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A review of laser radar measurements of atmospheric properties

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Abstract—A review is presented of the results of laser-radar measurements made on the atmosphere at altitudes between 10 and 100 km.

The theory of the method is discussed in relation to the scattering properties of the atmosphere. The variations in atmospheric temperature and density which are believed to occur are outlined and the limitations in the present knowledge discussed. A brief account is also given of the results of observations on atmospheric aerosols by methods other than the laser radar.

The results from laser radar studies are examined and critically discussed in relation to atmospheric models. Emphasis is placed upon the evaluation of this method for providing the information about (a) aerosol content, (b) atmospheric density and its variations, and (c) atmospheric composition.

1. INTRODUCTION

THIS review examines the various laser radar methods for making measurements of the parameters of the upper atmosphere, and discusses the latest results. Studies that are primarily concerned with the troposphere have been omitted. However some methods which have only been applied to the very lowest atmosphere are included when it is believed that they may in future be applied to higher altitudes.

The standard laser radar technique has been described by many authors (see for example, BAIN and SANDFORD, 1966). The principle is identical to that of the ordinary microwave radar in that a pulse of energy is sent out, the scattered signal is detected and its energy measured as a function of range. In the case of the laser radar the transmitted signal has usually been a pulse from a Q-spoiled ruby laser. This has a wavelength of 0·6943 μ and the energy, of the order of a few joules, is emitted in the space of a few microseconds. The beamwidth of the laser output is of the order of 20 mrad. and this is reduced to a few milli-radians by passing the light through an optical system. A narrow column of the atmosphere is illuminated by the pulse as shown in Fig. 1(a). The light scattered back from the atmosphere is collected by a telescope. It is then passed through a narrow-band filter to reject unwanted stray light of different wavelengths, and focussed on to a photomultiplier. The signal received from the stratosphere and above is so weak that individual photons are usually counted. The problem of measuring the signal amplitude from a range of heights reduces to one of counting the number of pulses from the photomultiplier in a series of gated intervals.

Assuming that the scattering is primarily Rayleigh scattering from atmospheric molecules, the scattered signal is proportional to the number density of molecules at the scattering height. It can thus be used to measure the atmospheric density, and consequently its pressure and temperature. In practice the atmosphere contains aerosols at various levels and the problem of distinguishing between the types of scattering centre will be examined in the remaining sections of this review, together

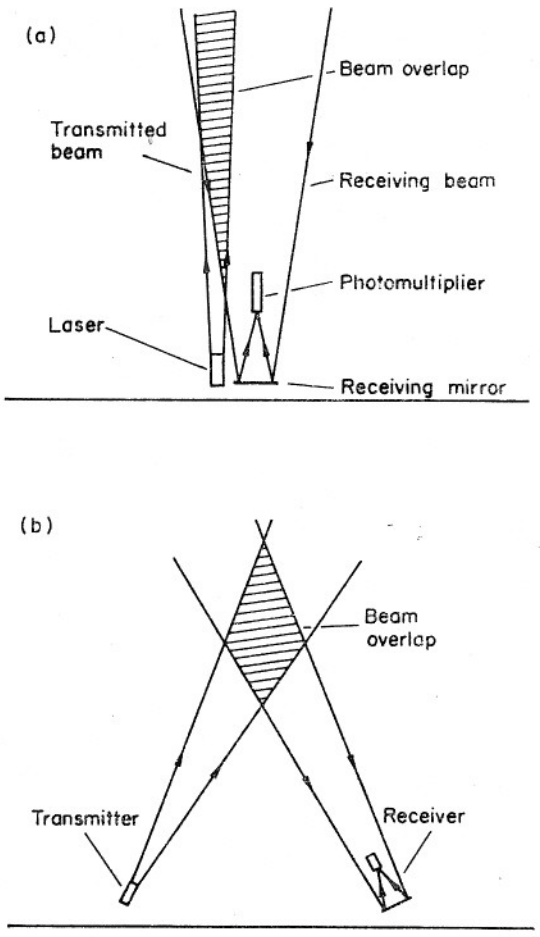

Fig. 1. Laser radar systems. (a) Monostatic, (b) Bistatic.

with the accuracy and reliability of results so far obtained. The experimental technique is less straightforward than has been so far described, in particular, various sources of noise are present, which gave considerable trouble during the early developmental stages of the technique.

An alternative less used technique is the bistatic arrangement shown in Fig. 1(b). The transmitter and receiver are separated and it is no longer necessary to use a pulsed light source. In this method it is difficult to study more than one height range, and there are considerable difficulties in alignment and in knowing the exact extent of the overlap of the beams. The bistatic arrangement has been used (NISHIKORI *et al.*, 1965) in conjunction with a non Q-spoiled ruby laser, which has an output energy which is an order of magnitude greater than that from a Q-spoiled laser. A possible future application of the bistatic arrangement is to the measurement of the doppler shifts caused by atmospheric scattering of signals from a continuous wave laser.

The basic laser radar method as outlined above, examines the scattering of the transmitted pulse by the atmosphere. Most work has so far been done using ruby lasers and examining the returned signals on the basis of Rayleigh scattering from the atmospheric molecules or Mie scattering from aerosols in the atmosphere. Some recent work has been done by selecting transmitted wavelengths to obtain (a)

resonant scattering from specific molecules in the atmosphere; (b) Raman scattering with a wavelength change by specific molecules in the atmosphere, and (c) resonant absorption by certain atmospheric constituents. The experimental technique used to measure the signal scattered from a certain height range is the same as for the earlier method.

2. Laser Radar Theory and Practice

2.1 *Theory*

2.1(a) *The basic scattering equation.* Scattering of the transmitted laser pulse may occur in the atmosphere from atmospheric molecules, aerosols, ions or free electrons. In any of these cases a single formula may be used to describe the received signal. It can be shown that the number of photons received from a height h and detected, when a single pulse is fired from a laser radar system, is given by (Kent, Clemesha and Wright, 1967; Sandford 1967a)

$$C = N_0 A T^2 \eta (\sum n_i \rho_i / h^2)\, \delta h \tag{1}$$

where N_0 is the number of photons in the transmitted pulse,
A is the photon collecting area of the receiver,
T is the atmospheric transmission,
η is the overall receiver counting efficiency,
n_i is the number density of scattering centres of type i at the height h,
ρ_i is the back-scattering coefficient of the scattering centre,
and δh is the height range integrated over.

This expression makes the assumption that the atmospheric properties do not change appreciably over the height range δh and that the transmitted pulse length is short compared with the integration time. It is also only applicable to the case of scattering through 180°, that is to a system with receiver and transmitter adjacent. ρ_i is the fraction of light back-scattered per unit path length, per unit solid angle, per scattering centre, per unit volume.

2.1(b) *Molecular scattering.* Three types of molecular scattering are possible. In two of these, Rayleigh scattering and resonant scattering, the wavelength of the scattered radiation is the same as that of the incident radiation. In the third, Raman scattering, a frequency shift occurs. Although this last case has been the least used it is potentially of considerable importance as it is the only one in which the scattered signal can be clearly distinguished from that caused by other scattering particles in the atmosphere. As most published material is concerned with results obtained by Rayleigh scattering the majority of this and the next section will be on this subject. Some discussion of the other two types will however be given.

Rayleigh scattering of light occurs in a gas when the wavelength of the incident light is remote from that of any absorption line. We may use the theory first developed by Rayleigh (1871) which predicts the back-scattering function ρ_R per molecule to be

$$\rho_R = \frac{4\pi^2(\mu - 1)^2}{n^2 \lambda^4}. \tag{2}$$

In this expression μ is the refractive index, λ is the wavelength used and n is the

molecular number density. The expression may be applied to a single gas or to the atmosphere where μ is the mean refractive index and n the total molecular number density. ρ_R is dependent upon the reciprocal of the fourth power of the wavelength and for a given wavelength may be considered a constant at heights below about 90 km in the atmosphere. Above 90 km dissociation of molecular oxygen occurs and ρ_R will begin to vary. The value of ρ_R in the lower atmosphere at the wavelength of ruby laser light (0·6943 μ) is

$$\rho_R = 1{\cdot}98 \times 10^{-32} \text{ m}^2 \text{ sterad.}^{-1}$$

Resonant scattering may occur when the wavelength of the incident radiation is close to that of an absorption line. A transition to an excited state occurs, followed, if the life-time of this state is short, by subsequent re-emission of the photon. The scattering cross-section will vary across the absorption line and, as experiments so far done have been carried out using radiation with a bandwidth greater than that of the absorption line, it is convenient to discuss an integrated back-scattering function $\int \rho(\nu)\, \mathrm{d}\nu$ where $\rho(\nu)$ is the backscattering function at frequency ν. It can be shown that

$$\int \rho(\nu)\, \mathrm{d}\nu = \frac{e^2}{16\, \pi \varepsilon_0 mc} f \tag{3}$$

where f is the oscillator strength of the transition,
e is the electronic charge,
m is the electronic mass,
ε_0 is the permittivity of free space,
and c is the velocity of light.

The value of $\rho(\nu)$ will depend on the oscillator strength of the transition and the line-shape but will be many orders of magnitude greater than ρ_R.

Raman scattering occurs when an exchange of energy takes place between the incident photon and the atom or molecule concerned. The scattered spectrum contains the original wavelength and other lines which may be shifted by as much as several hundred Angstroms. The Raman lines are always weak compared to the Rayleigh scattered line and the corresponding scattering function is normally lower by a factor of the order of 1000 (LEONARD, 1967).

2.1(c) *Other forms of scattering.* Scattering by atmospheric aerosols is known to be an important contributing factor to the scattered signal received from atmospheric heights below 30 km and may be important for heights above 70 km.

