

AN INVESTIGATION OF SOME TIME VARIATIONS OF
THE COSMIC RAY INTENSITY

THESIS

submitted by

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For the degree of

DOCTOR OF PHILOSOPHY

in The

UNIVERSITY OF LONDON

JUNE 1961

ABSTRACT

In recent years a large number of experiments have been performed to study the meteorological, and time, variations of the μ -meson, and nucleonic, components. These are now reasonably well understood. Little attention, however, has been given to the electronic component of the cosmic radiation. In particular, the way in which meteorological changes affect it has been obscure. It was therefore decided to make a study of these, and the time, variations associated with the electronic component. An experiment for studying these variations of the electronic component is described.

We need first to find the way in which changes of atmospheric pressure and temperature affect the electronic component. Regression analysis is used to study these relationships on a day-to-day, and on a month-to-month, basis. This analysis shows the soft component to be subject to two main temperature effects: a positive, surface temperature, effect, and a negative, upper atmosphere temperature, effect. A theoretical expression for the atmospheric effects of the electronic component, put forward by Dorman, is examined and found to be qualitatively correct.

The results of the above regression analysis are used to correct the diurnal variation of the electronic component for the effects of atmospheric pressure and temperature. It is then found to agree with the similarly corrected μ -meson component variation.

An attempt is made to correct for the effects of atmospheric temperature the results of an inclined telescope, diurnal variation, experiment

(iii)

performed in London during 1954-55, which originally appeared to show anomalous results. It is shown that after this correction has been applied the results are consistent with present concepts of the diurnal variation.

After a correction for the effects of atmospheric temperature variations has been made, the long term changes in the diurnal variation of the μ -meson component are investigated. The presence of a twenty-two year wave in the phase of the diurnal variation is confirmed. Comparison is made with the diurnal variation of the nucleonic component.

Finally, the rigidity dependences of the hard and soft components are compared during Forbush decreases and found to be not significantly different.

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CHAPTER I - INTRODUCTION

1.1. General Introduction

As a result of studies of cosmic ray intensity variations spread over the past twenty-five years, it has become increasingly evident that the intensity of the primary cosmic radiation incident upon the earth's atmosphere is strongly influenced by solar processes. Examples of these solar controlled variations are the eleven year variation of intensity that follows the sunspot cycle, and the daily variation of intensity, with a period of one solar day.

Since we know the primary cosmic ray intensity to be composed of positively charged particles, which are affected by electric and magnetic fields, the variations we observe must be due to the effects of electro-magnetic conditions in interplanetary space, particularly the conditions close to the earth, the entire electro-magnetic state being controlled by the sun's behaviour. Thus the cosmic ray variations afford us a means of investigating the changes in the interplanetary electro-magnetic state. This is of considerable importance because of the profound influence these electro-magnetic conditions exercise on a wide variety of phenomena, such as magnetic storms, aurora, the Van Allen radiation belts and the propagation of flare-produced solar protons.

While the primary cosmic radiation incident upon the top of the atmosphere is obviously of great interest to us, it is not in general possible to record it directly for any very long period of

time, so that we have to record the so-called 'secondary radiation' at the surface of the earth. Because of complicated nuclear interactions occurring within the atmosphere, the cosmic radiation measured at sea level differs considerably from the primary radiation. It is thus necessary to reconstruct the primary cosmic ray variations from the secondary variations recorded, a process that can be difficult.

The radiation observed at the surface of the earth can be divided into three components; μ -mesons, nucleons (i.e. neutrons and protons) and electrons. The μ -mesons are unstable particles with a rest lifetime of 2.22×10^{-6} seconds, after which they decay into an electron and two neutrinos. The electronic component mentioned above is due largely to the decay electrons and to 'knock-on' electrons created by the μ -mesons in their passage through the atmosphere.

The μ -meson component has been reliably recorded over the longest period - approximately twenty five years, and it was from these records that the eleven-year variation in intensity was discovered (Forbush 1954). Unfortunately the μ -meson component is seriously affected by an atmospheric ~~is~~ temperature effect. When the height of the mean μ -meson production layer varies with atmospheric expansion and contraction then due to the instability of the μ -meson it causes changes in the probability of the μ -mesons reaching ground level before they decay, thus introducing fluctuations into the secondary cosmic ray data recorded which have nothing to do with

the primary radiation. The effect is serious because little is known about periodic variations of temperature in the upper atmosphere, (Quenby and Thambyahpillai 1960) so that little can be done to correct a phenomenon like the daily variation for such temperature changes, unlike the varying absorption effect due to changes in atmospheric pressure for which correction is easily made.

The nucleonic component has not been recorded over so long a period as the μ -meson component, the neutron monitor only coming into general use about ten years ago. However, it enjoys several advantages over the μ -meson component. Since the proton and the neutron are effectively stable particles, the radiation recorded by the neutron monitor is unaffected by atmospheric temperature changes, and only corrections for a simple absorption effect need be made. The nucleons have their antecedents in the low rigidity end of the primary cosmic ray spectrum and so are more sensitive to electro-magnetic changes in interplanetary space than the μ -meson component.

1.2. The Daily Variation and Atmospheric Temperature Effects

The existence of the daily variation of the cosmic ray intensity has been established for some twenty-three years (e.g. Schonland et al. 1937, Forbush 1937, Duperier 1945). The variation in the μ -meson component has an amplitude of about 0.2% when measured close to sea level, with a time of maximum between 1200 to 1800 L.M.T.

In investigating the daily variation it is preferable to make directional measurements i.e. measure radiation coming from a given direction in space. It is possible to do this for the μ - meson component but not for the nucleonic component, the neutron monitor being an omni -directional device limited only by atmospheric absorption. Thus in the study of the daily variation the atmospheric temperature independence of the nucleonic component is of limited use.

If we wish to study the long term changes in the daily variation we need to make use of the μ -meson data that has been recorded over the longest period. To do this satisfactorily we need to make corrections for the atmospheric temperature effects. Since these effects are of such importance we shall discuss them in a little more detail.

Until very recently, the main difficulty in the elucidation of atmospheric temperature effects on the daily variation of the cosmic ray intensity has been ignorance of the diurnal variation of temperature in the upper atmosphere. This has been due to the large radiation and lag errors associated with the temperature sensor in the meteorological radio-sondes. To circumvent this difficulty and to establish the order of magnitude of the temperature correction required it was necessary to employ rather an indirect method of approach.

The first necessity was to establish the reality of the daily variation as a primary cosmic ray effect. This was done

by an experimental method that eliminated, to a first order of magnitude, atmospheric effects. Directional measurements (e.g. Malmfors 1949, ^{ELLIOT} ~~DOLBEAR~~ and ^{DOLBEAR} ~~ELLIOT~~ 1951¹) with geiger tube telescopes inclined at angles of up to 45° to the vertical were made. Pairs of telescopes were used, situated in a North-South plane. Since cosmic rays recorded by the telescopes passed through approximately the same slab of atmosphere, then to a first order of magnitude both are affected in the same way by changes in atmospheric pressure and temperature. Assuming a primary anisotropy to exist, then both telescopes will record it as well as variations due to atmospheric changes. However the two telescopes will not record this primary variation with the same phase since they scan different sectors of sky. Hence a simple vector subtraction of the daily variations measured by the telescopes should eliminate the common atmospheric effects and leave a residue due to the out-of-phase components of the daily variation.

The experiments of Malmfors (1949) and ^{ELLIOT} ~~DOLBEAR~~ and ^{DOLBEAR} ~~ELLIOT~~ (1951¹) observed such a residue and established the daily variation as an extra-terrestrial effect. Later measurements of the daily variation using neutron monitors, which record the atmospheric temperature independent nucleonic component, when compared with μ -meson measurements indicated that the atmospheric temperature correction to the μ -meson results was possibly of the same order of magnitude as the daily variation recorded by the μ -meson detectors.

While the above work clearly defined an effect to be corrected for, it did not suggest a method of obtaining this correction. In general there seem to be two experimental lines of approach. One is to try to make use of the measurements of the nucleonic component. This has recently been done by Quenby and Thambyahpillai (1960) and an apparently satisfactory result obtained. An alternative is to consider the electronic component. In later sections of this Chapter we shall see that there is good evidence for believing that the electronic component is unaffected by atmospheric temperature changes. It can also be measured by the same type of apparatus that measures the μ -meson component. This thesis is concerned with an experiment that utilises this fact in an endeavour to obtain an atmospheric temperature correction vector for the daily variation of the μ -meson component.

Before we proceed any further we shall briefly consider the main types of intensity recorders used in cosmic ray work.

1.3. Types of Intensity Recorders

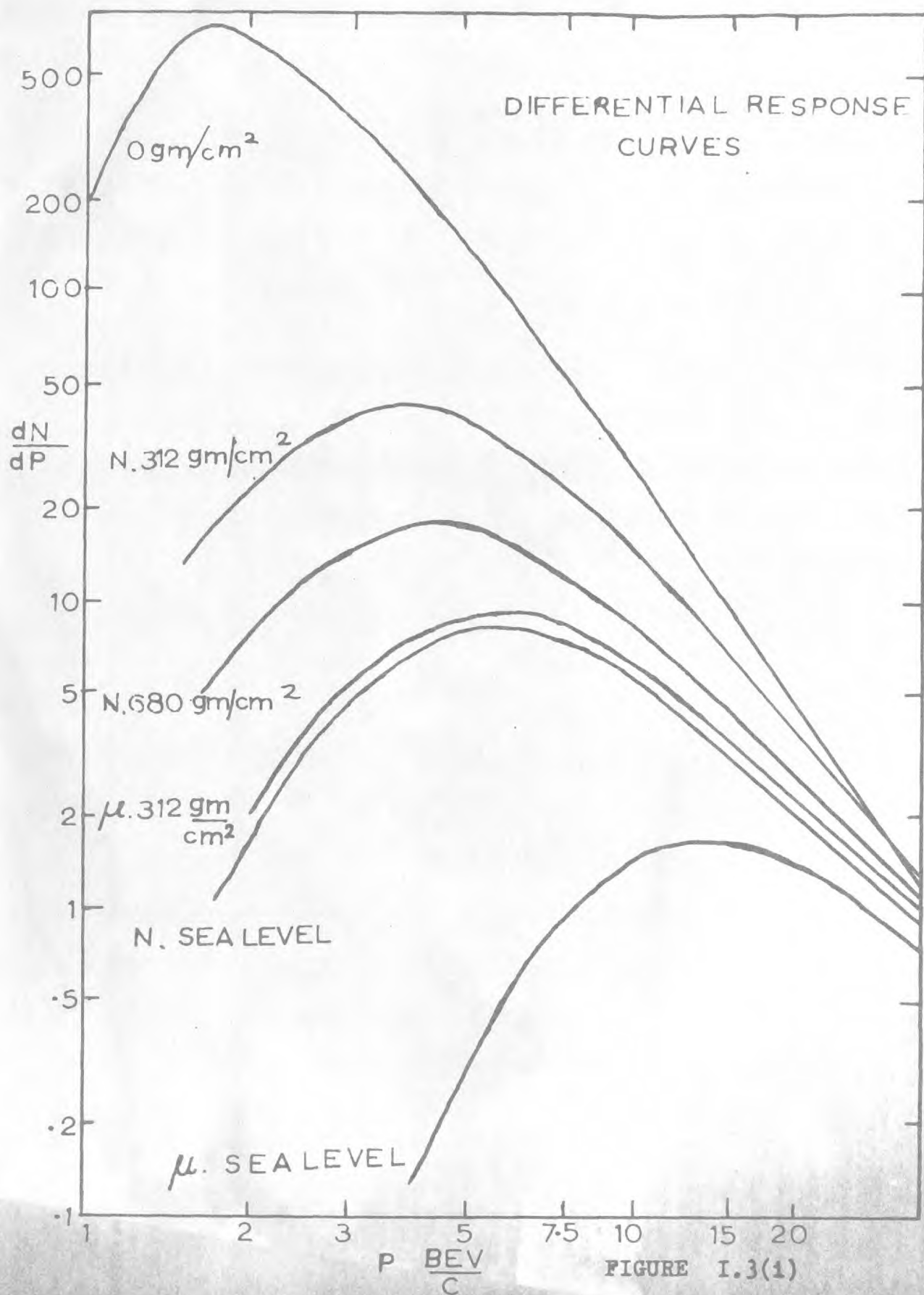
There are three main types of intensity recorders used in cosmic ray work. These are the neutron monitor, the ion-chamber and the geiger tube telescope. First let us consider the relationship between what the instruments record, deep in the earth's atmosphere, and the primary cosmic radiation incident

on the top of the atmosphere. In principle, if we know the processes occurring in the atmosphere, it should be possible to ~~directly~~ ^{directly} relate the primary cosmic radiation with the secondary radiation incident on the intensity recorders. This cannot be done at present in any very satisfactory fashion, so we use a 'net effect' approach.

By flying detectors on balloons at the top of the atmosphere we can measure the primary cosmic ray spectrum with reasonable accuracy. At the bottom of the atmosphere we record the secondary radiation continuously. By using the earth's magnetic field as a magnetic spectrometer and by comparing the two sets of measurements, then without knowing the intervening processes, it is possible to deduce the 'differential response' curves for various recording instruments. These curves show the rigidity response of a particular type of instrument at a given atmospheric depth. Such a set of curves has been constructed by Webber and Quenby (1959) using the most recently determined primary spectrum and greatly improved values for the magnetic 'cut-off' rigidities of cosmic ray particles at various latitudes. These curves are the most accurate at present available and are shown in figure 1.3 (i).

The Neutron Pile

The Simpson-type neutron pile is well known to monitor the nucleonic component of the cosmic radiation. (Simpson et al.



