THE TEMPERATURE OF THE IONOSPHERE IN THE AURORAL ZONE

by

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1. INTRODUCTION.

In the absence of man-made interference and atmospherics, it has been found possible to measure radio frequency thermal radiation from the ionosphere at medium frequencies.

The temperature of the ionospheric D-region has been previously measured by this method at 2 Mc/s in temperate latitudes (Pawsey, McCreddy and Gardner, 1951, and Gardner, 1954). In a companion paper (Gardner and Pawsey, 1953), the absorbing structure of the lower ionosphere was determined from D-region echoes to find the heights to which the temperatures applied. However, it was felt that conditions might be very different in an auroral zone. Echoes from the D-region (apparently) of much greater intensity and at only half the height of those observed by the above authors, were reported at Macquarie Island which is close to the centre of the auroral zone.

Experiments were carried out by the author at the Australian National Antarctic Research Station at Macquarie Island as a member of the 1956 party. The apparent D-region echoes mentioned above were found to be not from the ionosphere, however, but from sea waves (Dowden, 1957). This thesis describes radio noise experiments at a number of medium frequencies carried out at a site twenty miles from the Main Station where interference was negligible.

This thesis is divided into two parts. The first part describes observations of high accuracy at 2 Mc/s using a half wave dipole aerial to provide a direct comparison with Gardner's (1954) observations in a temperate latitude (30°S). The second part describes observations of lower accuracy at a number of frequencies (7 Mc/s., 5.6 Mc/s., 2.0 Mc/s. and 450 Kc/s.), using a terminated "long wire" aerial.

PART A.

2. Experimental Procedure.

2.(i) Site and Aerial.

The site of the experiment was near Hurd Point, Macquarie Island (55^{o}S, 159^{o}E, geomagnetic Lat. – 62^{o}) on peat slopes at the foot of the plateau which forms most of the island. The nearest source of man-made interference of any kind was the Station, twenty miles north. This was completely negligible at this distance. The nearest broadcast stations and atmospherics were several hundred miles to the north.

The aerial used for this part of the experiment was a half wave dipole (2 Mc/s) suspended from wires attached to various crags. The aerial itself was about 20^{o} off horizontal, about a quarter wave above the ground (120 feet) and more than a wave length from the cliffs of the plateau. The aerial passed directly over the recording hut from which it was fed by an 80 ohm shielded twin line. Far less trouble from winds was encountered than was expected, the wind run at the site being only about half that recorded at the Main Station.

2.(ii) Determination of Ionospheric Temperature.

The equipment used and the recording technique was similar to that described by Gardner (1954). The receiver was automatically connected to a dummy aerial for three minutes every seven and a half by a chronometer operated relay. The sensitivity of the recording system was checked with a diode noise generator three times a day. The impedance of the aerial was accurately rematched to that of the dummy usually once each day. *

The difference "t" between the noise level produced by the aerial and that by the dummy was read from the record in arbitrary units. A typical record showing this difference is in Fig. I (a). The temperature T_i of the relevant part of the ionosphere can be found from this by an expression of the form: $T_i = At + T_d.$

* See appendix for complete description of sensitivity calibration, matching, attentuation and aerial efficiency measurements.

The scale factor "A" is the reciprocal of the product of the sensitivity of the receiving system (in recorder units per degree K), the attenuation of the aerial feed and matching unit, and the "efficiency" of the aerial. T_d is the ambient temperature of the dummy aerial, provided this is not greatly different from that of the matching unit, feed and ground. It was found that the temperature in the recording hut was usually within a degree Kelvin of that of the outside air and ground.

The efficiency of the half wire dipole was calculated from the Sommerfeld and Renner (1942) analysis using measured values of ground conductivity and aerial height. The figure obtained was also checked by the impedance method (Gardner, 1954). Fortunately, for an aerial height of about a quarter wave length, the efficiency is nearly unity and varies only slowly with height and ground constants. The adopted value was 0.90 ± 0.05. The total attenuation of matching unit and aerial feed was easily measured by the noise generator.

The errors of reading, sensitivity calibration and matching and those due to errors in aerial efficiency and attenuation measurement could be expected to give both systematic errors and random erros of about 5° K.

3. OBSERVATIONS.

3.(1) General.

The seven observing periods from May to November (1956) are shown in the table below.

TABLE 1.

Period	Interval	Duration
MAY-JUNE	30th May - 15th June	17 days
JUNE-JULY	26th June - 16th July	21 days
JULY	20th July - 30th July	11 days
AUGUST	23rd Aug. - 31st Aug.	9 days
SEPT.-OCT.	26th Sept. - 17th Oct.	22 days
OCT.-NOV.	24th Oct. - 3rd Nov.	11 days
NOVEMBER	13th Nov. - 20th Nov.	8 days

The observations made in the second period (JUNE-JULY), showed strong interference from natural sources and so were not used for temperature measurement. During daylight hours random thermal noise was usually the only component seen on the "aerial" trace - similar in all but level to the "dummy" trace.

Occasionally, local cloud discharges produced small isolated spikes. Atmospherics propagated by the ionosphere from tropical latitudes were usually not observed until near sunset. Within an hour of their first appearance they were usually sufficiently strong to hold the recorder pen off scale continuously. A typical record showing this is Fig. 1 (a). The length of observing time (about sunrise to sunset) varied from about eight hours in MAY-JUNE to sixteen in NOVEMBER.

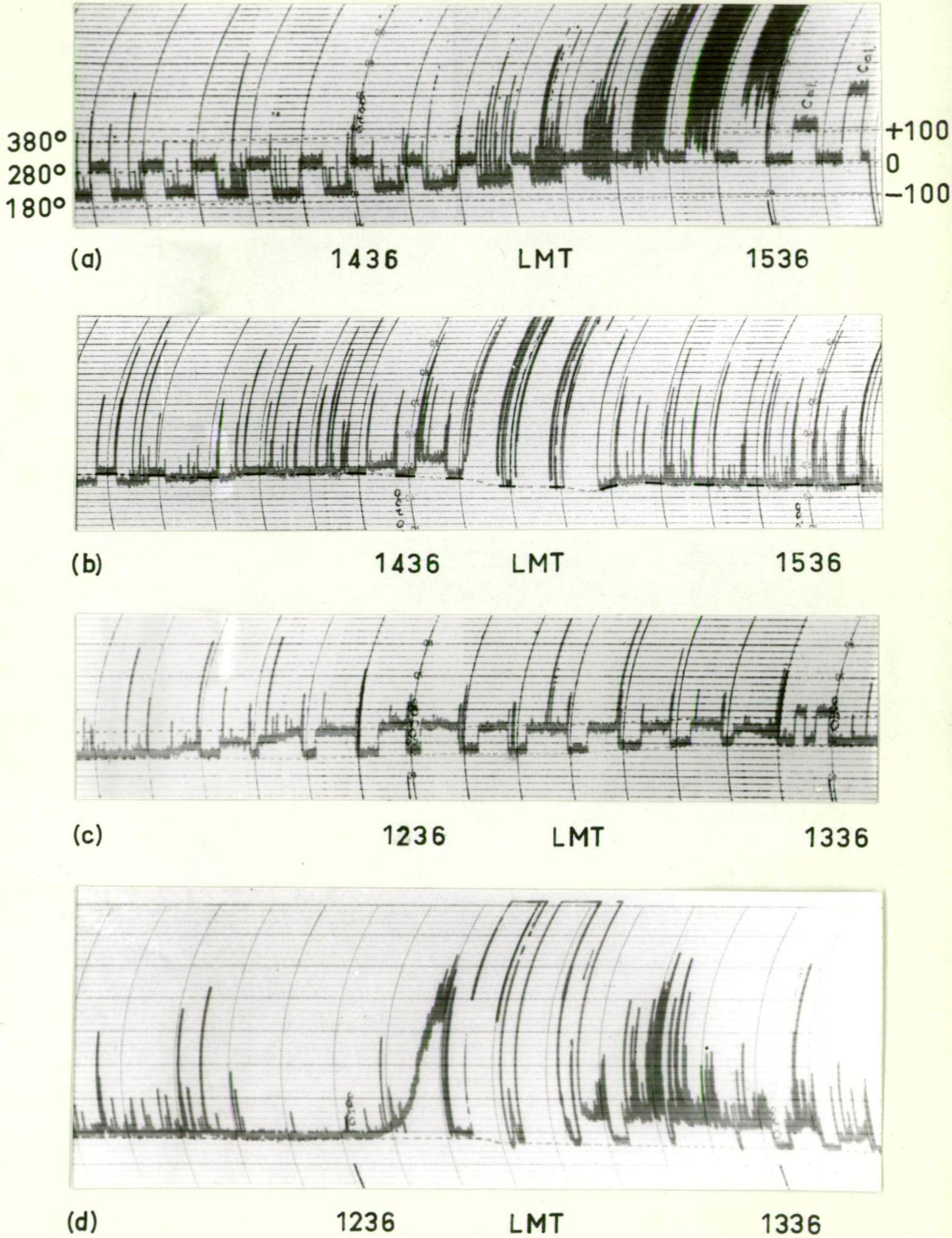


Fig. 1. (a) Typical record showing onset of atmospherics (June observations). (b) Tropospheric noise. The level gradually increases through the dummy level. During the off scale period a strong "crying" noise was heard (June). Other examples observed in August (c) and July (d). Snow storms occurred nearby during these observations.

3.(ii) OBSERVED TEMPERATURES.

Histograms of the observed temperatures in each of the observing periods are shown in Fig. 2. All observed temperatures have been incorporated except those obviously affected by atmospherics or precipitation static. These can be directly compared with temperate latitude observations (Gardner, 1954). For the undisturbed months both sets of observations show distributions of temperatures starting from about 200°K, rising to a maximum around 220° - 230°K then a slower tailing off to around 260° - 280°K. This undisturbed component of the distribution persists in most of the histograms. The temperatures of these first maxima (usually the modes), for each period in the Macquarie Island observations are usually within five or ten degrees of those for the corresponding periods, in the Urisino (Gardner, 1954) data. Again in agreement with lower latitude results, the temperatures observed in the winter periods are lower than those observed towards summer.