Unlike the situation described in Section 2.1(b) where we are dealing with effectively a single type of scatterer so that n_i and ρ_i are single parameters, we know that aerosols exist over a wide range of sizes, and hence we have a wide range of scattering coefficients. It is well known that aerosols exist in a continuous size distribution, and hence we must replace the $\Sigma n_i \rho_i$ term by the integral $\int \rho(a)\, \mathrm{d}n(a)$ where $n(a)$ is the number density of particles of radius 'a' per unit radius interval and $\rho(a)$ is the back-scattering coefficient for a particle of radius 'a'.

Stratospheric aerosols are known to have radii up to at least 1 μ (see for example the results of CHAGNON and JUNGE, 1961). This means that we are dealing with scattering from particles with dimensions of the same order as the wavelength of the

scattered light, and thus the simple Rayleigh theory does not apply. The theory of scattering by particles in this size domain is generally referred to as Mie theory. Mie theory in fact applies only to spherical particles, and we have no direct evidence to show that atmospheric aerosols, with the exception of liquid droplets in the troposphere, are spherical. However, as a first approximation, it is not unreasonable to use the results of Mie theory to estimate the effects of aerosols, lacking as we do at the present stage any detailed knowledge of the particles concerned.

An exact calculation of the volume back-scattering coefficient requires a knowledge of the particle size distribution. Measurements made in the stratosphere indicate that, over a limited range of particle sizes between about 0·1 and 1·0 μ, the distribution is approximately of the form

$$\frac{dn(a)}{da} = Ca^{-\gamma}$$

where γ is a constant between 2 and 4 (Chagnon and Junge, 1961; Newkirk and Eddy, 1964). Calculations have been made of the scattering coefficient at different values of γ and its variation with wavelength (Nakamura and Clemesha, 1969). Measurements at different wavelengths offer the possibility not only of distinguishing aerosol from molecular scatter but also of determining γ. Different size distributions also give rise to a differing variation of the scattering coefficient with scattering angle and bistatic experiments can be expected to yield useful information on these distributions (Palmer, 1969).

Discussion of aerosol scattering in the mesosphere is extremely difficult from a purely theoretical standpoint. We have very little direct knowledge of the size, shape or composition of mesospheric particles; moreover, the optical properties will depend critically on the presence of an ice coating on the particles which itself is dependent on the mesospheric temperature structure (Witt, 1968).

At heights above 70 km scattering from free electrons becomes a possibility. Thomson scattering theory (Bleaney and Bleaney, 1957) yields a back-scattering function ρ_e for a free electron which has a numerical value of $7{\cdot}94 \times 10^{-30}$ m^2. sterad.$^{-1}$. This is approximately 400 times greater than the corresponding value for an air molecule at a wavelength of 0·6943 μ. At a height of 100 km the total particle density is about 10^{19} m^{-3} whereas the electron concentration is about 10^{11} m^{-3}. The ratio between these two is such that at this height and lower the scattering from free electrons may be ignored in comparison with Rayleigh scattering from the atmosphere.

2.1(d) *Absorption studies.* As light passes through the atmosphere both scattering and absorption will occur. The intensity of the scattered light received from any height will depend on the total transmission losses which occur below that height. If the wavelength transmitted is inside an atmospheric absorption line, serious attenuation will occur and this may be measured by comparison of the scattered signal with that scattered at a neighbouring wavelength just outside the line. The variation of attenuation with height may be found and used to obtain directly the distribution of the atmospheric constituent causing it.

2.2 *Practice*

2.2(a) *Rayleigh scattering from the atmosphere.* In Section 2.1(a) a value was given for the back-scatter function of an atmospheric molecule for light from a ruby laser. It is possible in theory to substitute this value and the values for n into equation (1) and so determine the expected photon count. Alternatively one may, as is the object of a laser radar experiment, substitute ρ_R and C, determined experimentally and so determine n. This procedure makes a number of assumptions. The first is that Rayleigh scattering theory is valid and ρ_R does not vary. As mentioned above, this is likely at all heights below about 90 km; above this height consideration will have to be taken of the dissociation of O_2. A second assumption is that the equipment parameters and T, the atmospheric transmission are well known. This, at the present state of development of the technique, is not true and it is necessary to calibrate the equipment by measuring the signal scattered from an altitude where the atmospheric density is known. The altitudes chosen for the calibration are just above 30 km, on the grounds that the density there is well known, and that the signal returned from atmospheric aerosols is negligible. This last brings us to a third assumption, namely that for much of the atmosphere other forms of scattering, in particular that from aerosols, is small compared with molecular scattering. The experimental belief for this rests mainly on two foundations. The first is that for the height range 30–60 km the variation of signal with altitude agrees with that expected in the absence of aerosol scattering. The second is that in the region 15–30 km where aerosols are known to be concentrated, the excess scattering is easily observable but is not in fact enormous. This third assumption will be incorrect if there is a constant mixing ratio for aerosols in the height range.

The density of the atmosphere varies approximately exponentially with altitude with the consequent result that signal returns from great altitudes are very weak and even large improvements in receiver performance give a relatively small increase in maximum usable altitude. A further consequence of the variation is that δh, the height range over which integration is carried out, cannot usefully be more than half a scale height which is about 3 km in the height range under discussion.

Several laser radar systems are in operation (CLEMESHA, KENT and WRIGHT, 1967; SANDFORD, 1967b; MCCORMICK *et al.*, 1969). These differ in laser output power, firing rate and receiving mirror area. SANDFORD (1967a) has discussed various figures of merit which may be used to describe and compare these systems. We shall not consider these here but it is useful to take typical systems and determine the signal strength and accuracy of measurement to be expected of them. The parameters of two such systems are shown in Table 1(a) together with the product $(N_0 A T^2 \eta\, \delta h)$ which appears in equation (1). One of these is a first generation system such as has been in use for several years now, the other is a second generation system a limited number of which are now under construction and beginning to come into operation. Figure 2 shows the values for Rayleigh scattering from the U.S. standard atmosphere for 45°N for these systems.

In any determination of an atmospheric profile, an integration over a large number of firings is normally carried out. If, for any height range a total of x counts is obtained then the standard deviation of this value is equal to $x^{1/2}$. This means that in order to achieve statistical reliability at high altitudes a very large number

Table 1. Parameters and capabilities of typical laser radar systems

(a) Parameters

		First generation system	Second generation system
Energy per pulse		5 J ($N_0 = 1{\cdot}75 \times 10^{19}$ photons)	20 J ($N_0 = 7 \times 10^{19}$ photons)
Receiving area,	A	0·25 m^2	5 m^2
Atmospheric transmission,	T	0·7	0·7
Height integration interval,	dh	2 km	2 km
Overall efficiency	η	0·01	0·02
Product $N_0 A T^2 \eta$ dh		$4{\cdot}4 \times 10^{19}$ photons. m^3	$7{\cdot}0 \times 10^{21}$ photons. m^3
Firing rate		100 pulses/hr	200 pulses/hr

(b) Detected signal and accuracies for a 1-hr run

Height	First generation system		Second generation system	
	No. of detected photons	Standard error (%)	No. of detected photons	Standard error (%)
40	4500	1·5	$1{\cdot}4 \times 10^6$	1*
50	750	3·5	$2{\cdot}4 \times 10^5$	1*
60	150	9	$4{\cdot}8 \times 10^4$	1*
70	35	18	$1{\cdot}0 \times 10^4$	1*
80	6	40	2000	2
90			250	6
100			30	20

*Accuracy limited by maximum photon counting rates.

of firings will be required. In addition to statistical fluctuations, there will be errors due to the presence of noise in the system. This noise arises from two basic sources, (a) unwanted photons arriving in the receiver due to airglow and other sources in the night-sky, and (b) thermal noise inside the photomultiplier. Spurious signals caused by laser fluorescence or overloading of the photomultiplier, may also occur but these can be eliminated by careful experimental techniques. Estimates of the magnitude of these two sources of noise are also shown for both systems in Fig. 2. The photomultiplier noise is calculated on the assumption that the photomultiplier tube has been cooled so that the number of noise counts per second is 100. The sky-noise has been calculated on the assumption that it is entirely due to airglow, that the bandwidth of the receiving filter is 10 Å, and the beamwidth is 1 mrad. Under these conditions it is clearly seen that below about 80 km noise should not be an important factor. In practice this has not always been so. Early measurements were often made in ignorance of spurious signals such as those listed above and did not approach these rather idealized figures. In addition there may be other sources of radiation in the night-sky, such as the scattering of light from urban areas which can only be removed by a careful choice of site. As mentioned earlier Sandford (1967a) has

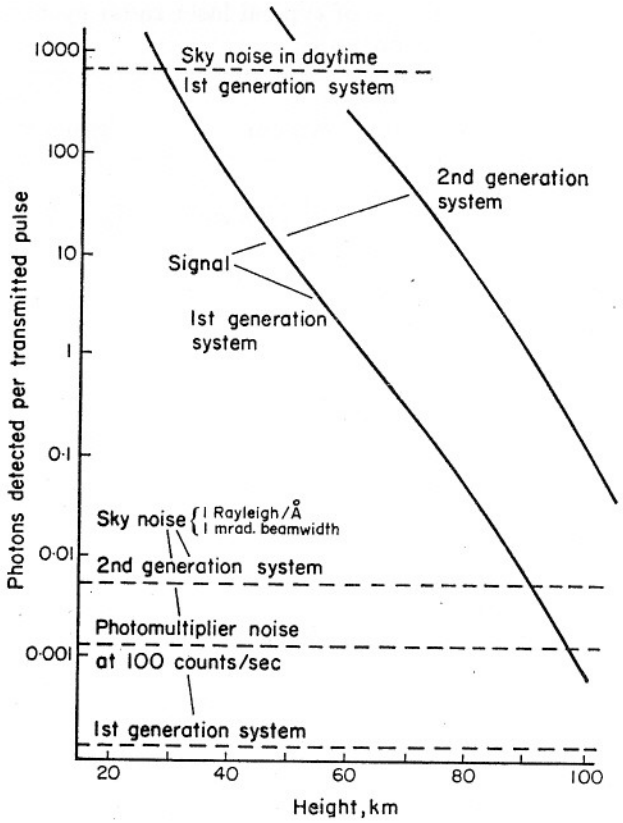

Fig. 2. Detected photons as a function of scattering altitude for typical laser radar systems whose parameters are given in Table 1(a).

discussed the performance of various laser-radar systems; included in his discussion is a consideration of the signal to noise ratio of these systems.