1953). It is an omni-directional instrument limited only by atmospheric absorption. One of its principle assets is that the secondary radiation it records is largely unaffected by variations in atmospheric temperature, (Section 1.5). Latitude surveys with neutron monitors show that the nucleonic component undergoes^a far greater latitude effect than the μ -meson component of the cosmic radiation. Figure 1.3 (i) gives the explanation of this. We see that the curves marked neutron monitor, for various atmospheric depths, have their main rigidity response in the 2 to 10 GV region. This is the rigidity range most affected by the earth's magnetic field. The neutron pile is thus a recorder of the low rigidity end of the cosmic ray spectrum.

The Ionization Chamber

This type of recorder is the one that has been longest used as a monitor of secondary cosmic rays. The Compton-Bennet model C chamber has been operated at various Carnegie Institute stations since 1937. It is also an omni-directional recorder, limited only by atmospheric absorption. Due to its 12 cm. thick lead shield, necessary to shield it from local radio-activity, it can record only the so-called 'hard' component, predominantly μ -mesons, of the cosmic radiation. Its main assets are simplicity, reliability and ruggedness. In figure 1.3 (i) those curves not labelled 'neutron monitor' are only labelled 'ionization

detector'. This is because, below a depth of 312 grms/cm^2 in the atmosphere, there is negligible difference in the rigidity responses of the ion-chamber and the counter telescope. Even at 312 grms/cm^2 depth there is only a significant difference in response below 5 GV.

From the ionization detector differential response curves we see that the ion-chamber responds to a higher mean rigidity than the neutron pile. At sea level this mean rigidity is probably in the range 20 to 30 GV.

The Counter Telescope

The neutron pile and the ionization chamber are omnidirectional devices. When attempting to determine the direction in space of the primary anisotropy causing the daily variation these instruments are of limited use. The counter telescope provides a device that measures cosmic radiation contained within a given solid angle, which is defined by the arrangement of the trays of geiger tubes that constitute the telescope.

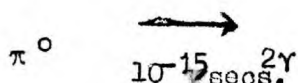
A telescope usually comprises three trays of tubes, and is most commonly used to measure the 'hard' component of the cosmic radiation. This is arbitrarily defined as secondary radiation capable of penetrating 10 cms. of lead, and is almost entirely composed of μ -mesons. The radiation absorbed in 10 cms. of lead is largely electrons with about ten per cent soft μ -mesons, and is known as the 'soft' component.

As indicated above the counter telescope has a differential rigidity response similar to that of the ionization chamber.

1.4. Theoretical Atmospheric Effects

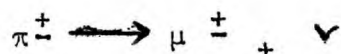
It is now well established that the primary cosmic ray particles are positive in charge, being some 91.5% protons, 7.8% α -particles and < 1% heavier ions (Ginsberg 1958). These positive particles have a mean free absorption path of about 140 grms/cm². Thus between the top of the atmosphere and the 200 mb layer over 60% of the primary particles have undergone nuclear interaction and been absorbed. These interactions rapidly split the radiation up into three distinct components. After the first collisions the radiation is composed of nuclear fragments, secondary nucleons and positive, negative and neutral π mesons. There are also a few, much rarer, heavy mesons.

The rest lifetime of the π^0 meson is less than 10^{-15} seconds (Rochester Conference 1960) and it almost immediately decays into two photons.



These high energy photons then 'cascade' in the atmosphere and become the so-called 'non-equilibrium' portion of the 'soft' component. The positive and negative π mesons have rest lifetimes of 2.56×10^{-8} seconds. There is a small probability of nuclear interaction but largely they decay into μ -mesons plus

neutrino, the decay occurring in the upper atmosphere.

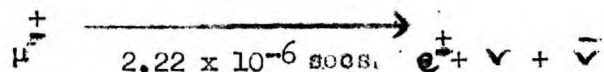


$$2.56 \times 10^{-8} \text{ secs.}$$

Finally the secondary nucleons cause further nuclear disintegration until the end products have had their energy degraded so much that they are no longer capable of nuclear interaction. These are the so-called 'disintegration product nucleons' with mean energies in the range 1 - 20 MeV.

Thus, deep inside the atmosphere we have three distinct components to consider. The electron-photon ('soft') component, the μ -meson ('hard') component which constitutes the main portion of the cosmic radiation at sea level, and the nucleon component, ultimately composed of protons and neutrons with the energies indicated above.

As we indicated in section 1.1, due to the instability of the μ -meson, atmospheric temperature effects are introduced into the secondary cosmic radiation recorded at sea level. The mode of decay of the μ -meson is



and this is the main source of the electrons recorded at sea level, those resulting from the π^0 decays having been largely absorbed in the atmosphere above two kilometres (Puppi 1956).

The atmospheric temperature effects resulting from the instability of the μ -meson were first discussed by Blackett (1938), and later by many other workers, notably Trefall (1955) and Dorman (1956). However before we discuss ~~this~~ work we shall first consider the atmospheric effects associated with the nucleonic and the electronic components at sea level.

The Nucleonic Component

In general the meteorological effects on cosmic radiation are most conveniently grouped under two headings, those due to changes in atmospheric pressure, and those due to changes in atmospheric temperature. Let us consider the effects due to changing atmospheric pressure.

The variation of nucleonic intensity with depth is found to be represented by the expression (Simpson et al. 1953)

$$N(h_0) = N_0 e^{-h_0/L} \quad \text{where } h_0 = \text{atmospheric depth of observation}$$

L = absorption m.f.p. of the nucleonic component

$$\text{Then } \frac{\delta N(h_0)}{N(h_0)} = - \frac{\delta h_0}{L} = \alpha_p \delta h_0$$

where $\alpha_p = - \frac{1}{L} = \text{Pressure coefficient.}$

For atmospheric depths less than 700 gms/cm^2 L is found to vary both with h_0 and latitude (Simpson and Faglot 1953). However at

mountain altitudes and sea level it is found to be constant with latitude and equal to (145 ± 9) grms/cm².

$$\text{Therefore } \alpha_p = - \frac{1}{(145 \pm 9)} = (-0.69 \pm 0.04) \% / \text{gcm} = (-0.68 \pm 0.04) \% / \text{mb.}$$

Calculation of atmospheric temperature effects are rather more difficult. If all the neutrons created in a neutron pile were due solely to secondary nucleons, then there would be no atmospheric temperature effects due to the relatively long mean lives of the nucleons. However a small number of neutrons are formed due to 'stars' with meson antecedents. Figure 1.4 (i), taken from Simpson et al. (1953), shows the contribution to the neutron pile counting rate with atmospheric temperature sensitive links.

Simpson has attempted to analyse the atmospheric temperature links shown. He comes to the conclusion that they cause a negligible effect.

Dorman (1956) has done a rather more careful analysis of these 'links'. He comes to much the same conclusions as Simpson except that he predicts a rather small atmospheric temperature effect (0.01 to 0.1%/°C) for a neutron monitor situated at sea level at the equator ($\lambda = 0$).

The Electron-Photon Component

The electron-photon component at sea level related to the decay of π^0 mesons in the upper atmosphere is often called

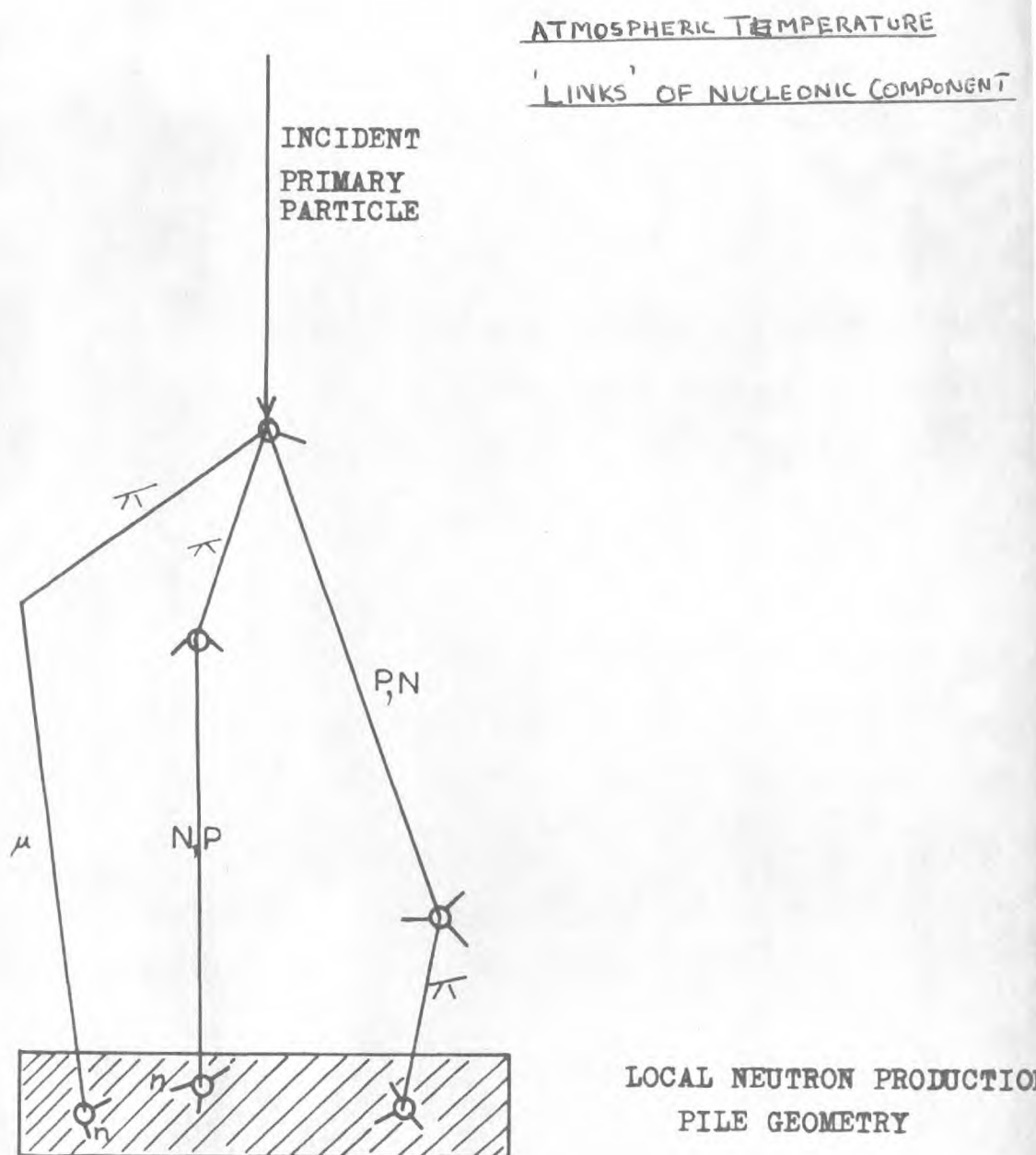


FIGURE I.4(1)

the 'non-equilibrium' portion of the soft component. Those electrons that result from the decay of the μ -meson, and the 'knock-on' or δ -electrons, caused by the passage of the μ -mesons through the atmosphere, are usually known as the 'equilibrium' portion, since they are in approximate equilibrium with the μ -mesons producing them. If they are sufficiently energetic when produced, the 'equilibrium' electrons will form cascades in the atmosphere.

At sea level most of the 'soft' component is due to the 'equilibrium' portion of the electrons and photons, those related to the decay of the π^0 mesons having been largely absorbed in the atmosphere above two kilometres. To deal adequately with the development of the soft component in the atmosphere the cascade theory of showers must be used. Fortunately it is possible to make a rough, first order of magnitude, approach in a much more elementary way. This assumes that the rate of production of electrons does not change appreciably over a distance in which the showers generated by these electrons are completely absorbed. (Rossi 1940). This treatment initially ignores such details as instrumental cut-off, but is probably not more than 20% in error. Dorman (1956) has used this simple approach to estimate the likely pressure and atmospheric temperature effects for the equilibrium portion of the soft component.

For the sake of simplicity we shall quote the results he obtains for each source, then put these results into one composite

expression for the whole of the soft component.

Decay Electrons

Using the initial assumptions listed above, and a further assumption that the μ -meson on decay yields approximately 1/3 of its energy to the decay electron, it may be shown that

$$\frac{\delta N_e^{\text{decay}}}{N_e^{\text{decay}}} = - \frac{\delta h_0}{h_0} + \frac{\delta T(h_0)}{T(h_0)} + \frac{\delta N_\mu(h_0)}{N_\mu(h_0)}$$

where N_e^{decay} = Total number of decay electrons; h_0 and $T(h_0)$ are the ^{atmospheric depth} ~~pressure~~ and temperature of the air at the observation level, $N_\mu(h_0)$ = Total μ -meson flux at the observation level.

Thus the variation of the decay electrons contains a portion repeating the variations of the meson component of both atmospheric and non-atmospheric origin. There are also additional meteorological effects i.e. a local pressure coefficient of -0.10%/mb. at sea level, and a local temperature coefficient of 0.36%/°C at sea level.

δ -Electrons

The factor expressing the energy lost by a μ -meson to δ -electrons on its passage through the atmosphere is a logarithmic one, and so depends rather weakly on the actual energy of the μ -meson. If we make the assumption that we can replace it by an effective, constant, value then we may write

$$\frac{\delta N_e^\delta}{N_e^\delta} = \frac{\delta N_\mu}{N_\mu} \quad \text{where } N_e^\delta = \text{Total number of } \delta\text{-electrons.}$$

Thus to a first approximation the variation of the δ -electrons reproduce the variations of the μ -mesons, both atmospheric and non-atmospheric.

