



**SOME ASPECTS OF V.L.F. PROPAGATIONS
BY
MEANS OF ATMOSPHERICS**

by
Pawel Rozenfeld

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SUBMITTED IN PARTIAL FULFILLMENT OF THE
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MASTER OF SCIENCE

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Comissão Nacional de Atividades Espaciais
São José dos Campos — SP — Brazil

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Pawel Rozenfeld

Abstract

A short review of the source and propagation characteristics of atmospherics is given. Basic concepts of mode theory of wave propagation are presented and the single-station and multi-station techniques used by earlier workers to study the propagation of atmospherics are discussed. Using the single-station technique the distances to origin and amplitude spectra of atmospherics are determined. It was observed that the amplitudes of nighttime spectra are always greater than the amplitudes of daytime spectra. There is a tendency for the peak frequency of atmospherics to vary with distance during the day and night. The results indicate that there is probably an energy maximum in the ELF band beyond the maximum observed in the V.L.F. band.

SOME ASPECTS OF V.L.F. PROPAGATION BY MEANS OF ATMOSPHERICS

1 - INTRODUCTION

The great reliability of communications at global distances in the VLF band led to increasing interest in the study of propagation characteristics in this band, but the large antenna dimensions and high power required present difficulties in VLF propagation studies. As a result there is lack of VLF transmitters that cover the entire VLF band.

Atmospherics provide an easy means for such a study.

The principal problem in the atmospheric propagation study is the lack of knowledge about the exact spectrum of the source (i.e., the parent lightning discharge that produces the atmospheric).

One way to overcome this difficulty is to record the same atmospheric at two or more stations simultaneously. Such a multi-station technique has been successfully employed by Taylor (1960, a,b), Jean, Taylor & Wait (1960), Croom (1964). Though the multi-station technique avoids unrealistic assumptions regarding the spectrum at the source it requires elaborate synchronizing arrangements and involves considerable expenditure.

It is possible, also, to study the VLF characteristics through atmospheric by the single-station method. Hales (1948) was the first who studied the single-station technique.

Horner & Clarke (1955), described the single-station technique using a direction finder to determine the direction of the origin of atmospheric. The disadvantages of this method are the same as that of the multi-station technique.

Clarke & Mortimer (1951) described an Automatic Atmospheric Waveform Recorder used by Horner & Clarke.

In a series of papers Hepburn (1967, 1958, 1959, 1960a, 1960b) studied the single-station method and by improving Hales' method

made it possible to determine the distance to the origin of each of the waveforms. Hepburn's method of distance determination is used now by all who use the single-station technique. Wadehra & Tantry (1966, 1967) and most recently Rao (1968) used this method.

2 - ATMOSPHERICS AND THEIR PROPAGATION

2.1 - The Lightning discharge

Lightning is a result of the ionization of the air which occurs between adjacent charge centers or between a charge center and the ground when the potential gradient between them reaches the breakdown value.

The intensity of the pre-discharge field is that required for the electrical breakdown of air at atmospheric pressure ie. about 10.000 V/cm. -

The exact mechanism of the lightning discharge is very complex. Schonland's work (1953) shows that the discharge is initiated by a weakly ionized slow moving tip streamer which advances from the cloud toward the ground a distance of about 50 meters with velocity of the order of 0.5 m/ μ sec. The tip is followed by a highly ionized leader whose velocity is 100 times greater producing a current pulse of the order of 300 A. The charge is then transferred from the cloud to the ionized channel which acts as a charge storage element. The tip again moves forward and the process repeats itself every 25 to 100 μ sec. This process radiates energy during the period which is of the order of 1msec, known as the predischage.

Once the leader reaches the ground the main or return stroke with average peak current values of 20.000 A starts upward at a velocity of order of 50m/ μ sec.

Sometimes the charge neutralization is not complete after the first return stroke and other return strokes occur in the original ionized path (Workman et al, 1960).

Horner (1964) points out that this description of the lightning phenomenon is not general and depend on the location where it occurs. The phenomenon may be much more complex than described above.

There have been some attempts to study the structure of the ionized path during lightning discharge by means of radar reflection (Horner, 1964). The reflections are presumed to be from the ionization produced by the discharge.

Cloud discharges are more frequent than discharges to ground. Typically the proportion r of discharges striking the ground is dependent on the geographic latitude, λ , of the storm and is given by

$$r = 0.1 + 0.25 \sin\lambda$$

Most of the energy from a lightning discharge is associated with the high current flowing in the return stroke.

Bruce & Golde (1941) have suggested the following mathematical model which describes reasonably well this current.

$$I = I_0 (\exp(-\alpha t) - \exp(-\beta t))$$

where t is the time in seconds and I_0 is the current at a time t , the origin of the time being taken at the start of the return stroke.

Values of the parameters in the above equation can be determined either by monitoring the current in objects directly struck or by observations of the electromagnetic field radiated during the discharge.

Bruce & Golde (1941) derived values of α and β by monitoring current while Morrison (1953) obtained them from records of radiated fields. These results are listed in the following table.

Author	$\alpha(\text{sec}^{-1})$	$\beta(\text{sec}^{-1})$
Bruce & Golde	4.4×10^4	4.6×10^5
Morrison	7.0×10^3	4.0×10^4

I_0 is assumed to be equal to 28,000A.

The differences are due to the fact that Bruce & Golde's current represents the current at the base of the channel while Morrison's records were obtained from radiation during the return stroke and imply some integration over the whole ionized path.

Bruce & Golde's expression for current describes inadequately the electrostatic field changes and some finer details of pulse shape and duration. So Hepburn (1952) suggested the following expression for current.

$$I = Ae^{-\alpha t} - Be^{-\beta t} + Ce^{-\gamma t}$$

where $A = 20\text{kA}$, $B = 25\text{kA}$, $C = 5\text{kA}$ and $\alpha = 5 \times 10^4 \text{ sec}^{-1}$, $\beta = 5 \times 10^5 \text{ sec}^{-1}$ and $\gamma = 7 \times 10^2 \text{ sec}^{-1}$. This current form gives a better approximation to the average observations.

Fig. 1 shows the three current forms proposed.

2.2 - Atmospherics

The ionized path produced by the return stroke to the ground can be represented by a vertical antenna of length $l(t)$, given by

$$l(t) = \int_0^t v_t dt$$

where t is the time after the start of the return stroke and v_t is the ascent velocity of the return stroke.

Experimental studies indicate that the velocity v_t for the first return stroke, could be expressed as:

$$v_t = v_0 \exp(-\gamma t)$$

where $v_0 = 8 \times 10^7 \text{ m/sec}$ and $\gamma = 3 \times 10^4 \text{ sec}^{-1}$.

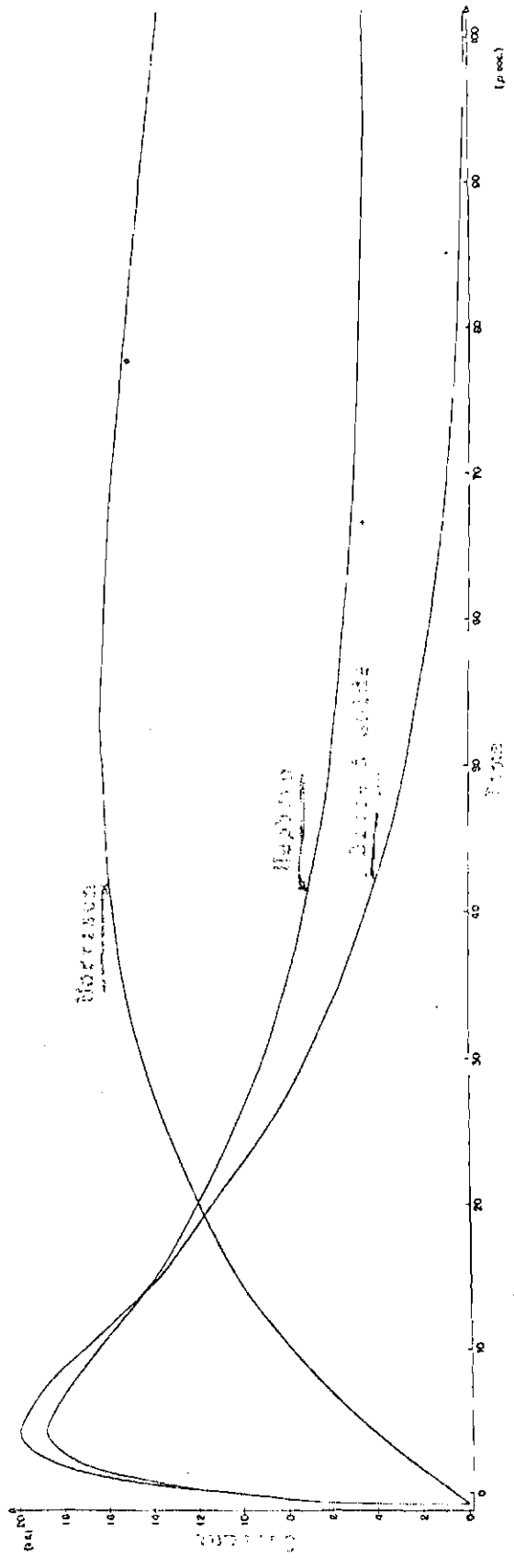


Fig. 1 - Discharge current variations (based on F. Anderson (1944) and S. Hepburn (1952)).

The electric field produced by such a vertical antenna is:

$$E = \frac{1}{2\pi\epsilon_0} \left[\int \frac{Mdt}{d^3} + \frac{M}{cd^2} + \frac{dM/dt}{c^2d} \right]$$

where: M is the magnetic moment of the radiator, i.e., instantaneous product of height times the average current along the antenna (m.A)

c is the velocity of light (m.sec⁻¹)

ϵ_0 is the permittivity of free space (F.m⁻¹)

d is the distance from the source (m)

E is the vertical electric field (V.m⁻¹)

The third term of the above equation is the radiation field, i.e., the atmospheric.

The Fourier components and the phase relations of these component frequencies of the radiation field are shown in Fig.2.

The atmospheric energy covers all frequencies of the radio spectrum but the radiation is most intense in the VLF and E.L.F. bands. It was observed that ground discharges have a greater relative content of V.L.F. than cloud discharges.

Watt & Maxwell (1957) present the characteristics of atmospheric noise from 1 to 100 KHz.

2.3 - Propagation of Atmospherics

There are two main methods of describing the propagation of V.L.F. radio waves. The first of them is by considering successive reflections of the original signal between the ionosphere and the ground together with the ground wave and is known as the reflection theory and the other is to consider the earth and ionosphere as forming a parallel-plate waveguide this being known as the waveguide theory.