Also shown in Fig. 2 is a figure for the sky-noise during daytime. It can be seen that using the same equipment this becomes a limiting factor at about 20 km. Some improvement can be obtained by use of polarized filters and narrower beamwidths but it is clear that it is impossible to obtain useful echoes from great altitudes during the daytime.

On the assumption that the main limitation to the accuracy of measurements below 80 km is the random nature of the signal it is useful to ascertain the accuracy with which the atmospheric density may be measured at the different heights. This, as outlined earlier, will depend on the total number of firings. Possible firing rates vary from system to system but a typical rate is about 200 per hr. At this rate values for the expected photon count and its standard error are shown in Table 1(b). The accuracy in the first generation system can be seen to vary from an acceptable ± 1 per cent at 40 km to an intolerable ± 30 per cent at 80 km. These values can be improved slightly by the use of higher repetition rates, longer integration times and larger height intervals but nevertheless the accuracy will only improve as the square root of the improvement in performance. The large second generation systems provide some two orders of magnitude improvement. The accuracy attainable with 1 hr's operation on these systems is about 1 per cent at 70 km falling to about 20 per cent at 100 km.

Some preliminary measurements have been made (SANDFORD, 1968) using a frequency doubled ruby laser. Such measurements offer a possibility of separating the scattering due to molecules and to aerosols, as the wavelength dependence differs in the two cases. So far the measurements have not reached any great altitude in spite of the increased molecular scattering at the shorter wavelength, as this is more than compensated for by the decrease in laser output energy.

2.2(b) *Other forms of scattering and absorption.* It is only very recently that either resonant or Raman scattering has been used to investigate the atmosphere and these methods, although clearly possessing considerable future potential, are still very limited.

Resonant scattering has been used at a wavelength of 0·5896 μ to measure atmospheric sodium (BOWMAN, GIBSON and SANDFORD, 1969). Sodium occurs in the atmosphere between 80 and 110 km and has strong resonance at 0·5890 and 0·5896 μ. These wavelengths may be attained using tunable dye lasers and it is found that the effective back-scatter function for a sodium atom is of the order of 10^{14} greater than that for an atmospheric molecule. Thus, even though the concentration of sodium atoms is very much less than the total molecular concentration at these heights, a detectable scattered signal may be found. The method is applicable to other minor constituents in the atmosphere possessing strong resonance lines and now that tunable lasers are available one may hope to see a considerable extension of this technique.

Raman scattering has been applied to the measurement of the concentration of N_2 in the lower atmosphere and atttempts are also being made to use it to measure the concentration of H_2O and O_2 (COONEY, 1968, 1969; LEONARD, 1967). Using a ruby laser at 0·6943 μ as transmitter a Raman shifted spectral line from N_2 is detectable at a wavelength of 0·8285 μ. As indicated earlier, the strength of this line is very weak compared to the Rayleigh scattered line and signals have not yet been detected from altitudes above 10 km. As with resonant scattering, this method has considerable future potential, enabling, as it does, an atmospheric profile to be obtained uncontaminated by scattering from aerosols.

Resonant absorption has so far only been applied to the measurement of water vapour in the lower atmosphere (SCHOTLAND, 1965). A water vapour line at 0·6943 μ lies very close to the ruby laser output wavelength and the latter may be thermally tuned to coincide with it. Comparison of the signal attenuation at 0·6943 μ and at a neighbouring wavelength enables one in theory to determine the water vapour concentration. Practical difficulties arise with the precise determination of the absorption line-shape and with the control and measurement of the laser output wavelength. So far measurements have only been made in the lowest few kilometres of the atmosphere but the method is capable of extension to greater heights and to other atmospheric species such as carbon dioxide, ozone, etc.

Measurements of the scattering from stratospheric aerosols have been made on the assumption that the return from the molecular atmosphere is calculable and may be subtracted from the total returned signal leaving the component due to aerosols alone. (FIOCCO and GRAMS, 1964; CLEMESHA *et al.*, 1966). As described in Section 2.2(a) a fit is made to a standard atmosphere at a height between 30 and 40 km and the signal expected from the molecular atmosphere at lower heights determined by

extrapolation. The value obtained in this way for the back-scattering function of the aerosols may be correct, or too low, but cannot be too high. Signal strengths received from stratospheric heights are normally considerable and integral rather than digital recording techniques may be used. Considerable accuracy is potentially possible and temporal and spatial variations can be determined. Bistatic methods are also being used to measure the variation of the scattering function with scattering angle and with polarisation. Comparison of this with the variations expected for different particle size distribution may enable a choice to be made between the possible alternatives.

Measurements of the scattering from aerosols in the mesosphere have been attempted by several workers. The signals are very weak and subject to contamination by weak sources of extraneous noise. The results are still somewhat controversial

2.3 *Summary*

A summary of the properties of the various methods described in the previous two sections is given in Table 2.

Table 2. Laser techniques for studying the atmosphere

Technique	Laser and wavelength used	Atmospheric constituent examined	Comments
Rayleigh scattering	Ruby 0·6943 μ and 0·3472 μ	Molecules	Original method and still the most common. Assumes no other form of scattering present. Range of applicability still doubtful but potentially capable of measuring atmospheric density variations.
Aerosol scattering	—do—	Aerosols	Useful method of studying the behaviour of stratospheric aerosols. Capable of considerable development. Application to mesosphere under debate.
Resonant scattering	Tunable dye laser (0·5896 μ)	Na	So far only applied to sodium at heights of 80–100 km. Capable of extension to other minor constituents using other wavelengths.
Raman scattering	Ruby (0·6943 μ) (return at 0·8583 μ)	N_2	So far only applied to molecular nitrogen and oxygen. Signals returned very weak, height range less than 10 km. Avoids confusion with other forms of scattering.
	Pulsed nitrogen (0·3371 μ) return at 0·3557 μ for O_2 and 0·3658 μ for N_2	O_2, N_2	
Resonant absorption	Tuned Ruby (0·6943–0·6944 μ)	H_2O	So far only applied to water vapour, height range only a few kilometres. Capable of extension to other constituents.

3. Atmospheric Properties

In order to provide a background for the discussion of laser measurements of atmospheric density and temperature, we give here a brief outline of the generally accepted information about the atmospheric parameters in the height ranges below 100 km.

3.1 *Density and temperature*

The average patterns of behaviour of the values of density, pressure and temperature at various heights are summarised in the U.S. Standard Atmosphere (1962). Supplements (1966) to this provide a survey of the variations at different latitudes and also show the seasonal changes. It is necessary to stress that these 'standard' atmospheres are only guide-lines to the probable average behaviours.

Below 30 km we have a reasonably comprehensive picture of the variations from the standard atmosphere from balloon-carried radiosondes. In the troposphere large fluctuations of temperature occur associated with weather formations. Above the troposphere which is the region of more interest of this review, the mean day-to-day variations in temperature are only 2° or 3° at low latitudes in all seasons. At higher latitudes in Summer the fluctuations are 3° or 4°C; in Winter the situation is complicated by stratospheric warmings which give two separate and distinct regimes and so lead to considerably larger fluctuations. Changes of density below 30 km are only a few per cent of the average values.

Above 30 km our knowledge of the day-to-day fluctuations and the diurnal variations in temperature and density is still very sparse. The majority of the changes involved are probably due to atmospheric motions caused by the thermal and gravitational tides produced by the Sun and the Moon, together with the effects of internal gravity and planetary waves which are propagated upwards in the atmosphere. The diurnal changes will be directly associated with the main tidal system of the solar radiation, and affected by the natural periods of the atmosphere. The distribution of the energy input over the height range will have a major effect upon the different harmonics which are initiated. Studies by Lindzen (1967, 1968) and Butler and Small (1963) provide theoretical estimates of the likely diurnal and semidiurnal tides in the atmosphere in the height range with which we are concerned. However the experimental values above 30 km are so sparse that it is difficult to draw adequate conclusions about what diurnal variations occur. Considerable doubt has recently been thrown (Theon *et al.*, 1967) on many diurnal range and day-to-day measurements of temperature in this height range, which have been made using thermistor techniques. Their own results indicate a considerable temperature oscillation with height which is in qualitative agreement with Lindzen's (1967) diurnal tidal picture. The amplitude of the oscillation increases with height. The temperature wave has a range of 40°C with a vertical wavelength of about 15 km at a height of about 70 km. Fluctuations from day to day will be about the same order of magnitude.

Above 30 km the information regarding diurnal variations of density is also quite inadequate. Observations indicate a diurnal range of 10–25 per cent at a height of 60 km with the maximum density in the afternoon. The measurements are difficult

and limited in number. Wavelike variations of density are also observed in the 60–70 km region with a range of about 10 per cent from normal (THEON *et al.*, 1967).

