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The Upper Atmosphere

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1. Introduction

The study of the earth's atmosphere is of interest to scientists from many disciplines and present day knowledge and understanding of the atmosphere, from the ground to the highest levels, is the result of the combined contributions of meteorologists, physicists, chemists, astronomers, geomagneticians, radio engineers and space scientists. This range of scientific disciplines has, not unexpectedly, been matched by an equally extensive variety of experimental approaches – from the pioneer probing of the first few kilometres with instruments carried up mountains, to the modern radio and space-vehicle techniques yielding data from distances out to, and far beyond, the limits of the atmosphere.

Modern scientific studies of the vertical structure of the atmosphere may be considered to have their origins in the seventeenth century with Torricelli's invention of the barometer. It was in 1649 that he succinctly described the total envelopment of the earth by an atmosphere in the phrase "We live submerged at the bottom of an ocean of the element air which, by unquestioned experiments is known to have weight". This picture of mankind living at the bottom of an ocean of air naturally led to curiosity about the "depth" of this ocean and indeed it was not long before scientific investigation of the vertical structure of the atmosphere was under way. Within a few years one of the newly invented barometers was carried up a mountain and the fact established that pressure decreases upwards.

The invention of the thermometer in the early 18th century was in due course followed by an attempt to measure the temperature change upwards using kites and in 1793, John Dalton from measurements on mountain tops, showed that up to an altitude of three miles at least the temperature decreased upwards "in nearly arithmetical progression and at the rate of 1°F for every hundred yards". The 19th century saw the development of balloons suitable for carrying aloft different types of meteorological instrumentation. In the present century the development of radio and radar have made possible the radiosonde whereby meteorological data from tracked balloons can be transmitted continuously back to the ground. The practical upper height limit with balloons is about 30 km and in the past thirty years or so the upward extension of this ceiling to hundreds and thousands of kilometres has been made possible with the development of rockets and satellites.

In addition to these direct *in situ* measurements great quantities of scientific data on the high atmosphere have also come from ground-based studies. Optical observations of the movement of clouds, of atmospheric refraction phenomena, of

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This was a Friday evening discourse at the Institution in 1980

meteors and of the aurora date back several centuries. The refraction of sound waves by the atmosphere has also been turned to good account, not only in quantitative measurements from the ground, but in rocket studies. In the present century ground-based radio studies have provided a wealth of information on the upper atmosphere, especially on the electrical properties of the atmosphere above about 60 km. In the past decade laser beams directed upwards from the ground have been used to study the high atmosphere.

These experimental studies have, over the years, been supplemented by much theoretical work to give the picture we now have of the physical state of the earth's upper atmosphere. The sum total of this knowledge is now very extensive and in this lecture I can only touch upon a few selected topics.

2. The Earth's Atmosphere and the Sun

The principal physical, chemical and electrical properties of the earth's atmosphere and particularly of the higher reaches of the atmosphere are very largely the result of its interaction with solar wave and particle radiation. The sun radiates electromagnetic energy at all wavelengths from the long-wave radio end of the spectrum to the shortest wavelength X-ray bands and some 8½ minutes after leaving the sun these radiations reach the vicinity of the earth and interact to a greater or lesser degree with the atmosphere. Fortunately for life on earth the ultraviolet and X-radiations are absorbed at various levels between about 40 and 200 km. The selective absorption of the ultraviolet and X-radiation by particular atmospheric constituents gives rise to the unique electrical properties of the high atmosphere by providing a series of ionised strata collectively known as the ionosphere. Other ultraviolet radiation is very effectively absorbed lower down in the atmosphere by ozone.

In addition to electromagnetic-wave radiations the sun also emits continuously streams of energetic electrically charged particles – mainly protons and electrons. This is the so-called "solar wind" – an outward flow of charged particles moving at velocities of a few hundred kilometres per second – and a few days after leaving the sun, sweeping past the earth. At times of violent storms in the solar atmosphere these streams of particles may be ejected with much higher velocities and reach the earth within one day. They interact with the earth's magnetic field in a complicated way, and with the gases of the high atmosphere, and their arrival at the earth is often signified by the occurrence of geomagnetic storms, by disturbances to long-distance radio-wave communication *via* the ionosphere, and by visible auroral displays at high latitudes. The frequency and intensity of solar disturbances, and the consequent emission of energetic charged particles, are closely associated with the occurrence of sunspots and the incidence of geomagnetic storms, ionospheric storms and auroral displays show characteristic time variations over 11 years and 27 days, respectively, related to the sunspot cycle and the period of rotation of the sun.

3. Physical Parameters of the Atmosphere

The principal physical parameters of the atmosphere are its density ρ, (or pressure p), its temperature T and its mean molecular mass m (which is determined by its chemical composition). These parameters are not completely independent of one another but are related by the usual gas law $p = \rho kt/m$, k being Boltzmann's Constant. Density and pressure are particularly interlinked because over any small height range dh the pressure change dp is given by $dp = -g\rho dh$ where g is the gravitational acceleration. Hence measurement of the vertical profile of pressure (or density) will itself give the vertical profile of density (or pressure).

In upper atmosphere studies, when direct accurate measurement of a particular parameter is difficult or impossible, this inter-relationship between the basic parameters is often used to deduce that parameter from measurements of the others.

In the following paragraphs we shall briefly discuss each of these parameters.

3.1. *Density*

Both density and pressure decrease exponentially upwards and the rapidity of the fall-off with increasing height is quite striking. Thus at the top of Mount Everest (*ca.* 8.8 km) the density of the atmosphere is reduced to about two-fifths of the density at sea-level. At 20 km density and pressure are reduced by a factor of ten, at 50 km by a factor of a thousand and at an altitude of 100 km the atmospheric density is only about one ten-millionth part of its value near the ground. Because of the upward fall-off in density we can say that although the earth's atmosphere extends upwards in ever more tenuous form to hundreds and even to thousands of kilometres, most of the mass of the atmosphere is in the first few kilometres – in fact three-quarters of the total mass is contained in the first 11 km, *i.e.* in a zone of thickness only about 0.2% of the earth's radius. Viewed on a world-wide scale the bulk of the atmosphere must therefore be seen to be contained in an extremely thin shell enveloping the whole earth.

Fig. 1 shows the change in the density of the atmosphere up to an altitude of 500 km. At great altitudes the density shows quite marked changes with time of day, seasonally, with solar activity and with latitude. Such changes are basically due to changes in temperature. Above about 400 km the density of the atmosphere is appreciably smaller than that in the best vacuum attainable in the laboratory.

Although estimates of the density of the upper atmosphere have long been available from ground-based measurements (optical observations of the aurora, of meteors and radio studies of the ionosphere) in recent years the most striking data have been obtained from rocket and satellite experiments. Pressure gauges on rockets and observations of the rate of fall of lightweight spheres have yielded reliable data over the height range 40–100 km. Measurements of the effect of atmospheric drag on satellite orbits have provided density data for the upper height range 200–1100 km.

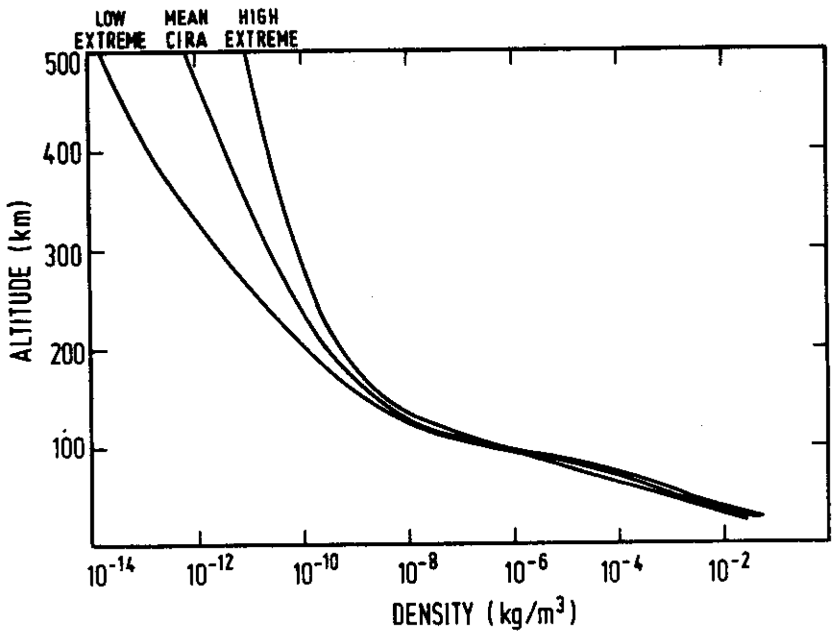

Fig. 1. Density/height variation up to 500 km. Mean curve refers to annual mean conditions for latitude 30° and average solar activity. Low and high extreme curves have a frequency of occurrence of 1% or less. (Committee on Space Research International Reference Atmosphere – CIRA).

3.2. *Temperature*

The variation in the average temperature of the atmosphere from ground level to 120 km is shown in Fig. 2. It will be seen that features of the vertical temperature structure in the first 120 km are the minima centred on 20 km and 85 km, the pronounced maximum at 52 km and the sharp positive temperature gradient in the 100–120 km height range. This characteristic variation in temperature with height provides a convenient basis for establishing the nomenclature (troposphere, stratosphere, *etc.*) shown in the diagram.

As stated earlier, the linear fall in temperature in the first ten kilometres or so was first noticed by John Dalton in 1793 and the lapse rate which he measured ("1°F in 100 yards") is within about 1% of the average value accepted to-day. Dalton is usually remembered as the founder of atomic theory and it is not so generally known that he was also keenly interested in the weather and in the atmosphere. In fact from March 1787 until his death in July 1844 he maintained a daily set of observations of the weather. In addition to his measurements on the fall of temperature with height, Dalton also established the existence of some connection between the earth's magnetism and the aurora and in 1802 he carried

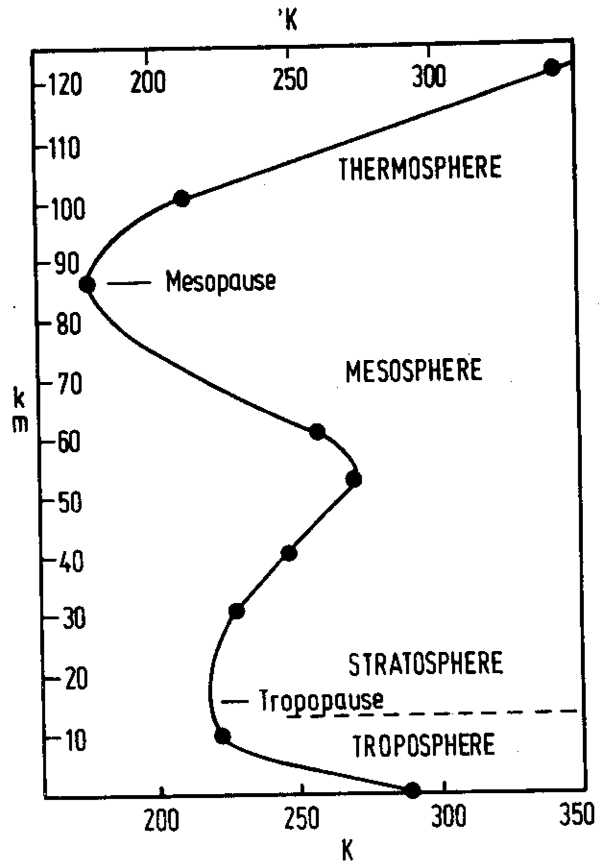

Fig. 2. Temperature/height variation up to 120 km.

out his first piece of major chemical analysis when he showed that air consisted in nearly 79% nitrogen and 21% oxygen.