Non-Equilibrium Electrons

The rest lifetime of the π^0 meson is so short, $\lesssim 10^{-15}$ seconds, that it decays almost immediately. Its decay m.f.p. in air is $\sim 10^{-5}$ cms. while its nuclear capture m.f.p. $\sim 10^4$ cms. The behaviour of the emitted photons is then determined largely by the mass of air traversed, and hence is not a function of atmospheric temperature. The intensity of this component reaches a maximum at about 14 - 15 kms, then decreases according to the expression

$$N_e^{\text{Non-Equilib.}} \sim e^{-h_0/L} \quad \text{where } L = 135 \pm 5 \text{ grms/cm}^2 \text{ (Barker 1955)}$$

$$\therefore \frac{\delta N_e^{\text{Non-Equilib.}}}{N_e^{\text{Non-Equilib.}}} = \frac{\delta h_0}{L} \simeq (0.74 \pm 0.08)\%/\text{mb. } \delta h_0$$

Composition of the Soft Component at Sea Level

Dorman endeavours to compute the relative proportion of each source of electrons at sea level. He arrives at the ratio

$$\text{Decay: } \delta : \text{Non-Equilibrium} = 0.5:0.4:0.1$$

He does not indicate how he estimates the proportion of non-equilibrium electrons, but his ratio of δ -electrons seems too high

from an experimental angle. While it seems justifiable to ignore instrumental cut-off in deriving the variation with atmospheric changes of each component, it is not so when estimating something to be actually measured. In relating the total number of decay electrons to the total μ -meson flux, Dorman obtained the expression

$$N_e^\delta \approx 0.15 N_\mu$$

However this includes electrons down to a minimum energy of 10^5 eV. Instrumental cut-off will vary according to set-up, but 10^7 eV seems a reasonable and general value (Rossi and Klapman 1942). Substituting this as a minimum figure in Dorman's expression we obtain

$$N_e^\delta \approx 0.05 N_\mu$$

This is in approximate agreement with the accurate value of $0.064 N_\mu$, calculated taking cascade theory into account (Rossi 1948, Barker 1955). Using the rest of Barker's values and taking into account the fact that about 10% of the soft component is soft μ 's of momenta < 240 MeV/c, we obtain the ratios
Decay: δ :Non-Equilibrium:Soft μ 's = 0.55:0.20:0.15:0.10

Using these figures, and collecting terms together, for the entire soft component we expect

$$\frac{\delta N_e}{N_e} = - \frac{0.17\%}{mb} \delta h_0 + \frac{0.20\%}{\circ C} \delta T_s + 0.75 \frac{\delta N_\mu}{N_\mu} + 0.1 \frac{\delta N_\mu}{N_\mu} \frac{\text{soft}}{\text{soft}}$$

..... (1.4) (1)

The Hard Component

The atmospheric effects of the hard component, can be divided into a simple absorption effect, and decay effects due to the instability of the μ -meson.

The main decay effect is the so-called negative atmospheric temperature effect, already mentioned in the introduction to this chapter. Due to the height variations of the mean μ -meson producing layer with varying atmospheric temperature, an increase in temperature (and height) causes a decrease in intensity at sea level.

There are several possible 'positive' temperature effects. The small, but finite, probability of the π -mesons suffering nuclear interaction before they decay, will depend on the local atmosphere density and thus on the temperature of the π -meson production layer. According to Dorman (1956) this effect is significant only in the range 40 to 60 GV and cannot exceed the magnitude $\frac{1}{T(h_\pi)} \text{ }^\circ\text{C}$ where $T(h_\pi)$ = Temperature of formation level of π -mesons.

A representative value of the upper limit in the 50 to 100 mb region is $0.5 \text{ }^\circ\text{C}$.

The more important positive temperature effect is the so-called 'mass redistribution' effect. It is possible that while the pressure and the height of the μ -meson producing layer remain constant, temperature changes may occur that will cause

redistribution of the mass of the atmosphere. Since the relativistic dilation of an unstable particle depends on its velocity, it is also a function of the distribution of a given thickness of absorber along its path. Particles will decay earlier if the absorber is concentrated near the start of their atmospheric path. Thus if, with other atmospheric parameters kept constant, the temperature of the upper atmosphere increases, then that of the lower atmosphere decreases, causing an increase in the probable life of the μ -mesons.

Theoretically there have been two main types of approach to the atmospheric effects problem. Dorman (1956) has set up expressions connecting the primary radiation with that observed at sea level, taking into account atmospheric processes. Applying variational calculus to these expressions with respect to the atmospheric variables, he obtains equations relating sea level changes in μ -meson intensity with changes in the pressure and temperature throughout the atmosphere. From these equations are constructed the so-called 'Density of temperature coefficients' functions. These functions express the contribution of atmospheric temperature changes at each depth in the atmosphere to changes in surface μ -meson intensity. Thus a variation δN_μ in μ -meson intensity due to atmospheric temperature variations $\delta T(h)$ is expressed as

$$\frac{\delta N_\mu}{N_\mu} = \int_0^{h_0} W_T(h) \delta T(h) dh$$

where $W_T(h)$ = Total Density of temperature coefficients function over the entire atmosphere, and $h_0 =$ surface level.

For the barometer effect, for vertical incidence, Dorman predicts a value of about -0.14 to $-0.15\%/mb$ at sea level. This includes the effect of absorption and also the decay of μ -mesons due to changes in pressure. It is not possible to give single values for the temperature effects predicted by Dorman, since his values are integrated over the entire atmosphere.

Trefall(1955) has approached the analysis with a view to simple, direct, comparison between his theory and the experimental results. The most commonly used experimental approach to the analysis of atmospheric temperature effects on the surface μ -meson radiation is to use the 3-fold regression equation introduced by Duperier (1949,) i.e.

$$\frac{\delta I}{I} = \beta_1 \delta P + \beta_2 \delta H + \beta_3 \delta T \quad \dots\dots 1.4.(ii)$$

where I = Cosmic Ray intensity; P = Pressure; T = Mean temperature of 100 - 200 mb layer; H = Height of given pressure level (usually 100 mb. layer).

Trefall sets up the theoretical analogue of this equation

$$\frac{dI}{I} = \left[\frac{1}{S} \frac{\partial S}{\partial T} + \frac{1}{w_d} \frac{\partial w_d}{\partial T} \right] dT + \frac{1}{w_d} \frac{\partial w_d}{\partial H} dH + \left[\frac{1}{w_d} \frac{\partial w_d}{\partial P} + \frac{1}{w_d} \frac{\partial w_d}{\partial P} \right] dP \quad \dots\dots 1.4.(iii)$$

where S = Number of μ -mesons produced at given atmospheric layer

w_d = Probability that μ -meson shall not decay before it reaches sea level

w_1 = Probability that μ -meson has sufficient momentum to be recorded.

Inspection of equation 1.4.(iii) shows that, with the exception of the decay term, the allegedly unique regression coefficients have several components.

The value of the positive temperature effect predicted varies with the reference level used, but a representative value, using the 100 mb level, is 0.06%/°C for 10 cms. of lead absorber in the detector. The barometer effect, the sum of the pure absorption effect and the μ -e decay, is predicted as -0.12%/mb for the same amount of absorber.

The negative temperature effect is harder to compute since it involves the choice of a temperature model of the atmosphere, but a representative value is -6 to -7%/km.

3 1.5 Measured Atmospheric Effects

The Nucleonic Component

The atmospheric absorption coefficient for the nucleonic component has been measured rather carefully (Simpson et al. 1953) and found to be $-(0.70 \pm 0.04)\%/mb$, in close agreement with the theoretical prediction. In the same paper an experimental analysis of the temperature effects is attempted. The analysis uses only two months of data and is a correlation between sea

level nucleonic intensity and the temperature of the 200-850 mb. layer. The values obtained are $(0.10 \pm 0.12)\%/^{\circ}\text{C}$ and $(0.02 \pm 0.15)\%/^{\circ}\text{C}$. With such large errors it is difficult to make any statement. However, Thambyapillai (1953) has correlated sea level ~~sea level~~^{nucleonic} intensity with the height of various pressure levels. He obtained very small correlation coefficients of fluctuating sign which were not significant. The evidence is not altogether satisfactory but seems to indicate that the sea level nucleonic intensity is effectively atmospheric temperature independent. No-one appears to have investigated the possibility of a small local temperature coefficient.

The Electron-Photon Component

The number of long term studies of the 'soft' component are rather limited. We shall quote those due to Trumpy (1955) and Dawton and Elliot 1953. We first consider the partial baro-coefficient. Trumpy and Dawton and Elliot obtained values in good agreement i.e. $(-0.34 \pm 0.01)\%/mb$ and $(-0.35 \pm 0.01)\%/mb$ respectively. Using their respective values for the partial barometer coefficient of the hard, which are again in good agreement, and equation 1.4.(i) we computed the theoretical value i.e. $-0.35\%/mb$. This value is obtained taking into account that some 10% of the soft component is actually soft μ -mesons absorbed in 10 cms. of lead. These soft μ -mesons have a barometer co-

efficient at sea level of $-0.9\%/mb$ according to Dorman. If we accept this figure then there is good agreement between the theoretical and experimental values.

It is rather difficult to ~~directly~~ ^{directly} assess the atmospheric temperature effects for the 'soft' component. However, we can see in a general way that the total effect should be close to zero. A direct regression between the hard component measured close to sea level and the surface temperature, keeping the pressure constant, is known to be unreliable since the values obtained vary seasonally (Hogg 1947), but is some indication of the total atmospheric temperature effect for the hard component. The mean yearly value, averaged over a period of years, of the temperature coefficient obtained by Hogg (1947) from a regression analysis based on daily cosmic ray totals and mean daily values of the pressure and surface temperature, was $-0.112\%/^{\circ}C$, but with considerable scatter about this value, the extreme values being -0.011 and $-0.286\%/^{\circ}C$ respectively. If we use this value and also assume that the value for the soft μ -mesons will certainly not be less than this, then substitution in equation 1.4.(i) indicates a total temperature coefficient for the 'soft' component of about $0.1\%/^{\circ}C$. However Rossi (1948) estimates that the computations involved in estimating the composition of the soft component at sea level may be as much as 10% in error. If this is so, then our value for the coefficient becomes $(0.1 \pm 0.03)\%/^{\circ}C$, and in

view of the wide scatter of the hard component value of $-0.112\%/^{\circ}\text{C}$, this cannot be considered significant.

This is, in general, reflected by the decay and positive temperature coefficients obtained by Trumpy (1955) and Dawton and Elliot (1953).

	<u>Decay Coefficient</u>	<u>Positive Temperature Coefficient</u>
Trumpy	$(-1.38 \pm 0.80)\%/ \text{km.}$	$(0.021 \pm 0.027)\%/^{\circ}\text{C}$
Dawton & Elliot	$(-1.05 \pm 0.48)\%/ \text{km.}$	$(0.021 \pm 0.020)\%/^{\circ}\text{C}$

There is good agreement between the two. The positive temperature coefficient cannot be considered significant, and the decay coefficient is rather small and of very doubtful significance.

The Hard Component

This is the most carefully studied component of the cosmic radiation. The usual method of analysis is by using Duperier's 3-fold regression formula. Thus while it is straightforward to compare experimental values with Trefall's predictions, only the barometer effect is directly comparable in the case of Dorman's analysis.

Dorman predicts a value of -0.14 to $-0.15\%/ \text{mb}$ for 10 cms. of lead absorber, apparently for the partial barometer coefficient.

Actual measured values are:	-0.127 ± 0.008	Trumpy and Trefall 1953
	$-0.125 \pm 0.012\%/ \text{mb}$	Dawton and Elliot 1953
	-0.129 ± 0.004	Lindgren and Lindholm 1960
Trefall's predicted value is $-0.12\%/ \text{mb}$, in reasonable agreement		

with the experimental results.

An example of the difficulties inherent in dealing with the model atmosphere implicit in equation 1.4.(ii) is illustrated by Trefall's attempts to calculate the negative temperature coefficient. One cannot assume uniform temperature variations throughout the atmosphere because of the third term in equation 1.4.(ii), so it is necessary to compute the negative temperature effect keeping the stratospheric temperature constant. This yields a value $\sim -6.7\%/km$ for 10 cms. of lead absorber. However, using the same sources of data as above, the measured value is $\sim (4.5 \pm 0.5)\%/km$. This is only a little over half the predicted value and implies something wrong with the stratospheric-tropospheric model used.

Finally we consider the positive temperature effect. Trefall predicts a total temperature coefficient of $0.06\%/^{\circ}C$. This compares reasonably with the measured values of

0.036 ± 0.021	Trumpy and Trefall
0.056 ± 0.018	Dawton and Elliot
0.039 ± 0.024	Lindgren and Lindholm

To test Dorman's model experimentally, using regression analysis, is rather difficult as it involves a considerable number of temperature variables that are probably intercorrelated. Lindgren and Lindholm have attempted to do so using an electronic computer for the computation. Their final conclusion was that

quantitatively Dorman's temperature density curve for a cubical μ -meson telescope at sea level is about 35% in error, but could be improved by adjusting the exponent of the assumed

π -meson spectrum used by Dorman. It then gives results comparable with the Duperier model.

Summing up; the atmospheric effects for the nucleonic and electronic components seem adequately described by present theory. Atmospheric temperature effects for the electronic component need more careful measurement for complete confirmation.

For the hard, μ -meson component, the situation is less well defined. As indicated above Dorman's curves have to be rather artificially adjusted to give comparable values to the experimental results. His value for the partial barometer coefficient seems rather high. Trefall's analysis gives good agreement between experiment and theory for the partial barometer effect and fair agreement for the positive temperature effect. However there is a large discrepancy between his theoretical, and the measured, decay coefficient. This may be due to the nature of the model atmosphere required by the regression equation used in the data analysis.