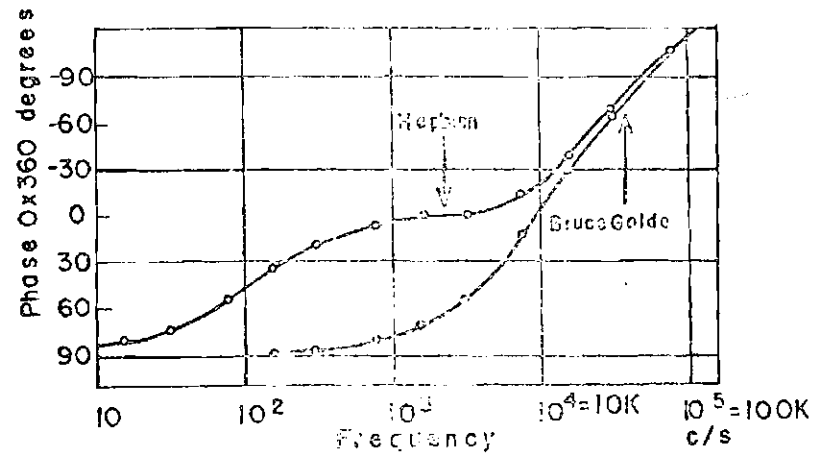
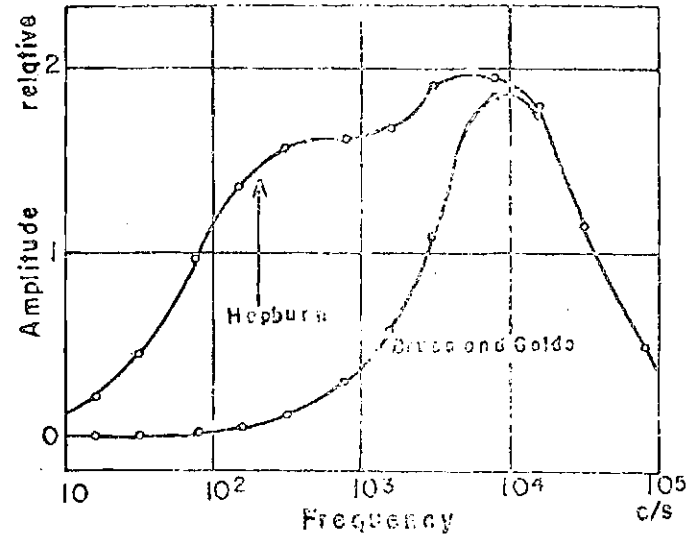


FIG.2 Radiation-field amplitude and phase spectra (After Hepburn 1957)

Horner & Clarke (1955) as well as Hepburn & Pierce (1954) had used the reflection theory to study the propagation. The other researchers have used the waveguide theory.

Wait (1962) has made a complete treatment of the waveguide theory applied to the ionosphere considered as stratified medium.

Hales (1948) has suggested the formal equivalence of the mathematical equations deduced from the two approaches and now it is beyond doubt that these two methods are equivalent and give the same results provided that a sufficient numbers of modes is considered.

The reflection theory is more useful for the study of atmospherics at short distances from the origin because the waveguide theory is more complex due to the mode interference at short distances.

At large distances from the origin, the waveguide theory is simpler because only one or two modes of propagation need be considered.

Wadehra & Tantry (1966) and Croom (1964) locate the limit between these two approaches at about 1500 km from the origin of the atmospherics.

As the aim is to study propagation at long distances we are going to study the propagation of atmospherics in the infinite wave-guide formed by the earth and ionosphere initially considered as perfectly conducting parallel planes with separation h .

Alpert et al (1967) had suggested that at distances greater than 3000 km the use of spherical waveguide theory is necessary.

2.4 - Waveguide Theory

The theory presented in this section is mainly taken from Wait (1962), and Davis (1965).

Waveguide theory will be explained based on a simple model for the earth-ionosphere cavity. It is assumed that earth and ionosphere are perfectly conducting planes.

We adopt a cylindrical coordinate system (ρ, ϕ, z) .

On this system the earth is the plane $z = 0$, the plane $z = h$ is the lower edge of the ionosphere, ρ is the radial coordinate and ϕ the azimuthal coordinate.

The source of the electric field is a vertical dipole located at the origin of the system.

Fig. 3 presents the arrangement used in the following discussion.

The perfectly conducting earth and ionosphere behave as if they were plane mirrors for the electrical wave and so at any point on the ground the signal would appear to come from the vertical dipole at the origin and a whole series of images produced by the ionosphere-earth system. These images are located just below the ground and at $z = \pm 2h, \pm 4h$ and so on and they all have equal sign and magnitude.

At any point on the ground for which the distance to the source is large compared to h , the electric field can be calculated by replacing the line of dipole image, by a continuous line source carrying an equivalent uniform current I_a .

Let \underline{s} be the height of the dipole and \underline{I} its current; the average current along the \underline{z} axis is given by:

$$I_a = \frac{s}{h} I$$

The field radiated by a line source of current is known and thus

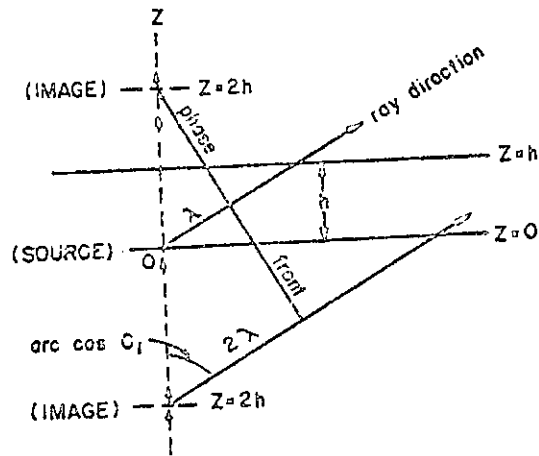


Fig.3 - Ray geometry corresponding to the 1^{st} mode between parallel conducting planes. (After Wait, 1962).

$$E_z = \frac{I_a \mu \omega}{4} H_0^{(2)}(k\rho) = \frac{\mu \omega I_s}{4h} H_0^{(2)}(k\rho)$$

where

$H_0^{(2)}(k\rho)$ is a Hankel function of the second kind of argument $k\rho$

μ is the permeability of the space

ω is the angular frequency

$$k = 2\pi/\lambda$$

For large distances ($\rho \gg \lambda$) it is possible to make an asymptotic expansion of the Hankel function and the expression for the field is:

$$E_z \approx \frac{1}{2} \left(\frac{\mu}{\epsilon}\right)^{1/2} \frac{I_s}{h(\lambda\rho)^{1/2}} e^{i\pi/4} \cdot e^{-ik\rho}$$

This field corresponds to the radiation directed broadside so the rays are parallel to the bounding walls and the angle subtended by the rays and the z axis is equal to 90° .

However, there will be another condition for which the rays emanating from each of the dipoles in the line of images are also in phase. This condition is

$$2hC = n\lambda$$

where C is the cosine of the angle subtended by the rays and the z axis and n is an integer. So, for each value of n there are 2 families of rays which have the same radial phase velocity (i.e., $= c/S$ where $S^2 = 1 - C^2$) but opposite vertical phase velocities (i.e., $= \pm c/C$). This set of waves each of which is called a mode, can be imagined to originate from an equivalent line source. However, the strength of this line source is IS where S is the sine of the angle subtended by the rays and the vertical direction.

To obtain the resultant vertical field this must be again multiplied by S . Consequently, the resultant field of all modes is obtained by summing over integral values of n from 0 to ∞ to give

$$E_z = \frac{\mu\omega I_s}{4h} \sum_{n=0}^{\infty} \delta_n S_n^2 H_0^{(2)}(k S_n \rho)$$

where

$$\delta_0 = 1 ; \quad \delta_n = 2 \quad (n = 1, 2, 3 \dots)$$

$$S_n = (1 - C_n^2)^{1/2}$$

$$C_n = n\lambda/2h$$

The term $n=0$, corresponding to zero mode discussed at the beginning of this chapter is included only once in the summation whereas the higher modes are included twice to include the image produced in $z = -2h, -4h$ and so on.

Just as for the zero mode it is possible to make an asymptotic expansion of the Hankel function and we have

$$E_z = \frac{1}{2} \left(\frac{\mu}{\epsilon}\right)^{1/2} \frac{I_s}{h(\lambda\rho)^{1/2}} e^{i\pi/4} \sum_{n=0}^{\infty} \delta_n S_n^{3/2} e^{-ikS_n\rho}$$

It was seen above that the phase velocity v_p in the radial direction is given by

$$v_p = c/S_n$$

If we let λ_g be the wavelength in the radial direction we have

$$\lambda_g = \frac{\lambda}{S_n}$$

where λ is the free space wavelength.

But $S_n^2 = 1 - C_n^2$ and $C_n = \frac{n\lambda}{2h}$ $\therefore S_n^2 = 1 - \left(\frac{n\lambda}{2h}\right)^2$

$\therefore \lambda_g^2 = \frac{\lambda^2}{S_n^2}$ or $\left(\frac{1}{\lambda_g}\right)^2 = \left(\frac{1}{\lambda}\right)^2 \cdot S_n^2$

So, $\left(\frac{1}{\lambda_g}\right)^2 = \left(\frac{1}{\lambda}\right)^2 \left[1 - \left(\frac{n\lambda}{2h}\right)^2\right]$

$\left(\frac{1}{\lambda_g}\right)^2 = \left(\frac{1}{\lambda}\right)^2 - \left(\frac{n}{2h}\right)^2$

This expression shows that for $\lambda > \frac{2h}{n}$, λ_g is imaginary and, hence, the mode is evanescent.

There is a minimum cut off frequency f_n below which waves will not propagate where

$$f_n = \frac{nc}{2h}$$

where

n - wave mode number

c - velocity of light

h - height of waveguide

The cut off frequency for the first mode of propagation is $f_1 = \frac{c}{2h}$ and adopting $h = 75$ km for the height of the daytime D layer, we obtain $f_1 \approx 2$ KHz. Taking $h = 90$ km for the height of the nighttime E layer we obtain $f_1 \approx 1,5$ KHz.

3 - MULTISTATION AND SINGLE-STATION TECHNIQUES OF STUDY OF ATMOSPHERICS

In this chapter we shall describe different approaches to the study of atmospheric, i.e. multi-station and single-station technique and some results obtained by them.