In noting the initial fall in temperature upwards and the subsequent maximum in temperature centred around 52 km (now known to be a consequence of very heavy absorption of solar ultra-violet radiation by ozone) it is interesting to recall an entry which Samuel Pepys made in his famous diary on Monday 4 June 1666. That summer there was a gun battle in the English Channel and Pepys, in his office at the Admiralty, wrote "It is a miraculous thing that we all Friday and Saturday and yesterday did hear most plainly guns go off and yet at Deal and Dover they did not hear one word of a fight nor think they heard one gun. This, added to what I have set down before, makes room for a great dispute in philosophy – how we should hear it and they not, the wind that brought it to us being the same that should bring it to them – but so it is!"

In the years since 1666 many other examples of abnormal sound wave propagation were reported, but the explanation in terms of upward refraction in the 0–10 km zone of negative temperature/height gradient and downward refraction in the positive gradient between 30 and 50 km, was not fully resolved until the 1920's when controlled sound-wave observations, experimental evidence from other sources, and theoretical work on atmospheric tidal oscillations, established beyond doubt the existence of a temperature maximum in the 50–60 km height range.

The lower atmosphere is largely transparent to solar radiation in the visible and near-infrared parts of the spectrum and consequently this part of the atmosphere is not heated much by direct sunlight. The energy of this ~~ionising~~ radiation is however absorbed by the surface of the earth and then re-radiated at much longer wavelengths and it is the absorption of this longer wavelength radiation by constituents such as water vapour and carbon dioxide which heats the first few kilometres of the atmosphere. (In trapping energy re-radiated from the land and sea surfaces these atmospheric constituents play a role analogous to the glass in a greenhouse and the term "greenhouse effect" is sometimes used to describe the phenomenon).

The atmosphere itself radiates energy upwards and downwards at infrared wavelengths and at any level the net upward flux is the result of downcoming energy radiated from the atmosphere above and upgoing radiation from the atmosphere and ground below. The presence or absence of clouds will, of course influence the net heat flux. In the troposphere there is a net upward flux of radiated energy and the resultant cooling of the atmosphere is compensated by a convective transfer of heat from the lowest levels adjacent to the warm surface of the earth. In this convective circulation the latent heat associated with the evaporation and condensation of water vapour plays an important role.

At altitudes above about 10 km and up to 20–25 km the net upward flux of radiated energy is roughly compensated by heating produced by the absorption of sunlight and the fall in temperature upward is halted. This constant temperature zone was first discovered around 1900 and it was appropriately designated as the "stratosphere". For many years subsequently, in the absence of evidence to the contrary, it was considered that this constant temperature persisted upwards to very high levels, *i.e.* that the "stratosphere" extended to very great heights. Later, as described below, the zone of constant temperature was found to be only about 20 km thick. At the same time the term "stratosphere" has been used to designate a greatly extended range of heights. Thus many workers now use the term "stratosphere" to indicate that part of the atmosphere from the top of the troposphere to the level of the temperature maximum – near 52 km – although only the lower part of this height range is a zone of constant temperature.

Higher up again the balance of heat flux is changed by a new factor *viz*., the absorption of solar ultraviolet radiation by ozone. At all levels in the atmosphere the ozone content is extremely small (its density is never more than about one

hundred thousandth that of the ambient atmosphere) but even this trace amount is sufficient to absorb all solar u.v. radiation of wavelength less than about 290.0 nm. The temperature maximum centred on 52 km is the consequence of this absorption of solar u.v. radiation. Ozone in the atmosphere is produced mainly by the recombination of atomic and molecular oxygen and its maximum concentration occurs around the 25 km level. However the absorption of the ionising solar radiation by ozone at and above 50 km is sufficiently large to produce maximum heating near that level. There is a greater density of ozone below 50 km but at these levels the increasing solar u.v. radiation is then greatly attenuated and consequently the rate of absorption of solar energy (and hence the temperature of the atmosphere) is smaller. Above 52 km the temperature of the atmosphere falls again and reaches a minimum of about 178K (−95°C) near 85 km (the so-called "mesopause"). This is the coldest spot in the earth's atmosphere and the low temperature is simply due to the absence of any significant heating mechanism – unlike the selective absorption of solar u.v. radiation at levels below and above 85 km. A major source of heat loss at the mesopause is the emission of infrared radiation at 15 μm by carbon dioxide.

Some of the first experimental evidence for the low temperature between 80 and 90 km (and for the high temperatures above 120 km or so) came some 40 years ago from radio-sounding studies of the ionosphere.

The thermal balance of the atmosphere above 100 km (the "thermosphere") is an exceedingly complicated process involving many different factors and it is convenient to consider separately the lower thermosphere (100 to about 250 km) where there is a large positive height gradient of temperature, and the upper thermosphere where there is practically no vertical gradient of temperature (Fig. 3).

In the lower thermosphere the principal source of heat arises from the absorption of solar ultraviolet and X-radiation which ionise or dissociate a good proportion of the atoms and molecules. A second important heat source arises from the entry into the high atmosphere of energetic charged particles. Other contributory sources of heat arise from the viscous dissipation of energy contained in tidal and other atmospheric movements and also from Joule heating effects by electric currents flowing in the ionosphere. The heat from these various sources will not necessarily remain at the place where it is initially deposited but will be distributed to different levels by conduction, convection, radiation and wind motions. The very marked heating of the atmosphere above 90–100 km (with resulting temperature gradients larger than at any level in the atmosphere) is mainly the result of photo-dissociation of molecular oxygen by solar ultraviolet radiation in the wavelength range 110.0–180.0 nm.

Above about 800–1000 km the atmosphere consists mainly of ionised helium, ionised hydrogen and electrons. Furthermore some of the principal sources applicable to the lower thermosphere are no longer of significance and thermal conditions at these high levels are mainly determined by conduction of heat parallel to the earth's magnetic field. The net result is that the upper

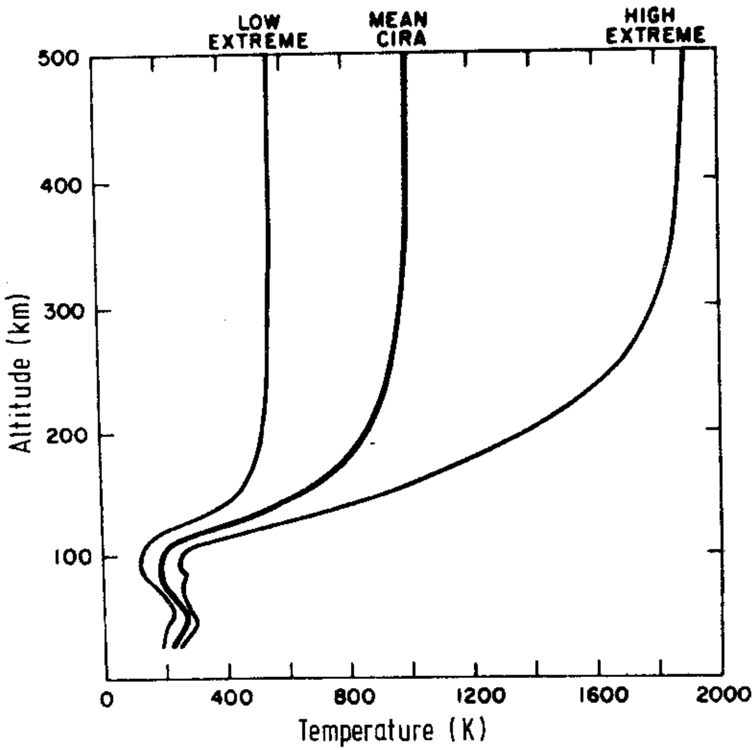

Fig. 3. Temperature/height variation up to 500 km. Mean curve refers to annual mean conditions for latitude 30° and average solar activity. Low and high extreme curves have a frequency of occurrence of 1% or less. (Committee on Space Research International Reference Atmosphere – CIRA).

thermosphere is an isothermal region with the temperatures at an upper limiting value which is in large measure determined by conditions in the lower thermosphere.

At all altitudes the temperature of the atmosphere shows some measure of change with time of day, with season, with latitude and with solar activity but the really large variations occur at levels above about 150 km (Fig. 3).

In speaking of the "temperature" in the thermosphere, and especially of temperature in the upper thermosphere, and comparing these values with temperatures lower down in the earth's atmosphere it is pertinent to remember the very significant differences between the atmosphere at these levels and that at, say, tropospheric levels. In the first place it is to be noted that it is at the lower thermospheric levels that really major changes in the composition of the atmosphere begin to be significant. Starting around 90 km with the large-scale

dissociation of molecular oxygen the change in composition continues upwards with the dissociation and ionisation of other constituents. With increasing height the number densities of lighter constituents and of ions and electrons become progressively more dominant until ultimately the atmosphere consists entirely of ionised hydrogen. (In this context we note too that the temperatures of the ions and of the electrons are sometimes quite different). Secondly it is to be noted that the density of the atmosphere at thermospheric levels is only the minutest fraction of that of the atmosphere near the ground – it is smaller by a factor of 10^7 to 10^{12}.

A variety of methods have been developed to measure temperature directly or indirectly at different levels in the atmosphere. The first 30 km or so are accessible to direct balloon sounding and present no insurmountable experimental problems. However with increasing altitude the decrease in gas density makes direct measurement ever more difficult and different approaches have to be employed. Up to about 100 km certain rocket experiments have provided good data. In the so-called grenade technique small grenades are fired at closely spaced intervals as the rocket ascends – the flash of each explosion is observed and also the time of arrival of the sound impulse from the exploding grenade is recorded with microphones on the ground. The velocity of sound over different height ranges is thus determined and from its variation with height the vertical profile of temperature can be deduced. This technique cannot be used much above about 100 km because the density of the atmosphere at these levels is too small to enable sufficient energy to be communicated to the atmosphere to give a measureable impulse at the ground.

