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AUTUMN COLLEGE
ON
THE TROPOSPHERE, STRATOSPHERE AND MESOSPHERE
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NOTES FOR THREE INTRODUCTORY LECTURES

W.J. GRANVILLE BEYNON
Department of Physics
University College of Wales
Penglais
Aberystwyth SY23 3BZ
Wales, U.K.

1984 Autumn Course on 'Troposphere, Stratosphere, Mesosphere'

Sir Granville Beynon

Outline Plan of Course

Three introductory lectures* will provide a summary of the main physical properties chemical composition, thermal structure, electrical properties etc., of the atmosphere from ground level to about 120 km - the level at which the composition of the atmosphere begins to change significantly. In Section 2, a group of 12 lectures deal with measuring techniques & also describe some measurements carried out with these various experimental methods. The techniques to be covered include ground-based methods (including optical, laser, radar etc.), balloons, radio-sonde, and satellite techniques. Section 3, totalling some 18 lectures, is devoted to the subject "Radio propagation in the Lower Atmosphere". This topic is of importance not only to radio engineers, but is also one of concern in environmental science, earth resources, pollution etc., Section 4 deals with the dynamics of the lower atmosphere, mass motions, tides, waves. Another major aspect of weather & climate - 'Numerical Modelling' - will be covered in Section 5. The basic principles of numerical weather prediction, general circulation models of the atmosphere, special features of the tropics & equatorial regions, monsoonal circulations will be included. Section 6 covers the detailed composition & complex chemistry of the atmosphere including the role of minor constituents. The 'Middle Atmosphere' (that part of the atmosphere between about 10-15 km to 100 km) has recently been singled out for a coordinated world-wide study and at the present time, and continuing well into the 1980's, scientists from many countries are participating in a cooperative study of problems of the Middle Atmosphere. This is the so-called 'HAP' project and it will be discussed in Section 7. Section 8 covers the important subject of pollution of the atmosphere - photochemical smog, the ozone & CO₂ problems, acid rain and the effects of global pollution. Finally in Section 9 we shall deal with atmospheric electricity & tropical thunderstorms.

*Briefly summarised below.

Physical Properties of the Atmosphere

If we apply the equation of state $pV = Nkt$ and the barometric equation $dp = -\rho \cdot dh$ we derive the pressure/height or density/height profile -

$$\rho = \rho_0 \exp \left(-\frac{(h-h_0)}{H} \right) \quad \text{where } H = kT/mg \text{ is the}$$

'Scale Height'. From the p/h (or ρ/h) expression the Scale Height can be defined as the height change in which the pressure (or density) changes by a factor of 'e'. Alternatively the scale height can be defined as the 'height of the homogeneous atmosphere' or the height the atmosphere would have if its density & pressure were uniform - if this was the case, then only some 8 km of atmosphere would be necessary to account for the observed pressure at the ground i.e. the 'scale height' of the atmosphere near ground level is about 8 km.

We note that $H \propto T/m$ or $H \propto T$ if m is fixed. H can be measured in various ways and if m is known, then H gives a measure of the temperature T .

Chemical Composition of the Atmosphere (Mean molecular mass)

The composition of the atmosphere is the result of chemical processes which took place after the formation of the earth & there may be a long-term variation in this composition though none seems to have been discovered yet.

Direct sampling of the lower atmosphere (the troposphere) shows that the four constituents N_2 , O_2 , A and CO_2 between them make up all but 0.003% of the atmosphere by volume. There are also minute fractions of inert gases & 'minor constituents' - the latter are often variable in amount, depending on local conditions. (see Table 1). The He content of the atmosphere is being continually replenished by an outward flow from radioactive sources in the solid earth & the small observed concentration may be ascribed to its escape at the top of the atmosphere (the 'exosphere') at which high temperatures exist. Fifty years ago many balloon experiments were carried out to sample the composition of the atmosphere up to altitudes of about 30 km but no conclusive evidence was found for any change from that observed in the troposphere. The temperature in the troposphere decreases upwards and this, together with other factors, results in continual mixing & hence we would expect that the composition would be maintained at least up to the tropopause. In the lower stratosphere the temperature is essentially constant and

one might anticipate diffusive separation to occur -with the lighter gases on top. In fact theoretical work by Chapman & Milne in 1923 suggested that this was to be expected and furthermore, there was strong indirect evidence from studies of the absorption of sunlight, of emission spectra of the aurora & of the night sky, indicating that up to 100 km at least, the dominant constituents of the atmosphere are still N_2 and O_2 as they are in the troposphere.

In the period 1920-30 & even in the 1950's there was still interest in this question as to whether the upper atmosphere consisted of light gases. However many experiments conducted first with balloons, and later with rockets, provided no evidence for diffusive separation with the lighter gases on top.