Taylor (1960) used a multi-station technique to study the daytime attenuation coefficients which were obtained from the waveforms received during the summer. His receiving equipment was located at Boulder, Colorado, Salt Lake City, Utah, Palo Alto, Calif and Maui, Hawaii. The equipment was composed of a 23 - ft vertical antenna, antenna cathode follower, bandpass filters composed of RC networks, 24- μ sec delay line, a wide-band amplifier (6-dB points at frequencies of 1 and 100 KHz) and an oscilloscope. As the oscilloscope was triggered by the waveform it was necessary to use a delay line in order to minimize the possible loss of the earlier part of the waveforms.

The direction finding system was used to determine the direction of arrival of atmospheric. The location of each atmospheric source as determined by triangulation using the direction of arrival indicated at each record. Atmospheric waveforms were located on the photographic records made at each station, and their times of reception and direction of arrival were thus determined.

It was assumed that only the first mode of propagation was important and so the amplitude spectrum may be expressed by the following equation deduced by Wait (1957b, 1958a):

$$|E(\omega, d)| \approx A(\omega) \left[\frac{a \sin d/a}{a} \right]^{-1/2} e^{-\alpha(\omega) d}$$

$|E(\omega, d)|$ - is the amplitude spectrum of the waveform

$A(\omega)$ - is the amplitude coefficient dependent upon the spectrum of the source

a - is the radius of the earth in km

d - is the great circle distance in km

$\alpha(\omega)$ - is the attenuation coefficient in nepers.

So, knowing the distance from each of the observing stations to the same lightning stroke and having its Fourier spectrum, it is possible to determine the attenuation coefficient in dB/1000 km by the following formula

$$\alpha(\omega) = \left[20 \lg \frac{|E(\omega, d_{\underline{i}})|}{|E(\omega, d_{\underline{j}})|} - 10 \lg \frac{\sin d_{\underline{i}}/a}{\sin d_{\underline{j}}/a} \right] \cdot \frac{10^3}{(d_{\underline{i}} - d_{\underline{j}})}$$

where the subscript \underline{i} denotes the far station and the subscript \underline{j} denote the near station such that $d_{\underline{i}} > d_{\underline{j}}$. It is seen that the term involving the source spectrum is eliminated from this expression.

Taylor made a study of the difference between east-west propagation and west-east propagation. In the east-west propagation the distances from the source of each atmospheric to the near station were between 1000 and 2400 km while the distances from the source to Maui (the far station) were between 4160 to 5200km.

In the west-east propagation, Maui was used as the close station and Palo Alto as the far station. The approximate average distance from the source to Maui was 6460 km and from the source to Palo Alto was 9160 km.

In both cases it was observed that the frequency of the peak amplitude of each spectrum increases as the propagation distance increases.

The attenuation - frequency curve for the east-west propagation and west-east propagation have the same shape i.e. the attenuation decreases from 3KHz to 4 KHz and after that increases and has its maximum at about 6 KHz. However, the values of attenuation for the east-west propagation are not equal to the values of attenuation for the west-east propagation. The difference between the attenuation for east-west minus attenuation for west-east propagation have the following characteristics: a signal propagating from west to east is attenuated 3dB/1000 km less between frequencies of about 3 to 8 KHz and 1dB/1000 km less between 10 to 30 KHz than a signal propagating from east to west. Taylor also concluded that propagation from west to east is

better than from east to west. Crombie (1960) have demonstrated that the earth's magnetic field introduce non-reciprocity in the propagation and that for propagation along the magnetic equator, the rate of attenuations is less for west-to-east propagation than for east-to-west propagation.

Croom (1964) studied the relative values of the nighttime attenuation coefficient also by the multi-station technique. The stations were located at Cambridge and London in England, Poitiers at France, Weissenau and Potsdam at Germany. A direction finder determined the position of the atmospheric source. As Taylor, Croom assumed that only the first propagation mode is important and used the same expression for the spectrum of the atmospheric. The frequent lack of a field strength calibration obliged Croom to normalize the atmospheric spectra with respect to a frequency f_0 . He chose initially 10 KHz as the normalizing frequency because this is the frequency at which atmospheric had maximum energy and this probably, would minimize, the amplitude spread of results. However, further analysis of a number of waveforms showed that 7KHz was a much better frequency to use for this purpose.

Croom analysed winter nighttime waveforms and obtained values of α_r for frequency intervals of 500 Hz between 5 KHz and 15 KHz.

Different waveforms recorders were used at each of the stations. They all had a flat frequency response over the range 500 Hz to between 12 and 15 KHz and all except Weissenau, recorded the waveforms on an oscilloscope, which was triggered by the waveform itself. Weissenau used a drum camera method of recording. The Poitiers records were of the rate of change of the electric field dE/dt rather than the electric field E .

The location of the receiving stations that had collaborated with Croom permitted him to obtain mean values of α_r , the relative attenuation coefficient for propagation over both land and sea paths.

The scatter of values of α_r from different waveform pairs was very large (about 4dB) and this may be attributed to the relative closeness of the observing stations.

At the same time Croom observed that in almost all the waveforms analyzed, a land path is also an east-west path and a sea path is a west-east path. So, any differences in results that he obtained are strongly dependent on variations in ground conductivity and the presence of the earth's magnetic field.

Although it was not possible to obtain any reliable information on the absolute value of attenuation, it appears to be a significant difference in the shapes of the attenuation frequency curves for sea (or west-east) propagation paths and land (or east-west) propagation paths. It appears that there is a minimum of attenuation at about 6 KHz for sea paths and from 6 to 11 KHz for land paths.

Croom observed the following behavior of the frequency of the maximum spectral components of nighttime atmospherics:

The peak frequencies lie between about 6 KHz and 8 KHz for west-east (over-sea) propagations and there should not be a noticeable progressive increase in the peak frequency with increasing distance for distances greater than 1000 km. For the east-west (land) propagation the peak frequency lies between 5.5 KHz and 10 KHz and there is perhaps an increase in the peak frequency with the distances of 0.5 KHz per 1000 km. The peak frequency in this case should, in general, be greater than the corresponding peak frequency for the sea path and there should be a greater scatter also.

In his study, Croom observed that the main source of errors in the measurement of the VLF attenuation is connected with the distance between a station pair. If the separation is of the order of only several hundred km the amount of attenuation being measured from the waveform pair is small and therefore liable to large errors. If, on the other hand, the separation is more than a few thousand km then the waveform received at the more distant station is very small in amplitude and therefore the resulting measured attenuation is again liable to large errors. From these considerations Croom suggested that the optimum distance between the stations is of the order

of 2000 km, which would give actual propagation distances of the order of 4000 km, since the source should be about 2000 km from the nearest station in order to avoid interference from modes higher than the first.

Let us now see a different approach to the problem using the single-station technique.

Obayashi (1960) presents one of them. A device called a radio spectroscopie is used to record continuously the frequency spectrum of atmospherics. Two such radio spectroscopes are used which covered the frequency ranges 1 to 10 KHz and 5 to 70 KHz. Spectroscopes scan the atmospheric VLF bands continuously, the outputs are displayed on an intensity modulated cathode ray oscillograph and recorded photographically. The photographic film is an intensity-modulated plot of frequency against time.

It is seen that the intensity of atmospherics is maximum at about 10 to 20 KHz and it decreases towards higher frequencies with remarkable undulating peaks. Obayashi noticed considerable diurnal and seasonal variations.

It was observed that short-distance atmospherics (probably less than 100 km) present a flat spectrum over wide frequency range. As consequence the strong selective absorption bands which appears in the records must be due to atmospheric propagation through considerable distances and may be largely influenced by the ionospheric conditions.

From these records it is seen that the nighttime spectrum shows strong absorption at about 2 to 3 KHz. The nighttime spectrum has a broad maximum at 5 to 50 KHz with several undulating peaks, while the daytime spectrum has a smooth double maximum at 15 KHz and 40 KHz. The daytime spectrum intensity is weaker than that of the nighttime. The effect of the transition between daytime and nighttime may be interpreted as a shift of the daytime maximum by about 15 KHz towards lower frequencies by the action of the local sunset while the effect of sunrise is the reverse; by the

waveguide theory this effect may be explained as the upward shift of the ionospheric height or a decrease of the conductivity in the lower ionosphere.

Obayashi observed that the effect of SEA (sudden enhancement of atmospherics) is to cause a sudden shift of the spectral pattern towards higher frequencies, through the amount of shift is variable for individual cases.

The SID (Sudden Ionospheric Disturbance) phenomenon causes an increase of the electron density in the lower ionosphere and its height is lowered considerably; these facts would eventually yield the explanation that the reduction of the field intensity in the lower VLF band is caused by the shifting of the cut off frequency towards slightly higher frequencies while in the frequency above the cut off the intensity is enhanced owing to the good reflection conditions at the lower edge of the ionosphere.

Obayashi could not determined the distribution of the originating sources of atmospherics. So, he assumed an average distance from the source and determined the curve of attenuation with frequency. The daytime curve has the following behavior: the attenuation decreases from 5 KHz to 10,5 KHz and after that increases up to 35 KHz, decreases a little up to 45 KHz and increases up to 60 KHz. The nighttime attenuation curves present lower attenuation than the daytime and show some undulations. However, the general shape is the same as that of the daytime attenuation curve.

Alpert et al (1967) presents another approach to the single-station technique. They used two wide-band receivers which have the bands 5 - 60 KHz and 3 Hz - 8 KHz and which record synchronously the high frequency part and the tail of atmospherics. The azimuth and the distance from the lightning are determined simultaneously with the help of a radiogoniometer system.

A spectral analysis of the waveforms is also made. The spectrum presents two maxima named the high frequency maximum and the low frequency maximum which correspond to VLF and ELF bands. It was observed that the daytime peak frequency changes with distance. The high frequency maximum moves from 6 KHz up to 9 KHz and the low frequency maximum moves from $f \approx 150$ Hz down to $f \approx 70$ Hz as the distance changes from 2000 Km to 6000 km. The amplitude ratio of the low frequency maximum to the high frequency maximum changes from 3 at 2000 km to 16.5 at 6000 km.

The nighttime mean amplitude spectra at all distances are approximately the same. The high frequency part of the spectrum is wider and more symmetrical than during the daytime and f_{\max} is not dependent on distance. The high frequency maximum lies in the frequency range 8 KHz to 9 KHz and the low frequency maximum at $f_{\max} = 150$ Hz. The ratio of the low frequency maximum to the high frequency maximum is approximately equal to 8.

In order to study under what conditions and at what distances different approximations are good and how often an ionosphere model is appropriate to the experiment, a method of signal synthesis was used i.e., a calculation of the waveform of the atmospheric by means of a theoretical synthesis of the waves forming the signal was carried out.

