conclusion, it appears that the presence of dust layers at heights above 30 km is not necessarily a rare event, although the concentration of the particles may normally be too small to be detected by lidar techniques. As yet, there is insufficient evidence to determine the origin of these particles.

4. CONCLUSIONS

The aerosol concentrations in the 20 km layer, measured at 23^{o}S between 1970 and 1974, were generally similar to those observed at other latitudes, although there is some evidence that the scattering ratios which we observed were slightly lower than at other locations. A major increase in the stratospheric dust loading occurred 6 months after the October 1974 eruption of Volcãn de Fuego in Guatemala. This delay in the trans-equatorial transport of dust to our latitude can be explained by the fact that, during southern summer, southward eddy transport would be inhibited by the mean meridional circulation of the atmosphere. Maximum scattering ratios were observed in August 1975, decreasing after this date to reach a minimum in March 1976. Since April 1976 the 20 km scattering ratio has oscillated about a mean value of about 1.15, but the integrated backscattering coefficient has increased by about 40% as a result of increased scattering from above 20 km. It has not been possible to identify this increase with any given volcanic event.

Observations of the upper stratosphere were made from July 1971 till December 1972. A strong scattering layer was observed at 47 km on October 19, 1971. Scattering was observed from a wide range of heights during subsequent months, and lidar profiles compatible with a purely molecular atmosphere were again observed only in May 1972. Excess scattering from heights above 50 km was again observed towards the end of 1972, and scattering layers at 30 km to 36 km were observed in October 1973. The source of these high altitude layers has not been identified, but there is some evidence for an extra-terrestrial origin.

ACKNOWLEDGMENTS

We are grateful to Dr. V.W.J.H. Kirchhoff, who helped with some of the observations, and to Mr. W.M. Lima, who assisted with the data reduction. This work was partially supported by the Brazilian National Fund for Science and Technology (FNDCT) under contract FINEP CT/271.

REFERENCES

Cadle, R.D., F.G. Fernald and C.L. Frush, Combined use of lidar and numerical diffusion models to estimate the quantity and dispersion of volcanic eruption clouds in the stratosphere: Volcān Fuego, 1974, and Augustine, J. Geophys. Res., 82, 1783-1786, 1977.

Cadle, R.D., C.S. Kiang and J.F. Louis, The global scale dispersion of the eruption clouds from major volcanic eruptions, J. Geophys. Res., 81, 3125-3132, 1976.

Clemesha, B.R., Transient Noise in Photomultiplier Tubes, J. Phys. E. Sci. Instrum., 10, 914, 1977.

Clemesha, B.R. and Y. Nakamura, Dust in the Upper Atmosphere, Nature, 237, 329, 1972.

Clemesha, B.R. and S.N. Rodrigues, The stratospheric scattering profile at 23^{o}S., J. Atmos. Terr. Phys., 33, 1119-1124, 1971.

Cunnold, D.M., C.R. Grays and D.C. Merritt, Stratospheric Aerosol layer detection, J. Geophys. Res., 78, 920-931, 1973.

Fegley, R.W., Event Notification Report, Event 7-77, The Center for Short-Lived Phenomena, 1977.

Fegley, R.W. and H.T. Ellis, Lidar observations of a stratospheric dust cloud layer in the tropics, Geophys. Res. Lett., 2, 139-141, 1975.

Fernald, F.G. and C.L. Frush, Lidar observations of the enhanced stratospheric dust layer associated with the eruption of volcano Fuego, EOS Trans., Am. Geophys. Union, 56, 366, 1975.

Fernald, F.G. and B.G. Schuster, Wintertime 1973 Airborne Lidar Measurements of Stratospheric Aerosols, J. Geophys. Res., 82, 433-437, 1977.

Ferry, G.V. and H.Y. Lem, Changes in stratospheric aerosols collected after the eruption of Volcān de Fuego, EOS Trans., Am. Geophys. Union, 56, 366, 1975.