We now know that the composition of the atmosphere only begins to change significantly at about 90-100 km. Above this level the energy of the ionising solar radiation is sufficient to ensure large-scale dissociation of O_2 . Rocket experiments, made some 25 years ago by Friedmann in the USA, showed that as the rocket ascended between 90 & 140 km the intensity of solar radiations in the 142.5 - 150 nm wavelength range, increased rapidly & that heavy absorption of these radiations ^{was} ~~were~~ giving rise to the transition from O_2 to O between these heights.

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, and I may briefly refer to the measurement of two such minor constituents which serve to illustrate two widely different techniques which have recently been employed.

Atomic oxygen in the upper atmosphere is an important minor constituent because of the key role which it plays in the neutral & 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 certain airglow emissions: it plays an important role in the thermal balance of the lower thermosphere & the vertical distribution of atomic oxygen is relevant to considerations of vertical transport by eddy & molecular diffusion. Hence

for these, & other reasons, it has become important to measure atomic oxygen concentrations in the height range 60 to 120 km & in recent years this has been successfully accomplished with a rocket-borne experiment. A specially constructed lamp provides an intense beam of ultra-violet radiation at the wavelengths of the oxygen triplet near 130 nm & is used in the measurement of both the absorption & resonance fluorescence of atomic oxygen over the height range 60 - 120 km. Atomic oxygen concentrations obtained with this type of rocket payload are shown in Fig.2.

Another, & in some ways surprising, minor constituent in the upper atmosphere, ^{from the high atmosphere} is sodium. The intensity of these sodium emissions/remains fairly constant throughout the hours of darkness, but during morning & evening twilight periods, it is enhanced by a factor of 50 to 100. Observations of the times at which this 'twilight flash' disappears have long been used to determine the height of the emitting layer & it is now known that the sodium atoms are contained between about 80 & 150 km, with a maximum concentration around 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. It may be noted that in addition to sodium, other metallic atoms ablated from meteors, & deposited around this level, are potassium, magnesium & 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. The sensitivity of this ground-based technique is emphasised by 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. Thus remote sounding techniques using infra-red 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, then these satellite infra-red emission measurements can be used to give the distribution of the emitting constituent & the distribution of ozone & water-vapour have been measured by this technique.

Fig.3 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.

Table 2 gives values of the mean molecular mass at various altitudes from ground level to 800 km. It will be seen that no significant change in mean molecular mass is observed in the first 90 km and thereafter it gradually decreases upwards. With increasing height, above 90 km diffusion gradually dominates over mixing tendencies and the lighter gases predominate more & more, until at very great heights atomic hydrogen, & finally ionised hydrogen, becomes the sole constituent. Fig. 4 is a rough block diagram representation of the change in composition up to about 4000 km.

Temperature in the atmosphere

Long been known from balloon flights that in the first 10 - 15 km temperature falls almost linearly at the rate of about 1°C in 165 m. (A lapse rate of this order was deduced in 1793 by Dalton from a study of temperatures at the tops of mountains - his figure was 1°C in 166 m.) The main atmospheric constituents in the lower atmosphere are essentially transparent to radiation in the visible and near infra-red, and consequently the troposphere is not heated very much by direct absorption - the energy of the incident radiation is taken up by the surface of the earth which then re-radiates energy at much longer wavelengths, & it is absorption of this long wavelength radiation (between 7000 & 15000 nm) by gases like H_2O , CO_2 and O_3 which heats the troposphere. ('Greenhouse effect'). The principal heat source for the troposphere is thus at the ground and the temperature distribution in this part of the atmosphere is determined by convection. As we move upwards the density of the atmosphere falls off, and so does the abundance of these absorbing constituents, and at altitudes of the order of 10-12 km a limit is reached beyond which convection ceases to dominate and the temperature is determined more by radiation considerations. At levels immediately above the tropopause, conditions are more or less isothermal with the temperature constant at some value of about 60° less than that at the ground. However at 30 km or so,

these isothermal conditions are disturbed by the selective absorption of certain solar wavelengths by O_3 and this produces a very large temperature 'bulge' between 30 & 60 km. The general heat balance in this part of the atmosphere is the result of a number of factors including:

- (i) absorption of solar radiation by O_3
- (ii) heating of the atmosphere by long-wave radiation from the earth's surface
- (iii) heating produced by radiation from O_3 .

At night, some cooling of the region by radiation upwards & downwards will be expected, but this is nothing like large enough to dissipate all the heat energy accumulated during the day and this temperature maximum centred around 50 km is pretty well maintained throughout the night. ('Stratopause').