Hourly mean temperatures for each observing period are shown plotted in Fig. 3. Only observations unaffected by atmospherics, as gauged by the repetition rate and amplitude of the atmospherics, have been used in calculating these means. The spread of observations in each period can be gauged from the corresponding histogram in Fig. 2. The diurnal variations shown are on the verge of significance. This can also be judged from the corresponding histograms. The SEPT/OCT. and the OCT.-NOV. periods were fairly undisturbed and also a large number of observations were made and so the slight diurnal effect for these two periods is probably significant. A slight rise in temperature of about five degrees occurs about noon. The temperatures near sunrise and sunset appear higher than those during the day by some ten degrees. This may be due to an error in judgement of the effect of weak atmospherics on temperature levels. Another possibility is that, as mentioned in a later section, on disturbed days observed temperatures are higher and the absorption of the D-layer is higher so that the onset of atmospherics occurs later. Hence on these days values unaffected by atmospherics can be made nearer sunrise and sunset (as happened in the NOVEMBER period) so that at these times the means may not be truly representative.

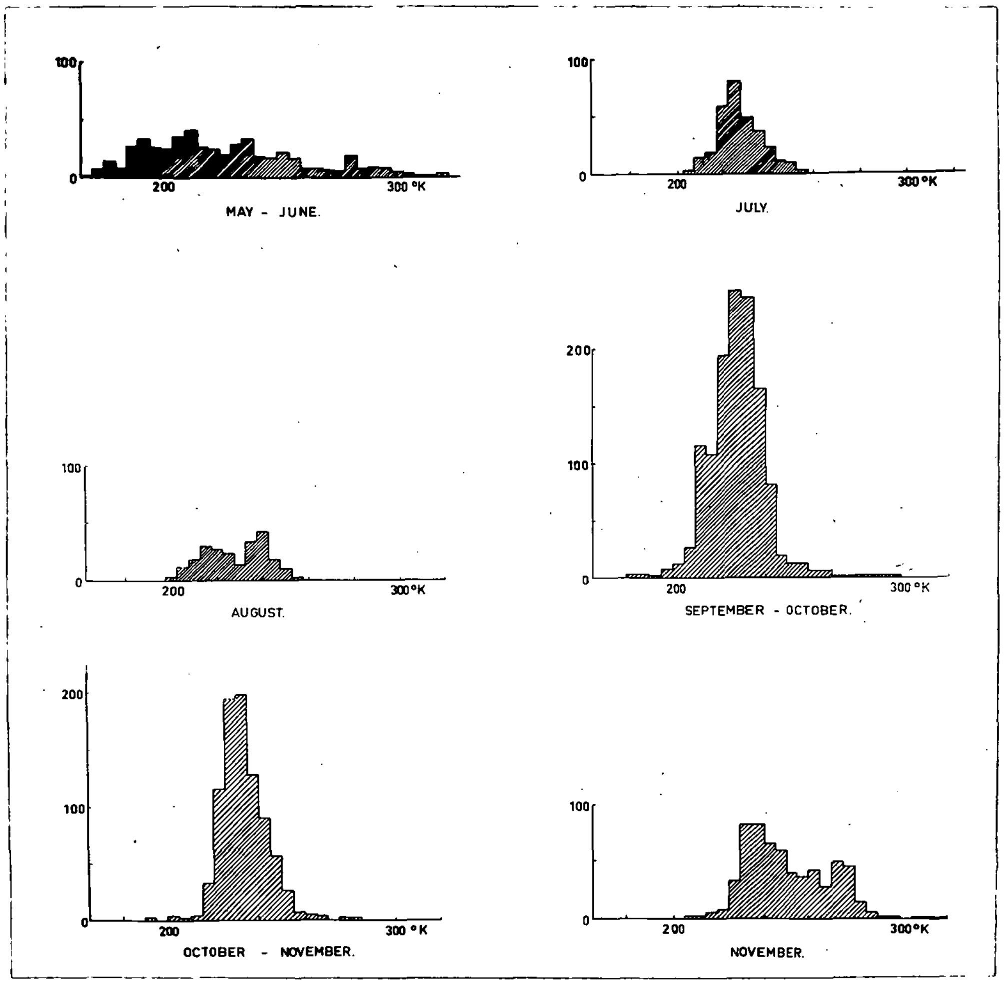


Fig. 2. Histograms of all observations of noise levels for the different observing periods. Only those obviously affected by atmospherics or precipitation static were excluded. The average atmospheric free time (length of observing time) in a day was much larger in the summer months than in the winter (see Fig. 3).

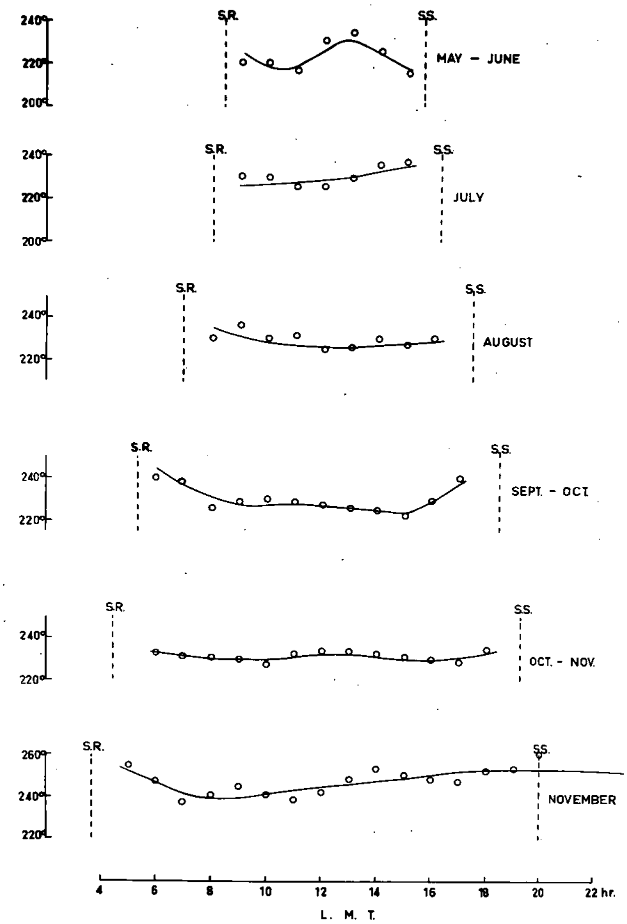


Fig. 3. Diurnal variation of ionospheric temperature for the different observing periods. Only observations unaffected by atmospherics have been considered. Times of sunrise and sunset are as indicated. The spread in observations can be gauged from the corresponding histograms (Fig. 2).

A much larger effect (about 50 degrees) at Urisino and a still larger one at Rankin Springs (Gardner, 1954) has been attributed to ionospherically-propagated interference. In which case the effect should be less at a site more remote from interference as at Macquarie Island.

3.(iii) TEMPERATURES OBSERVED DURING DISTURBED PERIODS.

During the November period some severe magnetic storms occurred. Two maxima are seen in the histogram (Fig.2.) suggesting a disturbed component. To separate this component histograms were computed of temperatures observed during quiet periods (three hourly periods for which the magnetic planetary index (K_p) was ≤ 1), three hourly periods for which $K_p > 6$, and during sudden ionospheric disturbances (S.I.D's). This is shown in Fig.4(a). The correlation of observed ionospheric temperature with K_p is quite significant and sufficient to show up in a Temperature - K_p plot as shown by the plotted histograms in Fig. 4(b). Up to about $K_p = 5$ the temperatures appear undisturbed but are increased by a further increase of K_p.

The effect of S.I.D's, however, does not appear significant. This may partly be due to competition with magnetic disturbances occurring at about the same time. Also by an unhappy chance the equipment was not operating during many of the large S.I.D's so that information is limited. At the Urisino site Gardner (1954) reported "a very definite rise of about 40°K during an S.I.D." This is not observed at Macquarie Island.

The relative effects of corpuscular and ultra violet disturbances on the D-layer can also be gauged from ionosonde records. During strong disturbances the absorption of the D-layer is increased so much that no echoes appear on the ionosonde, a condition often called a "blackout" (total ionospheric absorption). In low and medium latitudes blackouts are associated with strong ultra violet bursts (as indicated by S.I.D's), but the Macquarie Island ionosonde shows a much stronger dependence on corpuscular effects (K_p). To test this the tabulated hourly measurements from the Macquarie Island ionosonde records for 1956 (Dowden, 1957) were examined and the K_p value for each daylight hour for which total ionospheric absorption was reported was noted. The number of such backout hours for each K_p value is shown in the top histogram in Fig.5. The lower histogram shows the K_p - distribution for the whole year (1956).