CHAPTER II - THE APPARATUS

2.1. The Telescope

Figure 2.1 (i) is a schematic, block diagram of the apparatus. It comprises three trays of halogen quenched geiger tubes, a layer of lead 10 cms. thick, and immediately underneath the lead is a large tray of alcohol quenched geiger tubes, double banked for the most efficient possible detection. Also shown is the associated electronics and the recording apparatus. The key is given by

C.U. $\equiv$ Coincidence Unit

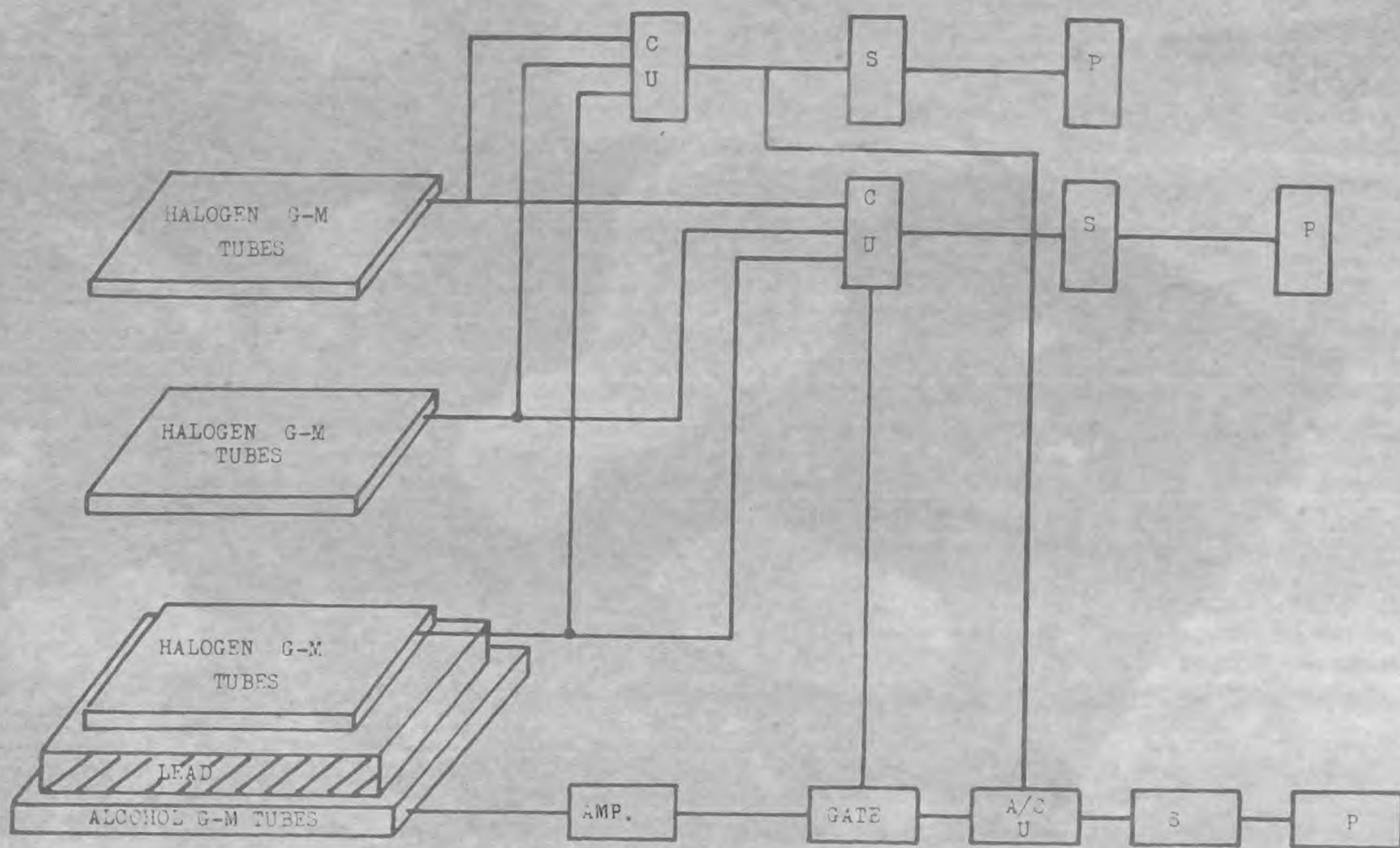
S $\equiv$ Scaler

P $\equiv$ Recorder and Printer

A/C U $\equiv$ Anti-Coincidence Unit.

Where there is any doubt as to the correct interconnection between the units a dot has been used to indicate the appropriate one.

The three trays of halogen quenched tubes are arranged in an almost cubical telescope geometry (60 x 55 x 63)cms³. They each have a sensitive area of 3×10^3 cms², and are 31.5 cms. apart, subtending a solid angle of approximately 3.3 steradians. The alcohol quenched tubes have a sensitive area of 9×10^3 cms² and due to their high efficiency and because they are double banked form a layer of detection efficiency of about 95-96%.



BLOCK DIAGRAM OF SOFT COMPONENT EXPERIMENT

FIGURE 2.I(i)

The operation of the telescope is as follows; the three top trays are connected in three-fold coincidence, using a conventional Rossi coincidence unit, and this measures the total radiation i.e. the sum of the μ -mesons and the electrons. The output of three-fold coincidence unit is scaled down by a factor of one hundred before being recorded and printed. This scaling factor is also applied to the other outputs.

The four trays are connected to a four-fold Rossi coincidence unit and this records all particles traversing the top three trays, the 10 cms. of lead and the bottom tray. This is the hard component and is almost entirely μ -mesons.

Finally the top three trays and the bottom tray are connected in anti-coincidence. Particles traversing the top three trays but not the bottom one i.e. they are absorbed in the lead, are recorded as anti-coincidences. The design and operation of the anti-coincidence unit will be considered in detail later.

2.2. Halogen Quenched Geiger-Müller Tubes

The use of an electro-negative halogen quenching agent in geiger tubes is a fairly recent development and some aspects of the mechanism involved are not yet fully interpreted. Compared with alcohol quenched tubes there are several advantages, and two main disadvantages. The halogen quenched tubes have low operating voltages, of the order of 300 - 450 volts, a theoretically infinite lifetime, can be used with a very simple R - C quenching

circuit, and have stable, but in our case rather short, plateaux. This latter is probably due in part to our rather low R i.e. $4.7 \text{ M}\Omega$ and to the difficulty of keeping the capacity across the tube to a very few pf's when operating a large number in a single tray. One of the minor weaknesses of the halogen quenched geiger tube is extreme sensitivity to the amount of capacity connected across it. (Van ^{DUREN}~~ZANEN~~ 1959).

The main weaknesses are a detection efficiency somewhat less than unity, depending on the partial pressure of halogen, in our case the efficiency being about 80%, and a variable time lag of some 5 to 15 μ secs. between the passage of the particle through the tube and the pulse appearing at the anode. This latter effect can be very serious in some applications i.e. fast counting.

The quenching mechanism is the usual one. The positive ion - sheath of rare gas ions stops the discharge, then, as it moves out towards the cathode loses its charge to the quenching agent, in our case Br_2 . At the cathode the ionized bromine is neutralized and dissociated before it has time to radiate excess energy and cause secondary emission. Instead of being steadily used up as in alcohol quenched tubes, the halogen quenching agent reforms after dissociation, so that if cathode materials that are unaffected by atomic halogen are used, the geiger tube should have infinite counting life. Tests seem to indicate a life in excess of 10^9 counts at least.

To ensure efficient quenching action in alcohol quenched tubes, rather high partial pressure of alcohol has to be used. This means high operating potentials in the range 1100 to 1200 volts. With halogen quenched tubes low partial pressures of halogen are adequate for good quenching action. Thus relatively low operating voltages can be used so that the threshold voltage of a halogen quenched tube is usually in the region of 300 - 400 volts. Plateaux of 100 volts or more have been reported for this type of tube (e.g. Franklin and Loosemore 1950) but for the tubes used in this experiment the plateaux were of the order of 50 to 60 volts long, although very stable. The slopes were of the order of 0.3 to 0.4%/volt.

Why halogen tubes show such long time delays relative to the passage of the particle through the tube before the voltage pulse appears at the anode is by no means certain. Van Zoonen (1953) supposes that the delay may be due to the slow build-up of ionization in the avalanche near the central wire anode. This is probably due to the slow build-up of secondary avalanches with respect to the primary avalanches. Experimental evidence that time lag decreases with overvoltage, and also with higher halogen pressures (Sherman 1956) supports this idea.

Several points of interest arise concerning the time-lag mentioned above. Since our counting rates were relatively low this was not a serious problem and in part worked to our

advantage. It was found necessary to make the resolving time of the Rossi coincidence units at least 16 μ seconds, but statistically this is not serious. The existence of the delay also saved putting in an additional delay line between the 3-fold coincidence unit and the A/C unit to ensure that the pulse from the bottom tray had adequate time to reach the A/C unit.

2.3. The Alcohol Quenched Geiger Tubes

The use of a quenching unit with each alcohol quenched geiger tube to prolong the life and reliable working of the tube is by now a well established technique (e.g. Elliot 1949). The quenching unit is a 'flip-flop' multivibrator which when triggered by the geiger pulse applies a sharp, negative, square pulse with a leading edge of -300 volts to the anode of the tube. The pulse used in our units was 0.8 milliseconds wide.

An inherent problem in the operation of twenty five quenching units in a single tray is 'multiple triggering'. The outputs of the units are A.C. coupled via ~~a~~ condensers of a few pfs to the output line. If three or more units are triggered simultaneously, the net pulse on the output line feeds back to trigger the entire tray, causing it to be paralysed for 0.8

m secs. To overcome this the anode loads on the quenching pulse half of the units were connected to the H.T. line via a common, 100 Ω resistor. The junction of the anode loads and the resistor was then A.C. coupled to the output line. The

magnitude of the single, negative going, pulses now appearing on the output line were 0.5 volts, eliminating the multiple triggering.

To provide a satisfactory pulse for operating the external multivibrator, this half-volt pulse is amplified and shaped in a head amplifier to give a sharp-edged, negative going, 30 volt pulse.

The reason for using alcohol quenched geiger tubes in the bottom tray, and for double-banking them, is because we need to detect all particles passing through the lead as efficiently as possible. The alcohol tubes are approximately 99% efficient (Rossi 1952) as opposed to the 80% efficiency of the halogen tubes. This need for high efficiency is why the quenching pulse is only 0.8 μ secs wide, compared with the usual 1.5 to 2.0 μ secs, thus keeping the paralysis time of the geiger tubes as short as possible. Another advantage of using alcohol tubes is that the time lag between the particle passing through the tube and the pulse appearing at the anode is $\leq 1 \mu$ sec (Rossi 1952). This is compared to the variable 5 to 15 μ secs. for the halogen quenched tube. Such a short time lag combined with the long lag associated with the halogen tubes saves the use of an additional delay line between the output of the three-fold coincidence unit and the input of the A/C unit (see section 2.2).

2.4. The Anti-Coincidence Unit

In figure 2.4.(i) is shown the circuit of the alcohol tray external multi-vibrator and anti-coincidence unit combined.

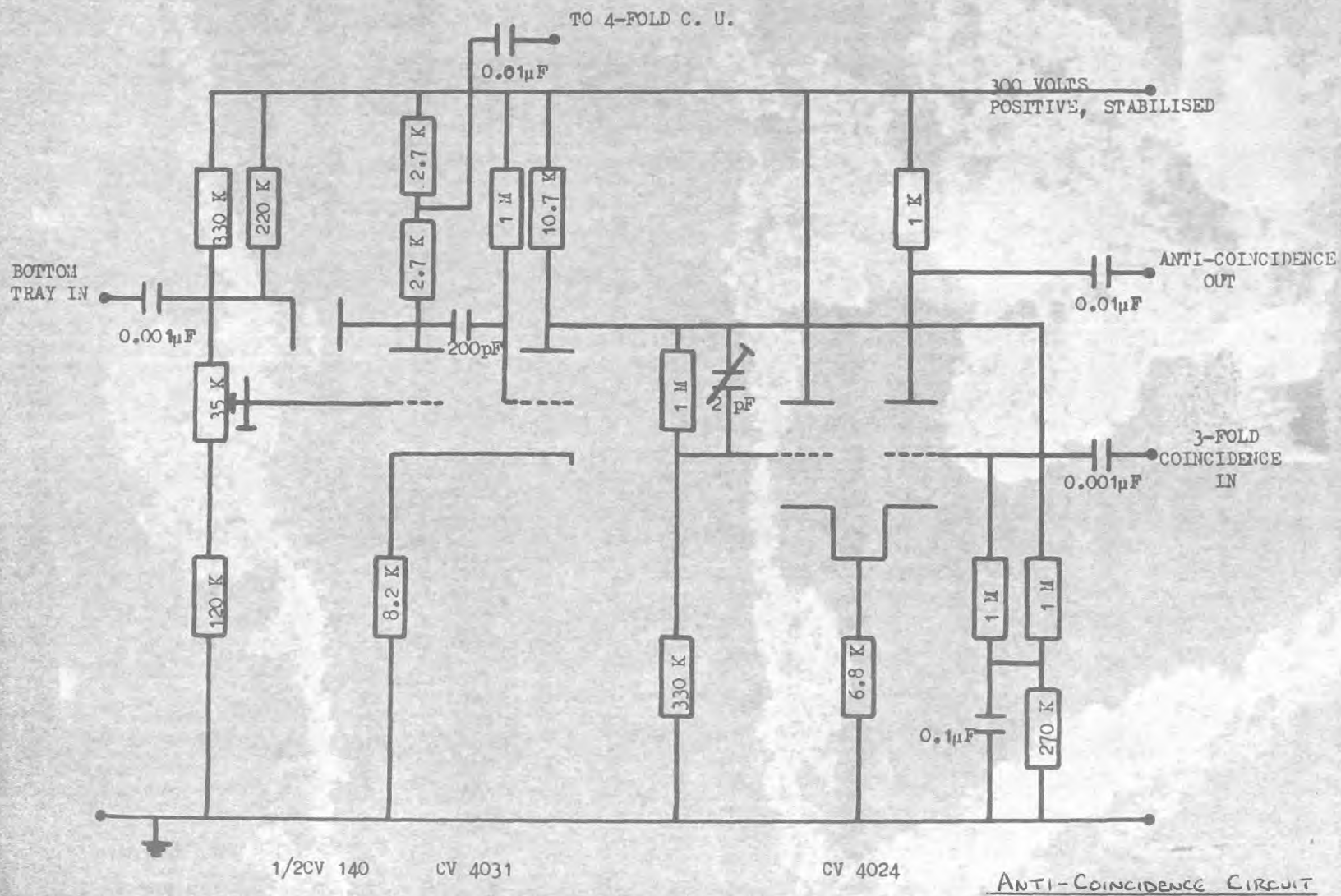


FIGURE 2.4 (1)