Above this level the effect of O_3 absorption decreases and the temperature falls again to a marked minimum of 173°K at 85 km ('mesopause'). Some of the earliest evidence for this minimum was based on the observation of 'noctilucent clouds' - the estimated temperature required at this level for the production of ice-crystals was about $160-200^{\circ}\text{K}$. Confirmation of this low temperature at this altitude came in 1939 from experiments using long (18 km) radio waves. These long-radio-waves are reflected from ionisation at this level and these radio measurements estimated that the scale height of the atmosphere at this level was about 25% less than that at the ground, indicating a temperature decrease of this order. Tidal theory also indicated the need for a low temperature at this level. Subsequently rocket measurements provided further confirmation of this minimum. Rocket experiments have provided good data on temperature in the high atmosphere. In some cases temperature has been deduced from measurements of pressure or density, but in other important rocket experiments the temperature has been deduced in a more direct manner. Thus in the so-called 'grenade experiment' temperature has been deduced from velocity of sound measurements made over different height ranges as the rocket ascends.

A summary Table of T for the international reference atmosphere is given in Table 3.

The rapid height gradient of temperature at, & just above the 100 km level, results from the very heavy absorption of solar u.v. radiation which takes place here. Above the 85 km level all u.v. radiation of wavelength less than about 1750 Å is gradually absorbed & an important fraction of the energy so obtained goes into heating the atmosphere.

The thermal balance of the atmosphere above 100 km (in the so-called 'thermosphere', is a complicated process involving many different factors & it is convenient to consider separately the lower thermosphere (100 to 250 km), where there is a large positive gradient of temperature, and the upper thermosphere where there is practically no gradient.

In the lower thermosphere, the principal source of heat arises from the absorption of solar uv & X-radiation which ionise or dissociate a good proportion of the atoms & 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 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 deposited initially, 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 photodissociation of molecular oxygen by solar u.v. radiation in the wavelength range 110 to 130 nm.

Above about 300 - 1000 km the atmosphere consists mainly of ionised helium, ionised hydrogen & 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 thermosphere at these levels 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 (at great altitudes) with solar activity. The really large variations occur above about 150 km. Fig. 5 shows temperature up to 500 km for latitude 30°, with a mean curve & extreme variations which can sometimes occur.

Atmospheric pollution

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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-di-oxide and ozone.

The exchange of carbon-di-oxide between the atmosphere and the oceans and the terrestrial biosphere (the 'carbon cycle') is of course critical to life. Carbon-di-oxide in the atmosphere also plays an important role in determining the rate at which the earth loses energy and an increase in CO₂ 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-di-oxide, and there is now a general consensus that the amount of CO₂ 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 was maintained, then the CO₂ content of the atmosphere would increase ten-fold by the year 2100. However interaction between the atmosphere and the oceans rapidly removes CO₂ 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 CO₂ in the atmosphere (and the oceans) needs to be carefully watched.

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 ultra-violet 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, I think, to be emphasised that because of the complexity of the problem, coupled with the lack of precise information on several aspects of upper atmosphere chemistry & 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 & for continuing research. Of course the scale of these pollution problems of the atmosphere - the CO_2 & O_3 problems, acid-rain etc., - these will be as nothing to the atmospheric pollution (or perhaps one should say atmospheric devastation) likely to follow a nuclear conflict. This is a 'pollution problem' which one might say is almost unthinkable, but as the 'pollution problem in extremis' it can never be excluded from our thinking.

Ionisation in the Atmosphere

The study of ionisation in the atmosphere falls into two broad areas. In the lower atmosphere (at tropospheric levels) the amount of ionisation present at any time is generally quite small and it is only under certain weather & cloud conditions do the concentration of electric charges become very large - as evidenced by lightning discharges. The subject of 'atmospheric electricity' will be covered in a special group of lectures in this course & I do not think I need consider it further in this introductory lecture.

The second broad field of ionisation in the atmosphere refers to ionisation at mesospheric & thermospheric heights. At altitudes above about 55 km ionisation becomes a regular & very important feature - at, and above this altitude, a number of different sources - solar u.v. & X-radiations, charged particles from the sun (the 'solar wind'), meteors & cosmic rays all contribute to produce, & maintain, a significant degree of ionisation in the atmosphere. This is the so-called 'ionosphere' - a term first suggested by the late Sir Robert Watson-Watt (the inventor of radar) in 1926.

The only part of the ionosphere which may be said to lie within the height range covered by this course is the lower ionosphere - extending from about 55 km to 110 km. The most important single parameter of the ionosphere is the electron density and in the past 50-60 years, the distribution of electron density with height has been

intensively studied, chiefly by radio sounding techniques. It is not difficult to show that for a radio-wave incident normally (i.e. vertically) on the ionosphere the electron density N required to reflect a radio wave of frequency f is approximately given by;

$$N = (4\pi^2 \epsilon_0 m / e^2) \cdot f^2$$

where ϵ_0 is the permittivity of free space
 $(8.85 \times 10^{-12} \text{ N}^{-1} \text{ m}^{-2} \text{ C}^2)$
 e is the electronic charge
 $1.60 \times 10^{-19} \text{ C}$
 m the electron rest mass
 $9.11 \times 10^{-31} \text{ kg}$

If we insert numerical values we get ;

$$N = 1.24 \times 10^{-3} \cdot f^2 \quad \text{where } f \text{ is in Hz \& } N \text{ in electrons/cc}$$

$$\text{Thus } N \propto f^2$$

Hence from this simple expression, we can infer that lower radio frequencies will be reflected at lower values of N i.e. low frequency radio waves will be reflected from the lower end of the ionosphere & higher frequencies from levels higher up, where N is larger.