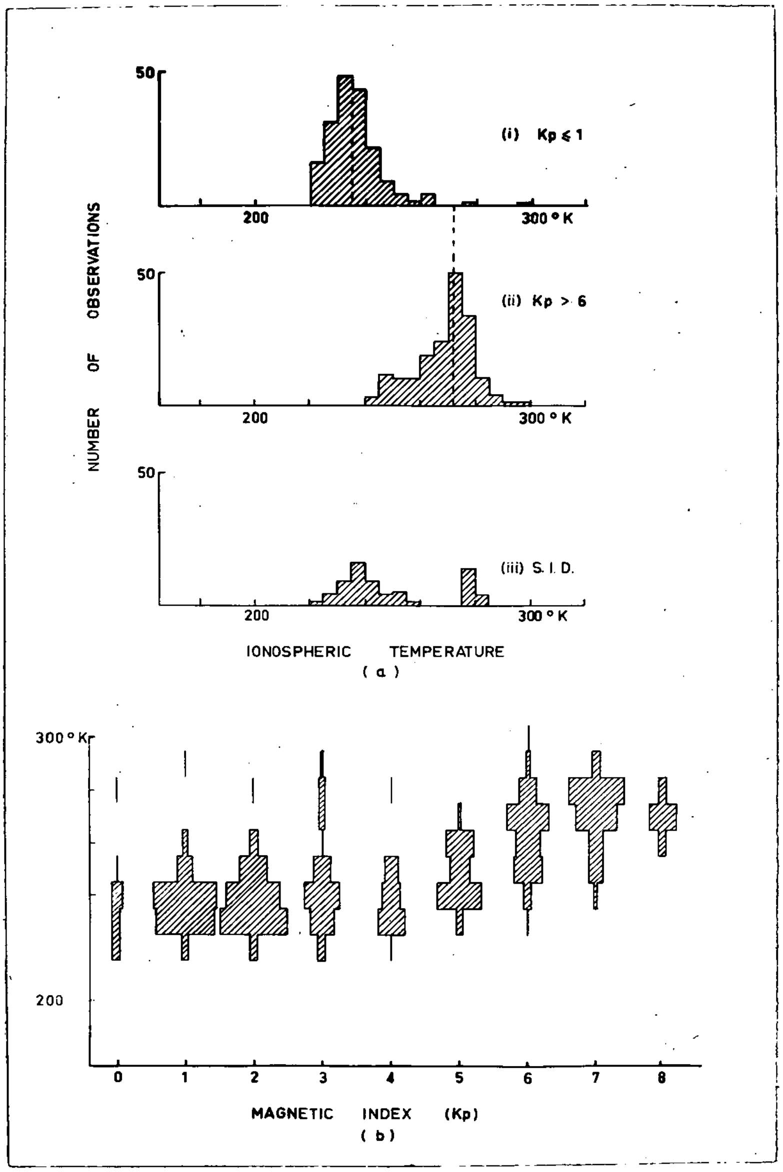


Fig. 4 (a) Histograms showing ionospheric temperatures during (i) three hourly periods for which $K_p \leq 1$, (ii) periods for which $K_p > 6$ and (iii) periods of S.I.D.'s.

(b) Histograms of temperatures for each value of K_p.

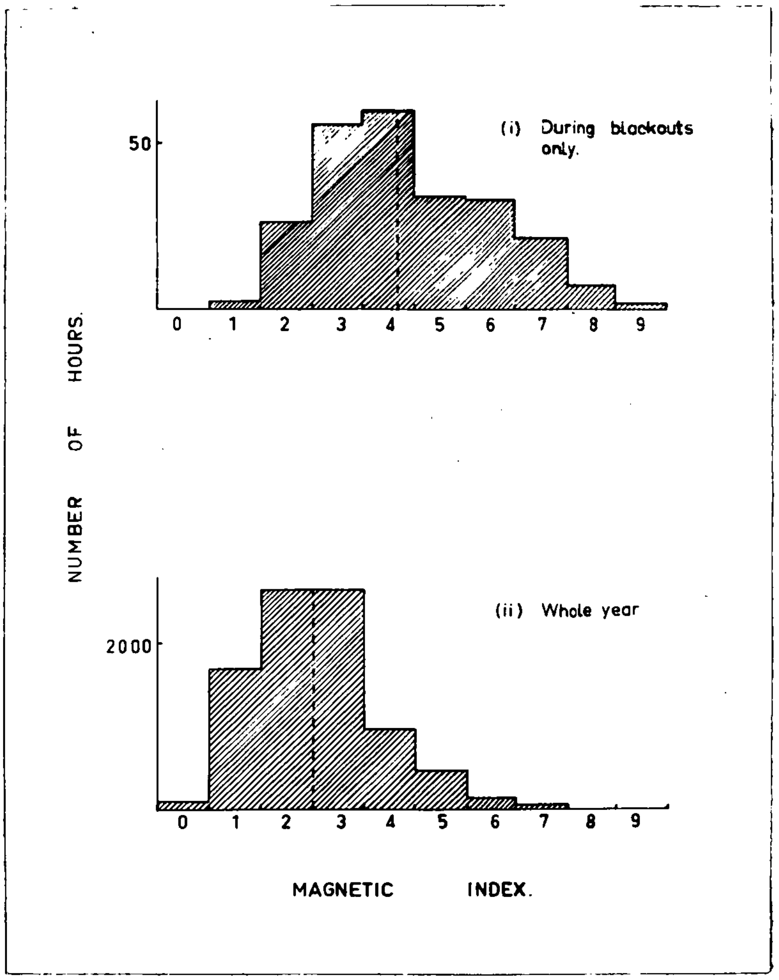

(i) During blackouts only.
50
0
1
2
3
4
5
6
7
8
9
NUMBER OF HOURS.
(ii) Whole year
2000
0
1
2
3
4
5
6
7
8
9
MAGNETIC INDEX.

Fig. 5.

3.(iv). NON THERMAL NOISE.

As mentioned above, owing to the remoteness of the site, no trouble was experienced with man made interference and comparatively little from atmospherics. The other forms of tropospheric noise listed by Gardner were observed, viz. rain static, cloud static and wind static (types B, C and E in Gardner). In addition, snow static was occasionally observed. This is aurally similar to rain static and is probably caused by the same mechanism. However, the intensity is very much higher, possibly approaching a volt per metre.

The "type D" noise mentioned by Gardner - random or "white" noise producing a slow increase in apparent temperature of up to 100°K - was experienced occasionally. It was usually recognised and so not included in temperature measurements. There was no known associated atmospheric disturbance at Urisino but at Macquarie Island this noise was only observed on days when snow storms occurred. Two examples are shown in Fig. I (c) and (d). In (d) a fairly sharp increase in noise level begins at about 1240 local time. Ten minutes later the pen is off scale during snow static. After the snow static ends the record shows this "type D" noise together with weak (probably local) atmospherics. This suggests that at Macquarie Island this noise may be distant snow static.

Another type of non-thermal noise similar to type D in character was observed in the winter. This produced an excess temperature of about 50°K to 100°K or more <u>for days on end</u>. Because of this the MAY-JUNE histogram (Fig.2) is strongly spread and shows temperatures above 300°K. In the JUNE-JULY series (which was not used for temperature measurement for this reason) on ten days out of the sixteen the apparent temperature <u>was above ambient</u> (280°K) <u>all day</u>. There was no associated tropospheric disturbance, but atmospherics appeared comparatively strong all day nearly every day of this series. Investigations of this noise at a number of different frequencies are described in part B of this thesis.

4. DISCUSSION.

Although a considerable volume of the ionosphere contributes to the measured temperature, only a small range of heights contributes appreciably. The emission takes place in a "non deviating" region. Hence the contribution from a layer of thickness Δx, at height x, will be given by the product of its electron temperature T, its emissivity and the attenuation along the path to the observer $\exp\left\{\int_0^x \kappa \, dx'\right\}$ where κ is the power absorption coefficient. From the reciprocal laws of emission and absorption the emissivity equals $\kappa \Delta x$ (Pawsey et al, 1951). The equivalent temperature T_i is obtained by integrating along the ray and is given by :

$$T_i = \int_0^\infty T \kappa \exp\left\{-\int_0^x \kappa \, dx'\right\} dx$$

This can be integrated if T and κ are known in terms of x. Gardner and Pawsey (1953) have published curves N_e, κ_o / N_e and $(\kappa_o - \kappa_x)/N$ as functions of height for Sydney where N_e is the electron density and the subscripts o and x refer to the ordinary and extraordinary propagation modes. From these we find that κ is negligible below about 65 km, has a subsidiary maximum at about 72 km and then increases approximately linearly from about 75 km. If we neglect the subsidiary maximum then, we can write

$$\begin{aligned} \kappa &= A\,(x - H) & & x > H \\ &= 0 \qquad (\text{i.e. } A = 0) & & x < H \end{aligned}$$

where H is about 75 km. The integrated absorption is then

$$\begin{aligned} \int_0^x \kappa \, dx' &= \int_0^H A\,(x - H)\, dx' + \int_H^x A\,(x - H)\, dx' \\ &= \left[\frac{A}{2} x'^2 + A H x'\right]_H^x \\ &= \frac{A}{2}(x - H)^2 \end{aligned}$$

On a typical day Gardner and Pawsey (1953) found that the integrated absorption for the ordinary ray up to a height of 80 km and back was about $3\frac{1}{2}$ nepers. For the single trip then

$$\int_0^{80} \kappa \, dx' = 7/4$$

$$\text{i.e. } \frac{A}{2}(80 - 75)^2 = 7/4$$

$$\text{i.e. } \qquad A = 7/50$$

The expression containing κ then becomes

$$\kappa \exp\left\{-\int_0^x \kappa \, dx'\right\} = Ah \exp\left\{-\frac{A}{2}h^2\right\}$$

where $h = x - H$. Differentiating, we find this factor is a maximum when:

$$Ah\cdot - Ah\cdot \exp\left\{-\frac{A}{2}h^2\right\} + A \exp\left\{-\frac{A}{2}h^2\right\} = 0$$

i.e. when $Ah^2 = 1$

or $h = A^{-\frac{1}{2}}$

$\approx 3\,\text{km}$

Suppose the ionosphere to be isothermal. The "T" can be taken outside integral and the remainder, summed over the range x = o to $x = \infty$, would then be unity. As seen above the maximum contribution to this integral is from the height 78 km (h = 3). It would be interesting to find the contribution of (say) the five kilometers about this maximum (say, the range 76 km - 81 km). The integral is then :

$$\int_{76}^{81} \kappa \exp\left\{-\int_0^x \kappa \, dx'\right\} dx = \int_1^6 Ah \exp\left\{-\frac{A}{2}h^2\right\} dh$$

$$= \left[\exp\left\{-\frac{A}{2}h^2\right\}\right]_1^6$$

$$= \exp\left\{-\frac{7}{100}\right\} - \exp\left\{-\frac{7.36}{100}\right\}$$

$$= 0{\cdot}93 - 0{\cdot}08 = 0{\cdot}85$$

Hence 85% of the temperature contribution comes from a five kilometer thick layer.