In order to calculate the signal $E(t,d)$ at a fixed distance \underline{d} it is necessary to compute the Fourier integral:

$$E(t,d) = \frac{1}{2\pi} \int_{-\infty}^{\infty} g(\omega,d) Q_0(\omega,0) \exp(i\omega t) d\omega$$

where $g(\omega,d)$ is the propagation function which defines the electrical field of the wave at a frequency $\underline{\omega}$ at distance \underline{d} from a source of constant intensity.

$Q_0(\omega, 0)$ is the spectrum of the source. For this calculation they used a standard source and Bruce & Golde's (1941) expression for $Q_0(\omega, 0)$. The propagation function for flat and spherical model waveguides has been calculated by different authors including Wait (1962).

After calculating theoretical signals $E(t, d)$ a number of its features were computed with the observed ones. The particular features compared were the number of the semi-periods n , the semi-periods of the oscillations of the signal τ_1, \dots, τ_n and the total duration of the signal T .

About 2000 daytime oscillograms were examined. For each of the chosen features ($n, \tau_1, \dots, \tau_n, T$) distribution curves were constructed and the most probable values and the limits of their variations at the level 0.8 and 0.5 of the distribution curves were determined. These 3 values were then compared with the corresponding values calculated theoretically. A rather satisfactory fit of the values n, τ_1, \dots, τ_n , and T with the experimental data was observed. The theoretical values coincide with the most probable values or lie within the ranges of their distribution at the level 0.8 and 0.5 of the distribution curves.

A very important conclusion from this study is the fact that at distances $d < 3000$ km the theoretical signals for a flat waveguide are in a good agreement with the experimental ones; for $d \gtrsim 3000$ km the use of the spherical waveguide theory is necessary.

Hepburn (1960) presented another approach to the single-station technique which was followed by some other workers.

He used an equivalent of the conventional sphere antenna. The signal cleaned from interferences was fed to both an amplitude-operated trigger unit and a 100 μ sec lumped constant delay line, which permitted examination of the field variation during the 100 μ sec immediately preceding the triggering of the device. The delayed signal was subsequently amplified and displayed on a cathode-ray oscilloscope, which was intensity modulated

for the duration of the 1 msec sweep by the appropriate trigger unit output.

The camera motor was activated by another output from the trigger unit for 1 sec after the recording of each waveform and so advance the film by 1 inch steps to space the traces along the film. 35 mm film was continuously exposed to the oscilloscope screen during the recording periods and recorded all the randomly occurring atmospheric for which amplitudes were sufficient to operate the trigger circuit. The trigger unit was inhibited during the film-advancing process.

Individual waveform source position estimates were available for many of the recorded traces by the spherics network of the British Meteorological Office.

The photographed waveforms were classified on the basis of the Hepburn's (1958) classifying system.

Hepburn's method of determining the distance to the origin of the atmospheric is based on a relation between the phase of signal components at the source and phase of the same signal components at arrival.

At the source the component frequencies of the signal are closely grouped in phase. By the waveguide theory they become progressively out of the phase as the propagation distance increases. This occurs at a rate depending upon the effective height of the waveguide. The final phase relations obtained as a function of frequency should then yield on analysis both propagation distances and guide height.

So, from the study of wave propagation by a waveguide treatment, Hepburn (1957) deduced the following relation between the delay T_n of a frequency f_n (i.e. arrival time relative to the start of the waveform) and the quasi-period $\tau_n = \frac{1}{f_n}$ of the same frequency in the trace, for a given storm distance d and guide height h :

$$T_n = \frac{d}{c} \left\{ \left[1 - \tau_n^2 \left(\frac{c^2}{4h^2} \right) \right]^{-1/2} - 1 \right\} \quad (1)$$

Fig.4 shows the values of T_n and τ_n they are measured from any waveform.

Expression (1) permits us to draw theoretical curves for any height of waveguide with distance d to the origin of the atmospheric as parameter.

By comparing an experimental plot of quasi-period against delay with the theoretical plot it is possible to determine the distance to the origin of the atmospheric as the value of d for the theoretical curve which has the best fit with the practical one. Some waveforms gave the correct distance for the source, while others presented serious discrepancies with the theoretical curves. These waveforms may originate from more complex sources. The theoretical implications involved in replacing the assumed ideal source with more plausible conditions have already been discussed in detail (Hepburn, 1959) and are shown to provide a possible explanation for the occurrence of irregularities.

Based on the spheric network data Hepburn assumed a standard ionosphere height of 85 km.

It is seen to be impossible to ascribe a definite value of h within the range 50-85 km for traces lacking their initial portions. Even when these are present, h values as low as 75 or 65 km depending upon the storm distance give traces indistinguishably different from those having $h = 85$ km and an appropriately chosen value of d .

So, the choice of the correct parameter values can only be made when there is independent evidence for the true value of one or the other.

In these circumstances the best approach appears to be the derivation of an average value of h , based upon those waveforms for

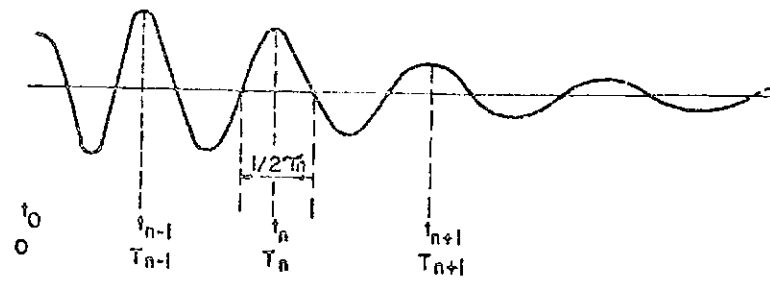


Fig. 3 - Waveform measurements after Hepburn

which good analysis and reliable spherics fixes are available and the assumed constancy of this parameter for other waveforms of undetermined or distant origin.

From the results available a constant average ionosphere height of 83 ± 2 km irrespective of time of day was the most reasonable conclusion.

A spectral analysis of the waveforms was made. The amplitude spectra present the following features: the occurrence of daytime amplitude greater than the amplitudes at 8 KHz for frequencies lower than 8 KHz was rare; lower frequency oscillations show relatively greater amplitudes at night than by day with correspondingly smaller amplitudes of frequencies greater than 8 KHz. These features were more pronounced with increasing propagation distance.

Attenuation coefficients expressed in dB/1000 km were estimated for day and night propagations referred to that at 8 KHz.

The attenuation varied little throughout the frequency range 4-14 KHz and had only slight diurnal variation.

The daytime attenuation shows an increase as the frequency is reduced, this behavior agrees with previous atmospheric measurements (Chapman & Macario, 1956) but its variation is much less rapid.

The nighttime attenuation always increases from 5 to 13 KHz which disagrees with the previous measurements (Chapman & Macario, 1956).

Wadehra & Tantry (1966) also used the single-station technique. An automatic atmospheric recorder was used with a suitable delay line to recover the initial portion of the waveform lost before triggering. A set of tuned narrow band receivers with a band-width of 150-200 Hz in the range of 2-27 KHz were used to record simultaneously the waveform and its energy spectrum with the help of a oscillograph specially designed and

constructed for this purpose. The record was taken on a 35 mm photographic film with the help of an automatic camera.

Two antennas were used; the first one 5 meters long for recording near atmospherics and the second one 14 meters long for recording distant atmospherics.

The source distances of atmospherics within 1500 km were determined by the formulae

$$h = 150 \frac{t_1 t_2 (t_1 + t_2)}{t_1 (r^2 - q^2) - t_2 (q^2 - p^2)}$$

$$d = \sqrt{\left[\left\{ \frac{1}{150} \cdot \frac{h^2 (q^2 - p^2)}{t_1} - 150 t_1 \right\}^2 - 4p^2 h^2 \right]}$$

where h is the height of ionosphere in km, d is the distance between the source and the receiver in km, $t_1 = t_q - t_p$ and $t_2 = t_r - t_q$ where t_p , t_q and t_r are the time delays in msec of three of p^{th} , q^{th} , and r^{th} order of reflections.

The source distances, for quasi-sinusoidal atmospherics separated by more than 1500 km from origin, were determined by Hepburn's (1960) method described above. From simultaneous records of atmospheric waveforms and their energy spectra the peak frequency of each atmospheric is determined.

Energy spectrum observations at the source during local thunderstorms revealed that in most of the cases atmospherics had a spectral peak at 5 KHz.

The theoretical curve which gives the vertical electric field of VLF radio waves at large distances from the source is given by (Wait, 1957b).

$$E_d = \frac{E_o \times 4 \times 10^{-(Ad/2 \times 10^4)}}{\sqrt{df}}$$

where E_o is the effective radiation field in V/m for a particular frequency at a distance of 1 mile from source; f is frequency in KHz; A is the attenuation factor in dB/1000 km.

The theoretical curve for nighttime conditions shows that initially there is a rapid increase in the peak frequency with distance up to about 1500 km, above which the rate of increase decreases slowly till the peak frequency becomes constant at about 3000 km and above.

Wadehra & Tantry's experimental curve for nighttime observations having either land path or mixed land-sea path quite closely fits in with the theoretical curve except for short distances up to 500 km where a large number of modes need to be taken into considerations. There is also a slight deviation of the experimental curve from the theoretical one at very large distances.

The nature of the theoretical curve for daytime conditions is similar to that of nighttime but the peak frequency at any distance is higher for day than for nighttime conditions.

The most recent work in this subject is by Rao (1968). He used four-stage R-C coupled amplifier having a response with 3dB points at 100 Hz and 90 KHz. The output of the amplifier was displayed on a cathode-ray oscilloscope, the intensity being controlled by a square wave generator triggered by the incoming atmospheric. The trace on the cathode-ray oscilloscope screen was photographed by a 35 mm camera operated manually. All the waveforms utilized were recorded during the nighttime. A direction-finder was not used.

The distance of origin of each of the waveforms has been determined by Hepburn's (1960) method.

A spectral analysis of waveforms was made in the range 3-15 KHz. For any particular distance the various spectra were normalized taking the mean of the amplitude expressed in dB. This was done in view of the results of various workers (Horner, 1964) that the amplitude variability of the atmospheric follows a log-normal law, though it has been found that the atmospheric originating in different storms give different standard deviations on any one frequency. It was observed that normalized spectra attain the maximum amplitude at about 9 KHz.