Experiment shows that the ionosphere consists of a series of 'layers' of ionisation in which the electron density generally increases upwards. The layers are designated by the letters D, E & F and are produced at levels of $< 90 \text{ km}$, $90-110 \text{ km}$ and about $180-220 \text{ km}$ and with peak electron densities of the order of 10^2-10^3 , 10^4-10^5 & $5 \times 10^5 - 10^6$, respectively. These are of course, only order of magnitude values, since the actual electron densities, at all levels of the ionosphere, will vary considerably with time of day, season & with the sunspot cycle, magnetic activity etc., The actual situation is also often more complicated than is indicated by a simple D, E, F subdivision - all the various regions or layers show different degrees of stratification, at different times.

Routine radio sounding of the ionosphere has now been established for more than fifty years and need not be described here. The small group of three or four stations which started these soundings in the early 1930's, developed in the I.G.Y. of 1957-58 into a world-wide network of stations, and although the number of stations currently operating is less than the IGY number (170), there are probably well over 120 which make routine soundings.

Incoherent Scatter Technique

In the past twenty years a very powerful radio technique for probing the ionosphere has been developed - this is the so-called 'Incoherent Scatter Technique'. In this, very strong radio signals (both pulse & CW signals can be used) radiate the ionosphere & give rise to weak scattered signals, and measurements of the amplitudes, frequency spectrum etc., of these weak scattered signals, enables not only the electron density to be measured over a great range of heights, but also many other ionospheric parameters, such as electron temperature, ion temperature, plasma velocities etc., Incoherent scatter installations are very powerful tools, but they are also very expensive, and there are only a few such installations in operation.

Originally these incoherent scatter radars were developed for the study of the ionosphere to great heights, but it has been shown that they can, with suitable technical modifications be adapted to study the atmosphere at mesospheric heights.

The production of an ionised layer in the upper atmosphere

The basic theory of the production of an ionised layer in the upper atmosphere by solar radiation was first given in 1931 by Sydney Chapman and an outline of this theory is given below. (Appendix 1).

Radio-sounding of the ionosphere readily yields information on the variation of electron density with height and with the altitude of the sun, and these experimental data can readily be compared with those predicted by the Chapman theory. The main result is given in expression (8) which gives the rate of electron production as a function of solar zenith angle χ and height h (measured in terms of scale-height H above & below h_0 the level of maximum production for overhead sun). From the plot of I/I_0 for different values of χ we conclude;

(i) the rate of electron production falls off with increasing χ i.e. for an ionised region of this kind (a so-called 'Chapman region') the electron production will be a maximum at noon & decrease symmetrically about noon. It will be larger in summer than in winter & will decrease with increasing latitude.

(ii) The level of maximum electron production increases with increasing χ i.e. it will be lowest at noon, lower in summer than in winter and lower at

low-latitude stations than at high-latitude stations.

(iii) The rate of electron production is not symmetrical about the level of maximum production - there is a sharper gradient on the lower side than on the upper.

The theory outlined above makes at least five simplifying assumptions :

- (a) the ionising radiation is monochromatic
- (b) the atmospheric density varies exponentially with height
- (c) the atmosphere is homogeneous (uniform composition) & static
- (d) the atmosphere is isothermal over the height range considered
- (e) 'g' is assumed constant

Assumptions (c), (d) & (e) taken together effectively mean that the scale-height of the atmosphere has been assumed constant over the height-range considered.

The theory gives a convenient expression for the rate of electron production but, of course what we measure in practice, is the electron density. The transition from electron production rate to electron density involves assumptions about the electron loss process. For recombination (or recombination-like process) we write $dN/dt = I - \alpha N^2$ and at noon (or any time when equilibrium conditions exist) we have $I = \alpha N^2$.

In radio-sounding experiments, when we measure the 'critical frequency', we measure N_{\max} - the peak density of the layer. If now in the theory, we make the assumption that the level of maximum production & the level of maximum density are the same, then we can set $I_{\max} = \alpha N_{\max}^2$. The theory gives $I_{\max} = I_0 \cos \chi$ so that $N_{\max} = (I_0 \cos \chi / \alpha)^{1/2} = (S_{\infty} \beta \cos \chi / e H \alpha)^{1/2}$

Since we assume α to be constant throughout the height-range considered, we have the result $N_{\max} \propto (\cos \chi)^{1/2}$

Since the 'critical frequency' $f_o \propto (N_{\max})^{1/2}$
 $f_o^2 \propto (\cos \chi)^{1/2}$

or $f_o \propto (\cos \chi)^{1/4}$

Measurements of the critical frequencies (electron-densities) of the normal region E and of region F1 of the ionosphere show a good measure of agreement with this prediction of the Chapman theory. These two regions of the ionosphere are closely solar-controlled and show the expected variations with solar zenith-angle and with variations in the intensity of the ionising solar radiations - solar cycle changes. (S_{∞} in the theory).