As mentioned earlier the interdependence of the basic physical parameters of the atmosphere means that, subject to certain assumptions, temperature can be estimated from measurements of another parameter such as, for example, density and at great altitudes (200 to 1200 km) temperature has been deduced indirectly from density measurements which themselves have been deduced from rocket and satellite "drag" studies. Temperature sounding of the upper atmosphere at a range of levels has also been very successfully carried out by remote sounding from satellites. Atmospheric constituents which absorb incoming selected solar radiation at infrared wavelengths will also radiate energy at these wavelengths and the intensity of this emitted radiation will depend on the vertical distribution of the gas constituent concerned and on the temperature. Subject to certain assumptions, measurement of the intensity of this emitted radiation at the appropriate wavelength by satellite-borne instrumentation can be used either to deduce the vertical concentration of the atmospheric constituent or the temperature profile. If the emitting constituent is uniformly mixed over the range of atmosphere under consideration as, for example are molecular oxygen and carbon dioxide up to about 90 km, then the intensity of the emitted radiation will depend only on the temperature profile. Carbon dioxide has a strong absorption band near 15 μm and the infrared emission of this molecule at this wavelength has been widely used in satellite experiments to measure temperature. Under clear conditions the measurement of temperature to the lowest altitudes by this radiometer technique presents no serious problems and in theory observations on about six different wavelengths permit the temperature profile from 0 to 50 km to be estimated. However difficulties which arise for lower atmosphere measurements when clouds are present, and the need to monitor surface characteristics of the ground, mean that in practice some twelve channels are required to cover this height-range accurately. Molecular oxygen has an absorption band near 5 mm and since the effect of clouds at millimetre wavelengths is significantly less than at infrared wavelengths satellite radiometer studies at this longer wavelength have been used successfully to measure temperature in the first 20 km or so of the atmosphere. At thermospheric altitudes satellite observations of atomic oxygen emission at 630.0 nm have yielded temperature measurements to an accuracy of about ±65 K for the height range 200–320 km.

Remote sounding techniques similar to those used for the earth's atmosphere have also been successfully used for the initial satellite studies of the atmospheres of Venus and Mars. Fig. 4 shows a temperature profile for the first 90 km of the atmosphere of Venus.

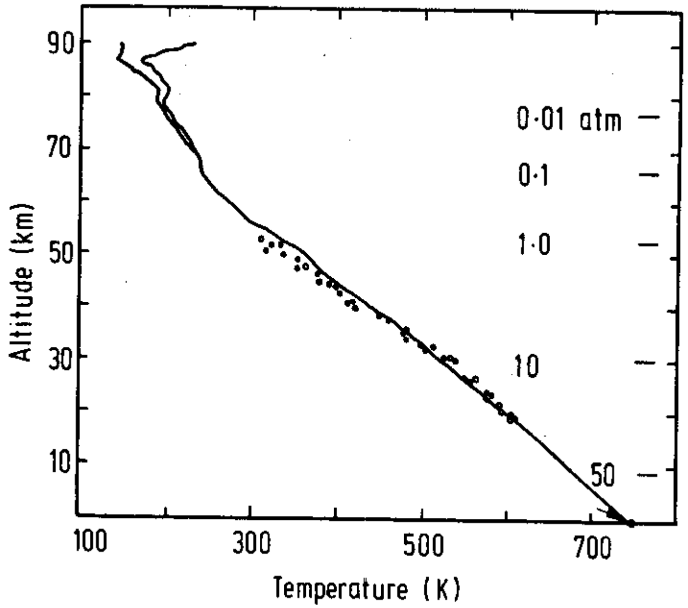

Fig. 4. Temperature profile on night side of Venus. (After S.I. Rasool and R.W. Stewart).

3.3. *Chemical composition*

In 1802, John Dalton in one of his first major pieces of chemical analysis showed that air consisted in some 78% nitrogen and 21% oxygen (a result which he reported in the course of a Friday evening discourse at the Royal Institution). In 1895 Lord Rayleigh showed that the nitrogen in the air had a slightly larger density than that of nitrogen extracted from its compounds and later he and Ramsay tracked this difference down to the presence in air of a heavier chemically inert gas to which they gave the name Argon. We now know that air contains nearly 1% Argon and that nitrogen, oxygen, argon and carbon dioxide are the principal permanent constituents of the atmosphere forming between them all but 0.003% by volume.

The mean composition of the troposphere based on large numbers of samples taken all over the world is shown in Table 1. In addition to the substances shown in Table 1 we have to include water vapour which though not strictly a "permanent" constituent can be present to a very variable degree and in addition there is a wide range of minor constituents such as ozone, hydrogen, xenon, methane, *etc.* which are also permanent constituents but present only in the minutest quantities.

TABLE 1

Chemical Composition of the Troposphere: Principal Permanent Constituents % by Volume

Nitrogen	78.084
Oxygen	20.946
Argon	0.934
Carbon dioxide	0.033
Neon	1.82×10^{-3}
Helium	5.2×10^{-4}
Krypton	1.14×10^{-4}

The composition of the atmosphere as we know it to-day is the result of chemical processes which took place over periods of millions of years after the formation of the earth some 5×10^9 years ago and there may be a long-term evolutionary change in this composition. Factors which influence the evolution and determine the ultimate constitution of a planetary atmosphere include (i) outgassing from the cooling surface of the planet itself, (ii) chemical reactions between the gases of the atmosphere and material in the crust of the planet and later, interaction between the oceans and the atmosphere, (iii) modification of the atmosphere by solar ultraviolet and *X*-radiation – photo-dissociation and photo-ionisation processes, (iv) capture of material from the interplanetary medium, (v) radioactive processes in the interior of the planet, (vi) escape of gases from the top of the atmosphere. The relative importance of these various factors may be expected to vary from one planet to another and to depend on factors such as distance from the sun and mass of the planet. Thus for a small planet near the sun (such as Mercury) the temperature will be so high and the gravitational attraction of the planet so small that practically the whole atmosphere will in time be lost. At the other extreme a massive planet like Jupiter, much further from the Sun and much cooler, will evolve very slowly and retain practically all its original atmosphere including the light constituents like hydrogen and helium. Planets at intermediate distances and of intermediate size will retain an atmosphere but perhaps lose nearly all the lighter constituents. The atmospheres of the planets of the solar system fall into two broad groups and the atmospheres of Mercury, Venus, Earth and Mars – the so-called "inner planets" are probably further along the evolutionary road than those of the outer planets Jupiter, Saturn, Uranus and Neptune.

A fluid in which the temperature decreases upwards will be unstable in that the cooler, denser, layers above will tend to displace the warmer, less dense, layers below. The troposphere has such a negative temperature/height gradient and this is consistent with (though not necessarily the cause of) the instability and continual convective mixing characteristic of this part of the atmosphere. In isothermal regions of the atmosphere in which there is no temperature gradient upwards, and especially in regions in which there is a positive temperature gradient upwards, one would expect much greater stability with little or no mixing and possibly even a tendency for gases to diffuse and separate out with the lighter constituents on top. Fifty or sixty years ago when the existence of a constant temperature zone (the stratosphere) has been definitely established there was considerable speculation as to whether this diffusive separation did in fact occur in and above the stratosphere. Theoretical work of that time showed that for a model atmosphere containing nitrogen, oxygen, argon and helium in the proportions actually found in the troposphere, if diffusive separation came into operation from about 20 km upwards then by about 150 km the atmosphere would already be 90% helium and some 10% nitrogen and that above 200 km it would consist entirely of the light gas helium. However, experiments carried out in the late 1930's in which air samples were recovered from the stratosphere provided no conclusive evidence for an increasing proportion of light gases at high levels. Later experiments in the 1950's with rockets also supported this conclusion. Furthermore spectroscopic evidence – from the absorption spectrum of sunlight and the emission spectrum of the aurora and of the night-sky strongly indicated that at high levels in the earth's atmosphere nitrogen and oxygen, and not light gases, were still the dominant constituents. It is now established that the composition of the atmosphere shows no significant change from ground level up to about 90 km, *i.e.* up to this level it is still 78% nitrogen and 21% oxygen. It should however be said that although the dominance of nitrogen and oxygen is these proportions remains at all levels up to and just beyond the mesopause the proportions of some of the minor constituents certainly change. Thus the ozone content in the atmosphere reaches a maximum around 25 km where the abundance is some 20–25 times what it is at the ground.

Above 90 km (in the lower thermosphere) the dissociation of molecular oxygen by solar ultraviolet becomes an important factor in effecting a major change in the chemical composition of the atmosphere (the absorption of solar energy associated with this process gives rise to the sharp temperature rise at this level noted earlier). In the stable thermal conditions of the thermosphere, gravitational separation of the atmospheric gases gradually sets in, with heavier constituents below and lighter ones above, until ultimately at some thousands of kilometres up the only constituents left are the light gases helium and hydrogen. Further out again hydrogen alone dominates and at a few earth radii (10 to 20 thousand km) the hydrogen exists only as ionised hydrogen *i.e.* it is a proton atmosphere. Gradually this proton atmosphere merges with the interplanetary "space" – a space which is itself continually and continuously replenished with protons ejected from the sun.

At the "top" of the earth's atmosphere (above about 600 km) the gas temperature is so high, the gas density so small, and the constituent gases so light, that the atoms acquire velocities which, in the collision-free environment, are sufficient to enable them to escape from the earth's gravitational field. This level in the atmosphere is appropriately named the "exosphere". In certain parts of the world substantial quantities of helium escape from the earth's crust into the atmosphere but the actual amount in the lower atmosphere remains at a very minute level because this light inert gas diffuses upwards and is ultimately lost at the top of the atmosphere.

3.3.1. *Minor constituents*

Although the main features of the chemical composition of the atmosphere are now fairly well established a great amount of data is still required on the number density distributions of several of the minor constituents. I shall refer briefly to the measurements of two such minor constituents, *viz.*, atomic oxygen and sodium, which serve to illustrate two widely different techniques currently being employed.

Atomic oxygen in the upper atmosphere is an important minor constituent because of the key role it plays in the neutral and ion chemistry of the atmosphere in the height range 60 to 120 km. Thus the electron concentration below 70 km is influenced by the role of atomic oxygen in detaching electrons from negative ions; around 100 km atomic oxygen is closely involved in the excitation of airglow emissions at 557.7 nm and 630.0 nm; it plays an important role in the thermal balance of the lower thermosphere and the vertical distribution of atomic oxygen is relevant to considerations of vertical transport by eddy and molecular diffusion. For these and other reasons, it has become important to measure atomic oxygen concentrations in the height range 60 to 120 km and in recent years this has been successfully accomplished with the rocket-borne experiment illustrated in Fig. 5. A specially constructed lamp provides an intense beam of ultraviolet radiation at the wavelengths of the oxygen triplet near 130 nm and is used in the measurement of both the absorption and resonance fluorescence of atomic oxygen over the