Fujiwara, M., T. Itabe and M. Hirono, Sudden increase of stratospheric aerosol after the eruption of Fuego volcano; lidar observations in Fukuoka, Rep. Ionos. Space. Res. Japan, 29, 74-78, 1975.

Gambling, D.J., K. Bartusek and W.G. Elford, A 12 month study of aerosols below 60 km, J. Atmos. Terr. Phys., 33, 1403-1413, 1971.

Giovane, F. and D.W. Schuerman, The solar occultation technique for remote sensing of particulates in the earth's atmosphere, 2, Skylab Results of a 48 km Layer, J. Geophys. Res., 81, 5383-5388, 1976.

Gras, J.L., Southern hemisphere mid-latitude stratospheric aerosol after the 1974 Fuego eruption, Geophys. Res. Lett., 3, 533-536, 1976.

Hofmann, D.J. and J.M. Rosen, Balloon observations of the time development of the stratospheric aerosol event of 1974-1975, J. Geophys. Res., 82, 1435-1440, 1977.

Kasahara, A., T. Sasamori and W.M. Washington, Simulation experiments with a 12 layer stratospheric global circulation model. I. Dynamical effect of the earth's orography and thermal influence of continentality, J. Atmos. Sci., 30, 1229-1251, 1973.

Manabe, S. and J.D. Mahlman, Simulation of Seasonal and interhemispheric variations in the stratospheric circulation. J. Atmos. Sci., 33, 2185-2217, 1976.

Meinel, A.B. and M.P. Meinel, A stratospheric dust-aerosol event of November 1974, Science, 188, 477-478, 1975.

Newell R.R., J.W. Kidson, D.G. Vicent and G. Boer, The general circulation of the tropical atmosphere, Vol. I, Cambridge, M.I.T. Press, 1972.

Pettifer, R.E.W., Signal induced noise in lidar experiments, J. Atmos. Terr. Phys., 37, 669-673, 1975.

Pinnick, R.G., J.M. Rosen and D.J. Hofmann, Stratospheric aerosol measurements III: optical model calculations, J. Atmos. Sci., 33, 304-314, 1976.

Remsberg, E.E. and G.B. Northam, Measurements of stratospheric dust over Virginia by laser radar, EOS Trans., Am. Geophys. Union, 56, 365, 1975.

Rosen, J.M., D.J. Hofmann and J. Laby, In-situ measurements of the recent increase in stratospheric aerosol, EOS Trans., Am. Geophys. Union, 56, 365, 1975.

Rosinski, J., Global deposition of extra-terrestrial particles during October, J. Atmos. Terr. Phys., 34, 487, 1972.

Rosinski, J., C.T. Nagamoto and M. Bayard, Extraterrestrial particles and precipitation, J. Atmos. Terr. Phys., 37, 1231-1244, 1975.

Rossler, F., The aerosol layer in the stratosphere, Space Research, 8, 633-636, 1968.

Russel, P.B., W. Viez, R.D. Hake and R.T.H. Collis, Lidar observations of the post-Fuego stratospherical aerosol, EOS Trans., Am. Geophys. Union., 56, 365, 1975.

Schuster, B.G., Detection of tropospheric and stratospheric aerosol layers by optical radar (lidar), J. Geophys. Res., 75, 3123-3132, 1970.

Telegados, K., The upper portion of the Hadley cell circulation as deduced from the 1968 French and Chinese nuclear tests, J. Geophys. Res., 76, 5018-5024, 1971.

Volz, F.E., The stratospheric dust event of October 1971, J. Geophys. Res., 79, 479-482, 1974.