Sporadic E or E_s ionisation

From time to time a thin stratum of ionisation, which maybe only hundreds of metres thick) appears at E-region heights and although it has some measure of regularity in, for example, at temperate latitudes it is more prevalent in summer than in winter & more prevalent in the day than at night, it is, in fact, very 'sporadic' both in time & in spatial extent. Sometimes it can be very intense & give rise to quite abnormal hf radio propagation exhibiting a very high reflection coefficient such as would arise from a layer with a very sharp height-gradient of ionisation. At high latitudes a form of E_s is associated with the occurrence of aurorae & no doubt particle bombardment is a major source of high-latitude E_s . Near the magnetic equator there is another distinctive type of E_s which can be associated with the equatorial electro-jet - plasma instabilities. One explanation for E_s at temperate latitudes is the so-called 'wind-shear' theory.

The D-region of the ionosphere

The part of the ionosphere below about 90 km is that in which, particularly during the daytime) substantial attenuation of Hf radio signals occurs & it is also the level at which LF, VLF waves, atmospherics etc., are reflected, or refracted. The D-region, like the normal E-region, shows a large measure of solar-control - it is most fully developed in the daytime and the electron density is much reduced at sunset, but there is also evidence that this lowest part of the ionosphere sometimes exhibits meteorological-like influences. Thus it has long been known from radio sounding experiments that the ionospheric absorption of frequencies in the 2 - 5 MHz range (absorption which is known to occur mainly in the D-region) is often quite abnormally large on individual days (or groups of days) during the winter period. This is the so-called 'winter anomaly' in ionospheric absorption.

Some years ago evidence was produced to show that this winter anomaly sometimes, but not always, was accompanied by enhanced temperature in the stratosphere ('stratospheric warmings') and more recently rocket experiments have clearly shown that the winter absorption anomalies are due to abnormal electron density in the D-region and that furthermore the increased electron density may be associated with enhanced nitric-oxide content at these levels at certain winter periods. Movements in the atmosphere on a planetary scale play an important role in the general circulation, particularly during winter, & it may be that this is the mechanism whereby, at certain levels, the composition, & hence the electron density, and hence the winter absorption of the radio waves, are affected. (Fig. 6).

Sources of the ionisation at D-region heights are:

- (i) Solar Lyman- α radiation 1216 Å ionising the trace constituent nitric oxide
- (ii) Solar X-radiation 1-8 Å ionising the 'tail' of the E-region
- (iii) cosmic ray radiation producing ionisation below about 70 km.

It is probable that solar X-rays are more important at solar maximum & Lyman- α important at solar minimum. We would thus expect sunspot-cycle (in addition of course, to diurnal & seasonal) changes in D-region ionisation and in the absorption of HF signals & this is what is observed.

APPENDIX 1 p2

CHAPMAN CONTINUED

Variation of $I(\chi, h)$ with h for fixed χ
level of maximum electron production given by
setting $\frac{dI(\chi, h)}{dh} = 0$

Then find I to be maximum when

$$e^{-R/H} = \frac{1}{H A P_0 \sec \chi} \quad (4)$$

Insert (4) in (3) gives

$$I_{\max}(\chi, h) = \frac{A S_0 \cos \chi}{H e} \quad (5)$$

Variation of $I_{\max}(\chi, h)$ with χ

For overhead sun $\chi = 0$ & let I_{\max} for $\chi = 0$
= I_0

Let corresponding $h = h_0$

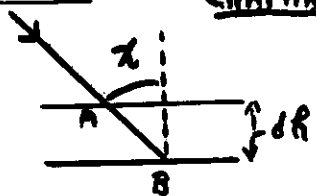
h_0 is then the level of maximum electron production
for $\chi = 0$

$$(4) \text{ then gives } e^{-R_0/H} = \frac{1}{H A P_0} \quad (6)$$

$$(5) \dots I_0 = \frac{A S_0}{H e} \quad (7) \quad \text{Rewrite (6) in terms of } I_0 \text{ and } h_0 \text{ gives}$$

$$\underline{I(\chi, h) = I_0 e^{1-z - \mu \chi e^{-z}}} \quad (8) \quad \text{where } z = \frac{h-h_0}{H}$$