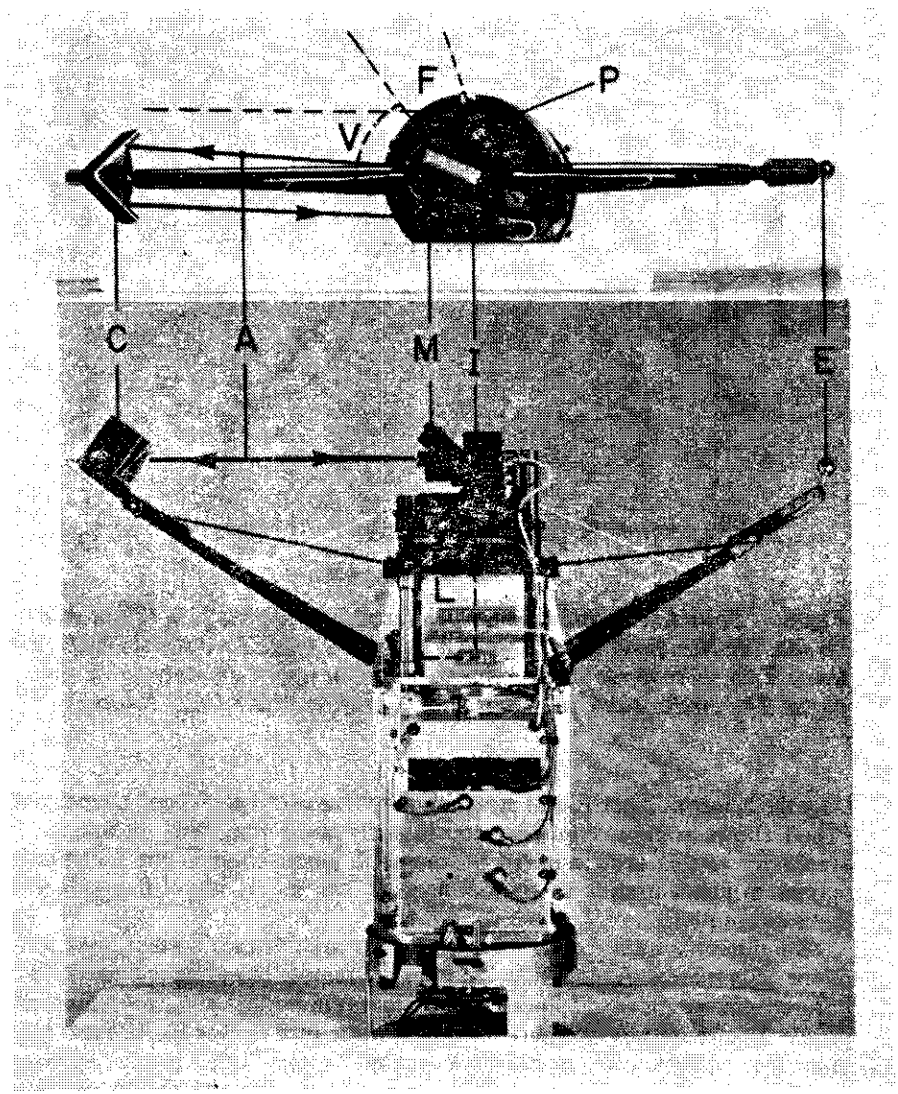

Fig. 5. Rocket payload to measure atomic oxygen. Top: plan view. Lower: side view. A: absorption path, C: corner reflector, E: electron probe, F: field of view of photomultiplier, I: ionisation chamber, L: lamp, M: mirror, P: photomultiplier, V: volume illuminated by lamp. (After P.H.G. Dickinson et al).

height range 60–120 km. The characteristics of the lamp are determined before the flight in a calibration experiment in which the absorption is measured for different known concentrations of atomic oxygen. The absorption is measured over a path length of 40 cm by reflection in a mirror and detected with an ionisation chamber photometer. In a separate simultaneous experiment the height variations of resonance fluorescence emissions at 130 nm stimulated in the ambient gas by the radiation from the lamp are detected by a suitable photomultiplier. These resonance fluorescence measurements only give relative values of the atomic oxygen concentration but provide a useful independent check

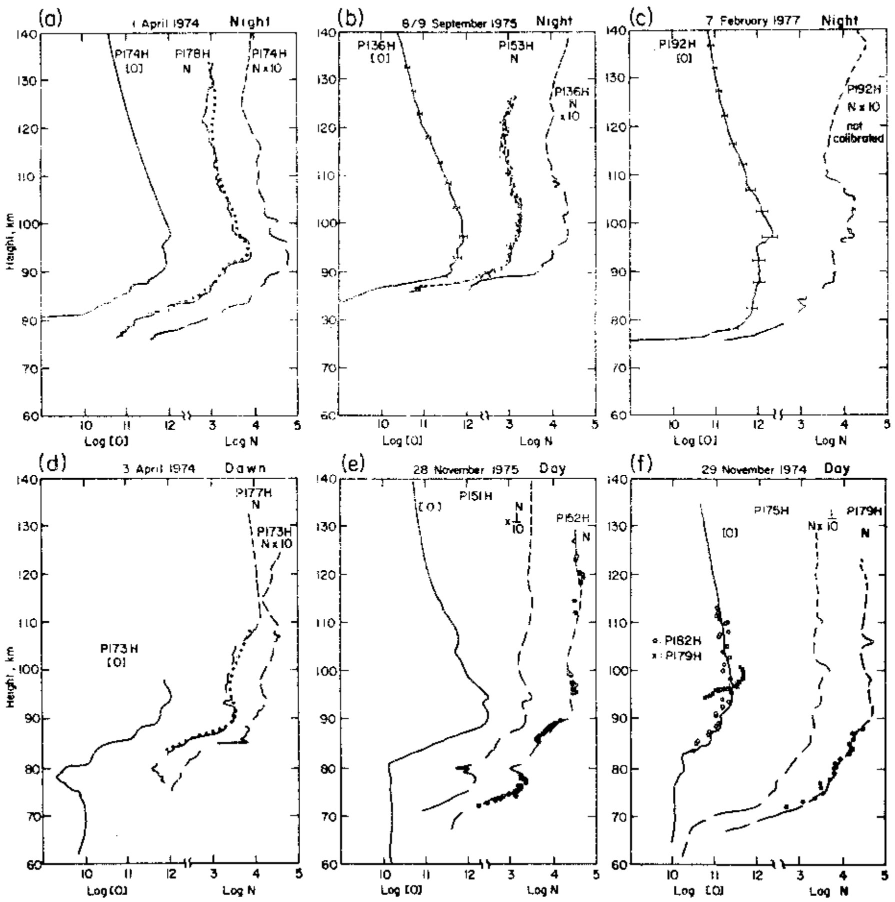

Fig. 6. Concentrations of atomic oxygen [O] and electrons N, both in cm^{-3} measured together at South Uist, Scotland on six occasions. (a-c) at night (d) at dawn (e-f) by day. (After P.H.G. Dickinson et al).

on the detailed shape of the vertical profile.

Atomic oxygen concentrations over the height range 80–140 km obtained with this rocket payload are shown in Fig. 6. Measurements of electron density made simultaneously with rocket payloads are also shown.

A second, and in some ways surprising minor constituent in the upper atmosphere is sodium. A prominent yellow line near 589.3 nm was observed in the night-time airglow spectrum by Slipher some fifty years ago and in 1938 accurate wavelength measured with a Fabry–Perot interferometer established beyond doubt that this indeed was a sodium line emission. The intensity of these emissions remains fairly constant throughout the hours of darkness but during morning and evening twilight periods it is enhanced by a factor of 50 to 100 – a phenomenon which is referred to as the sodium twilight flash. Observations of the times at which this twilight flash disappears have been used to determine the height of the emitting layer and it is now known that the sodium atoms are contained between 80 and 150 km with a maximum concentration about 95 km. It is believed that the main source of these sodium atoms is the ablation of meteors but a contribution from sea spray carried upwards from the troposphere has also been suggested by some workers. It may be noted that in addition to sodium other metallic atoms ablated from meteors and deposited around this level are potassium, calcium, magnesium and silicon.

In the past decade tuned laser beams directed upwards from the ground have been successfully used to study the distributions of some of these metals in the high atmosphere. A laser is tuned to the wavelength of the sodium lines at 589.0–589.6 nm (or 766.7 and 769.9 nm for potassium) and resonance scattering from the metallic atoms near 95 km is received back at the ground. By using a pulsed

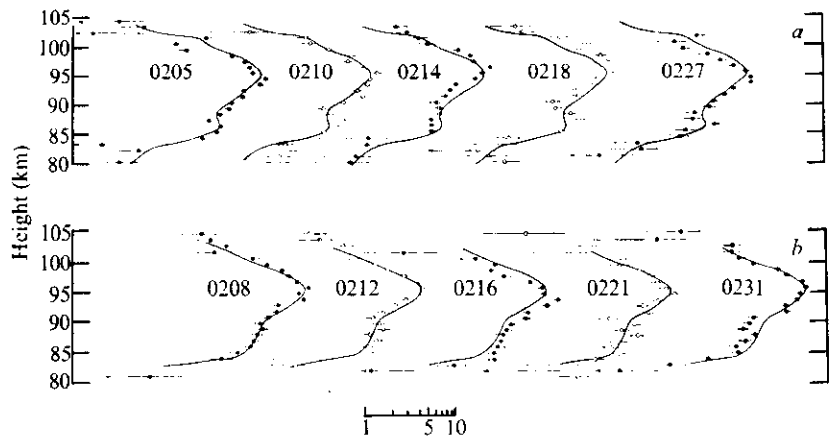

Fig. 7. Concentration of sodium between 80 and 105 km measured by laser technique. (After L. Thomas).

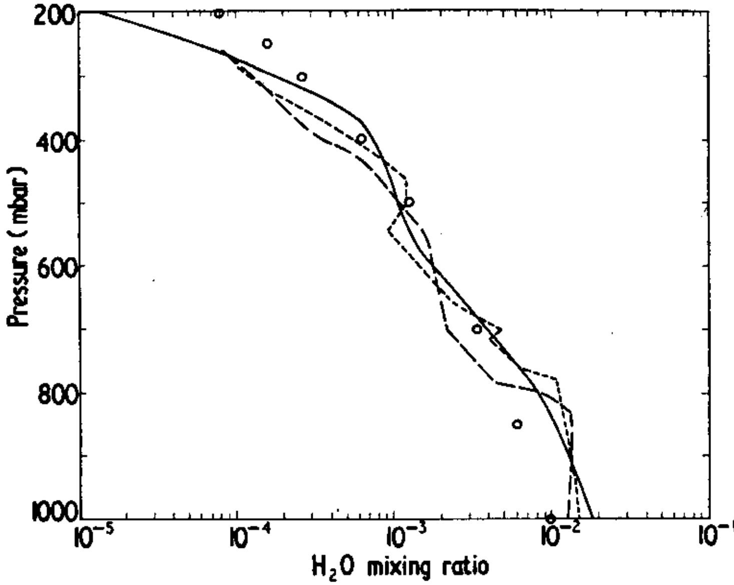

Fig. 8. Solid line: Water vapour profile over height range 0-12 km measured on satellite Nimbus 4 over 13°N 127°E at 1532 GMT on 8 April 1970. Broken lines: Two radiosonde measurements made at approximately same time and place. The circles show the "first guess" profile. (After W.L. Smith).

laser the height and horizontal distributions of the metal atoms can be measured. Fig. 7 shows some results obtained recently using this laser technique. The sensitivity of this ground-based technique is emphasised in the fact that the maximum concentration of sodium near 95 km is only about one part in ten thousand million.

Satellites have also been used successfully to study the distribution of certain minor constituents. As mentioned earlier the remote sounding techniques using infrared emissions can, subject to certain assumptions, give the temperature profile. In circumstances where the temperature profile over the relevant height interval is either known, or can be measured by some other means, then these satellite infrared emission measurements can be used to give the distribution of the emitting constituent and the distributions of ozone and water vapour have been measured by this technique. Fig. 8 shows a water vapour profile over the height range 0 to 12 km obtained by a satellite-borne radiometer together with profiles obtained from the usual meteorological radiosondes launched from the ground.

3.3.2. *Atmospheric pollution*

In recent years increasing concern has sometimes been expressed about the possibility that some of modern man's activities may produce serious permanent changes in the environment. In particular there has been concern about the possible influence of such activities on two minor but very important constituents of the upper atmosphere, *viz.*, carbon dioxide and ozone.