3.2 *Gaseous composition*

One important issue which relates to the scattering process is the composition of the atmosphere and its variation with height. It is generally accepted that up to about 90 km the effects of diffusive separation can be neglected. Our experimental information concerning the variation with height of the concentration of some of the minor constituents such as water vapour and carbon dioxide above 10 km is

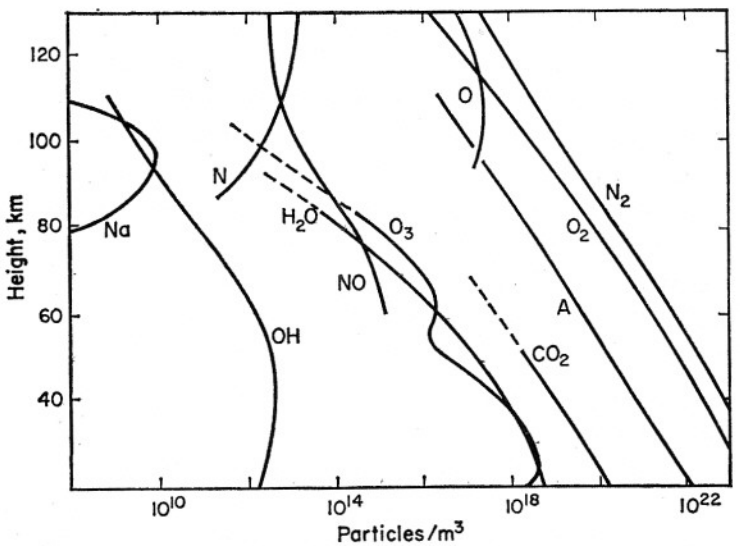

Fig. 3. Atmospheric composition between 20 and 120 km. altitude Sources—OH, HUNT (1966); N, NORTON (1967); O, CHAMPION (1968); CO_2, A, Handbook of Geophysics (1960); Na, HUNTEN and WALLACE (1967), DONAHUE and MEIER (1967); O_3, MAEDA (1968), REED (1968), NAGATA *et al.* (1968); NO, BARTH (1964), PEARCE (1968); H_2O SISSENWINE *et al.* (1968).

surprisingly poor. An ozone layer exists in the region above 20 km. The number density of ozone molecules at its maximum is of the order of 5×10^{18} m^{-3}. As this is only one ozone molecule to about 10^5 oxygen and nitrogen molecules, this means that ozone will not play a significant part in any non-resonant scattering experiments.

From 80 km upwards, the dissociation of molecular oxygen into atomic oxygen becomes an important process. The fraction dissociated increases rapidly upwards, becoming about 1 per cent at 100 km. Also from around 90 km the effects of diffusive separation appear. When reliable scattering measurements are obtained from above 90 km, it is clear that precautions must be taken in the interpretation of the data, and allowance made for any changes in the composition of the major scattering bodies.

Figure 3 shows the concentrations of some of the minor constituents in the 30–100 km region which have been suggested in recent publications. With the advent of selective methods using resonant scattering from a particular constituent, some of these minor constituents may be studied by laser radar. The power of these methods is underlined by the fact that sodium, one of the least of the minor constituents, is the first to be detected by this method.

3.3 *Aerosols—below 30 km*

3.3(a) *Vehicle-borne sample collection.* The most extensive measurements of this type have been made by JUNGE (1963), JUNGE and MANSON (1961) and CHAGNON and JUNGE (1961). These workers used both balloons and aircraft to carry sample collectors to heights up to 30 km. Above 10 km their results show that the concentration of particles varies as r^{-4} for particles with radii between 0·05 and 1 μ. There is a sharp cut-off at 1 μ and above this radius the concentration varies approxi-

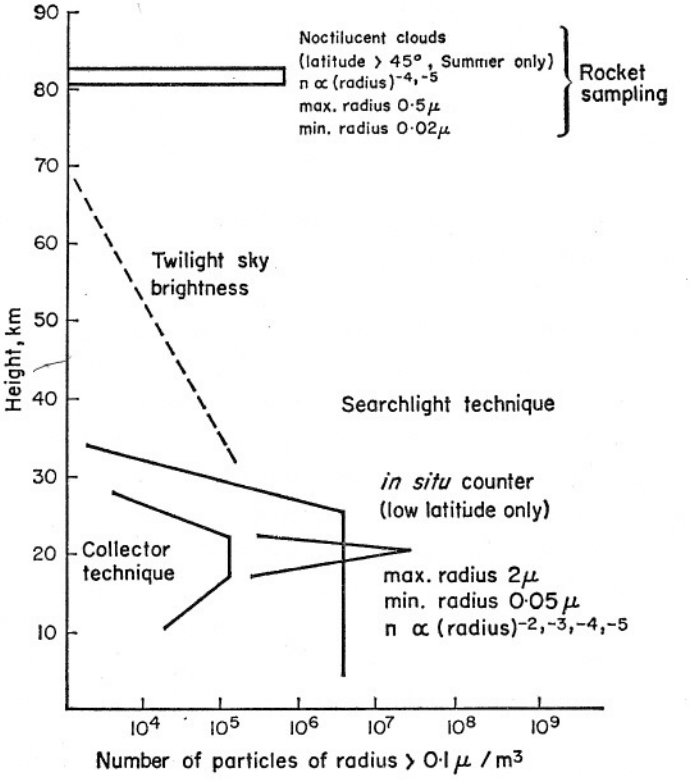

Fig. 4. Summary of aerosol properties, 10–90 km.

mately as the radius to the power of −7. Junge and Manson report no significant latitude variation of the 20 km aerosol.

The height variation of particle concentration, showing the well-known 20 km aerosol layer, is shown in Fig. 4 together with other estimates of aerosol concentration which are discussed below.

FRIEND (1966) has also collected stratospheric aerosols using an impact collector in a high flying aircraft. Whereas Junge and Manson reported a lower limit of radius below 0·1 μ, Friend found what amounts to a sharp cut-off at about 0·3 μ. The importance of this from the point of view of laser probing is that from the scattering theory for a wavelength of 0·7 μ as much as 50 per cent of the scattered light would have come from the particles smaller than 0·3 μ, using Junge and Manson's result. MOSSOP (1965) has reported results of a similar nature to Friend, indicating fewer small particles but also indicating a rather larger concentration of the bigger particles with radii greater than 0·5 μ.

All the collection measurements of stratospheric aerosols indicate that sulphur is the most prominent element present. The sulphur is probably in the form of a sulphate radical. There are indications that the positive ion is NH_4. It is suggested that these ions are formed in situ by photochemical reactions, possibly involving ozone. MOSSOP (1965) has found that almost every particle (sulphate aggregate) collected contains at least one foreign particle which he suggests assists nucleation of the ammonium sulphate from the gaseous reaction from which it is formed. These nuclei are mainly electron-dense materials but the investigations were inadequate to decide whether the particles were extra terrestrial or not.

3.3(b) *In-situ measurements.* ROSEN (1964) has measured aerosol concentrations in the stratosphere using a balloon-borne photoelectric counter. The measurements give a concentration of aerosols greater by an order of magnitude than those of Chagnon and Manson. This may indicate considerable condensation which would not be detected in the collector type experiment. Such condensation must be taken into account when calculating the scattering properties of the atmosphere.

ROSEN (1968) has also taken measurements at three different latitudes, 10°N, 45°N and 60°N, which show marked differences in the shape of the aerosol profile and in the maximum concentration. This is in direct contrast with Junge and Manson, who observed no appreciable latitude variation. At 10°N Rosen found the stratospheric aerosol to be almost entirely confined to a rather narrow layer between 16 and 22 km, with a peak concentration at 20 km of 3×10^7 particles m^3/having radii greater than 0·125 μ. At higher latitudes the maximum concentrations were much smaller and the sharp 20 km maximum was absent. Rosen also found a strong positive correlation with the ozone concentration, which he measured simultaneously.

3.3(c) *Remote measurements.* The two main techniques which have been used are the searchlight, notably by ELTERMAN (1966), and twilight sky brightness measurements by VOLZ and GOODY (1962). The laser radar method is a development of the searchlight technique, and the results from this method will be discussed in Section 4.

Elterman quotes his results in terms of an attenuation coefficient. This parameter, which will be proportional to aerosol concentration provided the distribution of sizes does not change, remains at around the same order of magnitude from 5 km up to 20 km, but falls rapidly above 20 km. Assuming a distribution of type reported by Manson, we may estimate that about 10^6 particles/m^3 of radius greater than 0·25 μ are needed to explain Elterman's results. This is also shown in Fig. 4 where it can be seen that the values fall between those of Junge and Manson and those of Rosen. It is likely that long term variations occur in the stratospheric dust content but the magnitude and cause of these is still under debate. Local increases certainly occur after volcanic eruptions but their geographical extent is still uncertain (MATSUSHIMA, 1968; ROSEN, 1968).

The twilight measurements of Volz and Goody extend from a few kilometres up to 70 km. Their results compared to those of Elterman suggest smaller concentrations of dust in the stratosphere by about a factor of 4. Furthermore, the twilight observations do not give any marked evidence for a sharp layer structure, and suggest that appreciable dust continues to be present above 20 km.

3.4 *Aerosols—above 30 km*

The twilight sky brightness measurements of Volz and Goody suggest that dust is present at all heights up to 70 km. The scattering is such that the order of magnitude of the aerosol extinction coefficient is 0·2 of the molecular extinction for red light from 30 km up to 65 km. Similar values for aerosol content are also suggested by Divari (1964) on the basis of photometric measurements of brightness of the twilight sky. More recently, Volz (1969) has indicated that this evidence is tentative and that too much reliance cannot be placed upon it.