CHAPMAN THEORY - LAYER FORMATION

Intensity at A = $S + dS$ " " B = S

$$\frac{dS}{S} \propto \rho(AB) \propto \rho dh \sec \chi$$

i.e. \propto mass of gas in column

$$\rho = \rho_0 e^{-R/H}$$

$$\frac{dS}{S} \propto \rho_0 e^{-R/H} \cdot dh \sec \chi$$

$$= A \rho_0 \sec \chi e^{-R/H} \cdot dh \quad (0)$$

$$\int_{S_0}^S \frac{dS}{S} = A \rho_0 \sec \chi \int_0^R e^{-R/H} \cdot dh \quad \text{where } S_0 = \text{intensity of incoming radiation at infinite depth } R = \infty$$

$$\log_e \frac{S}{S_0} = -A \rho_0 H \sec \chi e^{-R/H} \quad (1)$$

S - intensity - energy crossing unit area in unit time

$\left(\frac{dS}{dh \sec \chi} \right)$ units of energy absorbed per sec in unit volume.

Assume absorption of unit energy produces β electrons

Then $\beta \cdot \frac{dS}{dh \sec \chi}$ electrons produced per unit volume per sec

or 'Rate of electron production'

$$= \beta \frac{dS}{dh} \cos \chi = I(\chi, R) \quad (2)$$

say

From (1) above $\frac{dS}{dh} = S A \rho_0 \sec \chi e^{-R/H}$

$$(2) \text{ then becomes } I(\chi, R) = S A \rho_0 \beta e^{-\frac{R}{H} - A \rho_0 H \sec \chi e^{-\frac{R}{H}}} \quad (3)$$

Electron Production Rate (Chapman)

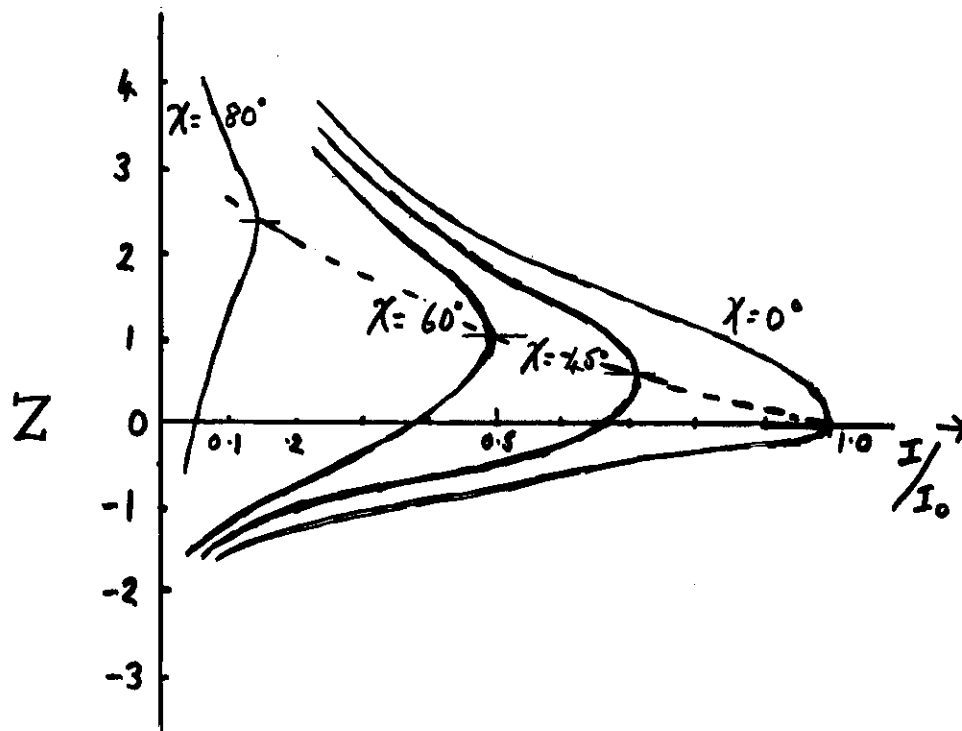


TABLE 1
TROPOSPHERE - COMPOSITION
PERMANENT CONSTITUENTS - % BY VOLUME

| | | |
|-----------------|--------|----------|
| N ₂ | 78.084 | } 99.997 |
| O ₂ | 20.946 | |
| A | 0.934 | |
| CO ₂ | 0.033 | |

INERT GASES

| | |
|----|-----------------------|
| He | 5.2×10^{-4} |
| Ne | 1.82×10^{-3} |
| A | 0.93 |
| Kr | 1.14×10^{-4} |
| Xe | 8.7×10^{-6} |

'MINOR' CONSTITUENTS

| | |
|------------------|---|
| H ₂ O | $10^{-3} - 1.0$ |
| O ₃ | $10^{-5} - 10^{-6}$ |
| CH ₄ | 1.5×10^{-4} |
| H ₂ | 5×10^{-5} |
| N ₂ O | 5×10^{-5} |
| NO _x | $5 \times 10^{-8} - 2 \times 10^{-6}$ |
| CO | $6 \times 10^{-6} \text{ to } 2 \times 10^{-5}$ |

FIG 2

ROCKET MSTS. ATOMIC OXYGEN

UCW. ABERYSTWYTH - R.A.L.