It must be remembered that the $\kappa = A(x - H)$ approximation is rather rough and that the ionosphere above Macquarie Island might be considerably different from that above Sydney. Nevertheless the above results indicate that the temperature observed from the ground is essentially that of a thin layer of effectively constant temperature. Also the above calculations have been based on the ordinary ray only. Actually two such layers will exist: one for the ordinary ray and one for the extraordinary ray. The temperature "seen" by a single horizontal dipole will then be an average one of these two layers. The behaviour of two thin layers in the following discussion will be very similar to that of only one layer, and so the extra complication of an extraordinary layer will be omitted.

Physically, the ionosphere is too opaque for the aerial to "see" beyond the layer and too transparent for it to "see" any ionosphere below the layer. To a first approximation, therefore, the measured temperature may be taken as that of a certain level of the ionosphere corresponding to the "optimum opaqueness" (i.e. where $K \exp\left\{-\int_0^x K\,dx'\right\}$ is maximum) of the ionosphere which depends on the electron density N and collision frequency ν. If either N or ν is increased the optimum opaqueness will occur at a lower level. Rocket measurements (Rocket Panel, 1952) show that the temperature of the ionosphere varies with height so that we would expect a variation of N or ν to produce a variation in measured temperature.

The diurnal variation of the height of the D-region has been investigated by Bracewell et al. (1951), with reflections of very long radio waves and found to fit the formula

$$h_\chi = h_{\chi=0} + A(t)\log_e \sec\chi$$

where χ is the altitude of the sun and A (t) is a constant having the value of about six during the equinoces. They also found a difference between the summer and winter noon heights and a sharp decrease in height during an S.I.D. (sudden phase anomaly). Using this, the expected diurnal height variation during an equinox at Macquarie Island together with summer and winter and S.I.D. effects is shown in Fig. 6. The temperature scale on the right was taken from the Rocket Panel (1952) temperature-height curve and applies approximately in the height range 53 Km to 74 Km (210° K to 270°

Consider the possibility that all the variation of observed temperature is due to variation in electron density (as indicated by height measurements at very low frequencies) alone. The predictions of this hypothesis and the experimental results are compared in Table 2 below.

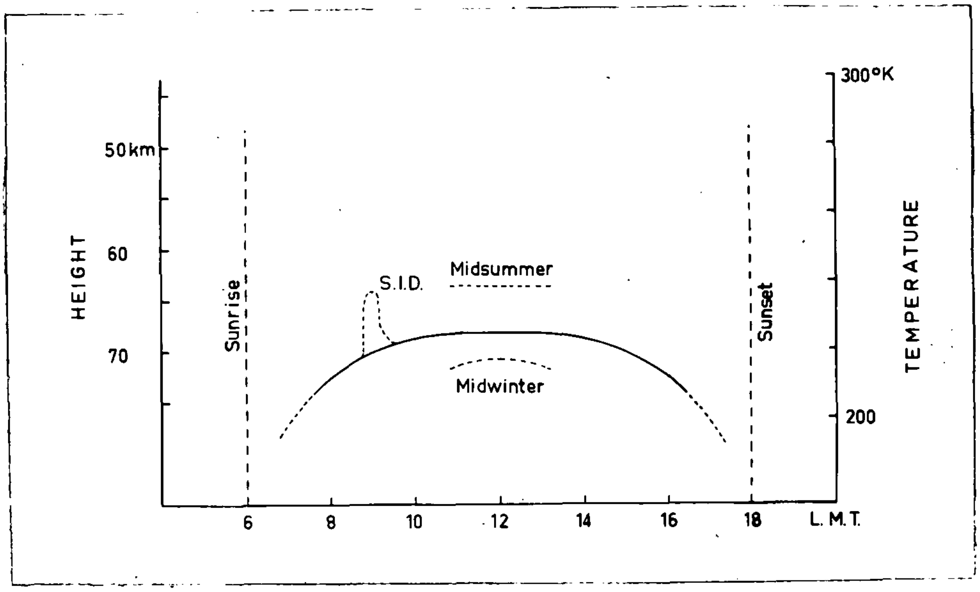


Fig. 6. Diurnal variation of the height of the D-layer at the equinoxes. Midday summer and winter heights are shown dotted. The temperature scale on the right was compiled from the Rocket Panel curve. The height decrease during a sudden phase anomaly indicates the temperature rise expected from an S.I.D.

TABLE 2.

Theory	Experiment
Midsummer temp. : 240°K. Equinox temp. : 225°K. Midwinter temp. : 215°K.	November : 235°K. Sept.-Oct.: 227°K. July : 225°K.
Noon temperatures. - Some 10° - 15° higher than early morning and late afternoon.	About 5° difference. (Oct. - Nov. series).
Rise of 15° - 20° during an S.I.D.	40° - 50° during strong magnetic activity.

The general agreement here is quite good. The predicted diurnal variation barely shows up probably owing to the atmospheric effect discussed in Section 3. A somewhat stronger temperature rise occurs during disturbed periods than that predicted. This was also the case during S.I.D's at Urisino (40°K). For large increases of ionization, however, the optimum opaqueness level, being a function of N and ν is depressed more than the isopycs* since the collision frequency ν increases exponentially with decrease in height, giving a stronger rise in measured temperature.

The large increases in measured temperature as well as the total ionospheric absorption condition ("polar blackout") is strongly correlated with the magnetic index K_p . It has been suggested (Mitra, 1948), that "polar blackouts" are due to the bombardment of charged particles causing a rise in temperature (in the lower ionosphere) which increases the collision frequency and thereby also the absorption.

The decrease in reflection coefficient due to absorption is :

$A = e^{-\rho}$ where $\rho = \int \kappa \, ds$ is the integrated absorption, up and down, in nepers.

* isopyc : surface through points of equal electron density.

Neglecting the effect of the magnetic field, and also making the simplifying assumption that most of the absorption occurs in a relatively thin stratum and that the various factors governing the absorption do not vary appreciably within this stratum, we find, for $f > \nu$:

$$A \approx \exp\left\{-\frac{C N_e \nu}{f^2}\right\}$$ C is constant.

At f_{min}, $A = A_0$ and echoes on lower frequencies are not recorded.

A typical value of f_{min} during daylight hours on undisturbed days is about 2 Mc/s. During severe storms f_{min} rises above 13 Mc/s, the upper frequency limit of the recorder.

If N_e' and ν' are the values during the "polar blackout" then :

$$A = A_0 = \exp\left\{-\frac{C N_e \nu}{2^2}\right\} = \exp\left\{-\frac{C N_e' \nu'}{13^2}\right\}$$

i.e., if $N_e' = N_e$ then the collision frequency must increase by a factor of about 40.

Now the collision frequency is proportional to the average velocity of the ions which is in turn proportional to the square root of the absolute temperature. Hence the observed increase in apparent temperature of the ionosphere during strong magnetic activity of about 15%, even if an actual temperature rise, could only increase the collision frequency some 8%.

In order to explain the apparent temperature variation, it was suggested above that the layer responsible for most of the temperature contribution moves to a height where the actual temperature is different. If the layer responsible for most of the absorption moves to a lower ~~highe~~ height (where the collision frequency is considerably greater) during disturbed periods, the absorption will increase.

It was shown by Gardner and Pawsey that most of the absorption occurs between about 80 and 90 kms. Very low frequency studies (Bracewell et al.) show a sudden decrease in reflection levels ("sudden phase anomaly") coincident with S.I.D's. The results of this paper show a considerable increase in apparent temperature which would correspond to a lowering of the height of the layer responsible for most of the temperature contribution to below 50 kms (using the Rocket Panel (1952) curve). Hence during severe magnetic storms it is not unreasonable to suppose that in high latitudes the isopyc which is normally around 85 km. is depressed to around 50 km. Now Gardner and Pawsey have shown that the collision frequency decreases approximately exponentially with height, and using their formula we get $\nu = 1 \times 10^6$ at 85 km and $\nu = 4 \times 10^7$ at 50 km. That is, the electron density of the absorption region remains essentially unchanged, but the collision frequency increases by a factor of about 40, which is the order of magnitude needed to explain the more severe "polar blackouts".

5. CONCLUSIONS.

The temperature of the lower ionosphere has been measured in an auroral zone, using similar techniques to those used by Gardner (1954) in a temperate region. The temperature range and seasonal variation of temperatures have been found to be the same as in temperate regions. Owing to the much lower intensity of atmospherics and complete freedom from man made noise, temperatures have been measured from sunrise to sunset, and the higher temperatures during early morning and late afternoon in Gardner's results were not apparent. The observed temperature variations can be similarly explained in terms of a fixed temperature height scale.

No large increases in temperature have been observed during S.I.D.'s, but during strong magnetic activity temperatures have been found up to 50°K above normal. This, and the very large increase in absorption ("polar blackout") experienced in auroral latitudes during magnetic storms is thought to be due to bombarding auroral particles causing a strong increase in ionozation and thus a descent of the region of absorption to levels of much greater collision frequency rather than a direct heating effect caused by these particles producing higher collision frequencies.

The persistent non-thermal noise experienced in midwinter will be discussed in Part B of this thesis.

PART B.

1. INTRODUCTION.

The measured temperature of the ionosphere, being effectively that of a relatively thin band of "optimum opaqueness", should be a function of the exploring frequency used, since the height of this band is dependent on frequency. This part of the paper describes attempts made to measure the temperature on frequencies of 7 Mc/s., 5.6 Mc/s., 2 Mc/s. and 450 Kc/s.