The attenuation coefficient was determined by the expression (Wait, 1958):

$$E(\omega, d) = \frac{A(\omega) \lambda^{1/2}}{h(a \sin \frac{d}{a})^{1/2}} \exp(-\alpha d) \cdot \Lambda$$

It was assumed that only one dominant mode is present and one can write

$$20 \lg E(\omega, d) + 10 \lg \left(a \sin \frac{d}{a}\right) = 20 \lg \frac{A(\omega) \cdot \lambda^{1/2} \cdot \Lambda}{h} - \alpha \cdot d$$

where

A (ω) describes the source spectrum

a is the radius of the earth

λ is the wavelength corresponding to ω

h is the height of the reflecting layer.

α is the attenuation coefficient and is a function of ω

Λ is the excitation factor.

The left side of the equation above for any one frequency should yield a straight line with slope α when plotted as a function of distance provided A (ω) is constant. However, the experimentally obtained points show a considerable amount of scatter but a mean value of α , for any one frequency, is calculated by taking the average of the gradient between different pairs of distances.

The curve of attenuation obtained in this way presented the following behavior: it increases from 3 KHz to 5 KHz, is undulating between 5 and 7 KHz with a minimum at 6 KHz, decreases from 7 to 9 KHz and undulates between 9 and 15 KHz with local maximum at 11 KHz.

4 - DATA COLLECTION AND ANALYSIS

In this chapter we shall describe our method of investigating the propagation of atmospheric.

4.1 - Circuit arrangement

The atmospheric is received by two loop antennas oriented at west-east and north-south the former one is referred in the following as antenna 1 and the other antenna 2.

The antennas are triangular and they have 18 meters base and 6,50 meters height.

The signal is amplified by an amplifier having a response with 3 dB points at 100 Hz and 100 KHz and voltage gain equal to 300.

Fig.5 presents the circuit arrangement. The output signal of the amplifier when it has sufficient amplitude to trigger the oscilloscope, is photographed by a camera operated continuously. The film speed is 20 cm/min.

Fig.6 shows typical received waveforms. The waveforms utilized in the present analysis have been recorded during the nighttime at about 21:00 LT. and during the daytime at about 15:00 L.T.

4.2 - Distance determination

The waveforms chosen for analysis were of the quasi-sinusoidal type. We did not use any direction finder.

The distance to the origin of each atmospheric was determined by Hepburn's (1960) method described above.

The expression used by Hepburn (1960),

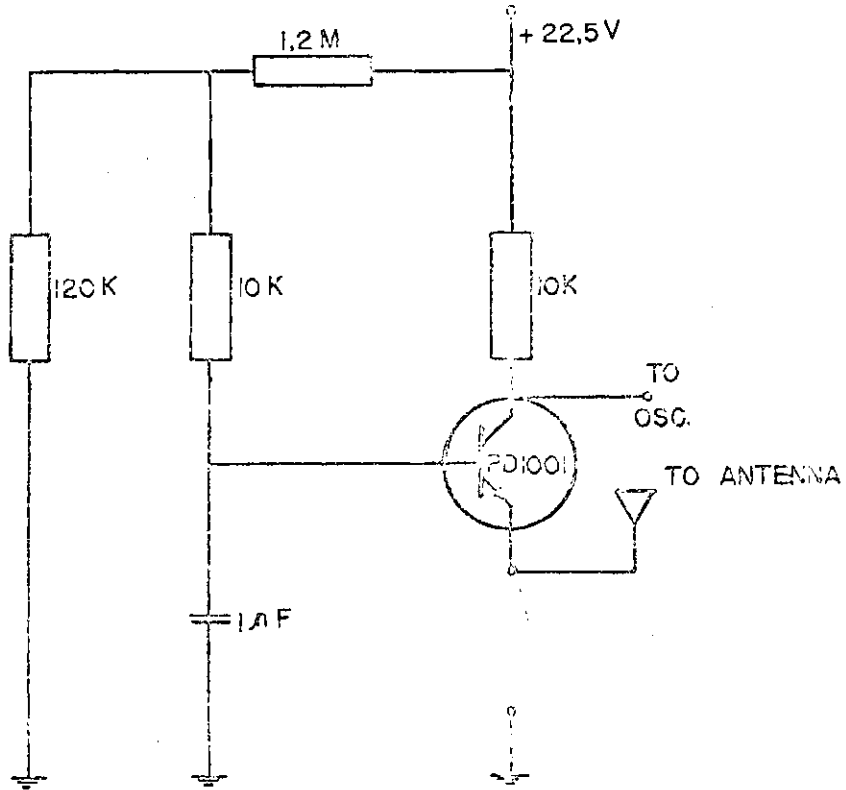


Fig. 5 - Circuit utilised in amplifying of atmospheric

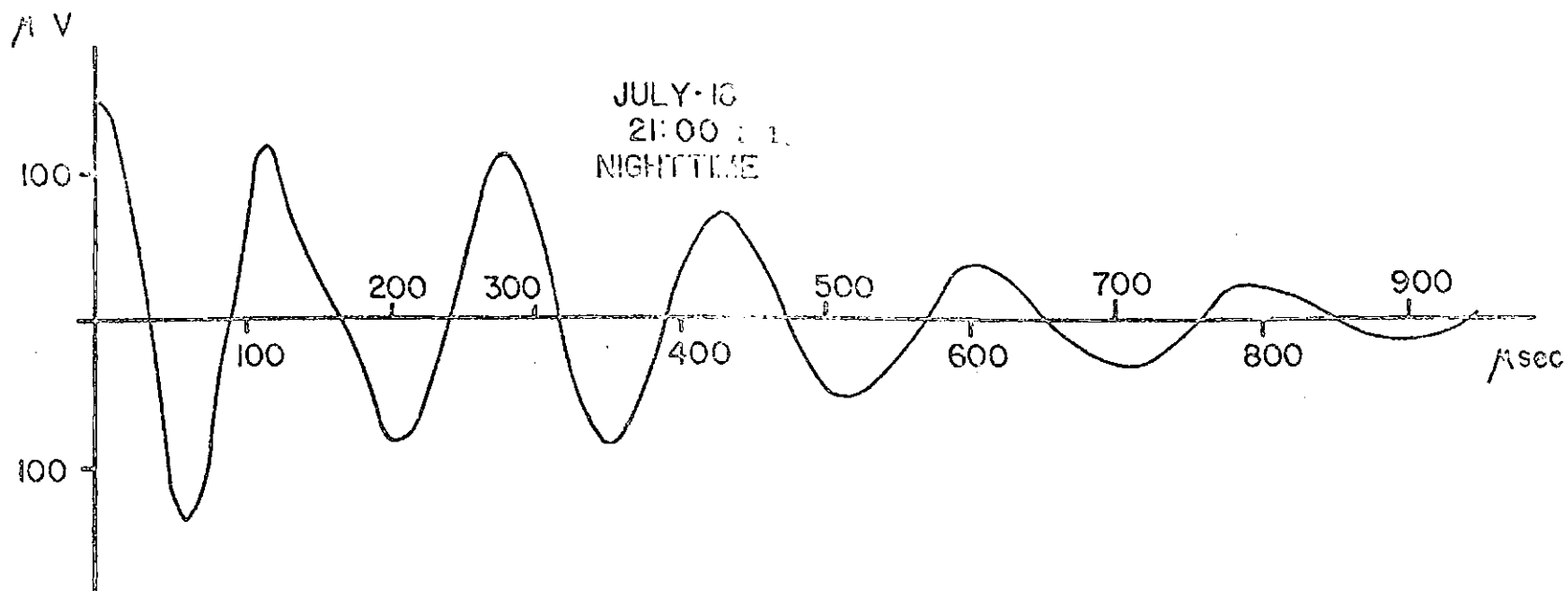
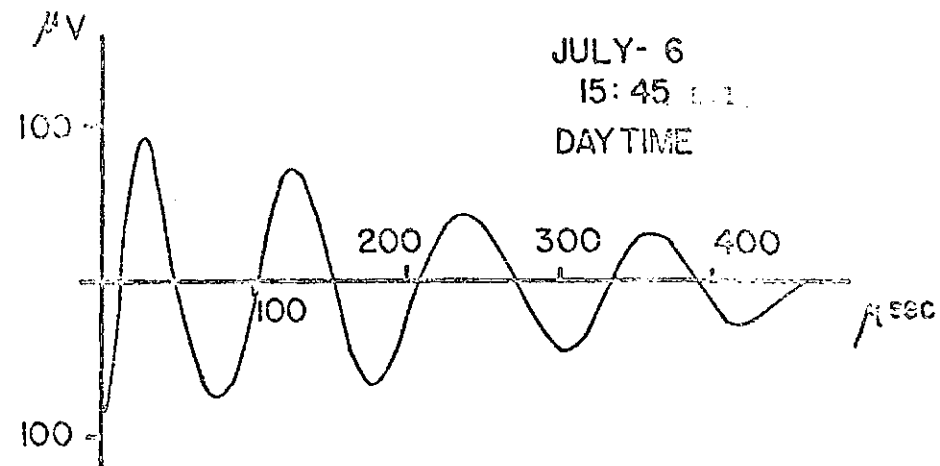


Fig. 6 - Typical Waveforms

$$T_n = \frac{d}{c} \left\{ \left[1 - \frac{\tau_n^2 c^2}{4h^2} \right]^{-1/2} - 1 \right\}$$

whete T_n and τ_n are in sec, c in m/sec, h in m and d in m, was modified, in order to make this expression easier to compute and it takes the form:

$$\tau_n = \frac{2.h.10^9}{c} \left[1 - \frac{1}{\frac{cT_n}{2} \left(1 + \frac{\tau_n}{d.10^9}\right)} \right]^{1/2}$$

where

- τ_n - quasi-period in μ sec
- T_n - delayed arrival in μ sec
- h - height of the waveguide in km
- c - velocity of light in m/sec
- d - distance to origin in km.

Theoretical curves were computed for the height of waveguide equal to 85 km for nighttime and equal to 70 km for daytime. These values of height were adopted because Hepburn's method, as was seen before, does not permit us to determine the distance and the height at the same time. The adopted values of height seem most reasonable for this case and are based on the results of the other workers. (Rao, 1968, Wait 1957a, b).

The distances were determined by the best fitting of the practical and theoretical curves. Fig.7 presents an example of such a determination. The dots present theoretical curves and the asterisks are the practical points. In this figure the practical points fit better with the theoretical curve for $d = 5000$ km. It is also seen that the precision in the determination of distances is worst for the greater distances than for the smaller distances due to the closeness of the theoretical curve as distance increases.