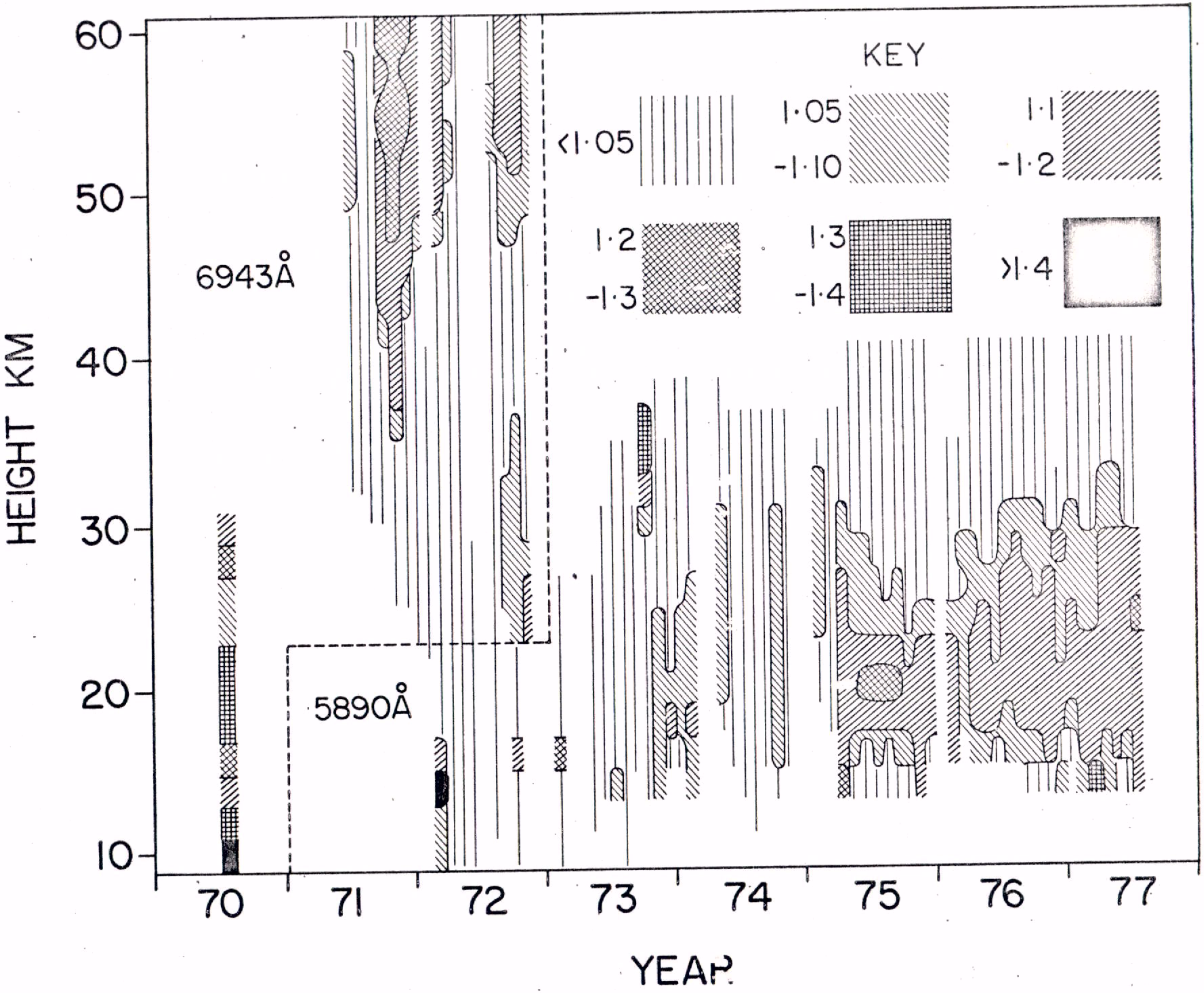
FIGURE CAPTIONS

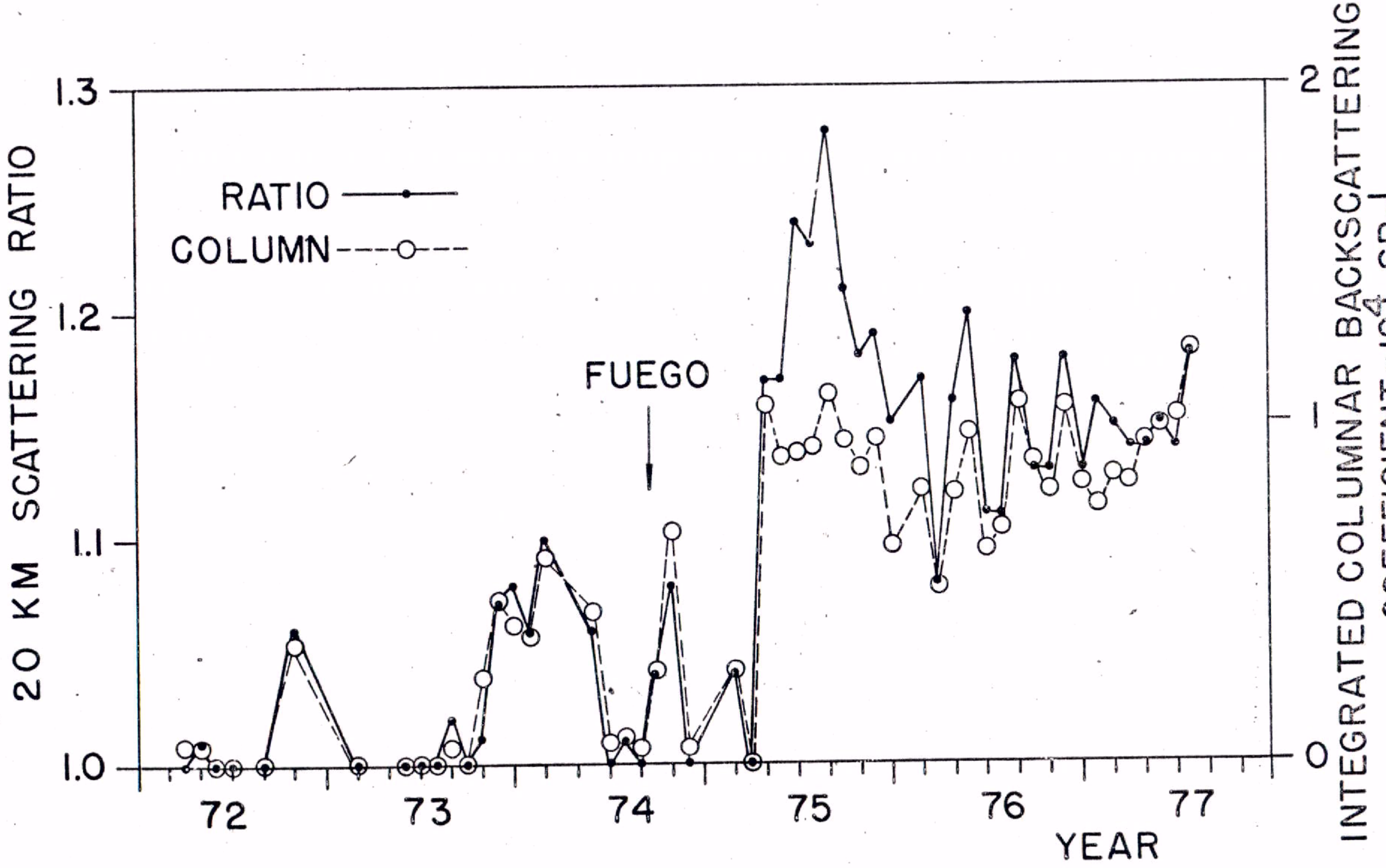
Figure 1 - Monthly average scattering ratios, 1970-1977.

Figure 2 - Monthly average 20 km scattering ratios and integrated columnar aerosol backscattering coefficient, March 1972-July 1977. The integral is taken from 17 km to 41 km.

Figure 3 - Weekly average scattering ratios, March 1975-July 1977.

Figure 4 - Weekly average 20 km scattering ratios and mixing ratios, March 1975-July 1977.