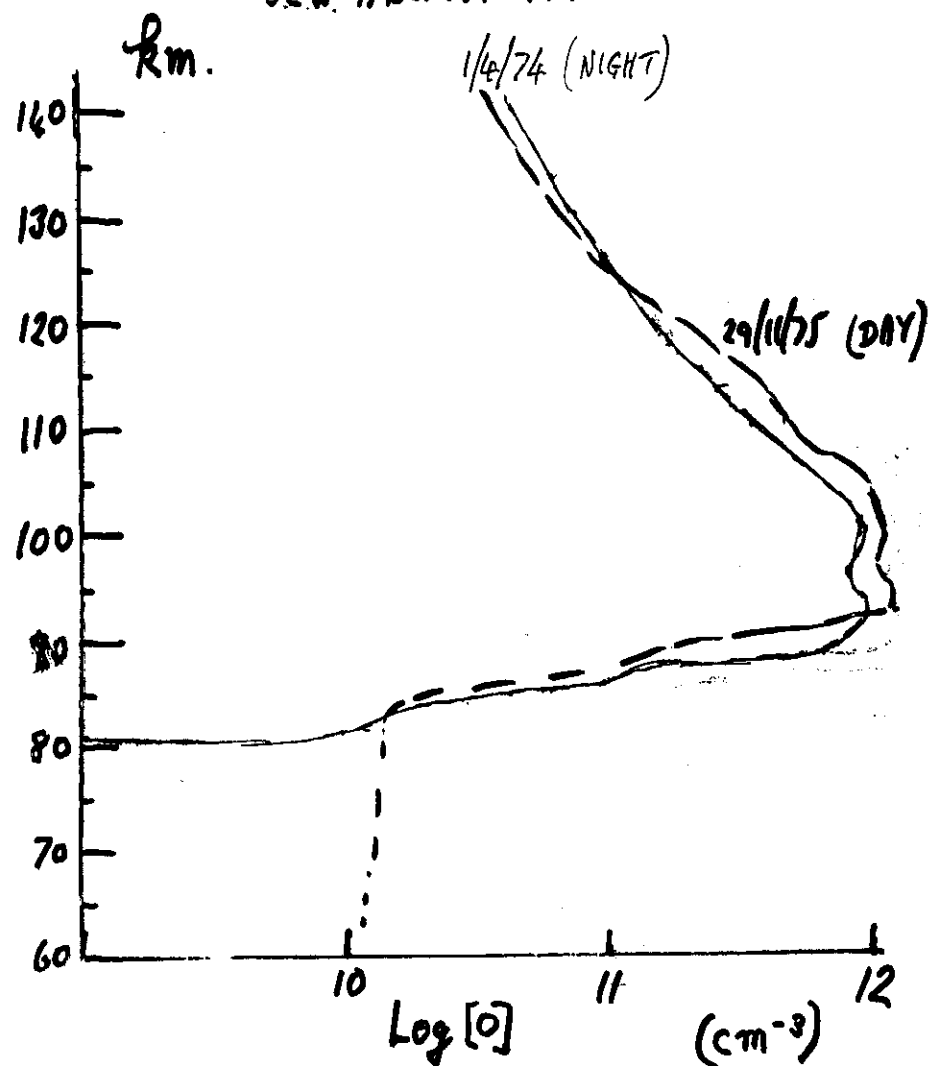


FIG. 3

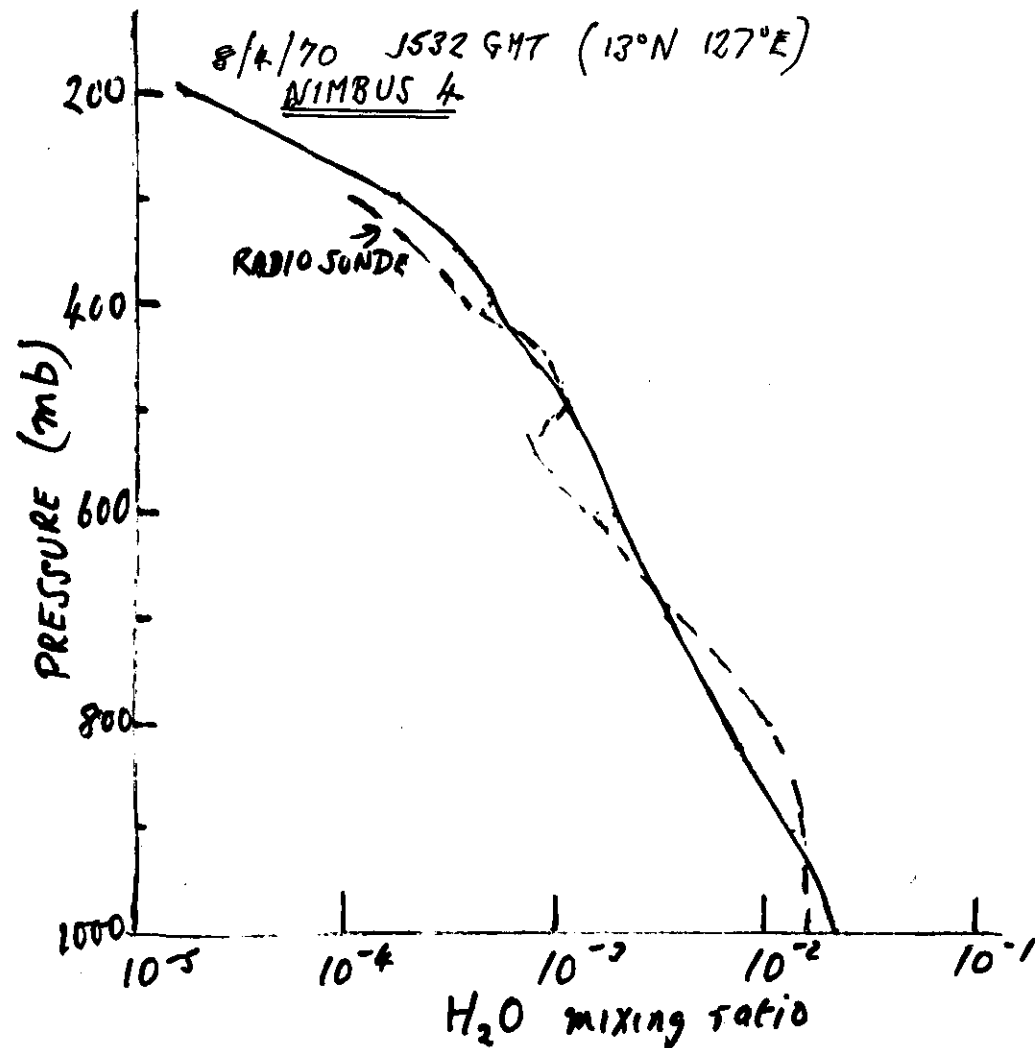
H₂O PROFILE 0-12 km

TABLE 2

MEAN MOLECULAR WT. \bar{m}

| ALTITUDE (km) | \bar{m} (a.m.u.) | |
|---------------|--------------------|--------------------------------|
| 0 | 28.97 | } NO CHANGE FOR FIRST 90 km |
| 90 | 28.97 | |
| 100 | 28.85 | |
| 120 | 28.60 | |
| 140 | 28.25 | |
| 160 | 27.90 | |
| 180 | 27.47 | |
| | | <u>Min</u> |
| 200 | 27.00 | 26.77 |
| 220 | 26.48 | 26.16 |
| 250 | 25.55 | 25.19 |
| 300 | 23.74 | 22.96 |
| 400 | 20.0 | 18.60 |
| 500 | 17.48 | 16.59 |
| 600 | 15.70 | 15.3 |
| 700 | 14.62 | 14.20 |
| 800 | 13.88 | 13.14 |

FIG. 4

| | \uparrow H^+ |
|------------------------------|---------------------|
| $R_m.$ | |
| 4000 | H |
| $10^4/cc$ 2000 | H^+ |
| | He |
| | H |
| | O |
| | N |
| 10^5 1000 | He |