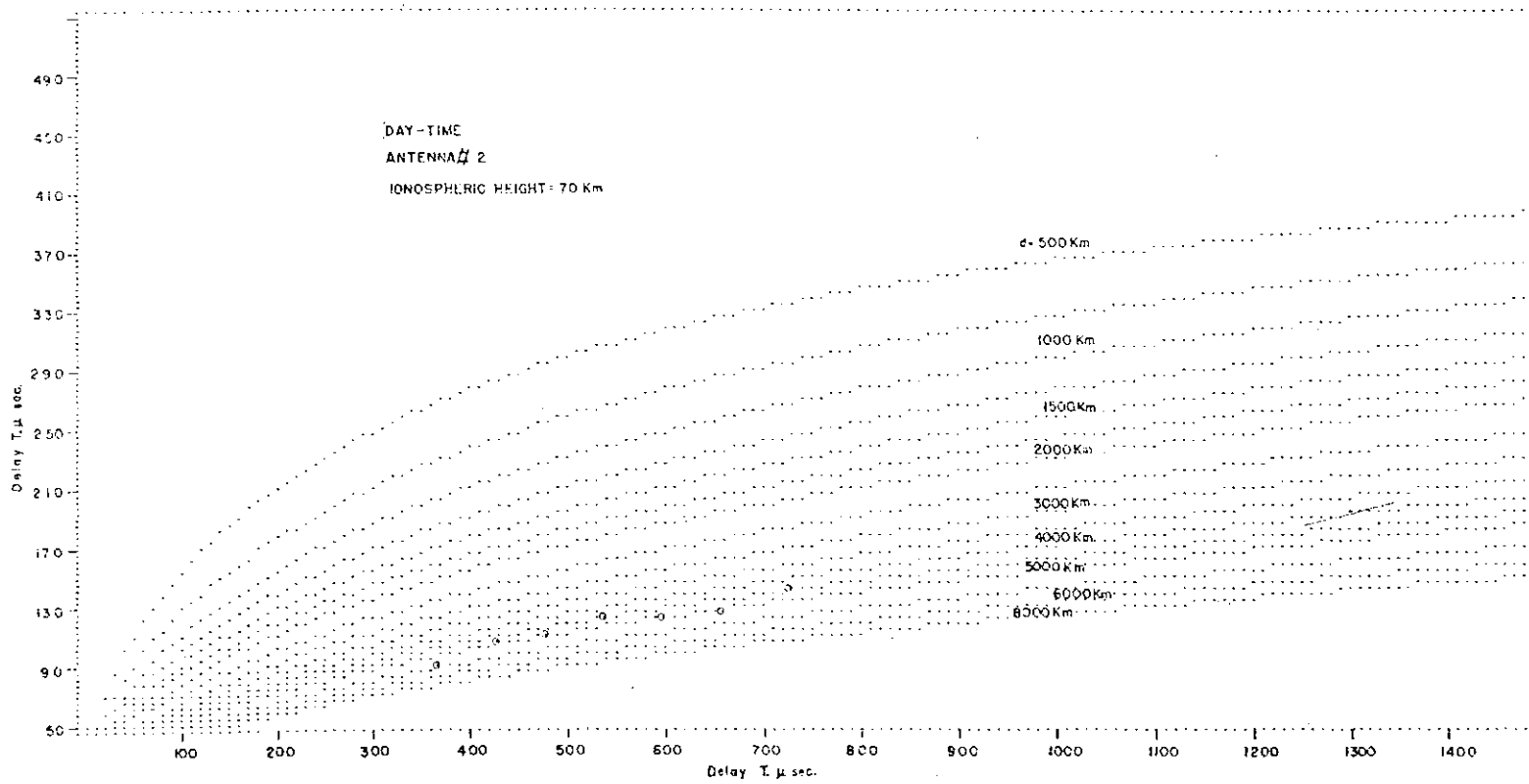


Fig. 7 - Distance Determination by comparison of experimental T_2 plots (Asterisk) with the theoretical curves.

4.3 - Fourier Analysis of Atmosphericics

Representing the atmospheric as a time varying function $E(t)$, Fourier spectrum $E(\omega)$ can be written as

$$E(\omega) = \frac{1}{2\pi} \int_0^T E(t) \exp(-i\omega t) dt$$

where ω is the angular frequency and T is the time length of received atmospheric.

In order to make possible this integration $E(t)$ was approximated in each semi-cycle by the following function:

$$E_i(t) = A_1^{(i)} \sin \omega_i t + A_2^{(i)} \sin 2 \omega_i t + A_3^{(i)} \sin 3 \omega_i t$$

where $\omega_i = 2\pi f_i = 2\pi \frac{1}{t_i}$ and t_i is the length of the i^{th} semi-cycle.

In the Appendix is presented the computer program which was used for determining the amplitude spectrum of the atmosphericics.

The amplitudes of each frequency of spectrum are expressed in dB over 1 $\mu\text{V-sec/m}$. The spectra were obtained in the range from 1 KHz to 50 KHz and in 1 KHz steps.

The spectra corresponding to the same distance were normalized as they were by Rao (1968).

Fig.8 presents some examples of such spectra. The energy content maximum of the atmospheric occurs in the range from 6 to 10 KHz depending on distance. In the VLF range (3 KHz to 30 KHz), the energy content of atmosphericics decreases as frequency increase. It is seen, also, from this spectra that the energy content of the atmospheric increases for frequencies lower than 3 KHz. This fact was also observed by Alpert et al (1967), and Taylor (1960b).