The presence of dust just above the mesopause is a necessary consequence of the influx of meteorites, and it is possible to establish a minimum flux of particulate matter through this region on the basis of meteor observations. Noctilucent clouds observed at high latitude in summer also indicate the presence of aerosols at heights near to the mesopause.

Direct measurements in high latitudes have been made by rocket sampling, by Soberman and Hemenway (1965). These measurements suggest a surprisingly high content of aerosols in the absence of noctilucent clouds. Other measurements by Carr and Gabe (1967) are orders of magnitude smaller and further observations are urgently needed.

Measurements made in the presence of noctilucent clouds show much higher particle concentrations. Particle concentrations in a vertical column through the cloud are of the order of 10^{11} m^{-2}. The particles are mainly of iron, nickel and silicon, and some show evidence of having been ice coated. The distribution of particle radii follows an inverse 4th or inverse 5th power law with a sharp cut-off at a lower limit of 0·02 μ radius. The height of noctilucent clouds is found to be 83 $\pm$ 4 km, corresponding closely to the height of the mesopause (Hemenway *et al.*, 1964).

It is now generally accepted (Paton, 1967) that noctilucent clouds occur at high latitudes in Summer when particularly low mesopause temperatures occur associated with a poleward meridional flow from the equatorial stratopause which diverges upwards at the noctilucent cloud latitudes. An increased number of particles arises in a narrow height range at the mesopause and the effective light scattering properties are increased by the crystallisation of ice onto the particles. Recently (Poultney and Silverberg, 1969) it has been suggested that dust associated with comets which orbit inside the earth may be present at certain times of the year in the upper atmosphere. This dust which is not associated with visible meteors, enters the Earth's atmosphere, breaks into particles about 1 μ in radius, and settles slowly. On the nights immediately following its entry it may be sufficient to produce an observable excess optical scattering in the region 60–80 km.

3.5 *Conclusions regarding aerosols*

We have tried to summarise our information about aerosol concentrations in Fig. 4. Below 30 km there is a growing body of data which contains some conflicting results. Comparison of results requires assumptions regarding size distributions, which can introduce major uncertainties. Nevertheless the general distribution of aerosol concentrations is beginning to emerge.

Above 30 km the information is quite inadequate. We do not know whether aerosols exist in appreciable quantities or not, except in the case of noctilucent

clouds which occur occasionally at high latitudes in summer at the height of the mesopause.

4. Results

4.1 *Above 30 km*

4.1(a) *Atmospheric density.* In this section we shall examine what information has been obtained about the atmosphere and its density using the laser radar method. Below 30 km the prevalence of aerosols provides additional scattering which masks the background of scattering from the atmospheric molecules. Above 30 km there must be a contribution to the scattering from aerosols, but, as is mentioned in Section 3 above, there is inadequate evidence to decide whether the contribution will be significant. It is worth noting that a small amount of aerosol content, particularly if the mixing ratio for aerosols is constant with height, could be overlooked by the laser radar measurements and would affect the strength of the signal. In this section we shall assume that the aerosol contribution to the returned signal is negligible above 30 km, but it is important that this question be more closely examined in the future.

In order to make absolute measurements of the molecular number density it is necessary to know the efficiency of the system accurately. This is as yet poorly known in all the systems in operation and changes occur, both over short and long periods. Consequently no absolute measurements have been obtained.

It is necessary therefore to use comparative measurements. The observed signals from heights where the density is known at around 30 km provide a reference point which allows calibration of the system and so determines the system efficiency. It is assumed that the aerosol contribution at this height is negligible. Calibration at 30 km height is more desirable than at 40 or 50 km because meteorological balloons can provide density values quite regularly up to this height. Measurements from most laser systems used to study the mesosphere have been restricted to heights above about 45 km. Consequently, the results are studied by comparison to an appropriate standard atmosphere. A calibration constant is selected which provides the best fit over the lowest 10 km of reliable measurements. The latest results using the laser radar method in the height range above 30 km which were available at the time of writing this review are shown in Fig. 5 from Maryland, U.S.A. (Silverberg and Poultney, 1967), from Winkfield, England (Sandford, 1968) and from Kingston, Jamaica.

Some discussion has already been given to the accuracy of the photon count in each height range, (see Section 2 above). The accuracies claimed are shown in Fig. 5. The results for Winkfield, England, indicate accuracies of better than 3 per cent up to 70 km and the data from Kingston, Jamaica, indicates the same accuracy up to 90 km or more. The Maryland data suggests about 3 per cent accuracy up to about 50 km. These accuracies are achieved in different ways. Winkfield have grouped together a much larger number of shots over a period of days than that used by Maryland. The Kingston system is very much larger than either, particularly in the receiving mirror system, and the accuracy shown in Fig. 5 for this station was obtained with only 300 shots in about 2 hr.

The results obtained for Winkfield, England, by Sandford, have provided the

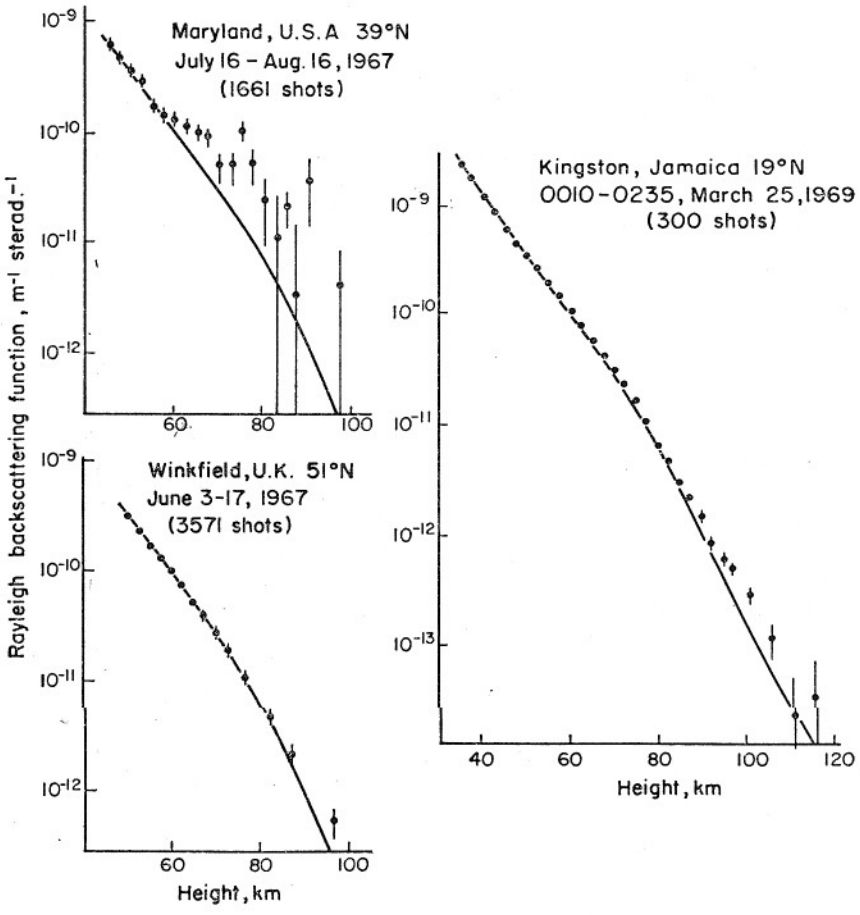

Fig. 5. Recent measurements of the Rayleigh backscatter function at altitudes between 30 and 120 km. ● experimental results; — fitted standard atmospheres.

best and most systematic collection of atmospheric density profiles. Unfortunately the lowest height range included is 50 km, which is rather high for comparison to regularly available density measurements. Nevertheless, Sandford does claim to be able to detect the expected seasonal changes of the density profile with height over the range 50–70 km. Figure 6 shows Sandford's data for two periods, July 16–August 5, 1967 and December 6–8, 1967. The method of presenting the results is rather unusual and must be studied carefully if it is not to be misleading. The data are shown as ratios to the Cospar International Reference Atmosphere 1965, with the calibration constant chosen for a best fit to the relevant monthly profile which is plotted in the same way.

4.1(b) *Atmospheric temperature.* Apart from the density and the density profile, it is also possible to determine the scale height and the atmospheric temperature from the scattering profile. In this application the absolute calibration is not so important. From the basic equations we may express the temperature T_z at a height z km as

$$T_z = \frac{\rho_1 T_1 + \dfrac{Mg}{R}\displaystyle\int_{z_1}^{z} \rho \, \mathrm{d}z}{\rho_z}$$

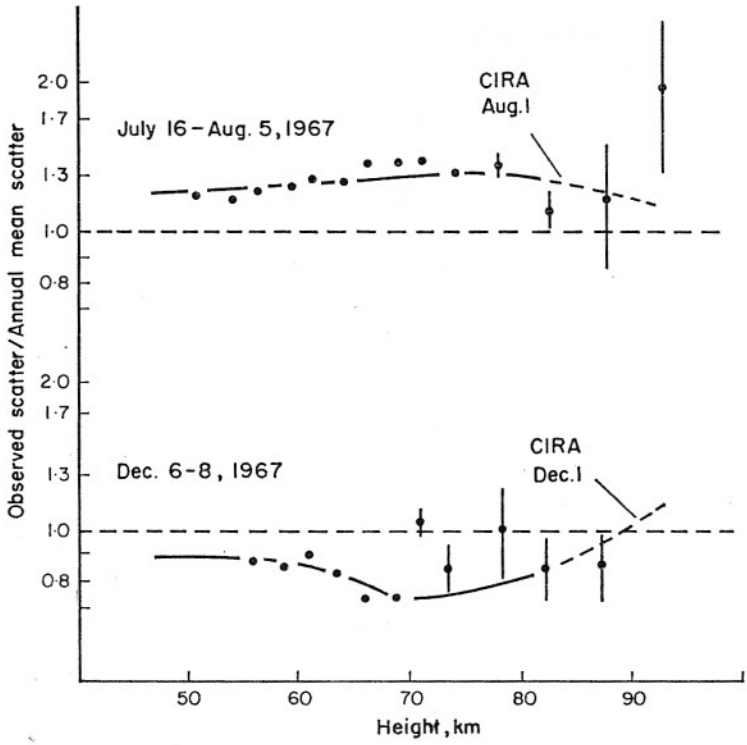

Fig. 6. Seasonal change in scattered signal observed at Winkfield, England.