The amplified pulse from the alcohol quenched geiger tubes is fed to one anode of the multivibrator via a diode. This pulse triggers a square negative pulse on the multivibrator, 32 μ secs. wide, part of which is tapped-off and fed to the four-fold coincidence unit. From the second anode of the 'flip-flop' a positive pulse, 120 volts high is passed to the anti-coincidence unit. This A/C unit comprises a double triode, one half of which acts as a cathode follower, the other half as a simple 'mixing' circuit. The 120V positive pulse from the anode of the multivibrator is suitably attenuated and fed to the grid of the cathode follower half, thus appearing at the common cathode of the two triodes as a 29V high positive pulse. To the grid of the second triode, wired as a simple amplifier, is applied the output of the three-fold coincidence unit, a positive pulse 10 to 15 volts high and 5 to 10 μ secs. wide. Now, if at the same time as a three-fold coincidence pulse appears at the grid of the second triode, a positive pulse from the multivibrator appears at the cathode of the triode, then the grid remains negative with respect to the cathode and nothing appears at the anode of the second triode. However, if no positive pulse is applied at the cathode then the triode conducts and a negative, 7 volt, pulse appears at the anode. This is the anti-coincidence pulse.

The three-fold coincidence input grid of the second triode is D.C.-coupled to the positive going anode of the multi-

vibrator to compensate for any D.C. drift due to ageing of components.

2.5. Errors

There are two sources of error to be considered. One source is due to chance coincidences and anti-coincidences caused by inefficiencies in the counting circuitry, and also to the effects of showers. The other is due to the statistical fluctuations of the counting rates.

The chance coincidence rate of a number of statistically independent counting rates $C_1, C_2, \dots C_n$, combined in a coincidence unit of resolving time T is given by

$$A = n C_1 C_2 \dots C_n T^{n-1}$$

For the four-fold rate, the hard component, this is negligible i.e. $\sim 10^{-3}$. To check the total effect due to chance coincidences, showers and pick-up, we displaced one of the three halogen tube trays from the geometry of the telescope. The four-fold rate was then found to be approximately 1% of its normal counting rate.

It is not possible to ~~directly~~ ^{directly} calculate the chance anti-coincidence rate since the output of the three halogen tube trays is not statistically independent of the bottom tray rate. However, with one of the three upper trays displaced, the anti-coincidence rate was found to be 3% of its normal

counting rate.

In general, cosmic radiation counting rates are found to approximate to Poisson distributions. The bi-hourly counting rates for the hard and soft components were 480×10^2 and 160×10^2 respectively. These correspond to standard deviations of $\sqrt{480 \times 10^2}$ and $\sqrt{160 \times 10^2}$, or 0.46% and 0.79% respectively. The latter becomes 0.80% when account is taken of the constant 3% error indicated above.

CHAPTER 3 - ATMOSPHERIC EFFECTS OF THE SOFT COMPONENT

3.1. Introduction

The nature of the atmospheric temperature effects of the hard and soft components of the cosmic radiation at sea level have been discussed at some length in Chapter 1. The object of the soft component experiment carried out at Imperial College was to check that the soft component has small atmospheric temperature effects associated with it, as reported by previous workers (Trumpy 1955 Dawton and Elliot 1953), to see if the surface temperature effect predicted by Dorman exists, and finally to use the information gained to try to obtain an atmospheric temperature correction for the daily variation of the hard component. The importance of such a correction has been discussed in Chapter 1.

The apparatus was initially installed in a hut on the roof of the old Physics building, with a roofing of about 2 grms/cm². The equipment ran satisfactorily from October 1959 onwards. The data used in the analysis therefore runs from October 1959 to December 1960. In June 1960 the entire apparatus was transferred to the new Physics building and was set-up under a light/aluminium roof, approximately 2 grms/cm². Due to the need to conserve laboratory space some additional absorber was included in the solid angle of the telescope causing a small

decrease in intensity. Using the results of the nucleonic component detector at Amsterdam, the geomagnetic latitude of Amsterdam being 54.5°N , the decrease was determined as 1.55% for the hard and 4.04% for the soft.

3.2. The Experimental Results

It is now well established (Duperier 1949, Bachelet and Conforto 1956, French and Chasson 1959) that a regression analysis based on a Duperier-type regression equation

$$\frac{\delta I}{I} = \beta_1 \delta P + \beta_2 \delta H + \beta_3 \delta T \quad \dots\dots 3.2.(i)$$

is adequate to determine atmospheric effects provided care is used in the selection of the data. The main cause of anomalous results is primary variations in the data, since the above equation is based on the assumption that they do not occur. These variations are particularly prevalent at times around sunspot maxima. Large disturbances such as Forbush decreases are easily detected but smaller variations over several days are often difficult to observe, particularly in the soft component with its large pressure coefficient.

There are two means of dealing with the problem. One is the method, recently used by Lindgren and Lindholm (1960), of employing the 4-fold regression equation

$$\frac{\delta I}{I} = \beta_0 \delta N + \beta_1 \delta P + \beta_2 \delta H + \beta_3 \delta T \quad \dots\dots 3.2.(ii)$$

where N = Pressure corrected nucleonic component intensity.

The use of N as a measure of primary cosmic ray variation is open to criticism but Lindgren and Lindholm showed that its use leads to a considerable reduction in the scatter of values obtained in an analysis using equation 3.2.(i). The main practical objection is the amount of computation involved in a 4-fold regression analysis.

A second, quicker, method, which requires a knowledge of the rough values of β_1 , β_2 and β_3 expected, was used in our case. The data was selected as carefully as possible and analysed using equation 3.2.(i). The values of β_1 , β_2 and β_3 obtained were inspected and those groups for which the values appeared anomalous were corrected for pressure variations, using the values of β_1 obtained. The values of β_1 are least affected by the presence of primary variations. From the pressure corrected data it was fairly simple to spot the days with large primary variations. They were removed and the data re-analysed. In general this procedure is open to objection, but in our case where we wished to check our results for the hard component with well established values of β_1 , β_2 and β_3 , and to compare them with the values obtained for the soft component, it is adequate.

Another point to be considered is which height to use in equation 3.2.(i) i.e. the 200 mb., 100 mb. or 50 mb. French and Chasson (1959), and Lindgren and Lindholm (1960) have

shown that with the present accuracy of radio-sonde data there is very little to choose between the results using either the 200 mb. or 100 mb. heights. The frequent failure of radio-sondes to reach the 50 mb. layer make it impracticable to use this height. We have therefore used the height of the 100 mb. layer and the temperature of the 200 to 100 mb. layer as our variables in equation 3.2.(i). Their values were obtained from the two radio-sonde flights made at 00 and 1200 hours at Crawley, close to London, by the Meteorological Office. The pressure values were obtained from the Royal Kew Observatory.

We thus performed a regression analysis of the daily totals of the hard and soft components of the cosmic radiation against the daily mean values of the pressure, the height of the 100 mb. layer, and the temperature of the 200 - 100 mb. layer. Later we did an analysis using the surface temperature in place of the temperature of the 200 - 100 mb. layer. The results of this analysis are shown in Tables 3.2.(a) and 3.2.(b). Table 3.2.(a) shows the values of the regression coefficients obtained for each group. Also shown are the multiple regression coefficients and the number of days in each group. The errors quoted were computed by standard statistical methods. Table 3.2.(b) contains the mean values obtained from each set of groups. They were obtained by weighting each value by $1/\sigma^2$ (Brunt 1933). The quoted errors were calculated from the scatter inside each set, similarly weighting each deviation.

Group	SOFT	HARD	Number Of Days
1	-0.330 ± 0.02	-0.113 ± 0.02	15
2	-0.385 ± 0.029	-0.154 ± 0.013	27
3	-0.313 ± 0.024	-0.117 ± 0.014	20
4	-0.330 ± 0.025	-0.093 ± 0.015	33
5	-0.290 ± 0.044	-0.115 ± 0.016	16
6	-0.330 ± 0.015	-0.146 ± 0.015	15
7	-0.277 ± 0.073	-0.104 ± 0.046	14
8	-0.360 ± 0.024	-0.098 ± 0.017	15
9	-0.374 ± 0.024	-0.140 ± 0.006	25
10	-0.302 ± 0.032	-0.111 ± 0.015	25

1	-4.64 ± 1.5	-6.64 ± 1.5
2	$+4.41 \pm 1.85$	-4.41 ± 0.80
3	-4.91 ± 1.06	-5.46 ± 0.61
4	-0.35 ± 1.5	-2.70 ± 0.89
5	$+0.38 \pm 6.27$	-7.16 ± 2.23
6	$+0.23 \pm 2.04$	-1.14 ± 2.04
7	-3.02 ± 4.40	-6.57 ± 2.79
8	$+0.96 \pm 1.34$	-6.04 ± 0.94
9	-5.72 ± 3.40	-3.70 ± 0.89
10	$+5.25 \pm 3.24$	-1.55 ± 1.52

RESULTS OF REGRESSIONANALYSIS OF HARD ANDSOFT COMPONENTS

Group	SOFT	HARD	R_s	R_{ph}
I	$+0.021 \pm 0.060$	$+0.138 \pm 0.060$	0.985	0.984
2	$+0.019 \pm 0.047$	$+0.025 \pm 0.020$	0.967	0.975
3	-0.063 ± 0.045	$+0.036 \pm 0.026$	0.978	0.972
4	-0.047 ± 0.028	$+0.002 \pm 0.017$	0.970	0.984
5	-0.082 ± 0.102	-0.006 ± 0.038	0.850	0.907
6	$+0.033 \pm 0.038$	$+0.027 \pm 0.039$	0.900	0.904
7	-0.284 ± 0.090	$+0.096 \pm 0.043$	0.927	0.914
8	-0.138 ± 0.034	$+0.082 \pm 0.024$	0.988	0.985
9	-0.070 ± 0.097	$+0.057 \pm 0.026$	0.946	0.972
10	$+0.247 \pm 0.053$	$+0.182 \pm 0.025$	0.925	0.938

I	$+0.070 \pm 0.082$	-0.093 ± 0.084	0.902	0.781
2	$+0.162 \pm 0.051$	-0.117 ± 0.023	0.888	0.824
3	-0.076 ± 0.073	-0.034 ± 0.042	0.886	0.577
4	$+0.066 \pm 0.076$	-0.132 ± 0.056	0.867	0.814
5	-0.015 ± 0.116	-0.042 ± 0.041	0.680	0.761
6	-0.054 ± 0.098	-0.097 ± 0.098	0.924	0.828
7	$+0.259 \pm 0.273$	-0.436 ± 0.134	0.743	0.682
8	$+0.022 \pm 0.116$	$+0.011 \pm 0.066$	0.875	0.852
9	-0.211 ± 0.116	-0.078 ± 0.032	0.832	0.854
10	$+0.348 \pm 0.099$	$+0.298 \pm 0.047$	0.769	0.798

TABLE 3.2 a(continued)

	SOFT	HARD
$\beta_{IP,HT} \text{ \%}/\text{mb}$	-0.336 ± 0.011	-0.129 ± 0.006
$\beta_{IT,PT} \text{ \%}/\text{km}$	-1.55 ± 1.07	-4.56 ± 0.45
$\beta_{IT,PH} \text{ \%}/^{\circ}\text{C}$	-0.028 ± 0.030	$+0.051 \pm 0.018$
$\beta_{IT_s,PH} \text{ \%}/^{\circ}\text{C}$	$+0.070 \pm 0.042$	-0.055 ± 0.038

TABLE 3.2 b

MEAN VALUES OF REGRESSION COEFFICIENTS

Examining the results we see that there is reasonable agreement between the values obtained for $\beta_{IP, H_{100}, T_{200-100}}$, $\beta_{IH_{100}, PT_{200-100}}$ and $\beta_{IT_{200-100}, PH_{100}}$ and the values quoted in Chapter 1. Before we examine these values in detail, and in relation to equation 1.4.(i), we shall go on to consider the seasonal variation, and then discuss the agreement between experiment and theory for the meteorological effects of the soft component.