The exchange of carbon dioxide between the atmosphere and the oceans and the terrestrial biosphere (the "carbon cycle") is of course critical to life. Carbon dioxide in the atmosphere also plays an important role in determining the rate at which the earth loses energy and an increase in CO_2 content will enhance the "greenhouse effect" mentioned earlier. Although this reduction in the loss of heat from the earth's surface would only be expected to increase the temperature by a degree or so, a more important secondary effect would come from an expected small increase in global cloud cover. The burning of fossil fuels is a direct and major source of atmospheric carbon dioxide and there is now a general consensus that the amount of CO_2 in the atmosphere has increased from about 290 parts per million in the last century to about 330 parts per million at the present time and that by the end of this century the figure will have increased to about 380 parts per million. The contribution of fossil fuel burning is between four and five times that of the contribution from the biosphere, and calculation indicates that if the present rate of increase in fossil fuel burning were maintained then the CO_2 content of the atmosphere would increase ten-fold by the year 2100. However interaction between the atmosphere and the oceans rapidly removes CO_2 from the atmosphere and it is estimated that as much as 60% of any increase would be removed from the atmosphere in this way. The rapid increase in fossil fuel burning of the past hundred years may of course not continue for another century but nevertheless the problem of increasing CO_2 in the atmosphere (and the oceans) needs to be watched carefully.

The second minor constituent of the high atmosphere which is the subject of general concern is ozone. The ozone content of the atmosphere is minute but it plays a critical role in screening the earth from harmful solar u.v. and X-radiation. The maximum concentration of ozone is located at an altitude of about 25 km, *i.e.* well into the stratosphere – in fact a large fraction of the total ozone in the atmosphere is concentrated in the stratosphere. The constant temperature conditions of the lower stratosphere and the positive temperature gradient upwards in the upper stratosphere tend to ensure that conditions at these levels are comparatively stable with little vertical mixing – unlike the continual vertical circulation processes which occur in the troposphere. Hence stratospheric pollutants of any sort are liable to stay in the stratosphere for long periods and so have ample time to upset the delicate chemical balance governing the concentration of minor constituents. Pollution of the troposphere is generally not too serious a matter and does little permanent damage because there is continual large-scale mixing and rain rapidly removes all soluble pollutants. However, in the stable conditions of the stratosphere pollutants stay around for very long periods and pollution effects thus tend to be cumulative. The pollutants which give rise to concern are those which act as catalysts in the destruction of ozone. Two possible sources of such catalysts which have been much discussed are: (i) The release of nitrogen oxides and water vapour by the engines of supersonic aircraft flying at stratospheric levels. With the present numbers of such aircraft this source is probably not a major danger but it could possibly become one if in the future supersonic travel became the normal form of air transport. Exhausts from rocket motors also introduce substantial amounts of water vapour into the high atmosphere. (ii) The chlorofluoromethanes (CFMs) which are extensively used as aerosol propellants and in refrigerants. These CFMs are accumulating in the atmosphere at rates close to their release rates since no significant destruction processes for these compounds on land, in the oceans or in the troposphere have yet been discovered. These compounds can be transported to and accumulate in the stratosphere where they can be decomposed by solar ultraviolet radiation and react with atomic oxygen to yield chlorine atoms which can then destroy ozone. It has been estimated that in the 25-year period 1948–73 the release of CFMs increased from 5 million pounds to 666 million pounds and with a predicted lifetime of many years in the stratosphere the possibility of this pollutant producing a significant reduction in the already minute quantity of ozone is naturally a matter of serious concern.

It is to be emphasised that because of the complexity of the problems, coupled with the lack of precise information on several aspects of upper atmosphere chemistry and dynamics, it is difficult at the present time to reach positive unequivocal conclusions about these pollution hazards. However we do well to recognise the urgent need for vigilance and for continuing research.

4. Ionisation in the Upper Atmosphere

A feature in the upper atmosphere which varies both with height and with time in an interesting manner and a feature which has been intensively studied by ground-based and space-vehicle techniques is the density of ionisation – or more loosely speaking the "amount of electricity" in the atmosphere. At low levels (in the troposphere) the amount of ionisation present at any time is generally quite small. There are always some ions and electrons present in the troposphere at all levels but normally there are not enough to make their presence very obvious. Under certain weather and cloud conditions the concentration of electric charges becomes very large as evidenced by lightning discharges, but apart from these exceptional local and temporary conditions there are never large numbers of free electrons and ions present permanently in the lower atmosphere. Ionisation only becomes a regular and a very important feature of the atmosphere at altitudes above about 55 km. At and above this altitude solar ultraviolet and X-radiation produce and maintain a significant degree of ionisation in the atmosphere. This part of the atmosphere beginning at about 55 km and extending up to the very highest levels is termed the "ionosphere" a term which was first suggested in 1926 by the late Sir Robert Watson Watt – the inventor of "radar".

The story of radio waves and the ionosphere has an historic connection with a Friday evening discourse delivered here 85 years ago. In 1887 Hertz produced the first man-made electromagnetic waves and in the 1890s many scientists were experimenting with these newly discovered radio waves. On the 1st January 1894, when he was only 37, Hertz died and exactly six months later, on Friday 1st June 1894, Sir Oliver Lodge gave an historic lecture in this theatre. Under the title "The Work of Hertz" Lodge not only paid warm tribute to the scientific achievements of Hertz but he also demonstrated for the first time the possibility of using radio waves for communication purposes – in the lecture he showed how radio waves could be sent from one end of this room to the other across the intervening space. This lecture was widely reported and in due course printed so that the text became available to workers the world over. It is said that a copy of this lecture found its way to a young Italian called Marconi and although he knew very little of the science of electromagnetic waves he quickly realised the commercial possibilities of using radio waves for long-distance radio communication and especially the possibility of communicating with ships where cable communication could not be used. In the late 1890s young Marconi came to Britain and with the help of his mother's business friends in London he set up the Marconi Wireless Telegraphy Company (Marconi's mother was Irish – a daughter of the wealthy whiskey distiller Jameson). A few years later, in 1901, Marconi successfully transmitted a radio signal from Cornwall to Newfoundland and this successful demonstration of radio wave propagation over thousands of miles was an epoch making discovery and clearly showed the possibilities of radio waves for communication purposes. Some scientists of the time tried to explain the propagation in terms of diffraction around the curved earth but this was disputed by others and in 1902, just one year after Marconi's experiment, the correct suggestion was made by two workers, Kennelly in the USA and Heaviside in this country, *viz.*, that the radio waves were being reflected back and forth between the ground (or the sea) and some kind of mirror reflecting layer in the high atmosphere, and Kennelly and Heaviside suggested that this reflecting mirror was a "layer of electricity".

Although over the years the ionosphere has come to be associated with the subject of radio science it is interesting to note that the first suggestion for these electrically conducting layers in the high atmosphere came not from radio experiments but from studies of the earth's magnetic field – and was made some twenty years before the Kennelly–Heaviside hypothesis – indeed it was made even before radio waves had been discovered.

A magnetised needle mounted horizontally on a vertical pivot will set along the magnetic meridian and if observed carefully it is found not to remain pointing to the magnetic North but to oscillate slowly twice daily about a mean position. The oscillation is very tiny – a matter of minutes of arc – but nevertheless it is quite real and regular and it is a phenomenon which has been known a hundred years and more. Similar diurnal variations are found in all geomagnetic elements. The first suggestions for the existence of ionisation in the high atmosphere was put forward to explain this regular diurnal variation in the earth's magnetic field.

It was in 1882 that the Director of Kew Magnetic Observatory, Balfour Stewart, published his hypothesis that electric currents in a conducting layer high in the terrestrial atmosphere might explain the well-known regular variations in the magnetic elements. It is now well known that it is electrical currents in the high atmosphere (in the ionosphere) which produce most of the daily variations in the earth's magnetic field and over the years notable contributions to our knowledge of the electrical properties of the high atmosphere have come from geomagnetic studies.

However, the main interest of the ionosphere has been in the way in which it profoundly affects the propagation of radio waves. Since the ionosphere reflects radio waves we can in turn use radio waves to investigate the ionosphere. If we use waves of appropriate frequency then echo signals will be reflected back from the ionosphere and the study of the characteristics of these echo signals – delay times, amplitudes, states of polarisation *etc.*, – these features of the returned waves can provide very considerable information about the medium in which they have been refracted and hence about the atmosphere itself up to several hundred kilometres. Systematic radio-sounding from the ground may be said to have started some fifty years ago with the pioneer experiments of Appleton in 1924. He described some of his early results in a discourse to the Institution on 29th April 1927 and in the following two decades a network of some 150 ionospheric stations (or "ionosondes") was set up at sites spread over the globe. The equipment at these observatories consists of a pulse transmitter the frequency of which can be continuously varied from about 0.5 MHz (wavelength 150 m) to about 30 MHz (wavelength 10 m). The transmitter and receiver are kept in tune by electronic means so that we can obtain a photographic record of the levels of reflection of all these various frequencies. The type of record obtained is known as an "ionogram" and a sample of three such ionograms is shown in Fig. 9. A valuable feature of these ionograms is that they enable the distribution of electrons in the atmosphere between about 55 and 300 km to be monitored conveniently from the ground. In particular the maximum electron densities in the various ionospheric layers can be deduced easily and accurately since the magnitude of the critical penetration frequency of a layer is proportional to the square root of the maximum electron density of the layer. (The designation of the principal layers by the letters E, F1 and F2 was first suggested by Appleton). Apart from the obvious practical value for long-distance short-wave radio communication, radio sounding of the ionosphere has, over many decades, provided a wealth of scientific information on the physical properties of the high atmosphere, on the interaction between solar wave and particle radiation with the earth's atmosphere, on magnetic storms and auroral phenomena. Thus radio studies have clearly established that the E and F1 layers are produced by solar ultraviolet and X-radiation and the critical penetration frequencies of these layers (indicated in Fig. 9) provides a quantitative measure of the intensity of these solar radiations at levels of 100–180 km. Forty years ago radio sounding had established that the intensity of solar u.v.