2. EXPERIMENTAL PROCEDURE.

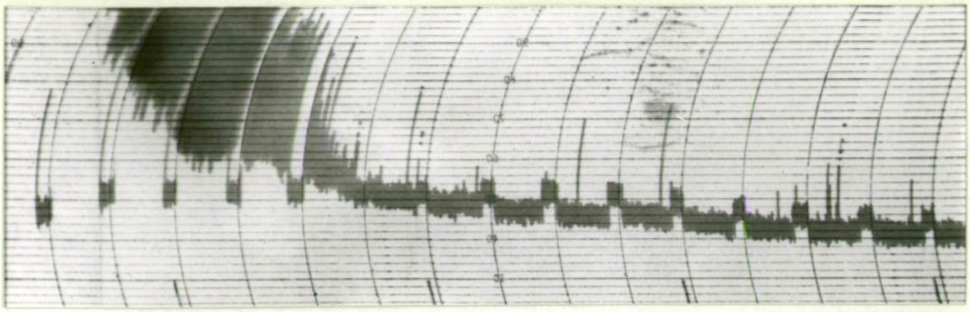
The site of the experiment and the equipment used was the same as that described in Part A of this paper. As only one equipment was available, only one frequency could be used at a time. Observations were therefore made over a period of about a month for each frequency. The same aerial was used for all frequencies : a 1200 foot "long wire" aerial, terminated by its characteristic impedance strung from the hut up a steep slope at an average angle of about 34^{o}. The average height of the wire above the slope was about 30 feet.

The efficiency of the long wire aerial was estimated in three different ways : by direct measurement at 450 Kc/s, from continuity with the half wave dipole results at 2.0 Mc/s., and from the Sommerfeld and Renner (1942) analysis.

The measurement was made by comparing the power received (p_o) when a noise generator was applied at the receiver end of the long wire to that (pt) when the noise generator was applied at the termination. Ideally, the impedance of the termination, that of the receiver and the characteristic impedance of the aerial will all be equal and purely resistive. The receiver and the termination are, therefore, interchangeable. When power 2 p_o is applied to the receiver end of the aerial, p_o will be absorbed by the receiver and p_o by the aerial. Of this, p_t is absorbed by the termination so that, neglecting losses, $p_o - p_t$ is radiated. Hence efficiency is :

$$\alpha = \frac{p_o - p_t}{p_o}$$

The three methods gave rough agreement. The Sommerfeld and Renner analysis which could only be expected to give rough agreement for a long wire aerial, was used mainly to find the way the efficiency should vary



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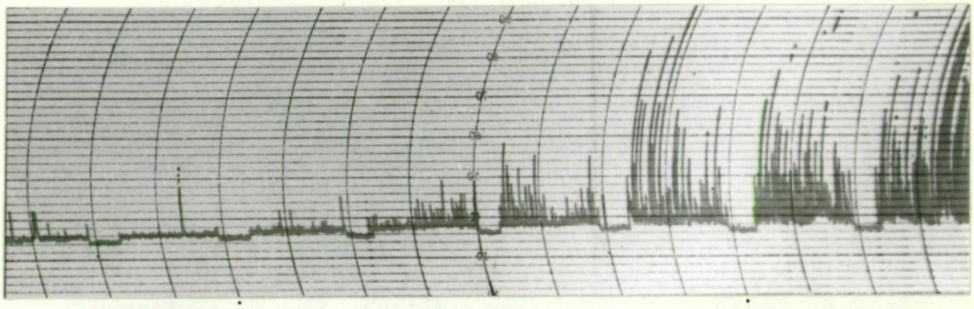
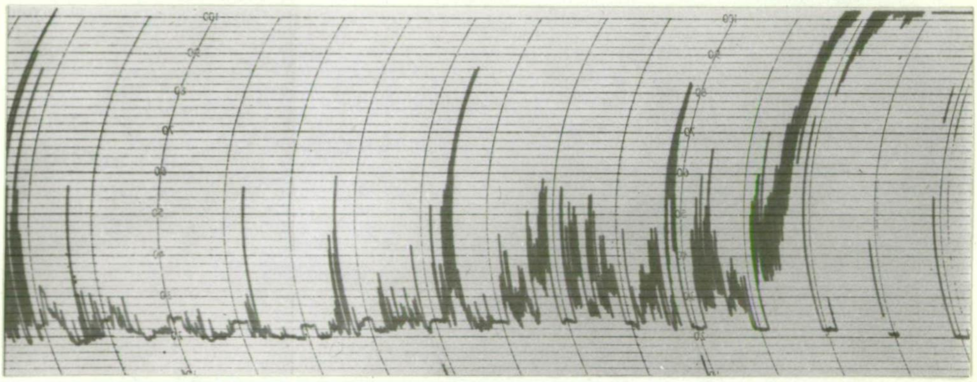


(b) 1506 LMT 1606



(c) 1136 LMT 1336

Fig. 7. (a) 5.6 mc/s record (March) showing decay of atmospherics some two hours after sunrise. Noise level is below that of dummy as on 2 mc/s observations. (b) 450 kc/s record (May) at the onset of atmospherics. The noise level was always above the dummy level (c) 7 mc/s record (April) showing onset of cosmic noise as foF2 gradually falls below 7 mc/s.

with frequency.

The aerial efficiency and average scale factor A (see Part A, section 2.(ii)) and the probable error of A is given in Table 3 below.

TABLE 3.

Frequency	Aerial Efficiency	Scale Factor "A"	Error in Temp.
450 Kc/s.	0.1	200	Factor of 2.
2 Mc/s.	0.3	6	5% (10–15°).
5.6 Mc/s	0.8	2	5% (10–15°).
7 Mc/s.	0.8	4	10% (20–30°).

The resulting percentage error in the difference temperature (measured ionospheric temperature less ambient ground temperature), is that of A. This difference is relatively small for the higher frequencies so that the measured ionospheric temperature is reasonably accurate. The equivalent temperatures measured at 450 Kc/s. were of the order of thousands of degrees. The receiver sensitivity was very low and the aerial efficiency was low and consequently not accurately known, so that at this frequency equivalent temperature estimates could be in error by a factor of two. At frequencies away from 2 Mc/s. (for which the equipment was designed), the attenuation of the matching unit was high, and the reduced output of the noise generator had to be taken into account.

3. OBSERVATIONS.

3.(i) 450 Kc/s. measurements.

A typical record at this frequency is shown in Fig. 7 (b). The difference temperature, as here, was always positive. The onset of atmospherics was not nearly as sharply defined as at 2 Mc/s., nor were daytime records as free from atmospherics. The equivalent ionospheric temperatures were remarkably high, ranging from about 300°K to about 2500°K. The accuracy of these estimates is not high, so that equivalent temperatures could be half or double those indicated. But even examination of the records alone show that the overall receiver gain was considerably less than it was at 2 Mc/s. and that the aerial temperature was always above ambient, so we cansafely say that the equivalent temperature at 450 Kc/s. is several times that of the <u>actual</u> temperature of the ionosphere.

The distribution of intensity of this non-thermal noise expressed as equivalent temperature is shown in Fig.8. As most of the expected error is in the scale factor A the errors will be systematic (and constant) rather than random, so that the shape of the distribution as shown should be fairly true. Only observations taken during periods for which the magnetic index was less than four or greater than six are included in this histogram (this was so about 90% of the time). This indicates that the high noise intensities are associated with magnetically quiet periods. On the other hand, atmospherics appeared stronger and more frequent on days of high noise intensities.

<u>3.(ii). 2 Mc/s. measurements.</u>

These were made (using the long wire aerial), immediately before the 2 Mc/s. half wave dipole measurements (Section A) to provide a comparison between the long wire and dipole measurements. Any differences in the results can, therefore, be attributed to the differences in the directivity of the two aerials. At this frequency the inclined long wire was $2\frac{1}{2}$ wave lengths long giving a horizontal major lobe in the southern direction, whereas the horizontal dipole had a vertical lobe.

The distribution obtained with the long wire aerial (Fig.8) shows a greater spread to higher equivalent temperatures compared with the dipole - obtained distribution (JULY, Fig.2). Some of this spread was day to day variations but variation of about a hundred degrees also often happened within about an hour.

<u>3.(iii) 5.6 Mc/s. and 7 Mc/s. measurements.</u>

There was no non-thermal noise of the type discussed above on either of these frequencies. The 5.6 Mc/s. histogram (Fig.8) is not significantly different from the 2 Mc/s. dipole one for SEPT.-OCT.(Fig.2). At 7 Mc/s the temperatures appear about ten degrees lower which is barely significant. Cosmic noise was sometimes apparent at 7 Mc/s when the critical frequency of the F layer (fO F 2) went below 7 Mc/s. The very high intensity of cosmic noise made it easily recognisable, so that it has not been included in the noise levels scaled for temperature measement. The trace in Fig.7 (c) shows the transition period when the critical frequency is slowly dropping through the operating frequency.