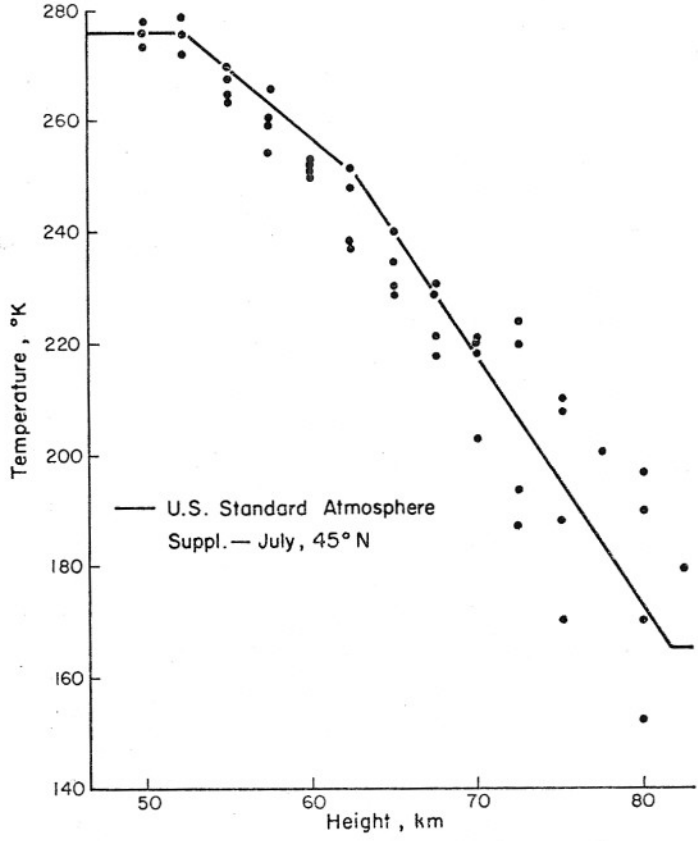

Fig. 7. Atmospheric temperatures deduced from laser radar measurements taken at Winkfield, England in Summer, 1967.

where ρ_1 and T_1 are the density and temperature at a height z_1 km at the top of the profile. T_1 is estimated approximately, and values of T_z are calculated step by step downwards from the height z_1. The value of T_1 chosen becomes rapidly of little consequence. The large amount of data collected at Winkfield has enabled Sandford to calculate a series of atmospheric temperatures which are shown in Fig. 7. The agreement with the accepted standard atmosphere temperatures is quite satisfactory.

4.1(c) *Deviations from the standard atmosphere.* The accuracy obtainable from the Kingston, Jamaica system together with its wide height-range of data collection makes it possible to examine the profiles of density with height for the possible existence of 'waves' or tidal effects.

Figure 8(a) shows the latest Kingston, Jamaica, profile presented as a ratio to the standard 15°N atmosphere from the U.S. Supplementary Atmosphere 1965. The data has been adjusted for a best fit in the 35–45 km height range. It will be seen that there is a definite pattern to the deviation from the standard atmosphere. In particular there is evidence of a wavelike structure in the region 70–90 km. In order to examine the reliability of this pattern the data was divided into two halves which are shown in the middle part (b) of the Fig. 8. Essentially the same structure is

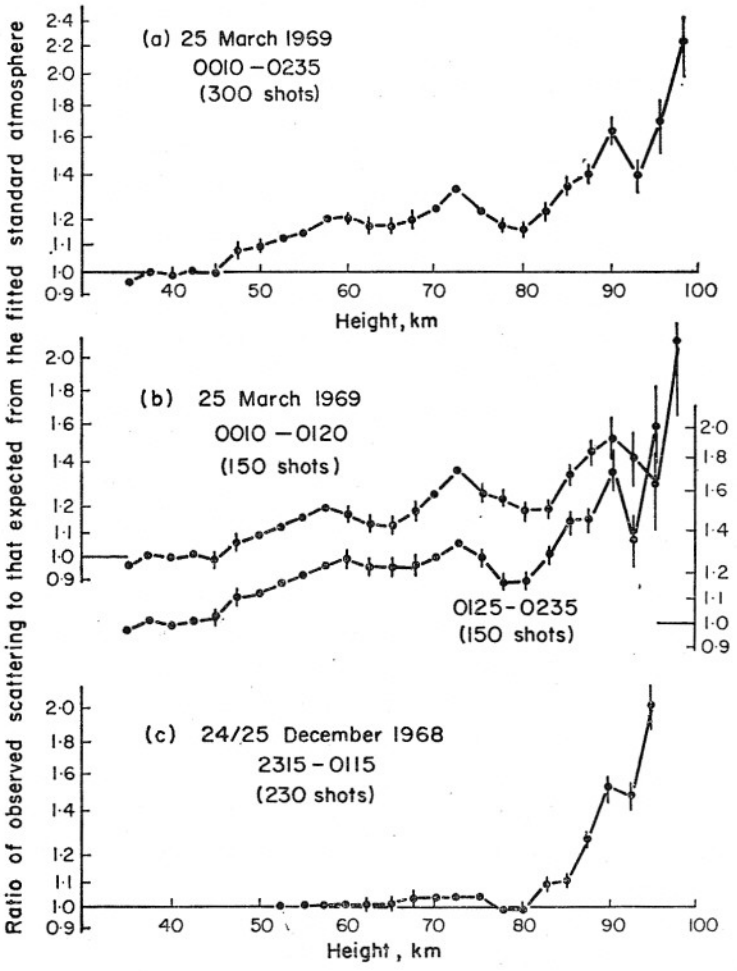

Fig. 8. Scattering profiles observed at Kingston, Jamaica.

present in both halves. No systematic changes can be seen, and it is clear that observations over a longer period are required if motion of the 'wave' or 'tide' is to be detected. Figure 8(c) shows a profile taken on a different occasion when the deviations from the standard atmosphere are very much less.

4.1(d) *High altitude aerosols.* The problem of whether it is possible to observe scattering of a laser beam by a high altitude aerosol layer at heights above about 60 km is one that has caused considerable controversy during the past five years.

This controversy has had two basic causes:

(1) That any excess photon count, over that expected on a Rayleigh scattering basis will, in the absence of obvious other causes, be attributed to scattering from aerosols.

(2) That the earliest laser observations very clearly had spurious sources of noise which were then unknown, and which only became obvious later. Successive experiments have succeeded in reducing or eliminating these spurious sources of noise with consequent reduction in the strength of the apparent 'aerosol scattering layers'.

As better and better profiles have been obtained so have the claims for the intensity of the scattering layers become smaller. Amongst the most recent results, BAIN and SANDFORD (1966) claim a 150 per cent increase in scattering around 71 km in Winter 1965. SANDFORD (1967b) measured an enhancement of 46 per cent of the light scattered from around 71 km in December 1966. A similar enhancement is shown in the December curve of Fig. 6 of about 30 per cent at the same height. Results from Maryland (SILVERBERG and POULTNEY, 1967) have on certain occasions also shown similar enhancements. Recently POULTNEY and SILVERBERG (1969) have been able to show that the observed enhancements are genuine by beaming their equipment off zenith and observing a corresponding increase in the range of the echoes. Moreover they suggest that the enhancements and those observed by Sandford occurred when a considerable cometary dust influx was expected, in agreement with their theory discussed earlier in Section 3.5. On the other hand an examination of the results from Kingston, Jamaica, does not show any enhanced scattering from similar heights which are suggestive of layer structure. The gradually increasing departure from the standard atmosphere with altitude shown in Fig. 8 could be due either to a general background of aerosols or to a small temperature departure from the standard.

The difference between these results may be due to the fact that the observations were made at different times or it may suggest a variation of the mesospheric aerosol content with latitude. The latter would not be inconsistent with the occurrence of noctilucent clouds which are also only seen at high latitudes. FIOCCO and GRAMS (1966) have looked for the aerosol layer using a laser radar in Sweden at times of noctilucent clouds and measured a seven times enhancement of the scattered signal from heights around 70 km. This height is however, 10 km below that at which noctilucent clouds are usually observed but is consistent with the results of Sandford and Poultney and Silverberg.

It is clear from the above that the experimental situation is at present unsatisfactory. According to DEIRMENDJIAN (1965) the volume back-scatter coefficient for visible noctilucent clouds should be of the same order as the molecular return from

about 50 km, allowing for the inverse square law. This means that echoes from noctilucent clouds should easily be detected. According to HEMENWAY *et al.* (1964) the particle concentration in the presence of noctilucent clouds is of the order of 10^3 times the concentration at high latitudes in their absence. In this case we should expect a scattering coefficient of the order of 10^{-11}–10^{-12} m^{-1} sterad.$^{-1}$ when no noctilucent clouds are present. This corresponds to the molecular coefficient at a height of about 90 km, which is still close to the limit of the systems presently operating.