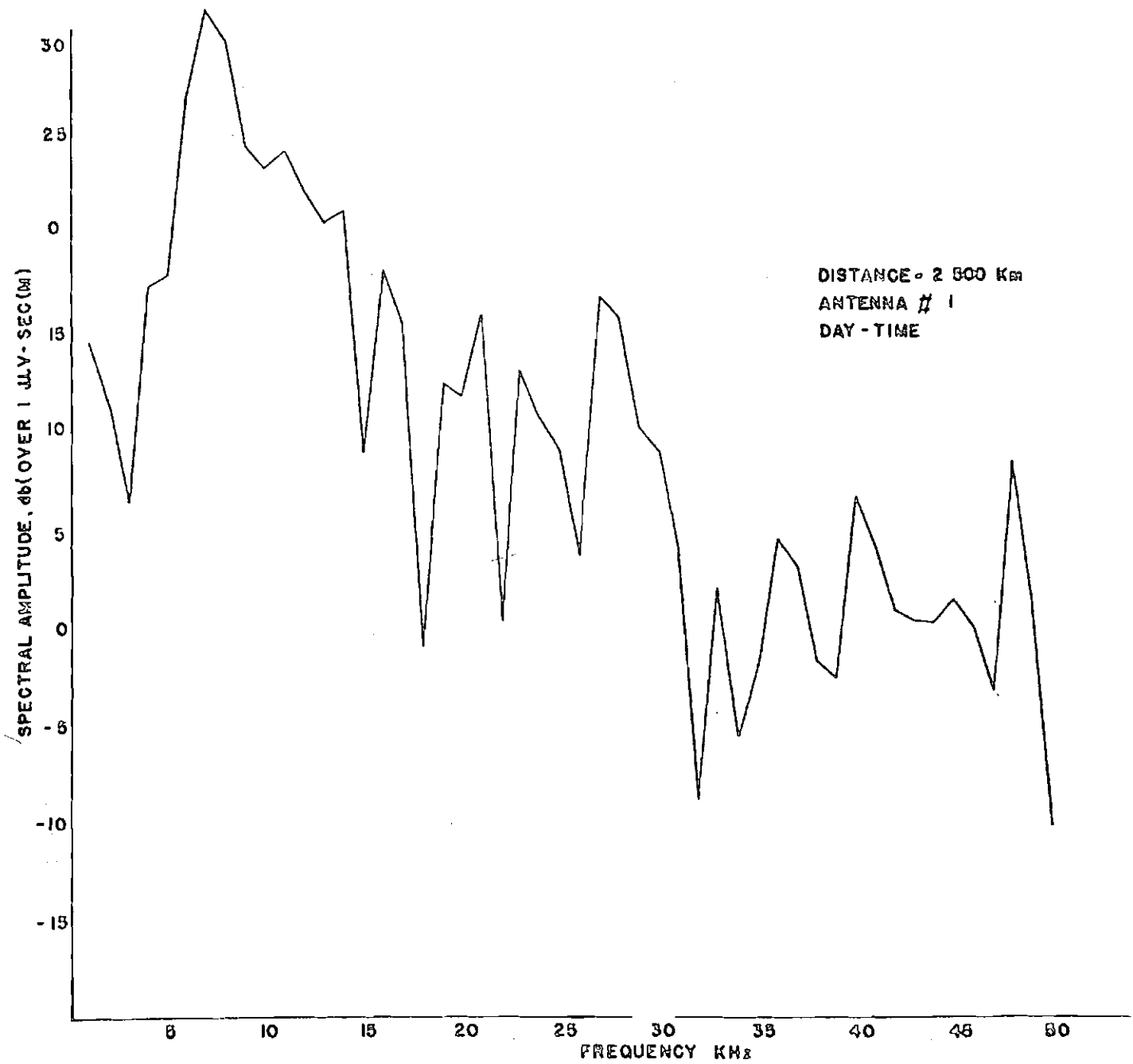


Fig. 8a - Normalized spectrum for 2500 km

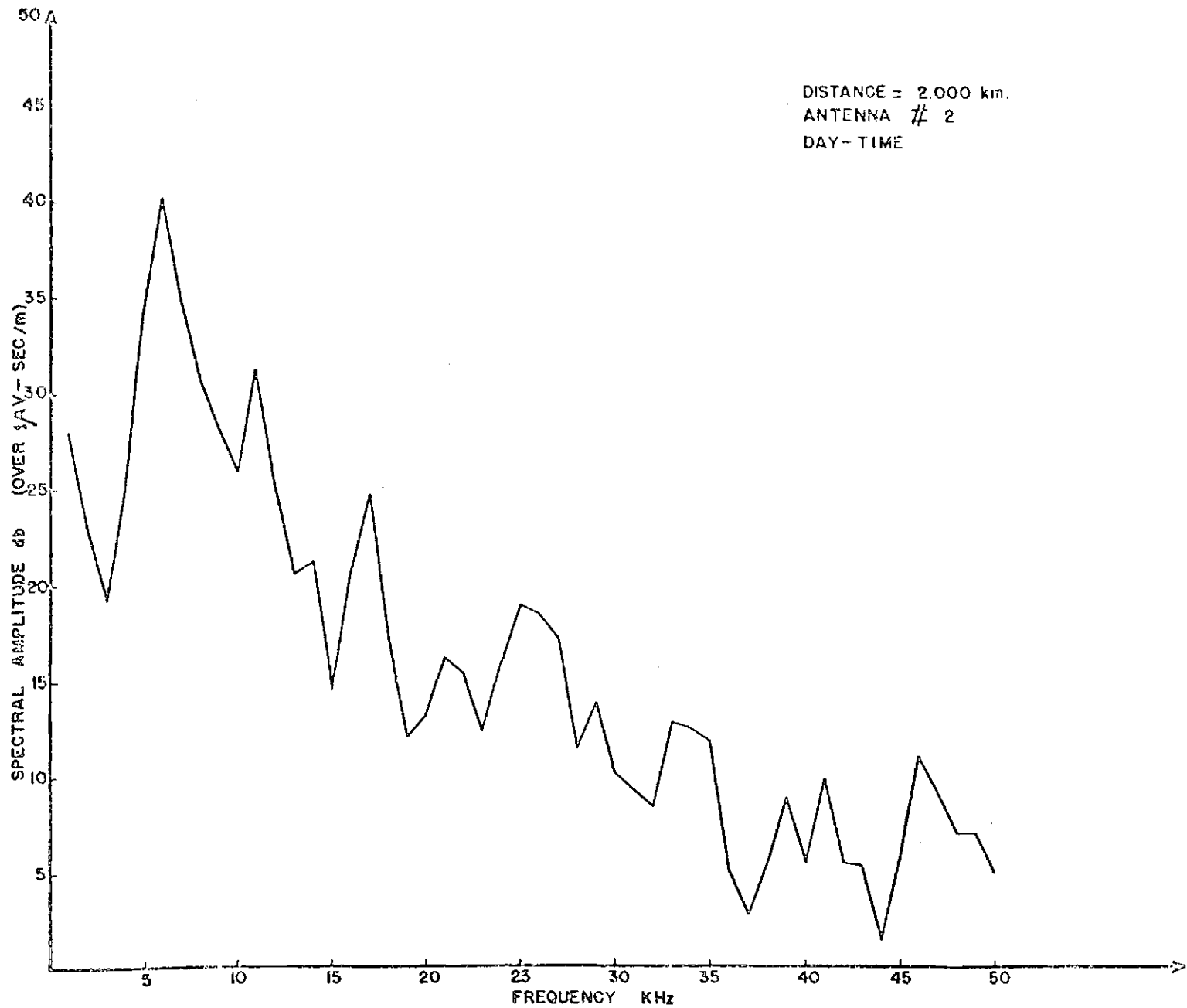


Fig. 8b - Normalized spectrum for 2000 km.

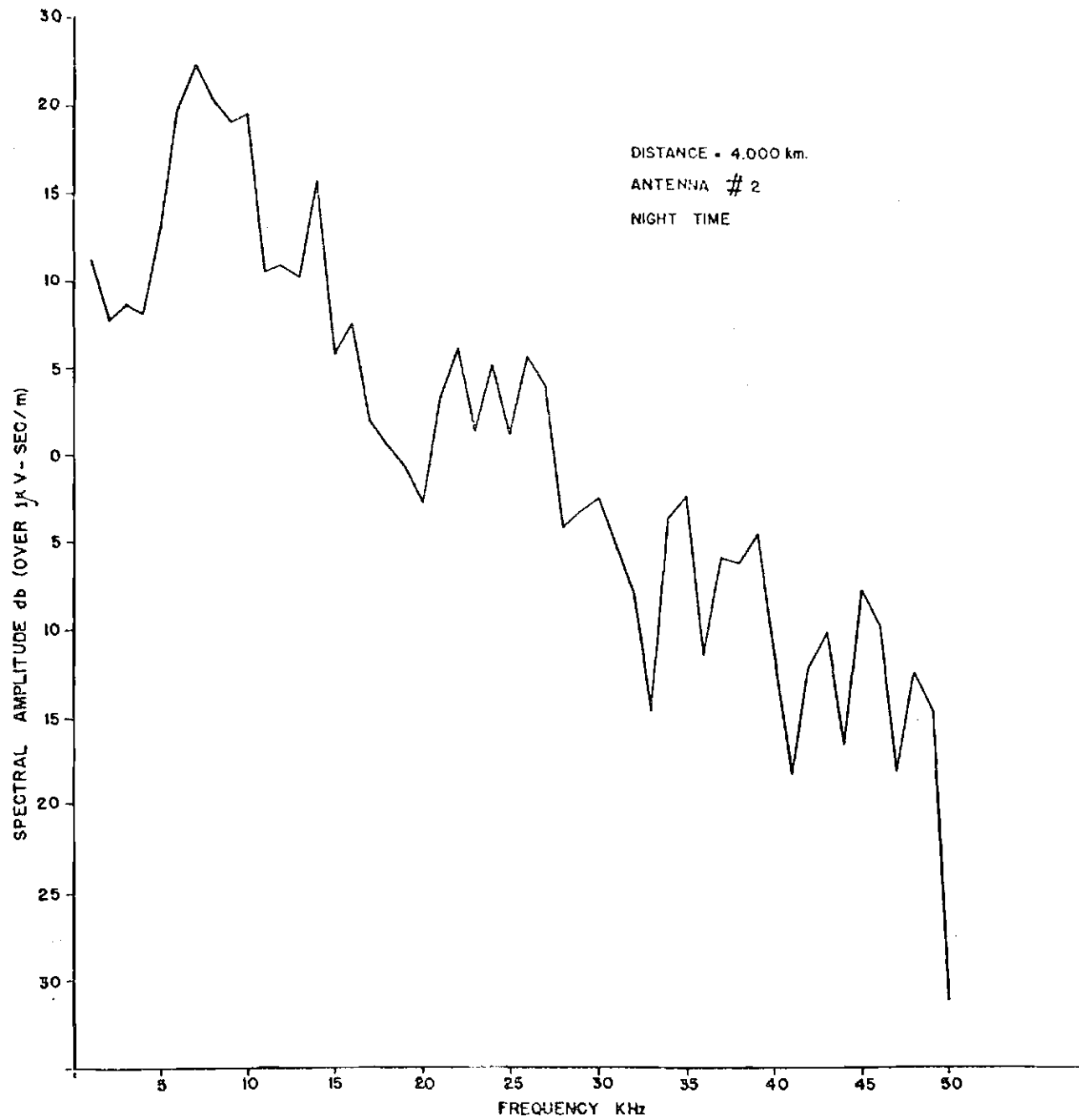


Fig. 8c - Normalized spectrum for 4000 km

5 - DISCUSSION OF RESULTS

We attempt to calculate the attenuation coefficient from the expression used by Rao (1968)

$$20 \lg E(\omega, d) + 10 \lg \left(a \sin \frac{d}{a} \right) = 20 \lg \frac{A(\omega) \cdot \lambda^{1/2} \cdot \Lambda}{h} - \alpha d$$

For each frequency the left side of above equation may be a straight line with slope α when plotted as a function of distance.

However the experimentally obtained points present a considerable amount of scatter. Fig.9 presents an example for determination of α for the frequency of 9 KHz. From this figure it is seen that it is impossible to fit a straight line for practical points. In his work, Rao (1968) also had referred to the scatter of the experimentally obtained points and a mean value of α was calculated by taking the average of the gradients between different pairs of distances. Although, even such a procedure did not allow the determination of α in this work.

This fact may be attributed to some effect at the source which makes the term $A(\omega)$ not constant or to the excitation factor Λ which depends on the height and properties of the reflecting layer, the electrical properties of the ground, the frequency and the order of the mode. It is possible that signals obtained were produced by interference between two modes and so the determination of α is more complicated than that assumed above.

The spectra of antenna 1 and 2 present the variation of the peak frequency with distance for nighttime and daytime with the same behavior as described by Wadehra & Tantry (1966), i.e. the peak frequency increases rapidly with distance up to 3000 km and then increases more slowly. In fig.10 we have the variation of peak frequencies with distance.

It was observed that nighttime amplitudes are greater than the daytime ones as was also observed by Obayashi (1960) and this indicates that