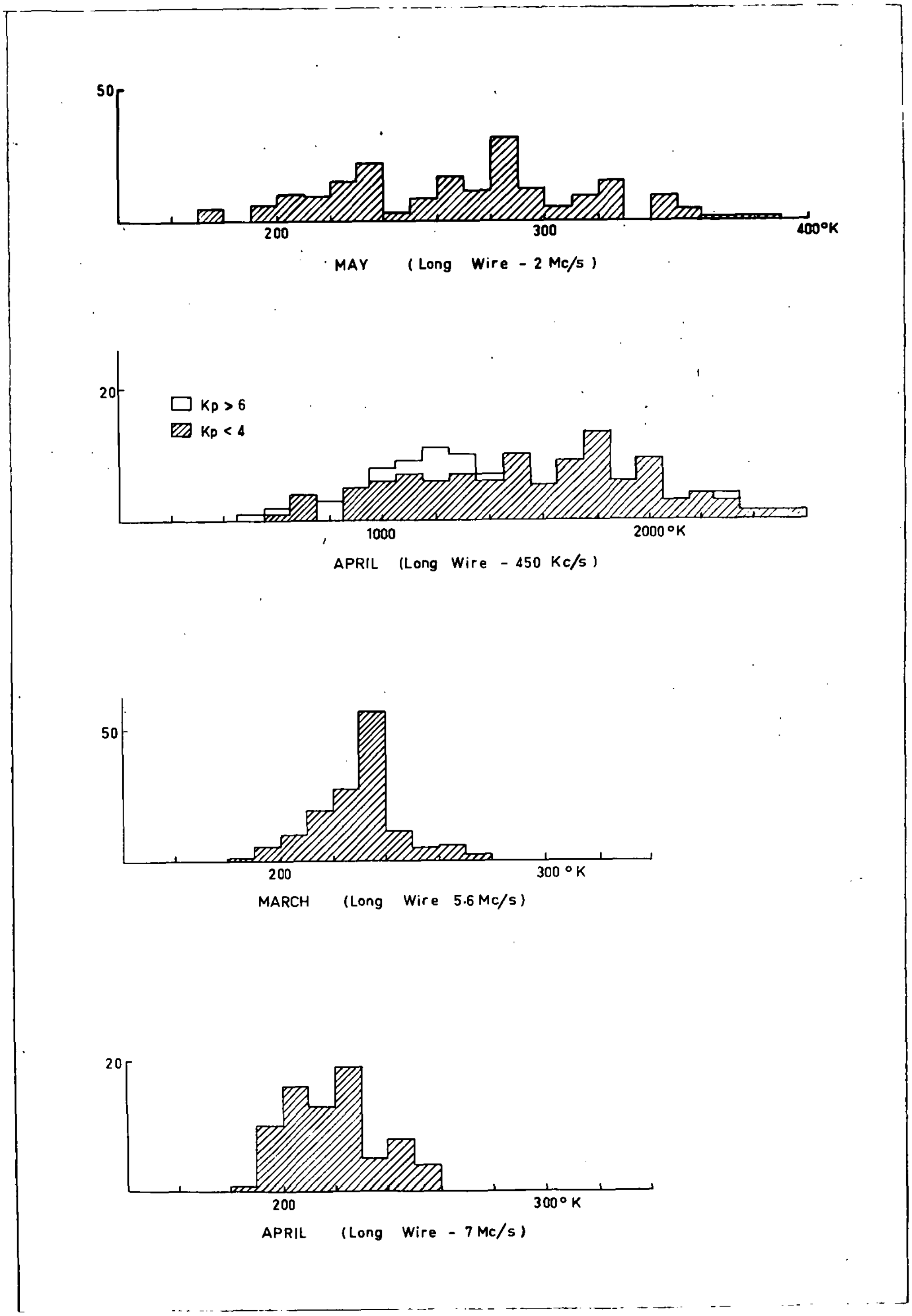

50
200
300
400°K
MAY (Long Wire - 2 Mc/s)
20
Kp > 6
Kp < 4
1000
2000 °K
APRIL (Long Wire - 450 Kc/s)
50
200
300 ° K
MARCH (Long Wire 5.6 Mc/s)
20
200
300° K
APRIL (Long Wire - 7 Mc/s)

Fig. 8.

While the critical frequency is close to the operating frequency the noise level is very unsteady, after this, the intensity is sufficient to hold the recorder pen off scale, corresponding to many thousands of degrees of equivalent temperature. The initial fluctuations are probably due to the very small area overhead through which cosmic noise can penetrate the ionosphere during this transition period, rapid fluctuations of position of this "hole" due to comparatively slight changes in the ionosphere, and the sharp lobes of the long wire ($8\frac{1}{2}$ wave lengths) aerial at this frequency.

Atmospherics were not troublesome at these frequencies near midday, or rather, during the period between the disappaarance of atmospherics in the morning and their recommencement sometime in the afternoon. However, the length of this period decreased with increasing frequency. Fig. 9 shows a contour diagram of the elevation of the sun at Macquarie Island as a function of diurnal and secular time. The times of disappearance and recommencement of atmospherics at the different frequencies are shown suitably marked. It is seen that these cut off times are grouped about a corresponding critical altitude which is dependent on frequency. When the sun is below this altitude the ionization and hence absorption of the D-region is not sufficient (at the frequency in question) to prevent atmospherics from lower latitudes arriving at Macquarie Island. This absorption limited cut off is quite sharp, particularly at the higher frequencies, and takes place within ten minutes or less, corresponding to a degree or so of sun's altitude.

4. DISCUSSION.

4.(i) Temperature measurement.

It was expected that measurements at frequencies higher and lower than 2 Mc/s would give information about the temperature of higher and lower levels of the ionospheric D-region. At 450 Kc/s. the apparent temperature is a thousand or so degrees whereas rocket measurements and other methods have indicated temperatures in the range 200°K to 300°K. Consequently, some natural non-thermal radiation of the "white noise" type must be present and the 450 Kc/s results are useless for temperature measurement.

At 5.6 and 7 Mc/s only thermal noise was present and the temperatures were much the same (within the expected error) as the 2 Mc/s measurements. At these frequencies the D-region is much more transparent so that there is no longer a thin layer of "optimum opaqueness" responsible for most of the temperature contribution. This situation is apparent from the principle of detailed balancing. Suppose the aerial is radiating power towards the ionosphere. At 2 Mc/s the absorption is very high and most of the power will be absorbed in a relatively thin layer. At 7 Mc/s though, probably only about half of this power would be absorbed on the ray's first trip through the D-region. If we assume that the reflection co-efficients of the F-layer and "ground" (in this case the Southern Ocean) are both unity, then all the power will ultimately be absorbed in the D-region but the depth of ionosphere responsible for (say) 90% of the absorption might include the entire D-region. The temperature measured would then represent an average one for the whole region. Further, if F1 were present and if fo F1 (or fx F1) coincided with the operating frequency, then considerable absorption would occur in the F1 layer so that the measured temperature would be partly that of the F1 layer. This probably did not happen (no evidence on the records) as the F1 layer is rarely present in April at Macquarie Island.

The long wire-obtained measurements at 2 Mc/s gave higher temperatures than did those made with the half wave dipole. An increased temperature is to be expected from an aerial having horizontal directivity since the absorption of the ray per unit of height increases and this lowers the optimum opaqueness layer to warmer heights. But the occasional large increases to nearly 400°K once more indicate the presence of non-thermal radiation which will be dealt with in the next section.

4.(ii) Non-thermal Noise.

The non-thermal noise mentioned above was complete unexpected and presents an interesting problem. It was thermal in character - purely random "white" noise and devoid of impulses - but obviously not in origin. Unlike the tropospheric noise discussed in Section A, it was present for days at a time on 2 Mc/s. and at all times on 450 kc/s., and it was not associated with any tropospheric disturbances. It appeared to increase when the D-region absorption was low in that it was strongest in midwinter, it was strong on days when atmospherics were present or when the atmospherics finished comparitively late in the morning and recommenced early in the afternoon, and it appeared less strong on magnetically disturbed days.

The possibility of plasma oscillations in the ionosphere (F region) was also ~~is~~ considered. However, this type of radiation would be

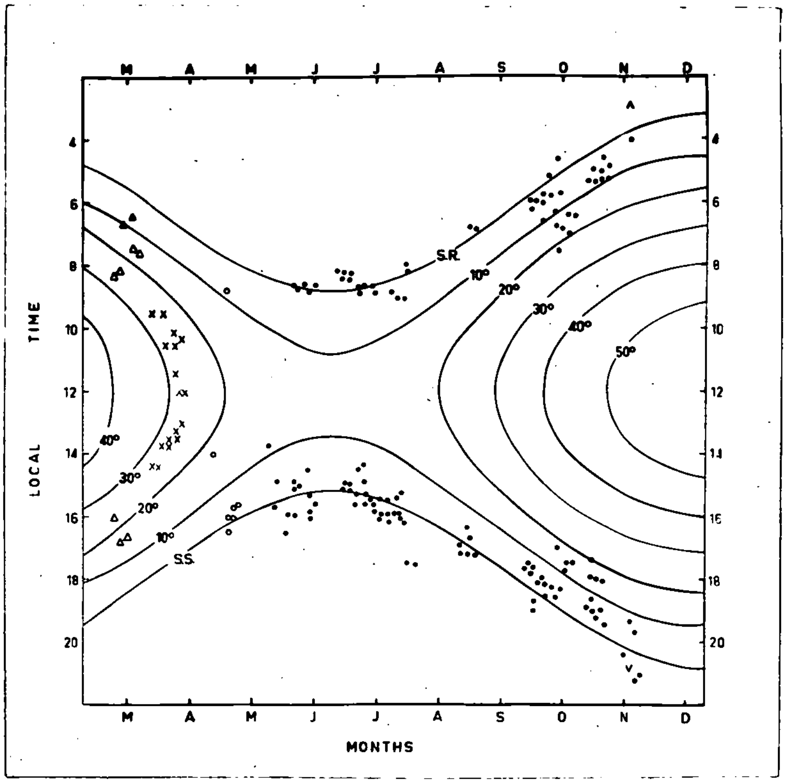


Fig. 9. Contour diagram of the elevation of the Sun at Macquarie Island, showing times of commencement and dissapparence of atmospherics at the frequencies: 450 kc/s (o), 2mc/s (•), 5.6 mc/s (Δ), and 7mc/s (x).