| | O |
| | H, N |
| 700 | O |
| | N_2 |
| | He |
| | N |
| 10^{10} 200 | N_2 |
| | O |
| | O_2 |
| 85 | N_2 |
| | O_2 |
| $2.5 \times 10^{19} cm^{-3}$ | O |

CONSTITUENTS LISTED
IN ORDER OF IMPORTANCE

TABLE 3

| MEAN TEMPERATURE / HEIGHT | |
|---------------------------|------------------|
| KM. | T (°K). |
| 0 | 289 |
| 10 | 222 |
| 30 | 228 |
| 40 | 247 |
| 52 (max. T) | 271 |
| 60 | 258 |
| 86 (min. T) | 178 |
| 100 | 212 |
| 120 | 343 |
| 140 | 799 |
| 160 | 1155 |
| 200 | 1227 |
| 300 | 1359 |
| 400 | 1436 |
| 800 | 1474 (1190-1830) |

VARIABLE
INCREASING
VARIABILITY
UPWARDS.

TEMP. GRADIENTS

100 - 120 R_m $\frac{\Delta T}{\Delta R} \approx \frac{131}{20} = 6.5 K/R_m$

120 - 160 $\frac{\Delta T}{\Delta R} \approx \frac{812}{40} = 20.3 K/R_m$

160 - 180 $\frac{\Delta T}{\Delta R} = \frac{35}{20} = 1.8 K/R_m$