4.2 *Measurements below 30 km*

4.2(a) *Raman studies on molecular nitrogen.* In Section 2 brief discussion was given of the use of Raman scattering to detect the concentration of molecular nitrogen and oxygen. This technique, although of considerable potential, is still confined to the lowest few kilometres of the atmosphere and no further discussion of the results will be given in this review.

4.2(b) *Scattering from the stratospheric aerosol layer.* The laser radar provides a convenient tool for the examination of the scattering and other properties of the stratospheric aerosol layer, and it is surprising that so little use appears to have been made of it for this purpose. The first observations of this kind were reported by FIOCCO and GRAMS (1964), working at M.I.T. These workers used a laser radar at a wavelength of 0·6943 μ and observed scattering over the height range 14–25 km. By assuming that the scattering at 25 km is entirely due to atmospheric molecules they calculated the ratio of total to molecular scattering as a function of height. The maximum value of this ratio was found to be 2·5 at a height of 19 km.

Other groups of workers (CLEMESHA *et al.*, 1966; GOYER and WATSON, 1968) have since reported observations of the aerosol layer, giving results very similar to those of Fiocco and Grams. Figure 9(a) shows three profiles obtained by these different groups of workers; the similarity between the observations is striking when

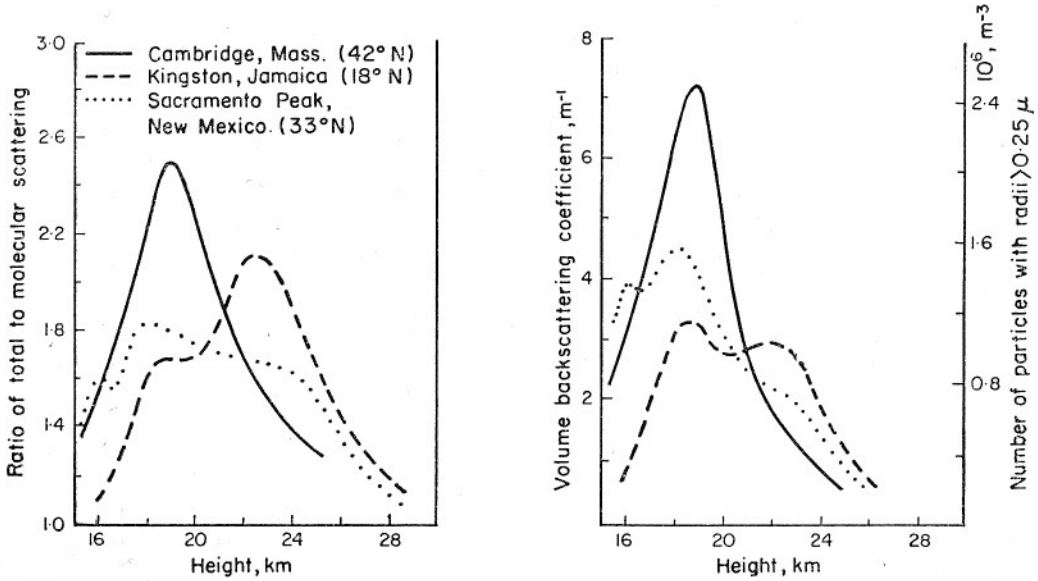

Fig. 9. (a) Scattering profiles for the stratospheric aerosol layer (b) Variation of the back-scattering coefficient and aerosol density with height.

one considers that they have been taken in different latitudes and at different times. Figure 9(b) shows the same data replotted in terms of a back-scatter coefficient rather than as a ratio to molecular scattering.

Calculation has been made of the aerosol concentrations necessary to explain the back-scatter coefficients shown in Fig. 9(b). For this purpose a refractive index of 1·33 has been assumed and an a^{-4} distribution of radii. It is found that the laser radar results require a concentration of between 10^6 and 3×10^6 particles/m^3 the size of the particles being greater than 0·25 μ. These values are much higher than those found by JUNGE and MANSON (1961). Somewhat better agreement can be made by considering an a^{-3} size distribution but a considerable discrepancy still remains. It appears that the laser results can only be explained in terms of a higher concentration of aerosols than has been found by Junge and Manson, or in terms of a different size distribution. The former hypothesis agrees with the results of ROSEN (1964). Alternatively GOYER and WATSON (1968) have shown that the results of MOSSOP (1965), in which a different form of distribution was found, can also give good agreement with the laser results.

GRAMS and FIOCCO (1967) have compared the data from a long series of measurements with various meteorological parameters and find very little correlation. When they use their data to obtain an average profile of the ratio of total to aerosol scattering they find a maximum of 1·9 at about 17 km. The r.m.s. deviation of this ratio is only about 0·3, indicating that the layer is remarkably constant. When autocorrelation coefficients were calculated for various time lags the correlation was found to drop to zero in about 5 days. No marked seasonal variation was observed. Attempts to correlate the aerosol return with wind velocity and direction and with the height of the tropopause all gave negative results. Similar attempts, although with less data, have been made by us at Kingston, also with negative results.

The M.I.T. workers also compared the day-to-day variation in the strength of their aerosol echoes with the day-to-day variation in ozone concentration and found a negative correlation. They found a maximum negative correlation coefficient of −0·6 at a height of 16 km. At first sight this result might seem to contradict that of ROSEN (1968) who observed a positive correlation between ozone and aerosol concentration. However, Rosen's positive correlation refers to the height distribution, not to the time dependence, and thus the two conclusions are not necessarily mutually contradictory.

KENT *et al.* (1967) have made observations of scattering from heights up to 30 km during daylight. This is made possible by the very strong echoes obtained from this region. It is clear then that 24 hr observations of the aerosol layer are possible under suitable weather conditions.

Recently a comparison has been made of the results of laser radar measurements of the aerosol layer and of downward directed infra-red irradiance from the same region, also presumed to come from aerosols (PILIPOWSKYJ *et al.*, 1968). Considerable agreement was obtained between the vertical profiles obtained by the two techniques. Difficulty was, however, encountered in fitting the results to a suitable aerosol size distribution and it was found necessary to postulate a large population of small aerosols to explain the infrared results. Other recent measurements have

been made by SCHUSTER (1969) who has recorded a series of scattered profiles at short time interval in order to observe variations in the layer and PALMER (1969) who has used a bistatic arrangement to measure the angular variation of the scattering function. An interesting feature common to these two sets of measurements is the observation of non-Rayleigh scattering at heights of 30 km and above.

4.2(c) *Transmission factor measurements.* In this paper we have limited the discussion to the atmosphere above 10 km. The lowest 10 km of the atmosphere do, however, affect the return from greater heights in so far as the laser radar signals suffer most of their extinction in this region. If we are to make an absolute determination of the scattering coefficient at a given height we must accurately know the transmission factor T for the lower atmosphere. Tables of extinction coefficients have been published by ELTERMAN (1964) but such tables only apply to an exceptionally clear atmosphere and may be seriously in error on a particular occasion. It is useful therefore to make measurements of T at the same location and, if possible, the same time as the laser measurements.

The best way of measuring T is to use the laser radar to obtain echoes from a given height at various zenith angles. Comparison of the echo strengths received at these different angles then enables the value of T to be calculated. SANDFORD (1967b) working at Winkfield in England, has made measurements of this type and has obtained values of T between 0·4 and 0·5 at the Ruby laser wavelength (0·6943 μ). These values are very much lower than those predicted by Elterman's model for a clear atmosphere, probably due to local atmospheric pollution near London, England. Similar measurements have been made in Kingston, Jamaica, which is an area relatively free from industrial pollutions. Values of T between 0·75 and 0·80 were obtained for clear nights free of haze and ground mist (OTTWAY, 1968). Similar values were also obtained at Kingston by a different experimental technique, in which the brighter planets were photographed at different zenith angles through a narrow-band filter centred on the ruby laser wavelength.

4.3 *Measurements on minor constituents*

4.3(a) *Resonant scattering studies.* As mentioned earlier, the only experimental results using resonant scattering techniques are those of BOWMAN *et al.*, (1969) on atmospheric sodium. Using a tunable dye laser at a wavelength of 0·5896 μ, the scattered photon count at heights between 60 and 160 km has been measured at Winkfield in England. The accumulated photon count for three nights in September 1968 are reproduced in Fig. 10. It can be seen that in spite of the high background noise level, there is a significant increase in the counts between 87·5 and 100 km.

Calibration of the system was carried out by comparing the excess signal received at great altitudes with that received by Rayleigh scattering from between 30 and 35 km. From approximate measurements on the bandwidth of the transmitted signal Bowman *et al.* were able to calculate the column number density of sodium atoms. A scale for the volume number density deduced from this is shown on the right-hand side of Fig. 10. It can be see that the values deduced from the laser results agree well with those shown in Fig. 3, which were based on rocket-borne photometer measurements.