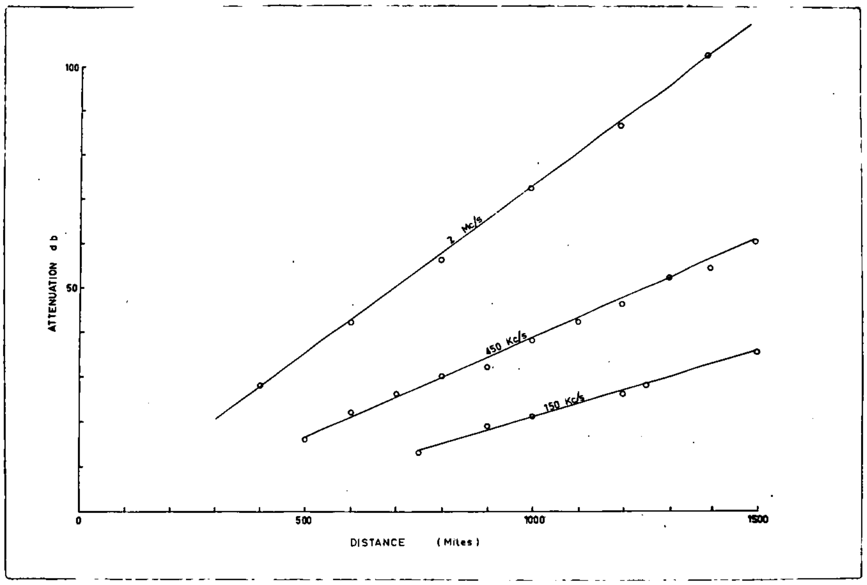


Fig. 10. Attenuation of plane ground waves over sea water. The linear relationship does not hold for distances less than shown.

very strongly attenuated in the D-region and one would expect stronger noise at the higher frequencies.

The long wire measurements at 2 Mc/s. indicated that the radiation was coming from low angles. This suggests ground wave propagated radiation. Accordingly, ground wave attenuation at 2 Mc/s., 450 Kc/s. and 150 Kc/s. was calculated as a function of distance by a method after Terman (1943) in which all factors (round earch, refraction, dielectric constant of air, etc.) are taken into account. It was found that the attenuation for plane waves (i.e. without inverse square of distance effect) plotted in decibels per thousand miles is remarkably linear (Fig. 10).

When it is noon at Macquarie Island in winter there is a region several hundred miles south where the sun's altitude is below the cut off angle at which the D-region becomes transparent as explained above. Now if conditions in the upper ionosphere (F region and ultra F region) are right there will be very strong cosmic radiation incident at this region. The equivalent temperature of cosmic radiation at 2 Mc/s. is approximately 10^7 ^oK (Reber and Ellis, 1956). Some of this radiation will be propagated away from this region in the ground wave mode and might eventually reach Macquarie Island in sufficient strength to give rise to the excess temperatures observed at 2 Mc/s in early June and at 450 Kc/s in late April - early May.

Consider the attenuation of 2 Mc/s plane ground waves from regions due south of Macquarie Island where the sun's altitude (ψ) is 0^o, 1^o, 2^o,.... 5^o at noon during early June ($\psi = 12^o$ at Macquarie Island at this time). The power attenuation of 2 Mc/s plane ground waves for distances greater than about 400 miles is at the rate of 78 db/K miles less 6 db (see Fig.10). For 2 Mc/s. the cut off altitude of the sun (ψ_c) is around 0^o (see Fig. 9). We allow (say) 10 db loss in the wave on changing from sky to ground wave propagation. The attenuation and resulting "leakage" temperature is given in Table 4 below.

TABLE 4.

ψ	Distance	Total Atten.	"Excess" Temp.
0°	750 miles	62 db	10°
1°	690 "	58 "	30°
2°	630 "	53 "	100°
3°	570 "	48 "	300°
4°	510 "	44 "	1000°
5°	450 "	39 "	3000°

This explains why "excess" temperatures of some 100–300 degrees were sometimes observed at 2 Mc/s in June. Also the days on which high temperatures were recorded (8, 9, 10, 11th June) were also days of early onset of atmospherics ($\psi_c = 3^\circ$– 5°).

At 450 Kc/s the situation is different. Ellis (1956) has shown that observations of cosmic radio emission at frequencies much below 1 Mc/s are unlikely. He points out that although low frequency waves could penetrate the lower ionosphere (D, E, F regions) in the ablique longitudinal extraordinary mode ("whistler" mode) this mode would be reflected at the $y = 1 - x$ level in the upper ionosphere (ultra F region). However, Reber (1956) reports night time temperatures of about 10^5 °K at 520 Kc/s. During the day the temperature was about 10^3 °K and the radiation appeared to be coming from the south. Ellis (1957) suggests that Čerenkov radio emission in the ultra F region caused by auroral particles approaching the earth should give flux densities of about $10^{-21} Wm^{-2}(C/S)^{-1}$. At 450 Kc/s this corresponds to equivalent temperatures of 10^6 °K.

Reber's daytime results are similar to ours at 450 Kc/s. If we assume a night sky temperature of 10^6 °K we might explain the daytime temperatures at Macquarie Island on the basis of ground wave propagation. At 450 Kc/s the onset of atmospherics was not as clearly defined and some atmospherics were usually present all day. Consequently, the cut off angle (ψ_c) is not definite. From Fig. 9 ψ_c extends from about -2° to $+18^\circ$. If we take $\psi_c = 10^\circ$ as a typical case, we have $\psi = 20^\circ$ at noon at Macquarie Island during April–May so that the nearest place where the night sky radiation can get under the D region will be 10° of latitude south. The attenuation for this distance (Fig.10) is about 20 db and allowing 10 db launching losses we get a total loss of 30 db. If the sky temperature where $\psi = 10$ is 10^6 °K this would give 10^3 °K at Macquarie Island.

The agreement is largely fortuitous but the above mechanism at least explains the southerly direction of arrival and the comparitively high temperatures in the day time. At still lower frequencies (150 Kc/s., say) we would expect this effect to be even more pronounced. This is qualitatively verified by recent measurements by Reber (unpublished) at these frequencies who reports quite a low ratio between night time and day time intensities from Tasmania.

5. CONCLUSIONS.

Only frequencies in the vicinity of 2 Mc/s can provide useful information about the temperature of the D region. At higher frequencies the absorption of the D region is less so that measurements can only be carried out in places and at times when the sun's altitude is higher, and even then the emitting region of the D region is no longer a thin layer of effectively uniform temperature. At lower frequencies there is more ground wave propagated interference. At the additional frequencies used no useful information on D region temperature was obtained.

It may be possible, with a suitably chosen frequency, to measure the temperature of the F1 region by this method when fo F1 (and fx F1) pass through this frequency. It would be necessary of course to have simultaneous ionograms to obtain accurate information on D region absorption, presence of sporadic E layer, fo F1 and fo F2. The F2 layer would presumably prevent interference from cosmic noise.

The non-thermal radiation encountered in the day time was probably ground wave propagated noise from near the twilight region of the earth. This might be cosmic noise from outer space (at 2 Mc/s) or Čerenkov radiation generated in the ultra F-region (at 450 Kc/s.).

APPENDIX.

Matching.

As shown in Fig.11 the receiver (i.e. R.F. amplifier, communications receiver, recording milliammeter) samples at point "A" the R.F. voltage of either the aerial via the matching unit or the dummy aerial (172 ohm resistor). The matching unit must be carefully adjusted so that the impedance of the aerial is accurately that of the dummy.

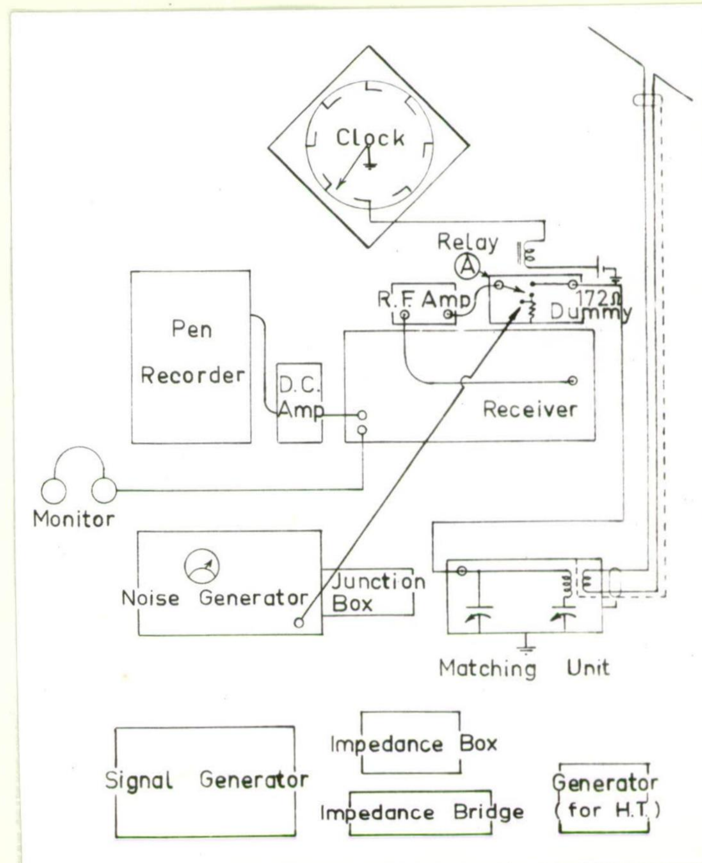

Clock
Relay
A
R.F. Amp
172Ω
Dummy
Pen
Recorder
D.C.
Amp
Receiver
Monitor
Noise Generator
Junction
Box
Matching Unit
Signal Generator
Impedance Box
Impedance Bridge
Generator
(for H.T.)

Fig. 11.

Then when the temperatures of the aerial and dummy are equal the R.F. voltages at "A" will be equal and there will be no difference in the trace level on the recording milliammeter. The matching was achieved with the impedance bridge as follows. With the signal generator as source and the receiver as detector, ~~a~~ short coaxial cable was connected from across the "unknown" arm of the bridge to point "A" in place of the receiver. The relay was then energised so that point "A" was connected to the dummy and the impedance bridge carefully balanced. The relay current was then turned off, connecting the bridge via point "A" to the aerial via the matching unit, and the matching carefully adjusted to give the balanced condition once again. By merely turning a switch the balance of the bridge with the dummy in circuit could be quickly rechecked without altering signal generator output or receiver gain. As seen from the co-axial socket at "A" the impedance of the aerial was then accurately that of the "cold resistor" dummy aerial.

Calibration.