3.3. The Seasonal Variation

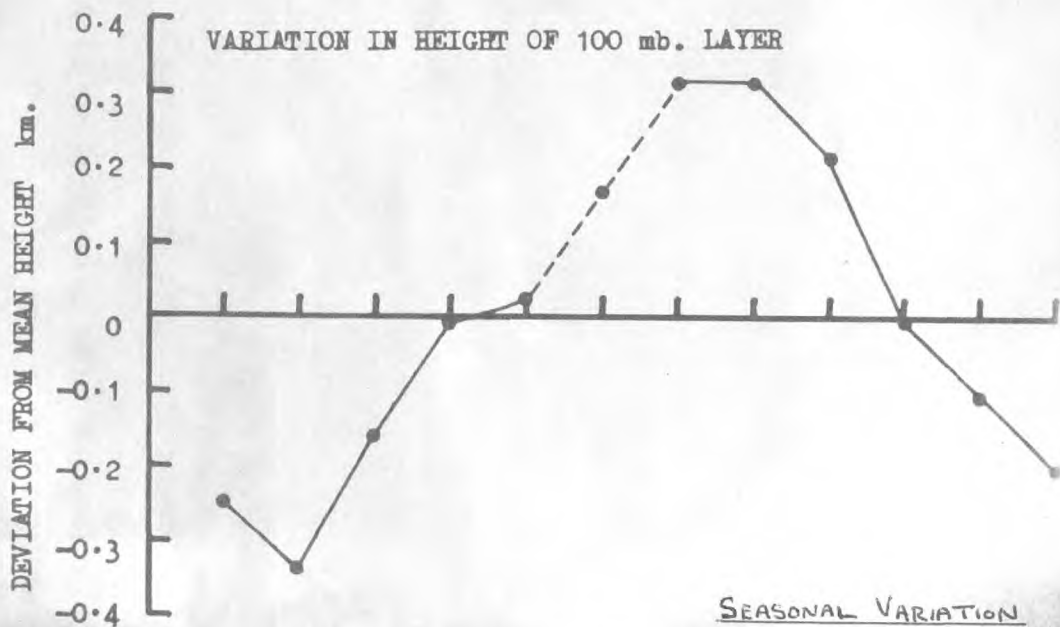
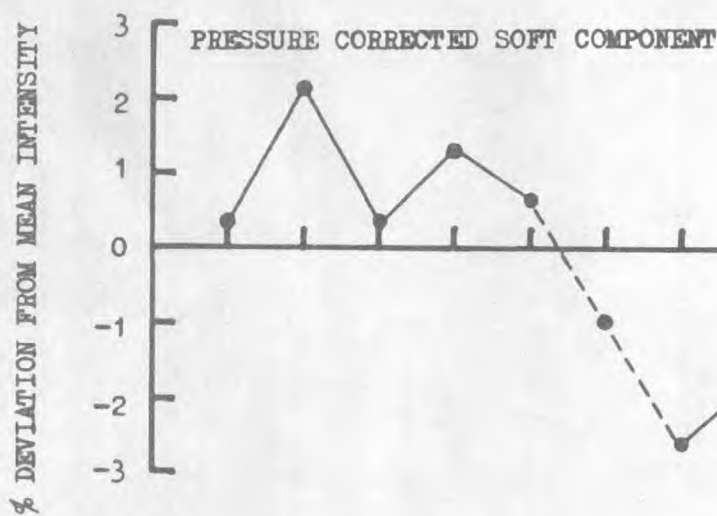
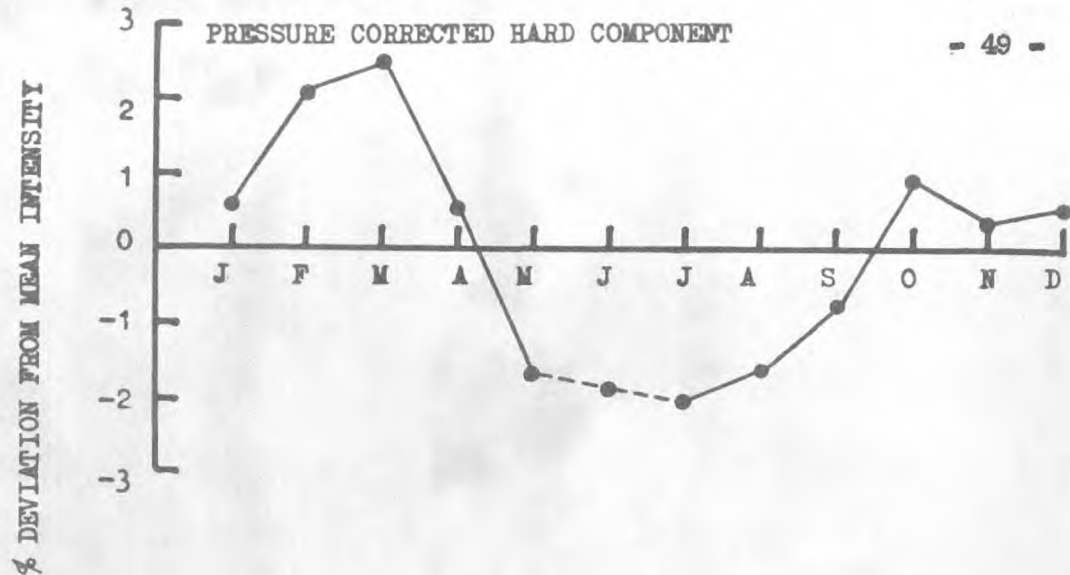
It is well known that the μ -meson component of the secondary cosmic radiation shows a marked variation over the course of the year (e.g. Dolbear and Elliot 1951). This variation is primarily due to the annual variation in atmospheric temperature. Such a seasonal change in intensity provides an alternative approach to the relation between atmospheric temperature and the intensity of μ -mesons and electrons at ground level.

An examination of the results of Dawton and Elliot (1953) shows that the seasonal changes of the 'soft' component vary from year to year (Dawton 1952). This is probably due to primary cosmic ray effects, since atmospheric temperature changes are fairly constant from year to year. Assuming these primary changes to be random, the more data used, the

less the primary cosmic ray effect. To utilise this point we have used some extra data, not employed in the regression analysis of section 3.2. This data was not used because it contained a small temperature effect that seriously affected the daily variation. However this was only of the order of 0.2 to 0.3% and was cyclic so did not seriously affect the general level of intensity. With this data our basic period becomes July 1959 to December 1960.

Figure 3.3.(i) shows the hard and soft components monthly totals, corrected for pressure variations. Unless indicated errors are less than the size of the plotting data. Also shown is the variation in the height of the 100 mb. layer. Due to the transfer of the equipment from the old to the new Physics building no data is available for June. The values have therefore been interpolated.

Figure 3.3.(ii) (a) and (b) shows the soft and hard monthly totals corrected for variations in the height of the 100 mb. layer, using the decay coefficients obtained in section 3.2. Figure (c) shows the variation in the temperature of the 200 - 100 mb. layer, left-hand scale, and the variation of the surface temperature, right-hand scale. Finally figure 3.3.(ii) (d) shows the monthly variation of the pressure corrected nucleonic component recorded at Amsterdam (Latitude 54.5°N geomagnetic) and at Mt. Norikura (latitude 25.6°N



SEASONAL VARIATION

FIGURE 3.3(1)

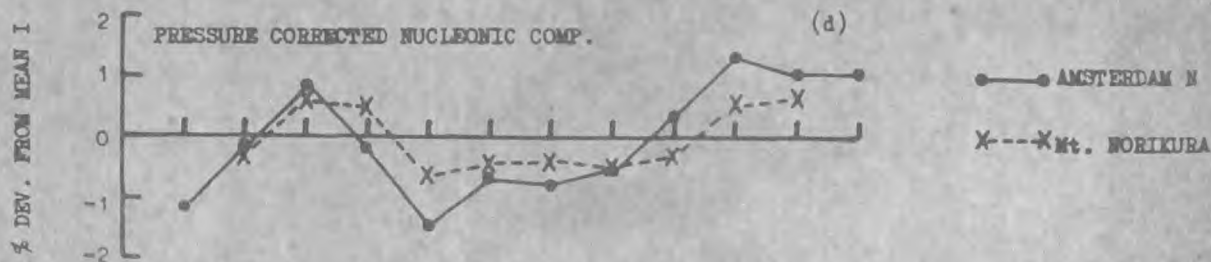
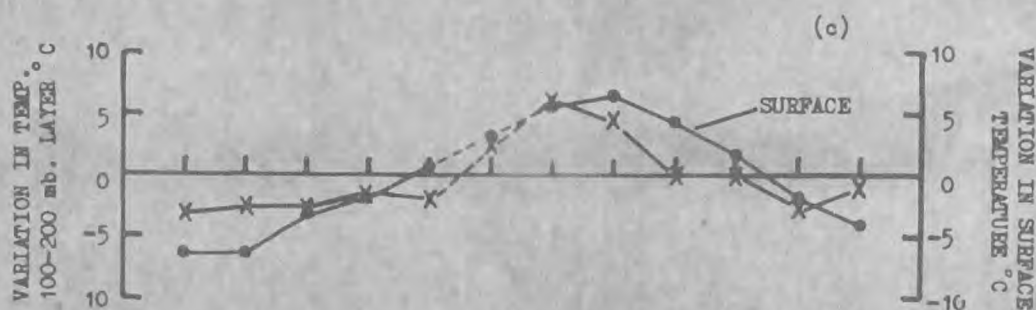
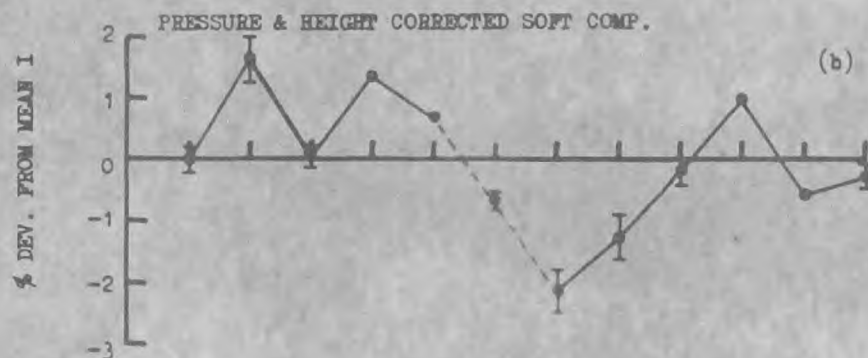
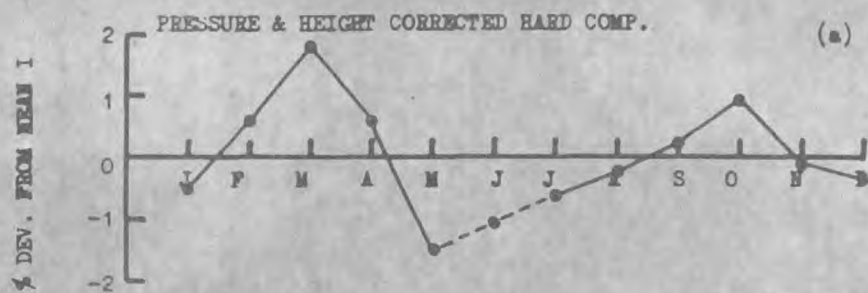


FIGURE 3.3(11)

SEASONAL VARIATION

geomagnetic). The nucleonic component is not subject to atmospheric temperature effects, but corresponds to a lower mean rigidity than the ionizing component (Chapter 1). However, the vertical cut-off rigidity at Mt. Norikura is of the order of 10 GV, so that the nucleonic component variation shown is a fairly high rigidity one.

An examination of (a), (b) and (d) shows that while the corrected hard component variation approximates to the changes observed in the nucleonic component, the corrected soft component variation is substantially larger than the primary cosmic ray variation indicated by the nucleonic data. Figure (c) shows that the temperature variations at the surface, and of the 200 - 100 mb. layer, are insufficient to explain the residual variations, if we use the regression coefficients obtained in section 3.2.

There are a number of possible explanations of the anomalously large variation, e.g. an instrumental defect, a different rigidity dependence to that of the hard component, or an incorrect decay coefficient. Due to the form of the instrumental set-up, an instrumental defect seems unlikely, since such a defect should appear in both the hard and the soft component data. To check this point we obtained monthly totals for the soft component by subtracting the hard component from the total component. The variation obtained in this way was not significantly different from that obtained using the

anti-coincidence measured soft component totals.

The appendix at the end of this thesis indicates that the rigidity dependences of the soft and hard components do not differ significantly during Forbush decreases or in a latitude survey.

Examination of figure 3.3.(i) shows an apparent marked negative correlation between the soft component data and the height of the 100 mb. layer. We therefore did a regression analysis between the mean pressure corrected monthly intensities of the soft component, the mean pressure corrected monthly intensities of the Amsterdam nucleonic component data, and the mean monthly height of the 100 mb. layer. The analysis was afterwards repeated using the pressure corrected hard component data in place of the soft component data. The results of the analysis are shown below.

	r_{I_{H100}, I_A}	$\beta_{I_{H100}, I_A} \text{ %/km.}$
HARD	-0.859	-5.27 ± 0.99
SOFT	-0.764	-4.67 ± 1.25

TABLE 3.3.A

There are two significant points to be noted. The value obtained from the hard component is not significantly different from that obtained by the regression analysis of

section 3.2. Also that the decay coefficient obtained for the soft component is now significant. This is in contradiction to the results of section 3.2. We thus have the situation where the soft component does not apparently suffer significant decay effects when considered on a day-to-day basis, but does so when considered on a monthly basis.

We first need to consider whether there are any other results consistent with this picture. An examination of Dawton and Elliot's data (1953) shows that the decay coefficient for the hard component is consistent when considered on a day-to-day or on a month-to-month basis. However, a regression analysis between the pressure corrected mean monthly soft component totals and the mean monthly heights of the 100 mb. layer yields a significant decay coefficient of $(-2.2 \pm 0.39)\%/km.$, as opposed to the value $(-1.05 \pm 0.43)\%/km$ obtained from the day-to-day regression analysis. The weakness of this analysis is that we do not know what the primary cosmic ray variations were during this period, so that it is possible that they would account for the difference between the two values. However, there does seem some grounds for supposing that a significant decay coefficient exists for the soft component when considered on a monthly basis.