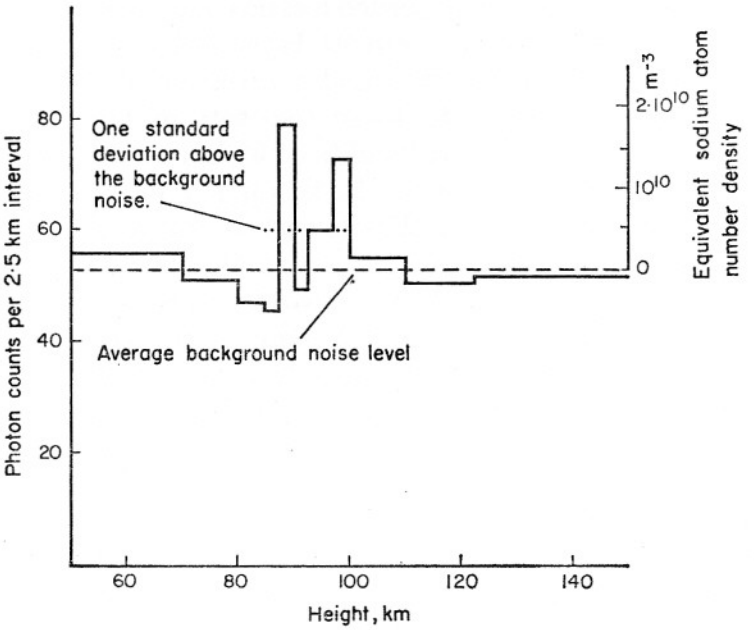

Fig. 10. Photon counts obtained at Winkfield, England using a wavelength of 0·5896 μ.

4.3(b) *Absorption studies.* As discussed in Section 2, the use of the resonant absorption technique has been confined to the measurement of the water vapour content. Since the results obtained by this method are still limited to the lower 10 km of the atmosphere, they will not be described any further in this review.

5. Conclusions

(1) The laser radar system provides a powerful and relatively cheap method for studying the height profile of the stratospheric aerosol layer and its variations with time during both day and night.

(2) Due to the presence of the stratospheric aerosol layer, the laser–radar system cannot—in its simple form—provide information about the atmospheric density and temperature below about 30 km.

(3) The more highly developed and powerful systems at present in operation are now able to record statistically significant signals at all heights below 90 km. This data can be used to provide estimates of the day-to-day and diurnal changes that occur from the standard atmosphere. The oscillations of density caused by 'tides' or 'internal gravity waves' can also be detected and studied.

(4) At high latitudes there is some evidence for scattering from aerosols around a height of 70 km. At heights other than this (apart from the stratospheric aerosol layer) the evidence from laser-radars for the presence of aerosols is highly controversial.

(5) The advent of tunable lasers has made it possible to observe resonant scattering by specific constituents in the atmosphere. Observation of the scattered signals offers a very powerful method for atmospheric study in the future.

(6) Absorption and Raman scattering measurements, both at present confined

to the troposphere, also offer interesting possibilities for the future study of specific constituents of the upper atmosphere.

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REFERENCES

BAIN W. C. and SANDFORD M. C. W.	1966	*J. Atmosph. Terr. Phys.* **28,** 543.
BARTH C. A.	1964	*J. geophys. Res.* **69,** 3301.
BLEANEY B. I. and BLEANEY B.	1967	*Electricity and Magnetism.* Oxford University Press, Oxford.
BOWMAN M. R., GIBSON A. J. and SANDFORD M. C. W.	1969	*Nature, Lond.* **221,** 456.
BUTLER S. T. and SMALL K. A.	1963	*Proc. R. Soc.* **A274,** 91.
CARR M. H. and GABE H. J.	1967	*J. geophys. Res.* **72,** 4007.
CHAGNON C. W. and JUNGE C. E.	1961	*J. Met.* **18,** 746.
CHAMPION K. S. W.	1968	*Meteorological Investigation of the Upper Atmosphere* (Edited by QUIROZ R. S.), p. 47–56. Amer. Met. Soc.
CLEMESHA B. R., KENT G. S. and WRIGHT R. W. H.	1966	*Nature, Lond.* **210,** 184.
CLEMESHA B. R., KENT G. S. and WRIGHT R. W. H.	1967	*J. appl. Metr.* **6,** 386.
COONEY J. A.	1968	*Appl. phys. Lett.* **12,** 40.
COONEY J. A.	1969	Raman scatter research. Presented at the *2nd Lidar Conf., Brookhaven, U.S.A.*
COSPAR INTERNATIONAL REFERENCE ATMOSPHERE	1965	North-Holland, Amsterdam.
DEIRMENDJIAN D.	1965	*J. geophys. Res.* **70,** 743.
DIVARI N. B.	1964	*Geomag. & Aeronomy* **4,** 688.
DONAHUE T. M. and MEIER R. R.	1967	*J. geophys. Res.* **72,** 2803.
ELTERMAN L.	1964	*Appl. Optics* **3,** 745.
ELTERMAN L.	1966	*Appl. Optics* **5,** 1769.
FIOCCO G. and GRAMS G.	1964	*J. Atmos. Sci.* **21,** 323.
FIOCCO G. and GRAMS G.	1966	*Tellus* **18,** 34.
FRIEND, J. P.	1966	*Tellus* **18,** 465.
GOYER G. G. and WATSON R. D.	1968	*Bull. Am. met. Soc.* **49,** 890.
GRAMS G. and FIOCCO G.	1967	*J. geophys. Res.* **72,** 3523.
Handbook of Geophysics	1960	Macmillan, New York
HEMENWAY C. L., SOBERMAN R. K. and WITT G.	1964	*Tellus* **16,** 84.
HUNT B. G.	1966	*J. geophys. Res.* **71,** 1385.
HUNTEN D. M. and WALLACE L.	1967	*J. geophys. Res.* **72,** 69.
JUNGE C.	1963	*Air Chemistry and Radio-activity.* Academic Press, New York.
JUNGE C. E. and MANSON J. E.	1961	*J. geophys. Res.* **66,** 2163.

KENT G. S., CLEMESHA B. R. and WRIGHT R. W. H.	1967	*J. Atmosph. Terr. Phys.* **29,** 169.
LEONARD D. A.	1967	*Nature, Lond.* **216,** 142.
LINDZEN R. S.	1967	*Q. Jl R. met. Soc.* **93,** 18.
LINDZEN R. S.	1968	*Proc. R. Soc.* **A303,** 299.
MCCORMICK P. D., SILVERBERG E. C., POULTNEY S. K., BETTINGER R. T. and ALLEY C. O.	1969	*J. Atmosph. Terr. Phys.* **31,** 185.
MAEDA K.	1968	*Annls Géophys.* **24,** 173.
MATSUSHIMA, S.	1968	*J. geophys. Res.* **73,** 3085.
MOSSOP S. C.	1965	*Geochim. cosmochim. Acta* **29,** 201.
NAGATA T. T., TOHMATSU H. and TSURUTA H.	1968	*Space Research VII.* pp. 639–646. North-Holland, Amsterdam.
NEWKIRK G. and EDDY J. A.	1964	*J. Atmos. Sci.* **21,** 35.
NISHIKORI K., ISHIDA T., UCHIKURA K., MURANAGA K., ICINOSE M., MASUDA Y., NAGATAKE T., HIRONO M. and IGARASHI T.	1965	*J. Rad. Res. Lab. Japan* **12,** 312.
NAKAMURA Y. and CLEMESHA B. R.	1969	*The Interpretation of Laser Radar Returns from Atmospheric Aerosols.* Relatoria Cientifico Lafe-86 C.N.A.E. São José dos Campos, Brazil. April 1969.
NORTON R. B.	1967	Thesis, University of Colorado.
PALMER E. P.	1969	Particle size distribution. Presented at the *2nd Lidar Conf., Brookhaven U.S.A.*
PATON J.	1967	*Bull. Inst. Phys. Lond.* **18,** 67.
PEARCE J. B.	1968	*Symp. Phys. & Chem. of the Upper Atmosphere.* Waltham, Massachusetts.
PILIPOWSKYJ S., WIENMAN J. A., CLEMESHA B. R., KENT G. S. and WRIGHT R. W. H.	1968	*J. geophys. Res.* **73,** 7553.
POULTNEY S. K. and SILVERBERG E.	1969	*Mesosphere aerosol concentration: theory and experiment.* University of Maryland Tech. Rep. No. 964.
LORD RAYLEIGH	1871	*Phil. Mag.* **51,** 107, 274, 447.
REED E. I.	1968	*J. geophys. Res.* **73,** 2951.
ROSEN J. M.	1964	*J. geophys. Res.* **69,** 4673.
ROSEN J. M.	1968	*J. geophys. Res.* **73,** 479.
SANDFORD M. C. W.	1967a	*J. Atmosph. Terr. Phys.* **29,** 1651.
SANDFORD M. C. W.	1967b	*J. Atmosph. Terr. Phys.* **29,** 1657.
SCHOTLAND R. M.	1965	*Optical Sounding II Final Report—I.* New York Univ. School of Engineering and Science.
SCHUSTER B.	1969	High altitude aerosol probing. Presented at the *2nd Lidar Conf., Brookhaven.*
SILVERBERG E. C. and POULTNEY S. K.	1967	*Optical radar back scattering from the mesopause region during July and August* 1967. University of Maryland Tech. Rep. No. 765.
SISSENWINE N., GRANTHAM D. D. and SALMELA H. A.	1968	*Humidity up to the mesopause* A.F.C.R.L. Rep. No. 68-0550.
SOBERMAN R. K. and HEMENWAY C. C.	1965	*J. geophys. Res.* **70,** 4943.

Theon J. S., Nordberg W., Katchen L. B. and Horvath J. J.	1967	*J. Atmos. Sci.* **24,** 428.
U.S. Standard Atmosphere	1962	U.S. Government Printing Office, Washington D.C.
U.S. Standard Atmosphere Supplements	1966	U.S. Government Printing Office, Washington D.C.
Volz F. E. and Goody R. M.	1962	*J. Atmos. Sci.* **19,** 385.
Witt G.	1968	*Tellus* **20,** 98.

Reference is also made to the following unpublished material:

Ottway M. T.	1968	Private communication.
Sandford M. C. W.	1968	Private communication.
Volz F. E.	1969	Private communication.