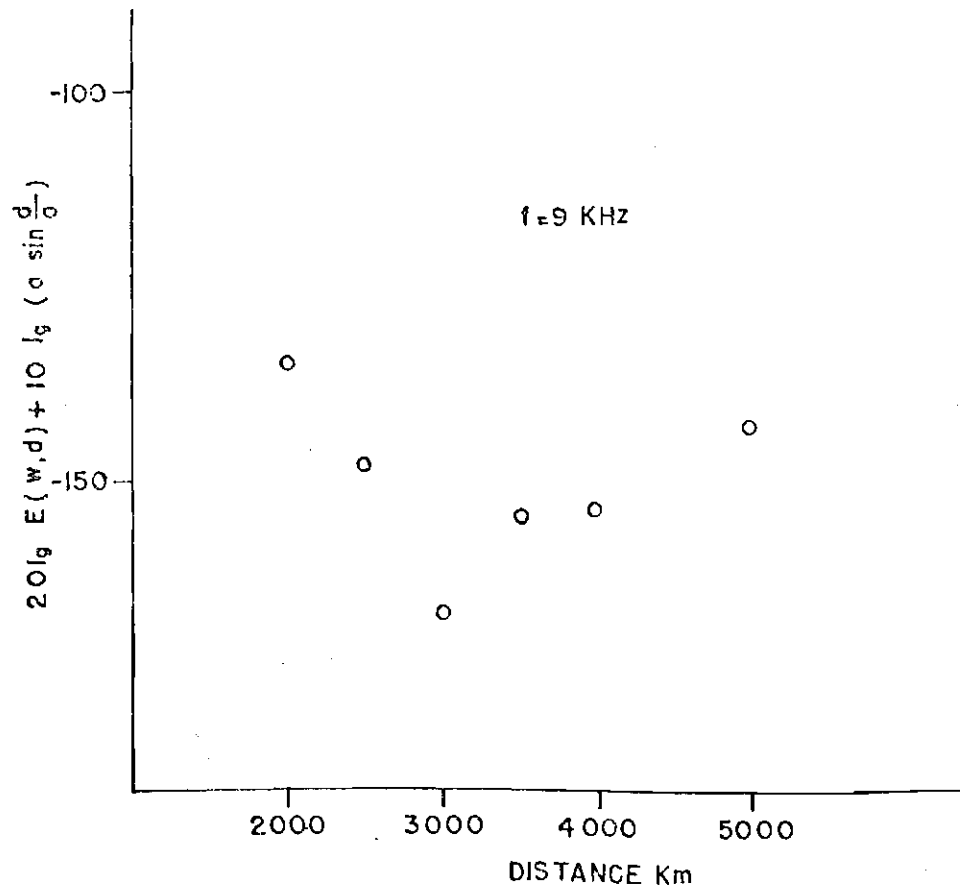


Fig. 9 - Determination of α

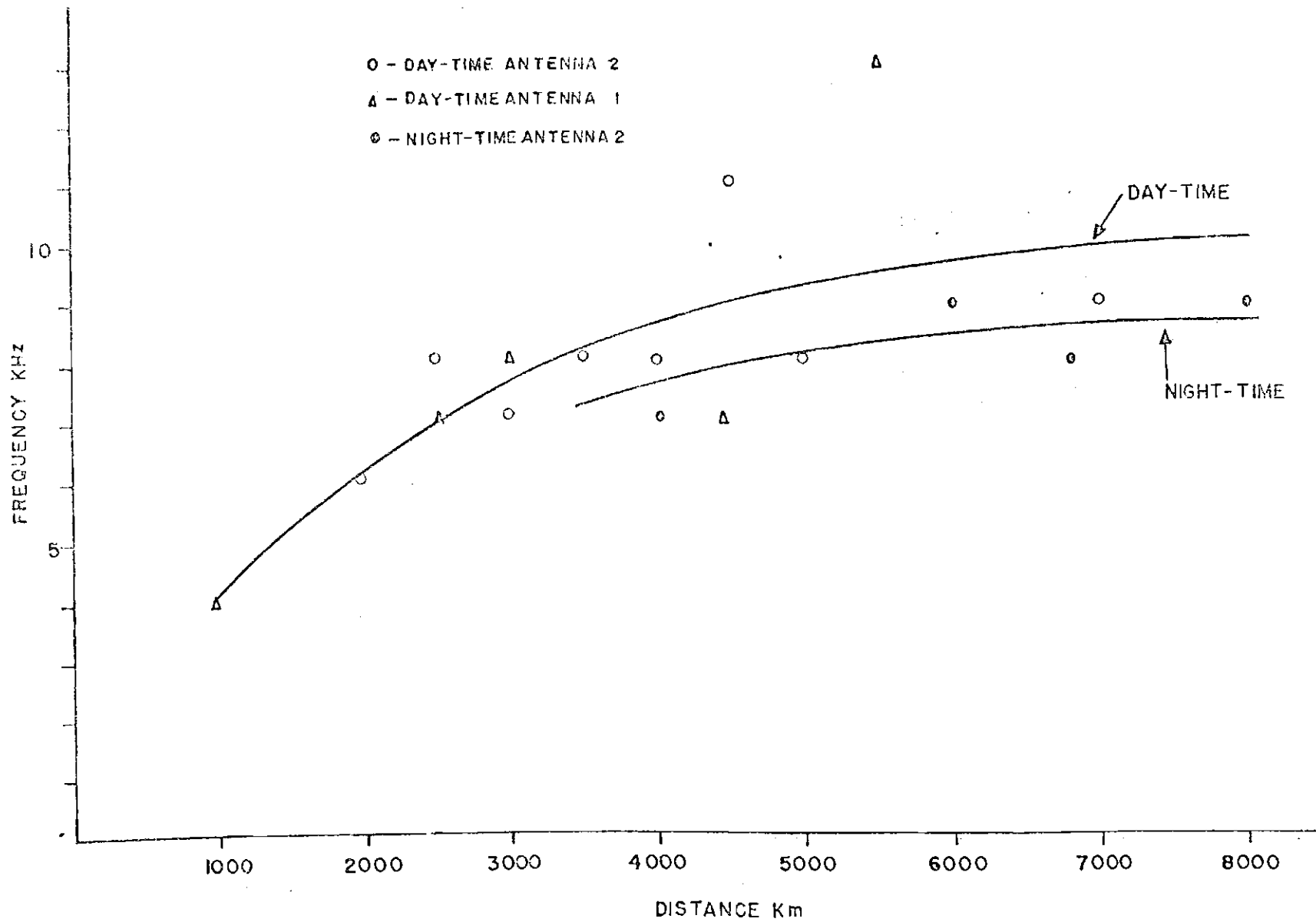


Fig. 10 - Variation of peak frequency of radio of atmospheric with distance

the attenuation at night is lower than during the day.

We intended, by using two loop antennas oriented at west-east and north-south, to study possible difference, in propagation in these directions. This difference could arise due to different paths of propagation caused by greater range of propagation over sea in one of them than in the other. However such a study proved very difficult; the location of our receivers relatively to sea was such that it was difficult to detect any difference in propagation paths. Very few atmospherics were recorded from antenna 1 and antenna 2 which present the same distance to the origin. When such records were available the frequency of the peak amplitude of the corresponding spectra were not the same and was difficult to make conclusions about the difference in the north-south and west-east propagation.

6 - CONCLUSIONS

1. The spectra of both antennas show a tendency for the peak frequency to vary with distance during the nighttime as well as the daytime.
2. The amplitudes of nighttime spectra are always greater than the amplitudes of daytime spectra denoting lower attenuation at night.
3. There is probably, an amplitude maximum in the ELF range.

APPENDIX

Let $F(t)$ be a function defined in the interval $[0, t_{\max}]$. Its Fourier spectrum may be written as

$$g(\omega) = \frac{1}{2\pi} \int_{-\infty}^{\infty} F(t) e^{-j\omega t} dt$$

or

$$g(\omega) = \frac{1}{2\pi} \int_0^{t_{\max}} F(t) e^{-j\omega t} dt$$

The above expression may be written as a sum of two components $g_1(\omega)$ and $g_2(\omega)$, so

$$g(\omega) = g_1(\omega) + g_2(\omega)$$

where

$$g_1(\omega) = \frac{1}{2\pi} \int_0^{t_{\max}} F(t) \cos(\omega t) dt \quad (1)$$

and

$$g_2(\omega) = j \frac{1}{2\pi} \int_0^{t_{\max}} F(t) \sin(\omega t) dt \quad (2)$$

Generally, our function $F(t)$ have a shape of a dampened sine wave. So, each semi-cycle of this sine wave can be approximated by the function

$$F(t) = A_1 \sin \omega_0 t + A_2 \sin 2\omega_0 t + A_3 \sin 3\omega_0 t$$

where $\omega_0 = \pi/\tau$ and τ - lenght of the semi-cycle in msec.

The values of the coefficients in the above expression can be determined by the values of the ordinates in each semi-cycle which corresponds to time $\tau/3$, $\tau/2$ and $2\tau/3$.

We can integrate expressions (1) and (2) in each semi-cycle and add them in the interval $[0, t_{\max}]$.

The amplitude spectrum of F (t) is

$$|F(t)| = \sqrt{g_1^2(\omega) + g_2^2(\omega)}$$

The following computer program determine the amplitude spectrum of F (t).

The input data necessary for this program are:

- IPER - data collection period
- IANT - antenna number
- NOME - name of the waveform
- N - number of semi-cycles in the waveforms
- TO - initial time, if there is displacement of waveform relatively to origin of time
- T(k) - length of each semi-cycle in milsec
- F1(k)- ordinate corresponding to time $\tau/3$ in each semi-cycle in mVolts
- F2(k)-ordinate corresponding to time $\tau/2$ in each semi-cycle in mVolts
- F3(k)-ordinate corresponding to tome $2\tau/3$ in each semi-cycle in mVolts
- XM(IF) - amplitude spectrum in dB over 1 μ Volt-sec/m

```
C   SPECTRUM ROSENFIELD CHAE 1969
C   TIME IN MILSEC,AMPLITUDES IN MILVOLTS
    DIMENSION A(3,20),F1(20),F2(20),F3(20),T(20),XM(50),TO(20),T1(20),
    T2(20),X(50)
    READ(5,1112)IPER
1112 FORMAT(I2)
    READ(5,1112) IANT
    READ(5,1006) NOME
1006 FORMAT(I5)
    IF(IPER.EQ.02) GO TO 500
    WRITE(6,1115)
1115 FORMAT(1X,14HPERIODO DIURNO)
    GO TO 20
    500 WRITE(6,1116)
1116 FORMAT(1X,15HPERIODO NOTURNO)
    20 WRITE(6,1117) IANT
1117 FORMAT(1X,6HANTENA,1X,I2)
    WRITE(6,1007) NOME
1007 FORMAT(1X,4HNOME,1X,I5)
    PI=3.1415926
    SQRT3=1.7320508
    READ(5,1000) N
1000 FORMAT(I3)
    READ(5,1001) TO(1)
1001 FORMAT(F5.3)
    READ(5,1002) (T(I),I=1,N)
1002 FORMAT(16F5.3)
    DO 7 K=1,N
    READ(5,1003)F1(K),F2(K),F3(K)
    F1(K)=F1(K)/300.
    F2(K)=F2(K)/300.
    7 F3(K)=F3(K)/300.
1003 FORMAT(3F8.2)
    DO 2 J=1,N
    A(1,J)=(F1(J)+F3(J))/SQRT3
    A(2,J)=(F1(J)-F3(J))/SQRT3
    2 A(3,J)=A(1,J)-F2(J)
    DO 3 IF=1,50
    FIF=2.*PI*FLOAT(IF)
    X(IF)=IF
    T1(1)=0.
    T2(1)=T(1)-TO(1)
    G1=0.
    G2=0.
    DO 4 J=1,N
    DO 5 I=1,3
    W=FLOAT(I)*PI/T(J)
    U=A(I,J)/(W**2-FIF**2)
    C=COS(W*TO(J))
    S=SIN(W*TO(J))
```

```
CT2=COS(W*T2(J))
ST2=SIN(W*T2(J))
CT1=COS(W*T1(J))
ST1=SIN(W*T1(J))
CWT1=COS(FIF*T1(J))
SWT1=SIN(FIF*T1(J))
CWT2=COS(FIF*T2(J))
SWT2=SIN(FIF*T2(J))
P1=ST2*SWT2-ST1*SWT1
P2=CT2*CWT2-CT1*CWT1
P3=ST2*CWT2-ST1*CWT1
P4=CT1*SWT1-CT2*SWT2
G1=G1+U*(C*(W*(-P2)+FIF*(-P1))-S*(W*P3+FIF*P3))
5 G2=G2+U*(-C*(W*P4+FIF*P3)+S*(W*P1+FIF*P2))
T1(J+1)=T2(J)
TO(J+1)=T2(J)
4 T2(J+1)=T2(J)+T(J+1)
XM(IF)=SQRT(G1**2+G2**2)/(2.*PI)
3 XM(IF)=20.*ALOG10(XM(IF)/.001)
CALL GRAF(X,XM,50)
WRITE(6,1120)
1120 FORMAT(1H1)
STOP
END
```

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