3.4. Discussion

We need to consider a situation where the decay

coefficient obtained for the hard component is the same whether obtained by a day-to-day or month-to-month regression analysis, while a negligible decay coefficient for the soft component is obtained from a day-to-day regression analysis, and a significant decay coefficient obtained from a month-to-month analysis.

It has already been suggested in the introduction (section 1.5) that the apparent atmospheric temperature independence of the soft component when analysed on a day-to-day basis is due to an approximate equality between two temperature effects. A positive, surface temperature, effect and a negative, decay, effect. The positive effect is due to the probability of decay of a μ -meson depending inversely on the local density of the air, and the negative effect is due to the μ -mesons effectively in equilibrium with the electronic component.

A hypothesis of this form offers an explanation of the difference between the month-to-month and day-to-day decay coefficients for the soft component. If the change in height of the 100 mb. layer relative to a change in surface temperature were larger from month-to-month than from day-to-day, then the negative, decay, term would predominate, causing the seasonal variation to show a significant decay effect.

Duperier (1948) has pointed out that such a relationship between the upper atmospheric temperature and the surface temperature is certainly so, since the atmospheric heating

processes of conduction and convection are more efficient the longer the period for which they act. By a regression analysis between the monthly mean values of the height of the 100 mb. layer and the surface temperature T_s , he obtained the regression coefficient $\beta_{H_{100}T_s} = (0.045 \pm 0.002) \text{ kms}/^\circ\text{C}$. Since he used data from over a period of two years, he assumed pressure effects to be negligible. Using our data we obtained the values

$$\beta_{H_{100}T_s} = (0.048 \pm 0.003) \text{ kms}/^\circ\text{C}.$$

and

$$\beta_{H_{100}T_{s,P}} = (0.050 \pm 0.003) \text{ kms}/^\circ\text{C}.$$

in good agreement with Duperier, and confirming his assumption about the pressure. A similar analysis of our data for the day-to-day changes yielded a value

$$\left[\beta_{H_{100}T_{s,P}} \right]_{DD} = 0.015 \pm 0.003 \text{ km}/^\circ\text{C}$$

Thus

$$\frac{\left[\beta_{H_{100}T_{s,P}} \right]_{MM}}{\left[\beta_{H_{100}T_{s,P}} \right]_{DD}} = (3.3 \pm 0.45)$$

while the ratio of the appropriate soft component decay coefficients is (3.0 ± 1.26) . Thus the difference in the day-to-day value and the month-to-month value is consistent with the way in which the atmospheric temperature variation depends on height

Let us consider the surface temperature coefficient more closely, and also the values of $\beta_{T_s, P, H_{100}}$ we obtained

in section 3.2. Using our decay coefficients and the values of $\beta_{H100TS,P}$ we can determine the cosmic ray intensity-surface temperature relationships in $\%/^{\circ}\text{C}$. If we do this for both the hard and the soft components, then the difference should give the positive temperature term. In an attempt to get a significant result we shall make use of all the data available to us. As a form of check we shall also use the seasonal results, as well as the day-to-day values, although for Dawton and Elliot's results there is some doubt about the seasonal value of the decay coefficient for the soft component.

Table 3.4.A shows the results of the calculations. The first two columns show such experimental data as is available. The next two show the calculated values, and the last column is column 3 - column 4.

Examining the London results first, we see that there is good agreement between the experimental values and the values calculated from the decay coefficients for the hard component and also for the month-to-month value for the soft component. The agreement between the calculated and experimental day-to-day value is rather poor.

The values of the positive temperature coefficient agree reasonably well with each other, except the value computed from Dawton and Elliot's month-to-month data, and the difference here is not significant.

		EXPERIMENTAL $\%/^{\circ}\text{C}$		CALCULATED $\%/^{\circ}\text{C}$		POSITIVE COEFFICIENT $\%/^{\circ}\text{C}$
LONDON		$\beta_{I_s T_s, PH_{100}}$	$\beta_{I_H T_s, PH_{100}}$	SOFT	HARD	
	DD	0.070 ± 0.042	-0.055 ± 0.038	-0.023 ± 0.020	-0.068 ± 0.020	0.045 ± 0.028
		$\beta_{I_s T_s, I_{AM}}$	$\beta_{I_H T_s, I_{AM}}$			
	MM	-0.213 ± 0.067	-0.249 ± 0.048	-0.234 ± 0.077	-0.263 ± 0.065	0.029 ± 0.10
DAWTON AND ELLIOT 1953	DD			-0.016 ± 0.010	-0.060 ± 0.018	0.044 ± 0.020
	MM			-0.11 ± 0.026	-0.255 ± 0.038	0.145 ± 0.046
TRUMPY 1955	DD			-0.021 ± 0.016	-0.067 ± 0.019	0.046 ± 0.025

TABLE 3.4,A.

DD = Day-to-Day

MM = Month-to-Month

To obtain a significant value for the positive temperature term we combined all the five values shown in Table 3.4.A., weighting them by $1/\sigma^2$ (Brunt 1933), as in section 3.2. This yielded a value

$$T_P = (0.052 \pm 0.015)^\circ\text{C}.$$

Having obtained an experimental value for this term, let us re-examine the theoretical expression obtained by Dorman (section 1.4). In section 1.4 we obtained an expression

$$\frac{\delta N_{\text{soft}}}{N_{\text{soft}}} = -0.17 \frac{\%}{\text{mb.}} \delta P + 0.20 \frac{\%}{1^\circ\text{C}} \delta T_S + 0.75 \frac{\delta N_\mu}{N_\mu} + 0.1 \frac{\delta N_{\mu\text{soft}}}{\delta N_\mu \text{ soft}} \dots 1.4(i)$$

This is based on the assumption that the soft component constitutes 25% of the total radiation. An examination of our equipment showed that the bottom, alcohol quenched, tubes were only approximately 96% efficient due to dead-time errors and poor tube geometry. This means that 14% of the 'soft' component was actually hard component, unrecorded as such due to the inefficiencies of the bottom tray. Taking this into account equation 1.4.(i) becomes

$$\frac{\delta N_{\text{soft}}}{N_{\text{soft}}} = -0.08 \frac{\%}{\text{mb.}} \delta P + 0.17 \frac{\%}{^\circ\text{C}} \delta T_S + 0.82 \frac{\delta N_\mu}{N_\mu} + 0.14 \frac{\delta N_{\mu\text{soft}}}{N_{\mu\text{soft}}}$$

..... 3.4.(1)

This equation yields a value of $-0.33\%/mb.$ for the absorption coefficient, in satisfactory agreement with the experimental value $(-0.336 \pm 0.011)\%/mb.$ However the positive temperature term is a factor of approximately three greater than the value we deduced from the decay coefficients. Thus we need to re-examine the basis from which equation 3.4.(i) is derived.

The positive temperature term is based solely on the local temperature effect of the decay electrons. It is conceivable that the 'knock-on' electrons could have a local temperature effect, which would be directly proportional to the density of the air and hence act in opposition to the positive temperature effect. However, Rossi and Klapman (1942) have considered this point when computing the number of 'knock-on' electrons with energies $> 10^7$ eV generated in equilibrium with the μ -meson component at sea-level. While there is some uncertainty due to the difficulties of making suitable approximations for the various energy loss processes at energies small compared with the critical energy in air, they concluded that the density effect was negligible, due to the low average energy of the 'knock-on' electrons, which are thus quickly absorbed.

Since the 'soft' component contains about 5% shower electrons which have a negative temperature effect of $(-0.38 \pm 0.11)\%/^{\circ}C$ (Hodson 1951), this would reduce the positive

temperature term to approximately $0.15\%/^{\circ}\text{C}$. However, there seems no further reason for reducing its magnitude, so we must assume some error in the reasoning used to obtain equation 3.4.(i).

Summary

A regression analysis of the atmospheric temperature effects associated with the intensity of the soft component shows that there are two main effects: a positive surface temperature effect associated with the formation of the decay electrons, and a negative, decay, effect associated with the μ -mesons which are in equilibrium with the electronic component. The resultant effect depends upon the way in which the atmospheric temperature varies with height. This variation with height is a function of the period of time over which it is considered. Thus the results of a day-to-day regression analysis differ from those of a month-to-month analysis. An examination of a theoretical expression for the atmospheric effects of the soft component put forward by Dorman (1956) shows that it predicts the right absorption coefficient but that its positive temperature^{term}/is too large by a factor of three.

CHAPTER IV - THE DAILY VARIATION

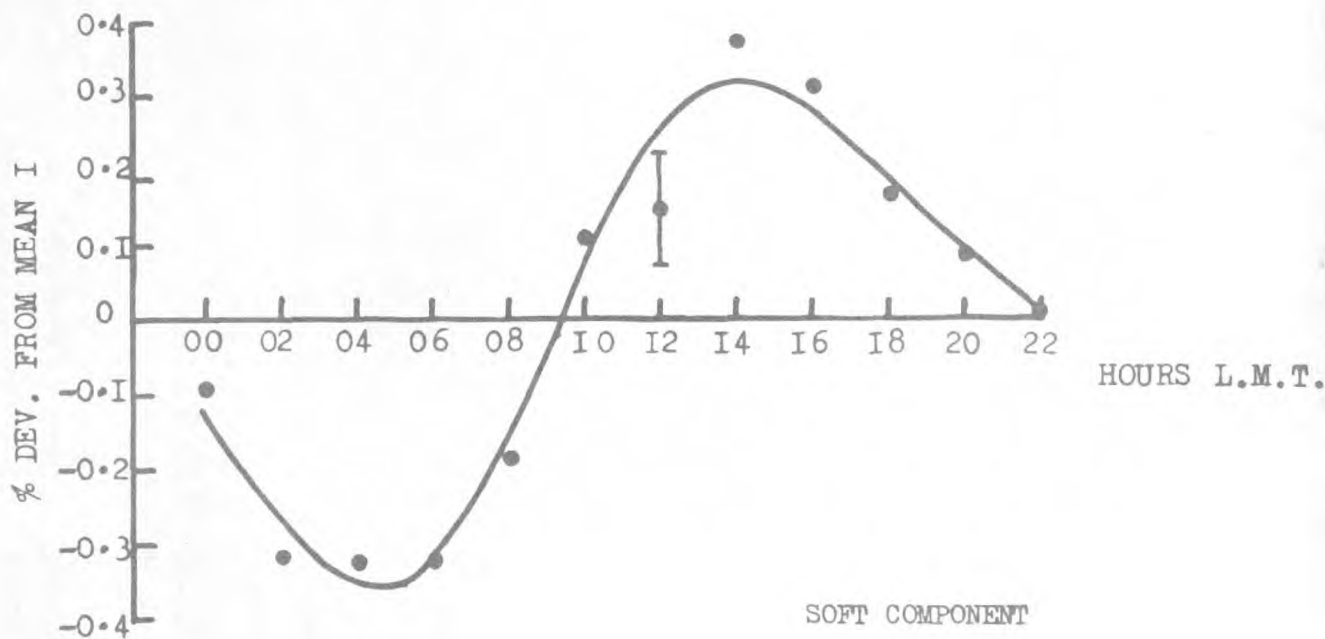
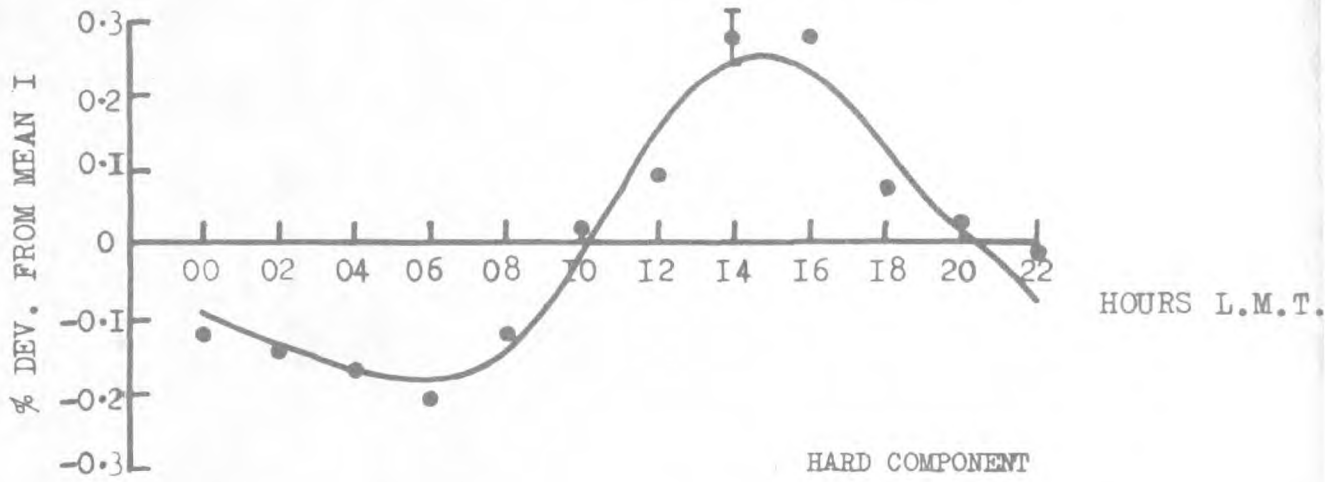
4.1. Introduction

In Chapter I a brief history of the daily variation of the cosmic ray intensity was given. In this Chapter we shall present the results of the daily variation of the hard and soft components, and discuss whether it is possible to use these results to deduce an atmospheric temperature correction vector.

Before presenting the experimental results let us consider the method of analysis employed on the daily variation. This variation is normally roughly sinusoidal in shape and when subject to Fourier analysis is found to be adequately represented by the sum of a 24-hour, and a 12-hour, wave. The amplitude of the 12-hour wave is very much smaller than that of the 24-hour wave. The amplitude and phase of the two Fourier components are normally displayed on harmonic dials.