KEY
<1·05
1·05 -1·10
1·1 -1·2
1·2 -1·3
1·3 -1·4
>1·4
6943Å
5890Å
HEIGHT KM
60
50
40
30
20
10
70
71
72
73
74
75
76
77
YEAR



RATIO
COLUMN
FUEGO
20 KM SCATTERING RATIO
1.0
1.1
1.2
1.3
INTEGRATED COLUMNAR BACKSCATTERING COEFFICIENT x10^4 SR-1
0
1
2
72
73
74
75
76
77
YEAR

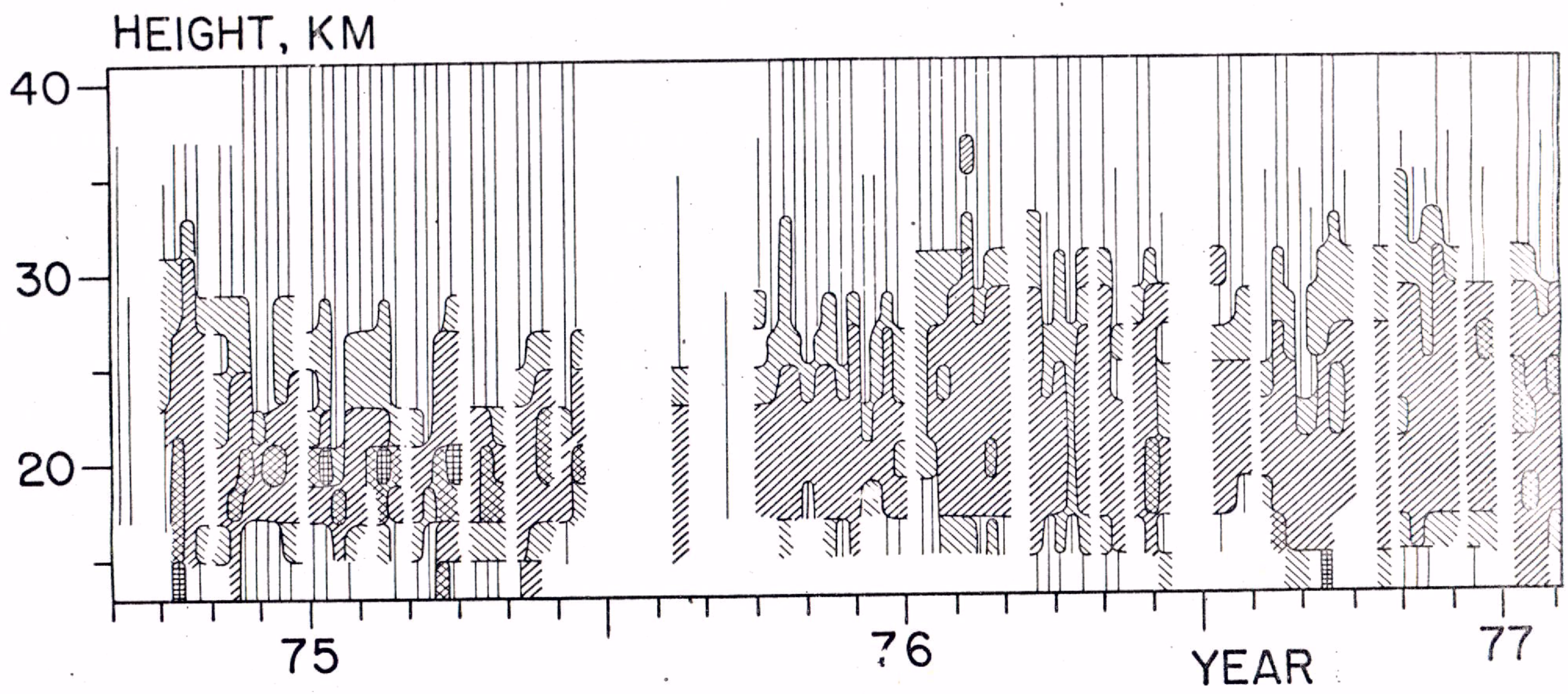
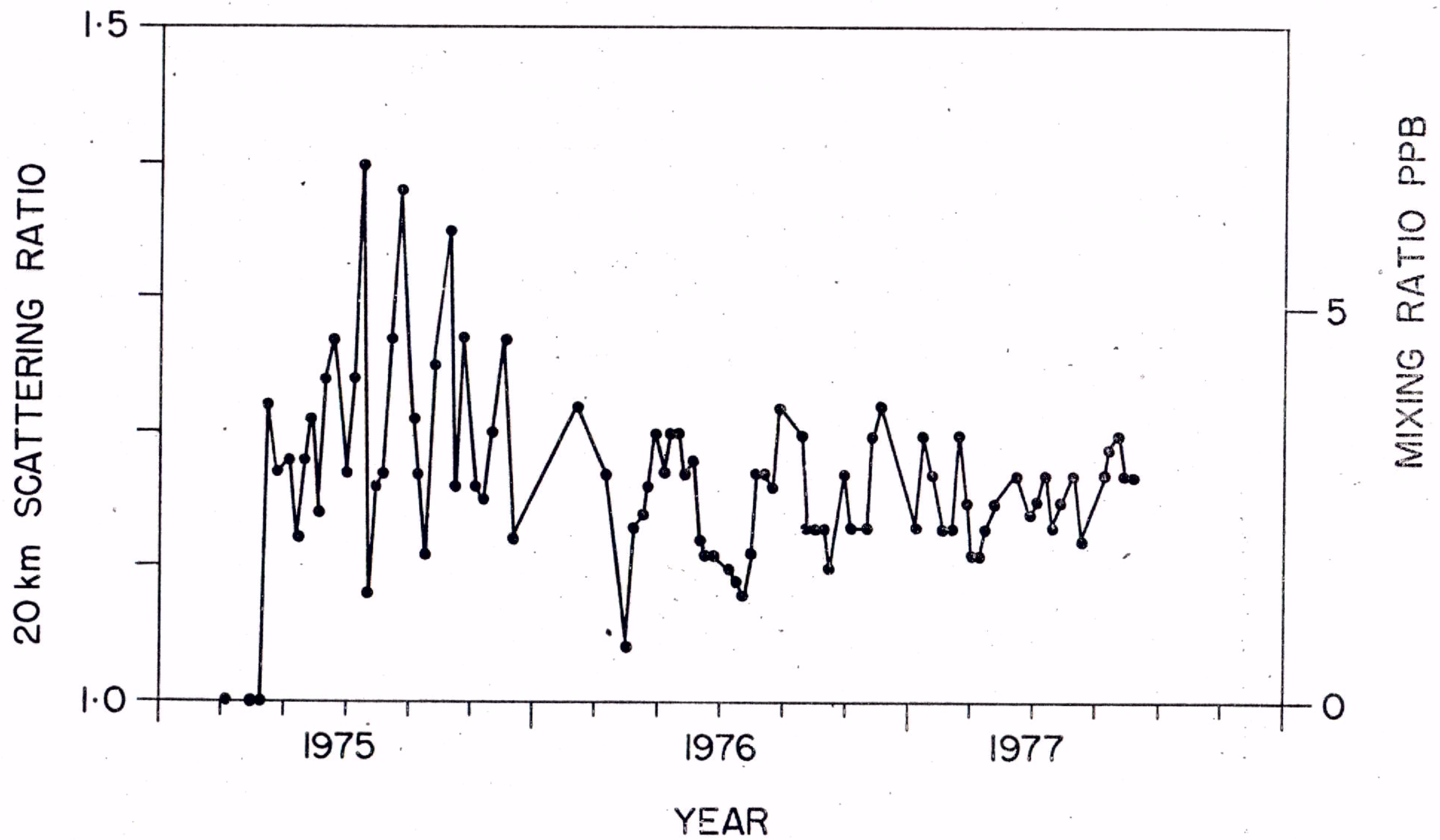

HEIGHT, KM
40
30
20
75
76
77
YEAR

Fig 4


1·5
1·0
20 km SCATTERING RATIO
5
0
MIXING RATIO PPB
1975
1976
1977
YEAR