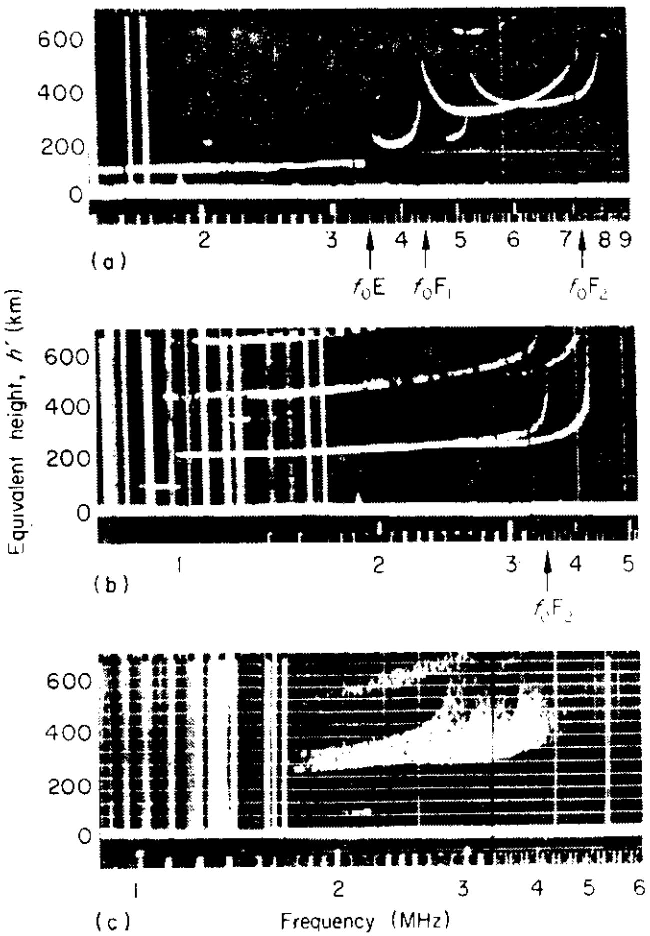

Fig. 9. Sample ionograms for Aberystwyth (51°N, 4°W) showing ordinary ray critical penetration frequencies, foE, foF1 and foF2. (a) Summer day near noon; (b) Equinox, early morning, 0500 h GMT; (c) Disturbed ionosphere.

radiation changes very markedly over the sunspot cycle. (By contrast the intensity of solar radiation in the visible spectrum reaching the ground shows no measurable variation at all over the solar cycle). Fig. 10 shows seasonal and sunspot cycle variations in the critical frequencies of the E, F1 and F2 regions for latitude 50°N over the period 1931 to 1946. These ionosondes have the advantage that echoes can be obtained with pulse transmitters of very modest power but two limitations of the technique are (a) since no echoes are received

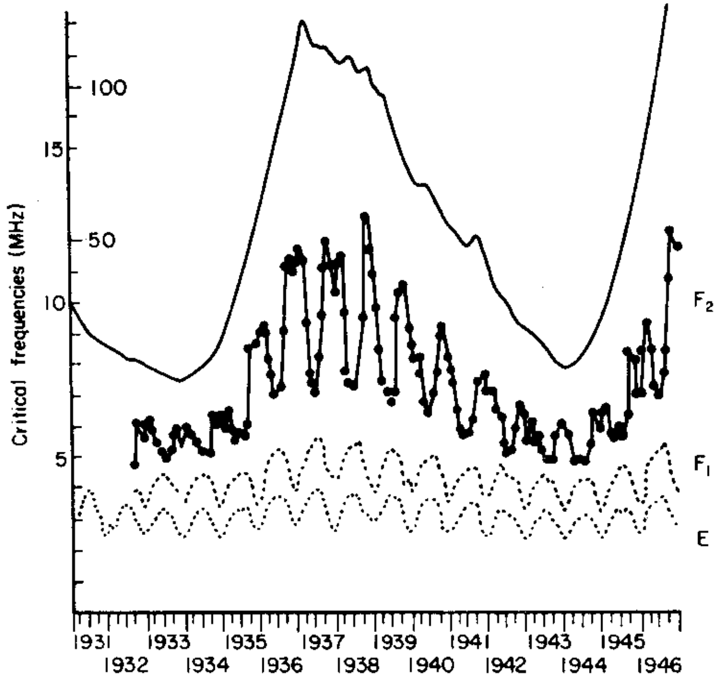

Fig. 10. Seasonal and sunspot cycle variations in the critical frequencies of the E, F1 and F2 regions for latitude 50°N. The upper continuous line shows the sunspot number.

back at frequencies above the critical frequency (the signals penetrate through the ionosphere and go off into space) the equipment cannot provide data on the "top-side" of the ionosphere, *i.e.*, on that part above the level of the peak electron density of the F2 layer (this level can be anything between about 200 and 350 km); (b) during the daytime at least the ionosonde will not give any reflection from the lowest part of the ionosphere (below about 80 km) because at these levels the comparatively high gas density results in the electrons making very frequent collisions with the neutral gas molecules, with the result that the energy of the incident radiowave is dissipated as heat and no echo signal results. (This daytime absorption of radio waves in the lower ionosphere is the cause of signals from medium-wave broadcast stations being very weak or not detectable at all at distant places until after sunset).

In 1949 some information on the top-side of the ionosphere was obtained using radar reflections from the moon and later rocket experiments provided isolated samples of data on electron density (and other parameters) to levels of 500 km and beyond. More recently top-side radio-sounding satellites and geostationary satellites have provided a wealth of data on the ionosphere above 300 km and upwards to thousands of kilometres. Small rockets have also been used to study the lower end of the ionosphere (the so-called D-region). In a lecture at the Royal Institution it is of especial interest to recall that in many of these rocket and satellite radio sounding experiments the technique used to measure electron density makes direct use of one of Faraday's most famous and important discoveries – the Faraday magneto-optical effect. Quite early in his career Faraday was convinced that it should be possible to detect some effect of an electric field on light and over a period of twenty-three years he returned five or six times to search for such an effect. In 1845 he returned to the problem yet again – mainly as a result of some correspondence he had with the young Professor

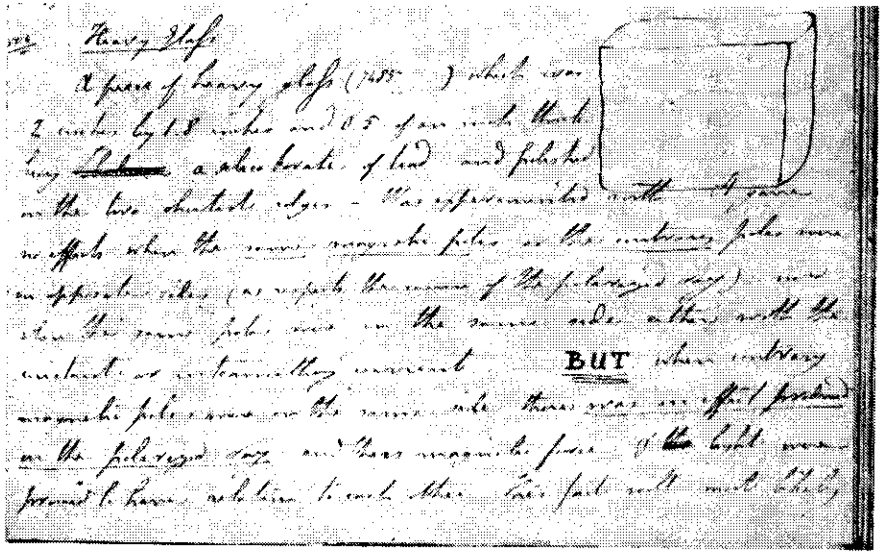

Fig. 11. Extract from Faraday's diary for 13th September 1845 – recording his discovery of a magneto-optical effect.

William Thompson (later Lord Kelvin). Again repeated attempts came to nothing and in fact another 25 years were to go by before Kerr established the electro-optical effect that had eluded Faraday for so long. However on 13 September 1845 Faraday finally had some success – not with an electric field but with a magnetic field. A copy of the relevant paragraph in Faraday's diary recording this successful conclusion to his 23-year search is shown in Fig. 11. His prophetic words about the possible importance of this discovery are to be noted. To-day, 134 years later, this same Faraday magnetic-optical effect (with radio waves in place of light waves, the earth's magnetic field replacing the laboratory electro-magnet and the ionosphere as the medium in place of Faraday's piece of

heavy glass), is extensively used in both rocket and satellite radio studies of the ionosphere.

In the past two decades a number of "top-side" radio-sounding satellites have been launched and one of the most successful has been the Canadian satellite Alouette I. Launched into a circular orbit about 1000 km above the earth this

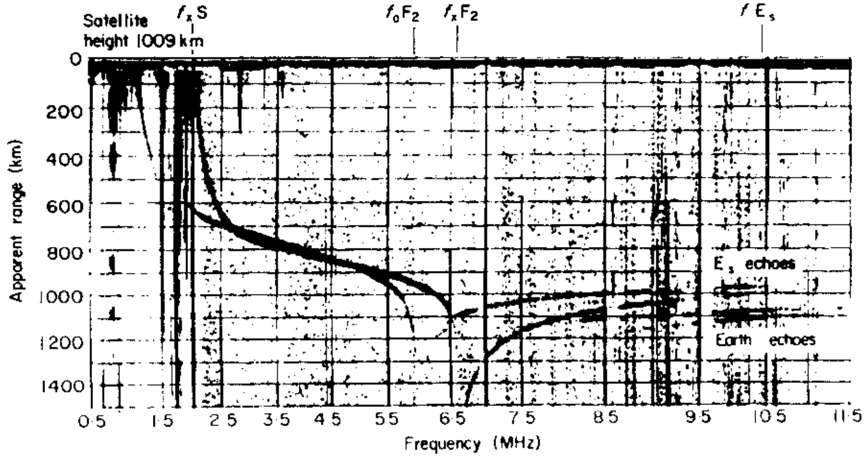

Fig. 12. Sample top-side ionogram. Satellite Alouette I 1652 GMT 29th September 1968. (Echo signals at frequencies above 6.5 MHz are produced by reflection from ground after penetration of ionosphere from above).

satellite carried a complete miniature version of the ground-based ionosonde and in its lifetime it produced several million high quality top-side ionograms. A sample top-side ionogram is shown in Fig. 12. These top-side sounding satellites have extended the range of the radio-sounding technique up to levels of a thousand kilometres and more.

4.1. *Incoherent scatter sounding*

In recent years a most remarkable, and powerful, new approach to radio sounding of the upper atmosphere from the ground has been developed – it is known as the "incoherent scatter" technique.

In 1906 J.J. Thompson showed that free electrons are capable of scattering electromagnetic radiation and for direct back scatter a single electron has a cross section of about $10^{-28} m^2$. This is a very small scattering area and in conventional radio sounding of the ionosphere it was hardly to be expected that there would be much chance of detecting scattering from individual electrons, nor indeed from very large numbers of electrons way up in the ionosphere. However, recent developments of very powerful radar detection systems have brought within

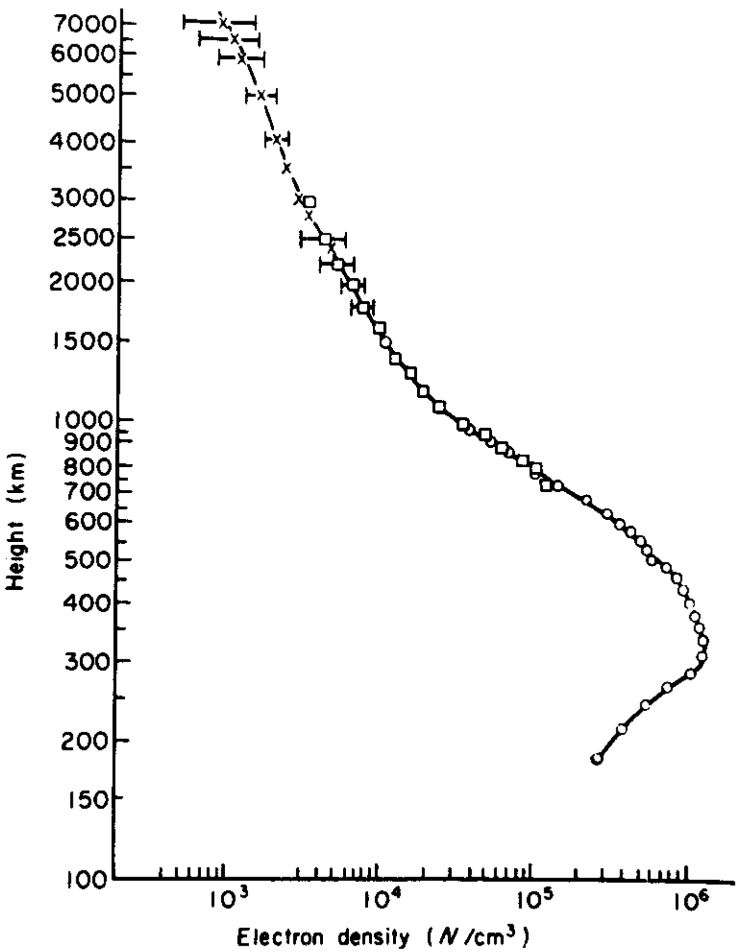

Fig. 13. Electron density profile over height range 200–7000 km obtained with incoherent scatter sounding equipment. Lima, Peru 25th April 1962, 1400 EST. (After K.L. Bowles).