The receiver was calibrated by connecting a saturated diode noise source across the dummy aerial. A schematic circuit diagram of this noise source is shown in Fig.12. The effective temperature Ta of the dummy aerial resistor then becomes :-

$Ta = T + eIR/2k$ degrees Kelvin where T is the ambient temperature and R the resistance of the dummy, and I is the current through the diode. If R is chosen 172 ohms this becomes :-

$Ta = T + I'$ degrees Kelvin where I' is the diode current I measured in microamperes.

It was found that at least for temperatures up to 200°K the deflection of the record trace was linear with temperature (or diode current) so that temperatures below ambient (T) were easily found. Connecting the noise source across the dummy did not appreciably change its impedance - that is, this did not produce a change in trace level. This was checked on ~~a~~ number of occasions by recording a trace of about half an hour duration during which the noise source was left connected and disconnected for alternate two minute periods. Incidentally, this same method was used to check the effect on the trace level of several other factors. One of these was the D.C. current in the "aerial" - "dummy" change over relay. This was checked by mechanically jamming the relay in the "dummy" position and

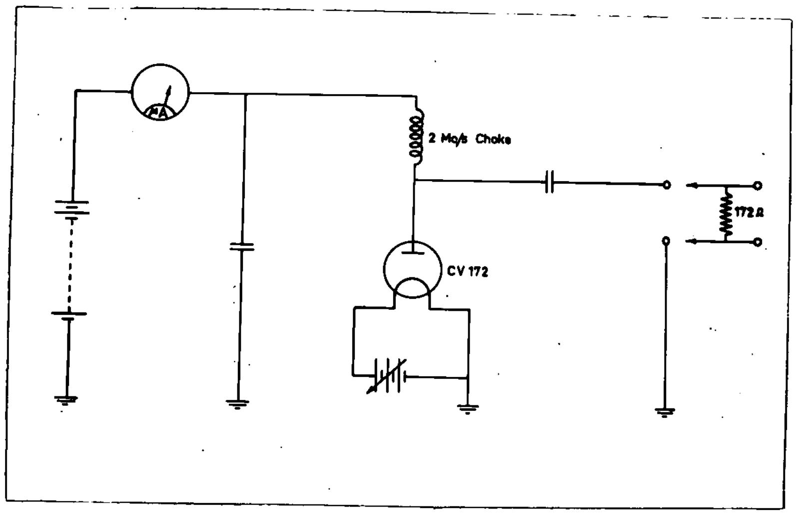


Fig. 12. Noise source.

applying current to the relay for alternate two minute periods. None of these factors were found to give spurious changes in trace level.

Attenuation measurements.

Since the aerial was sampled at point "A" it was necessary to find the total attenuation of the matching unit and feeder line between this point and the aerial itself. Fortunately the ambient temperature of the matching unit and that of the feeder line were always very neary the same so that only the total attenuation needed to be taken into account. To do this, the aerial was lowered and disconnected from the shielded twin line feeder.* In place of the aerial at the end of the feeder an 87 ohm resistor was connected. Thus terminated, the feed was then carefully matched to the 172 ohm dummy by the matching unit. As the 87 ohm resistor had nearly the same impedance as the dipole aerial, the matching unit settings were much the same as they were in normal temperature measurement. Reference temperatures were obtained from the noise source through a balance-unbalance transformer. The attenuation of this transformer was cancelled out by also using it when the noise source was connected to the 172 ohm dummy. A schematic diagram is shown in Fig. 13.

The addition of the transformer affected the matching and hence recorder trace level slightly, so this shift was measured from traces obtained for zero diode current. For diode current I' microamperes the effective temperature of the 87 ohm resistor becomes :-

$$T_R = T + I' \cdot \frac{87}{172}$$

For attenuation constant η the temperature (Ta) at "A" is :

$$Ta = \eta T_R + (1-\eta)\, T.$$

$$\text{i.e. } Ta - T = \eta I' \cdot \frac{87}{172}$$

The power attenuation constant (η) was found to be typically 0.6. It increased slightly with time owing to seasonal changes in average humidity at Hurd Point. The adopted values for the 2 Mc/s half wave dipole measurements were $\eta = 0.57 \pm .02$ up to 31/8/1956.

and $\eta = 0.64 \pm .01$ from 25/9/1956.

* This and later descriptions of techniques refer to 2 Mc/s half wave dipole measurements.

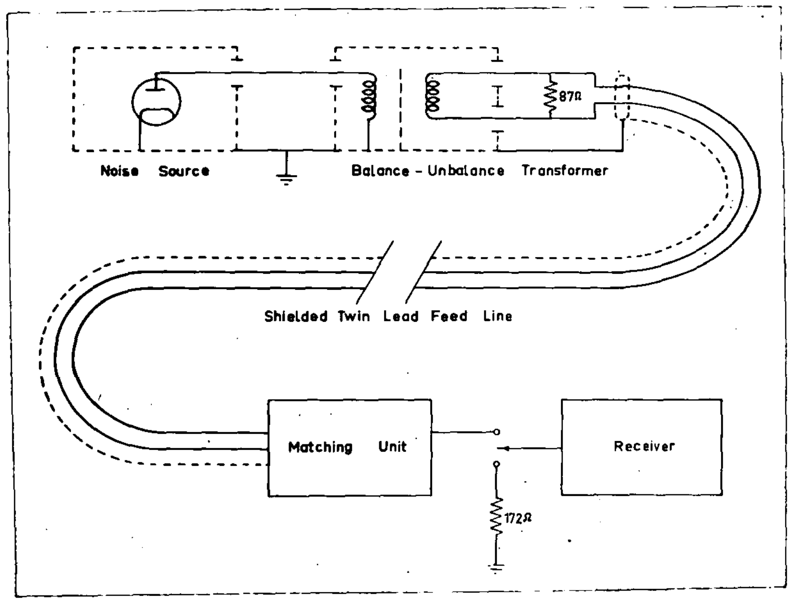


Fig. 13. Attenuation measurement.

Aerial Efficiency.

As mentioned earlier in this thesis, two methods were used to obtain the efficiency of the aerial. The more direct one, the impedance method (Gardner, 1954) gives the efficiency as the ratio of the output (or radiation) resistance to the resistive component of the input impedance of the aerial. For sufficiently large heights (greater than $\lambda/8$, say), the output resistance is very nearly the radiation resistance in the presence of a perfectly conducting earth. It is only necessary then to find the input impedance at the dipole.

There were no calibrated impedance measuring gear available at Macquarie Island at the time so the following method was used. A screened impedance box was constructed for which the reactive and resistive components were variable and graduated. The aerial could then be matched to the 172 ohm dummy as explained above and then when the aerial was lowered, the impedance box substituted for the aerial and adjusted to give the matched condition again. The resistive component of this impedance box would then be that of the aerial at the height for which it was matched. This impedance box was accurately calibrated on return to Australia. Since the feeder line was screened twin lead, its impedance would not have been affected by the close proximity of the ground etc., when removed from the aerial.

In actual practice, the settings of the matching unit and the height of the aerial (from the sag in the feeder line), were noted each time the aerial was matched (once or twice a day). When the aerial was lowered for measurements these settings were repeated and so it was possible to find what the resistive component of the aerial impedance had been at the times (and heights) when the aerial was matched. Table 5 below gives the aerial efficiencies obtained by this method. The radiation resistances were read from a graph giving this as a function of height above a perfectly conducting earth.

TABLE 5.

Aerial height in feet & λ	Number of measurements	Resistive comp. of input imp.	Radiation resistance	Aerial efficiency
110 (.22 λ)	8	91 Ω	76.5 Ω	0.84
127 (.26 λ)	9	90	88	0.98
122 (.25 λ)	9	93	86	0.92
121 (.25 λ)	12	90	84	0.93

The second method uses the Sommerfeld and Renner formula which gives the aerial efficiency as a function of height and ground constants. The conductivity of the ground at Hurd Point was found by measuring the D.C. resistance between test electrodes driven into the ground. These were galvanised steel pipes 6 cm. in diameter which were driven into the ground to a depth of 90 cm. It was found that the resistance between electrodes was independent of separation distance provided this was greater than some five or ten feet. The separations used were of the order of hundreds of feet.

The conductivity "σ" can be found as follows. Suppose the electrodes are of length "l" and radius "a" and that the end of the electrode is a hemisphere of radius "a". The lateral resistance of the (hemispherical and cylindrical) co-centred shell of thickness "Δ r" at distance "r" is given by :-

$$\Delta R(r) = \frac{\Delta r}{\sigma}\left(2\pi r^2 + 2\pi r l\right)^{-1}$$
$$= \frac{\Delta r}{2\pi\sigma}\left[r(r+l)\right]^{-1}$$

The total resistance between the electrode and a co-centred hemispherical conductor at infinity (i.e. "earth") is then :-

$$R = (2\pi\sigma)^{-1}\int_a^\infty \left[r(r+l)\right]^{-1} dr$$
$$= (2\pi\sigma)^{-1} \log\frac{a+l}{a}$$

Hence

$$\sigma = (2\pi l R)^{-1} \log\frac{a+l}{a}$$

The resistance between two such identical electrodes is then 2 R. Six electrodes were used and the value "R" for each found by measuring the resistance between all the combinations of single ones, pairs, triplets, etc. The values "R" obtained were typically about 100 ohms, though some were almost half this. Also star cross section fencing stakes driven into the same depth but otherwise of smaller dimensions gave lower values. This suggests that the electrodes used were not making good contact with the first layer of earth.

Using the value R = 100 ohms substituted in the expression above we get $\sigma \approx 0.6 \times 10^{-2}$ mho/metre. From the Sommerfeld and Renner analysis for an aerial height of $\lambda/4$ and ground constants K = 5 and $\sigma = 10^{-2}$ we get aerial efficiency 0.88. For the same height and ground constants K = 15 and $\sigma = 4 \times 10^{-2}$ we get 0.93. This is in reasonable agreement with the results of the impedance method so the value of 0.90 $\pm$ .05 was adopted for the 2 Mc/s half wave dipole temperature measurements.

A c k n o w l e d g e m e n t s

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