range the possibility of detecting such scattering at the ground. Because of the random nature of the motions of the electrons the scattered signals have more or less random phases and this scatter is termed "incoherent scatter". With this new technique it has proved possible to measure electron densities not merely to two or three hundred km but out to 7,000 km and more. (Fig. 13). However, this remarkable upward extension of the electron density profile is not the most

striking feature of the incoherent scatter technique. Due to the fact that in the ionosphere the movement of the electrons is influenced by that of the ions it is found that the scattering process is rather more complicated than if the effect of the ions were not there and the result is that the technique provides information not only on electron density but also on electron temperature, ion temperature, ion mass, plasma drift velocity, electric field strength and many other parameters. Fig. 14 shows an example of electron temperature measurements for the height range 150 to 650 km for a 24-hour period in winter made by the incoherent scatter technique. The sharp changes in electron temperature associated with sunrise and

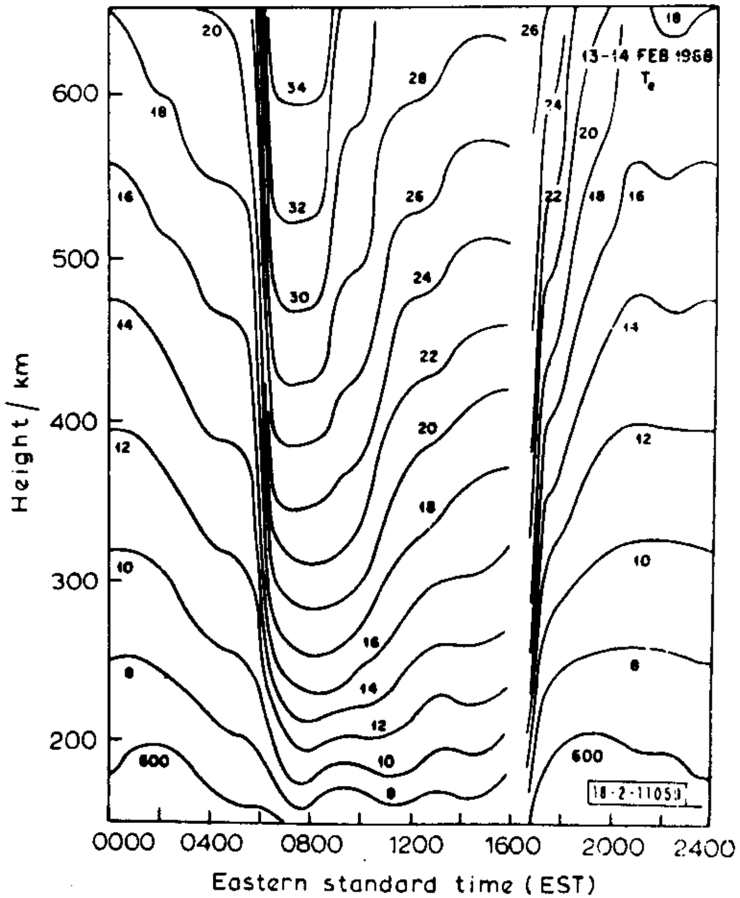

Fig. 14. Diurnal variation of electron temperature measured with incoherent scatter sounding technique. Millstone Hill (42.6°N 71.5°W) 13–14th February 1968. (After J.V. Evans).

sunset are clearly shown. Just as with Faraday's 1845 discovery of magnetic rotation, so too J.J. Thompson's 1906 discovery of electron scattering has, 70 years later, found a very important large-scale application in this new ground-based radio technique for studying the high atmosphere out to many thousands of kilometres.

4.2. *Movements in the ionosphere*

For many years radio studies of the ionosphere concentrated on determining the vertical structure of the various ionised layers, on studying the diurnal, seasonal, sunspot cycle, latitudinal and longitudinal variations in that structure and with the application of these studies to practical radio communication problems. Moreover in recent years much attention has been given to studies of the dynamics of the ionosphere and a number of different techniques have been developed for this purpose.

One such ground-based technique depends on the principle that if, for any reason, the level of reflection of a radio signal in the ionosphere changes, then for the period when the level is actually changing we may expect to observe a Döppler shift in the frequency of the received signal. Such changes in the level of reflection may result from one of several possible causes. Winds in the neutral atmosphere at atmospheric levels, or temperature changes consequent upon ionospheric or magnetic storms, can cause movements in the reflecting layer. Sudden changes in the intensity or character of solar ultraviolet and X-radiation, such as occur during solar flares, can produce marked temporary changes in the electron density/height profile which in turn may give rise to rapid changes in the level of reflection of a radio signal. Disturbances in the lower atmosphere (in the troposphere) due, for example, to natural or man-made explosions can generate wave motions which can be transmitted upwards to the ionosphere. In order to study the full three-dimensional character of ionospheric movements this Döppler technique employs three transmitters spaced some tens of kilometres or more apart with the reflected signals being simultaneously received at a single

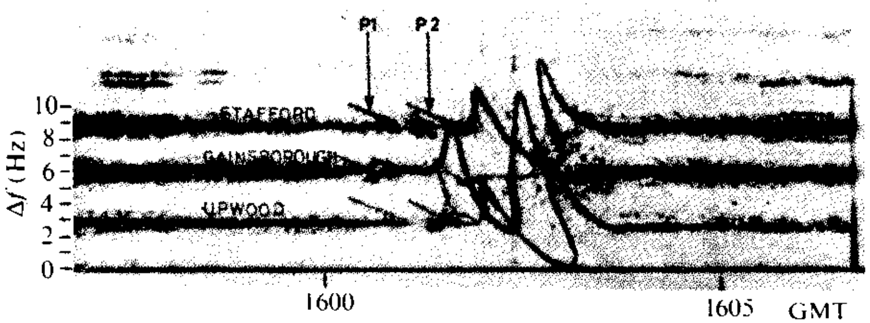

Fig. 15. Döppler frequency shifts showing large-scale disturbances in the ionosphere following a major explosion near the ground. (After T.B. Jones).

site. Fig. 15 shows an example of Döppler frequency shifts observed in radio signals reflected from the ionosphere following a man-made explosion at ground level. Simultaneous observations at three spaced ground stations of amplitude changes in radio signals reflected from the ionosphere have also been used to measure the horizontal movements near the reflecting level.

4.2.1. *Meteor trails*

In the height range 80–100 km movements in the ionised trails of incident meteors observed both visually (in the case of very bright meteors) and by radar, have provided a wealth of data on winds at this level. Diurnal, semi-diurnal and prevailing-wind components with magnitudes in the range 5–40 ms^{-1} have been observed and the amplitudes and phases have been interpreted in terms of solar and lunar tide oscillations in the atmosphere. Fig. 16 shows an example of a semi-diurnal oscillation in meteor wind data.

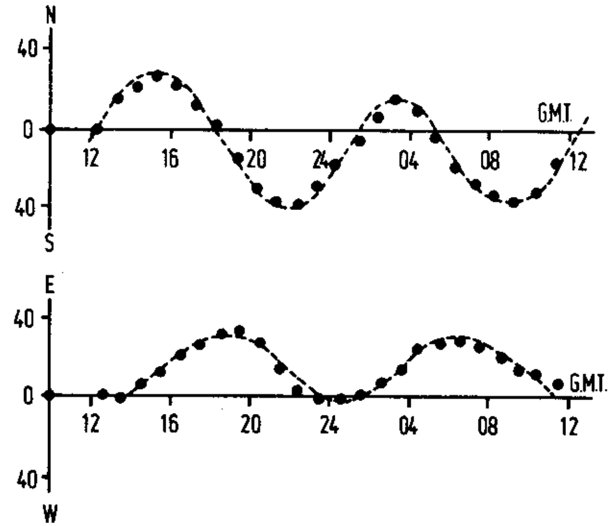

***Fig. 16.** Semi-diurnal wind components (altitude ca. 100 km) measured by the radar-meteor technique September 1953. Ordinates: ms^{-1} (After J.S. Greenhow and E.L. Neufeld).*

4.2.2. *Chemiluminescent trails*

In recent years some interesting rocket experiments which reveal the detailed and different character of ion and neutral gas movements in the ionosphere have been carried out using chemical contaminants such as sodium, barium, strontium and trimethylaluminium (TMA). When released in the ionosphere these contaminants react strongly with selected atmospheric constituents to produce luminous trails the movement of which can be studied from the ground. These experiments have served to underline the fact that the movements of the neutral and the ionised constituents in the high atmosphere can be quite different – the motion of ions being profoundly influenced by the magnetic field. Thus in one experiment it was observed that near 200 km the neutral cloud associated with the ejected contaminant was spherical but that the ion cloud was markedly elongated along the direction of the earth's magnetic field. It was also noted that whereas the component of motion of the neutral gas and the ions along the magnetic field directions were approximately the same, the velocity components across the field direction were very significantly different, that of the neutral gas being very much larger than that of the ions.

4.2.3. *Thermospheric winds*

In the thermosphere the large daily variation in temperature gives rise to horizontal pressure gradients which in turn result in winds with speeds of several hundred metres per second. A type of motion commonly observed at F-region heights are the so-called "travelling ionosphere disturbances". The wavefronts of these disturbances, which take the form of quite large perturbations in the electron density of the region, often have a lateral extent of several hundred kilometres, move at speeds of 300 to 600 kilometres an hour and can be tracked by ground-based radio techniques over distances of many thousand kilometres. These T.I.Ds are believed to be a form of atmospheric gravity wave resulting either from

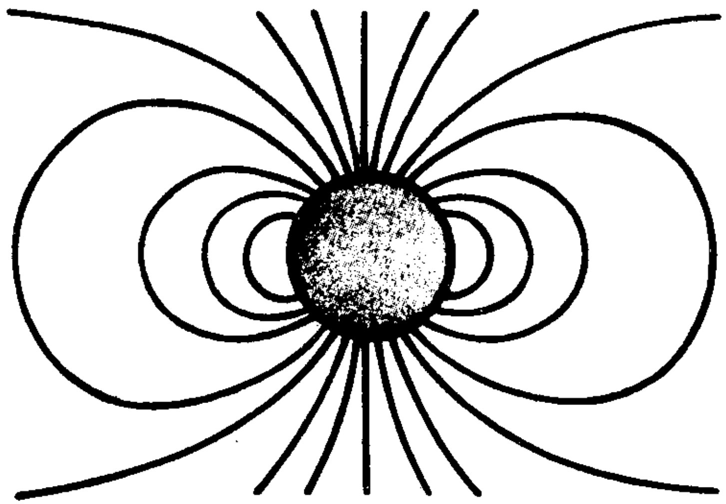

Fig. 17. Magnetic field lines surrounding a uniformly magnetised sphere.

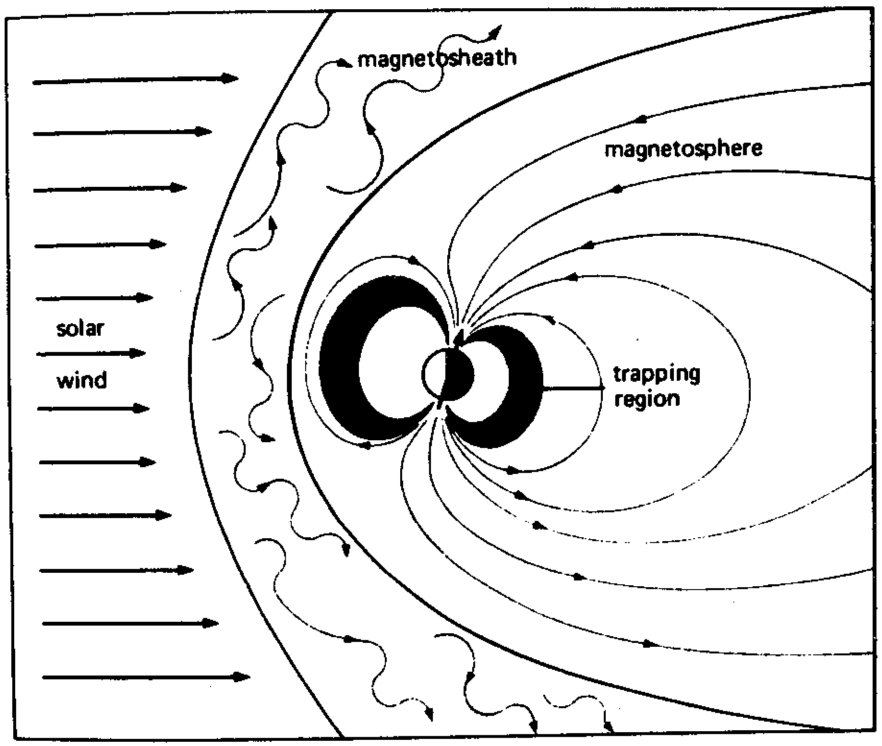

Fig. 18. Magnetic field lines of the earth distorted by the solar wind. (After J.A. Ratcliffe).

the breakdown of tidal movements or waves which have their origin in disturbances at tropospheric levels and which penetrate upwards to the ionosphere.

5. The Outer Limit of the Earth's Atmosphere

In discussing the changes in the earth's atmosphere as we move upwards, the pertinent question again arises – "Where does the earth's atmosphere end?" Until about 25 years ago, that is until we had artificial satellites and deep space probes, we used to believe that the space between the earth and the sun was a complete and perfect vacuum. Furthermore, as far as the magnetic field of the earth was concerned we thought that outside the earth the magnetic field was of the form that one would expect if the earth were just a large magnet (Fig. 17). However, soon after we had artificial satellites this simple model of the earth's magnetic field had to be very substantially modified and we now know that things are much more complicated. The earth is completely enveloped in the so-called solar wind, that is, the unending stream of protons and electrons thrown outwards from the sun at velocities of hundreds and sometimes of thousands of kilometres per second, and our picture now of the space outside the earth is represented in Fig. 18. The sun is away to the left of the diagram and the solar wind moving forward encounters the outer limits of the earth's atmosphere and of the earth's magnetic field. It is moving at supersonic speeds relative to the earth's atmosphere and consequently has the usual shock front which one associates with bodies moving at supersonic velocities and the magnetic field lines on the sunward side of the earth are compressed but on the dark side they are spread out, drawn out, to several earth radii. On the earth side of the shock wave there is a turbulent region in which the particles of the solar wind are shown moving irregularly in many directions and inside this we have another boundary line which is termed the "magnetopause". The magnetopause has a long ellipsoidal shape and the volume enclosed inside it is known as the "magnetosphere". In the magnetosphere the dynamics of the earth's atmosphere, which at these levels consists entirely of protons and electrons, is controlled by the earth's magnetic field. Outside the magnetopause the solar wind is in control – inside the magnetopause the earth's magnetic field controls the movements of the charged particles. Hence the magnetopause can be said to define the outer limit of the earth's atmosphere. This level is about 5 to 10 earth radii, that is about 60,000 km, on the day side of the earth but it is 50 or 60 earth radii out on the night side, *i.e.*, 300,000–400,000 km. The distance of the moon from the earth is about 384,000 km so that the night-time tail of the earth's atmosphere extends outwards to and possibly just beyond the moon. Of course the atmosphere at these extreme limits is a very tenuous ionised atmosphere with little physical resemblance to the atmosphere near the earth's surface.

In Fig. 18 two black segments or sectors are marked "trapping regions". In general most of the particles in the solar wind stream past the earth – the paths of those which come nearest to the earth obstacle are seriously distorted by the earth but most of them get away. However, some do get trapped by the earth's magnetic field in these black sectors and an interesting feature of these trapped particles is that they actually bounce back and fore between the Northern and Southern hemispheres of the earth. It is well known that the effect of a magnetic field on an electrically charged particle is to cause it to spiral or precess around the magnetic field line. At these levels in the atmosphere, some thousands of kilometres up, electrons spiral very rapidly with radii of a few hundred metres or so whereas ions which are much heavier spiral much more slowly and with radii of the order of perhaps 10 km. Now as we move from low to high latitudes the magnetic field lines crowd together and the strength of the magnetic field increases from low to high latitudes. The net effect is that the "pitch angle" of the spiral varies as shown in Fig. 19. The effect is similar to that when a spiral spring is pushed against an obstacle. When the spiral is at right angles to the magnetic field line then the particles may start going back along the field line direction. In this way trapped particles (electrons and protons) spiral rapidly from one hemisphere to the other.

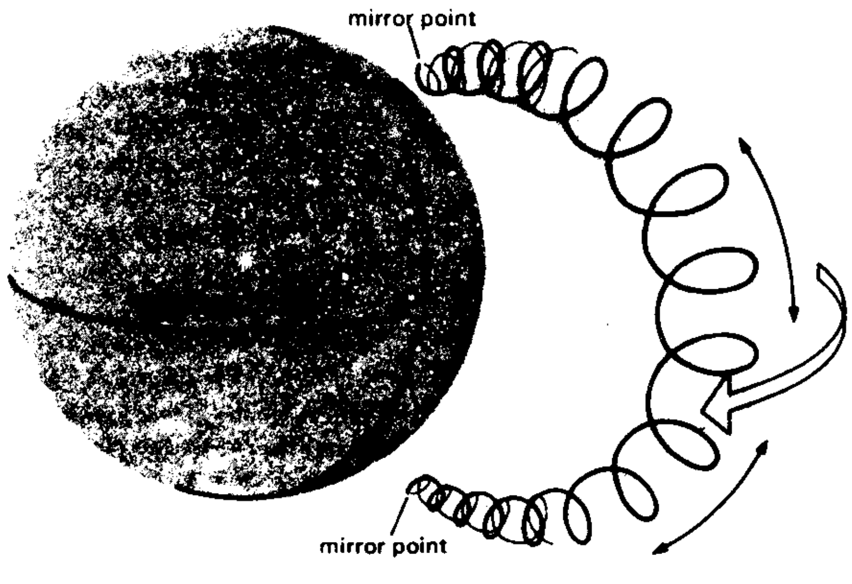

Fig. 19. Trapped particles spiral between mirror points in the two hemispheres. (After J.A. Ratcliffe).

It is found that electrons take some seconds to bounce from one hemisphere to the other and protons rather longer. It is known that many phenomena which occur in the very high atmosphere are closely linked from one hemisphere to the other by the magnetic field lines. Locations in the two hemispheres which are linked in this way are referred to as conjugate points. There is now a considerable body of experimental evidence to support this geophysical linkage between the two hemispheres along the magnetic field. Thus the occurrence of the aurora at northern and southern high latitudes is strongly correlated. Certain parameters of the ionosphere which show clear changes at sunrise or sunset, such as the electron temperature, can be observed at the conjugate point in the opposite hemisphere. Man-made changes in the ionosphere at one location such as the injection of energetic electrons have been detected at the magnetically conjugate point in the other hemisphere. A striking illustration of a magnetically conjugate phenomenon is the so-called "whistling atmospheric" or "whistler". First noticed by users of long telephone lines towards the end of the last century these "whistlers" were frequently observed again during World War 1. In those days radio was still in its infancy and forces on the two sides of the front line used to lay down wires on the ground directed towards the enemy lines in the hope of picking up some information from the other side. In the course of this work some of the observers noticed peculiar whistling sounds – long audible notes of descending pitch. The origin of these "whistlers" remained a mystery until the 1950's when they were

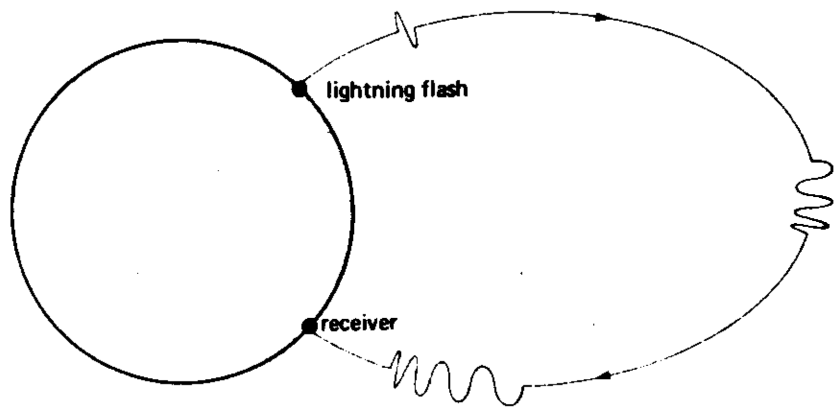

Fig. 20. Production of a "whistler" (After J.A. Ratcliffe).

identified as being very long-wave radio signals ("atmospherics") generated by lightning flashes occurring not locally but in the opposite hemisphere of the earth – in fact at the conjugate point at the other end of the magnetic field line (Fig. 20). These whistling atmospherics have now been studied very extensively and we know that they are produced by very long wavelength radio-waves radiated by lightning flashes which propagate up through the atmosphere into the magnetosphere, travel along field lines out to several earth radii and come down to earth again. The radio signal radiated from a lightning flash contains a wide band of frequencies and during propagation through the ionosphere along the magnetic field line the higher frequencies travel faster than the low frequencies. The higher frequency components of the signal thus arrive earlier at the receiver and this gives the signal its characteristic whistling sound of decreasing pitch. Whistlers can be reflected back and fore several times between the two hemispheres and studies of whistlers are now providing quite valuable information on the earth's atmosphere at altitudes of several earth radii.

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