4.2. Experimental Results

Figure 4.2.(i) shows the daily variation of the hard and soft components of the cosmic radiation, plotted as % deviations from their mean intensities. Both have been corrected for the effects of atmospheric pressure variations, using coefficients of $-0.129\%/mb.$ and $-0.336\%/mb.$ respectively. Estimated errors are indicated. We see that the two variations are very similar except that the soft is of greater



THE DAILY VARIATION

FIGURE 4.2(1)

amplitude. The thick lines indicate the sums of the 24-hour and 12-hour waves.

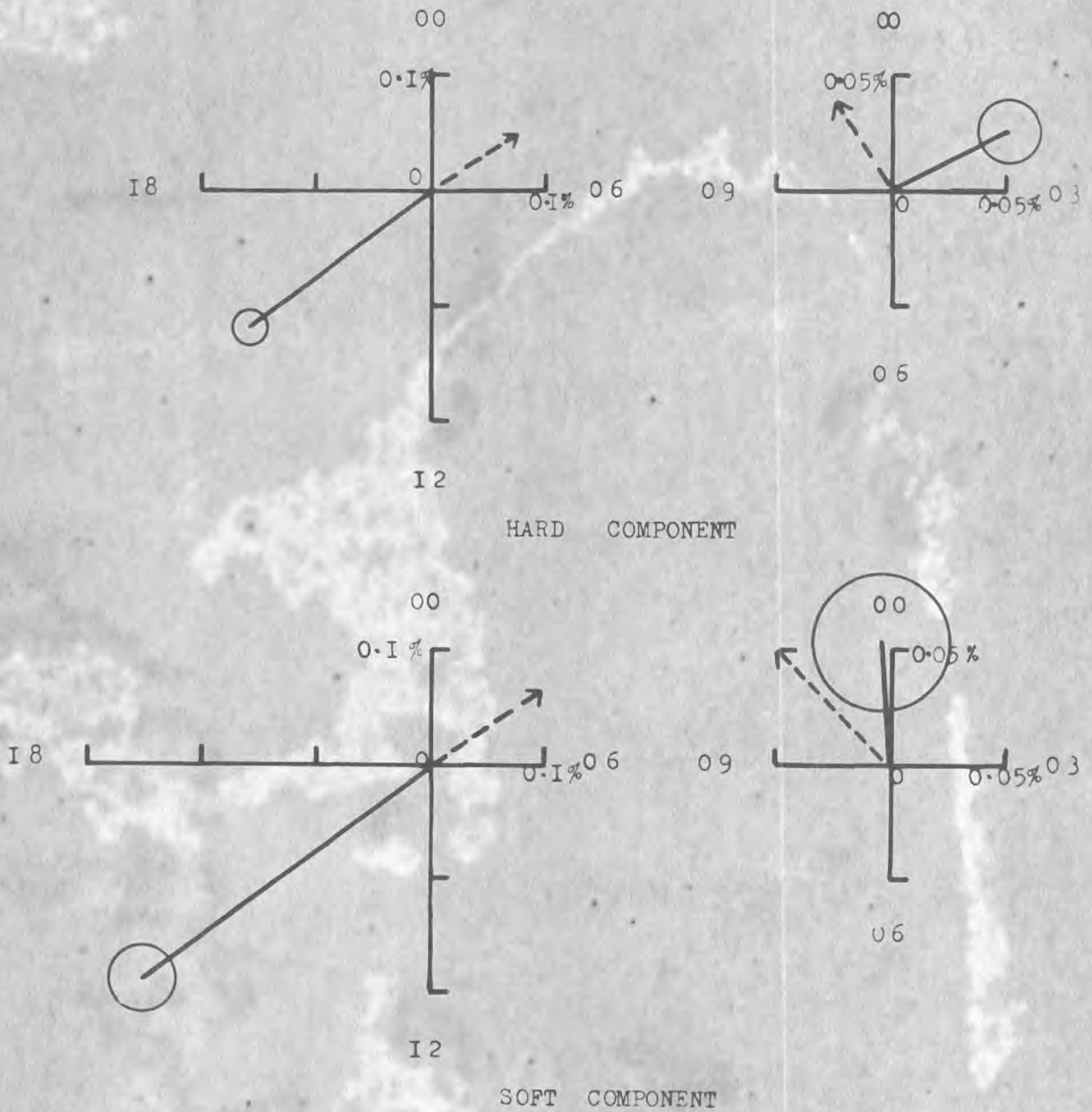
In figure 4.2.(ii) are shown the amplitudes and phases of the 24-hour and the 12-hour waves, for both components. They are represented on harmonic dials. Table 4.2.A shows the amplitudes and phases.

	<u>HARD</u>		<u>SOFT</u>	
	<u>Amplitude</u>	<u>Time of Maximum</u>	<u>Amplitude</u>	<u>Time of Maximum</u>
24-HOUR	$(0.20 \pm 0.01)\%$	1530 hrs $\pm$ 16 mins.	$(0.32 \pm 0.03)\%$	1530 hrs $\pm$ 20 mins
12-HOUR	$(0.06 \pm 0.01)\%$	0205 hrs $\pm$ 28 mins.	$(0.06 \pm 0.03)\%$	1155 hrs $\pm$ 80 mins

Table 4.2.A.

On all of the harmonic dials there are dotted lines, with arrowheads on. These indicate the time of maximum of the equivalent Fourier components of the atmospheric pressure variation.

Let us first consider the semi-diurnal i.e. the 12-hour, waves. The soft component wave is not significant, but the hard component variation certainly is. However, since the pressure effects are three hours out of phase with the hard component variation it is difficult to see how it could be due to atmospheric oscillations, unless the nodal surface, above which the oscillations are reversed in phase, is fairly wide and coincides with the μ -meson production layer. This does not appear so. Wilkes (1949) indicates the nodal surface at a height of about 30 kms., while the height of the main μ -meson



DIURNAL AND SEMI-DIURNAL
VARIATIONS

FIGURE 4.2(ii)

production layer is about 16 kms.

Considering the diurnal waves of the two components we see that the ratio of the amplitude of the soft component variation to that of the hard is as $(1.54 \pm 0.11):1$ and that the times of maxima are the same. This is in agreement with the results of Trefall (1960).

4.3. The Atmospheric Temperature Effects of the Soft Component Diurnal Variation

While the discussion of section 3.4 seems satisfactory in explaining the meteorological effects of the soft component, it raises difficulties in dealing with the soft component diurnal variations since we now need to know the relationship between the diurnal variation in the height of the 100 mb. layer and the surface temperature diurnal variation.

Due to radiation errors affecting the temperature sensor of the radio-sonde, there is considerable difficulty in deciding the diurnal variation in the height of the 100 mb. layer. However, Quenby and Thambyahpillai (1960) have determined a net correction vector for the μ -meson effects of this variation, which is a vector $(0.11 \pm 0.02)\%$ in amplitude, acting at 0530 hours L.M.T.

The diurnal variation of the surface temperature has an amplitude close to 3°C , occurring at about 1430 hours L.M.T. Using the experimentally determined positive temperature term

$(0.052 \pm 0.015)\%/^{\circ}\text{C}$, this corresponds to a vector of amplitude $(0.156 \pm 0.045)\%$ acting at 1430 hours L.M.T. If we combine this and the Quenby-Thambyahpillai vector the resultant will give the atmospheric temperature correction vector to the soft component diurnal variation. This is a vector of amplitude $(0.116 \pm 0.049)\%$ acting at 1130 hours L.M.T.

We are now able to correct the diurnal variation of the soft component for temperature effects. Figure 4.3.(i) shows the temperature correction obtained above, while figure 4.3.(ii) shows the hard component corrected using the Quenby-Thambyahpillai temperature vector, and the soft component corrected using the vector shown in (i). We see that there is no significant difference in phase or amplitude between the two components. This seems reasonable since appendix I indicates that one might expect the same rigidity dependence for the soft and hard components up to fairly high rigidities i.e. ~ 200 GV in the case of Forbush decreases.

As a further check on the soft component temperature correction vector, we can utilise the diurnal variation results of Dawton and Elliot (1953). Figure 4.3.(iii) shows the diurnal variations obtained and figure 4.3.(iv) shows the hard and soft components corrected as in figure 4.3.(ii). Again, within the limit of the rather large errors, there is no significant difference between the diurnal variations measured

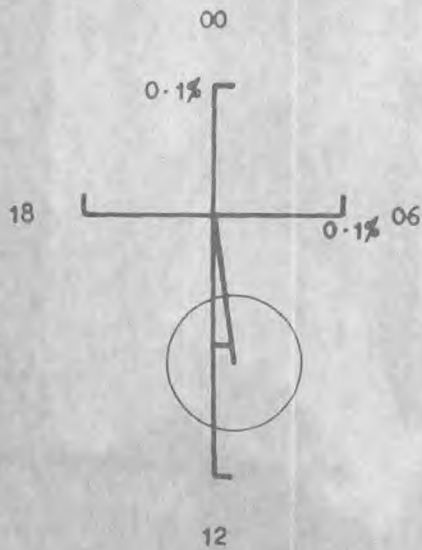


FIGURE 4.3(i)

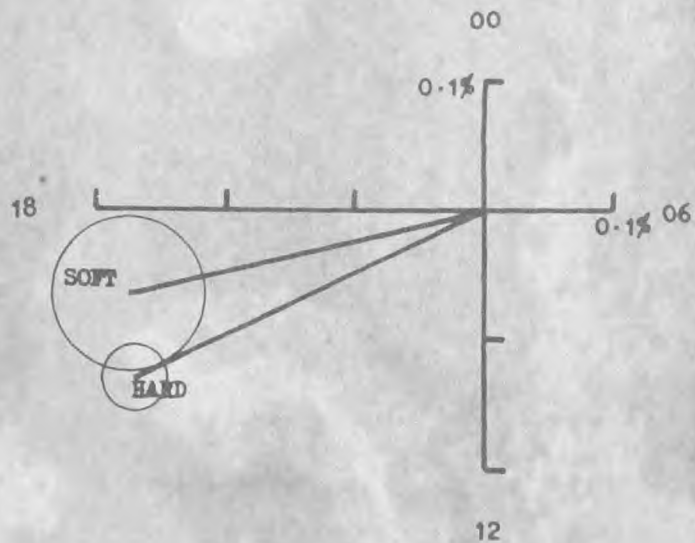


FIGURE 4.3(ii)

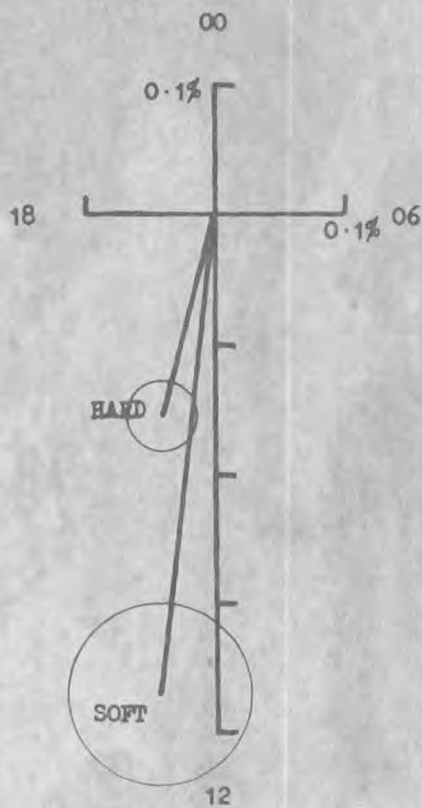


FIGURE 4.3(iii)

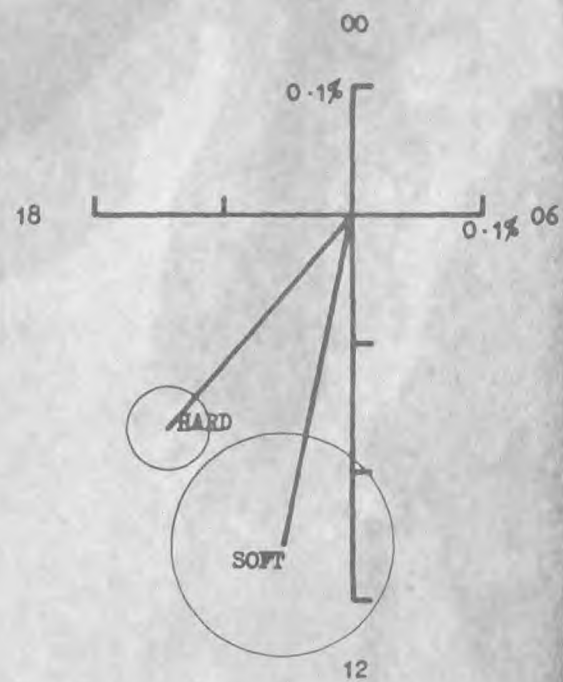


FIGURE 4.3(iv)

DIURNAL AND ATMOSPHERIC TEMPERATURE
CORRECTED DIURNAL VARIATIONS

by the two components.

Thus we can sum up by saying that while the soft component variation cannot be used to ~~directly~~ ^{directly} obtain an atmospheric temperature correction to the diurnal variation of the hard component it can be used as an indirect check. on the vector obtained by Quenby and Thambyahpillai (1960). The result is consistent with this vector.

4.4. Possible Latitude Dependence of an Atmospheric Temperature Correction Vector

In Chapter 6 we shall apply an atmospheric temperature correction vector to ion chamber results, from various stations, over a twenty-two year period, to investigate long term changes in the daily variation. One station will be Huancayo and the other two will be Christchurch and Cheltenham, i.e. one equatorial and two high latitude stations. The question then arises as to whether a latitude effect occurs for Quenby and Thambyahpillai's (1960) vector, which was deduced for Huancayo. This point is also relevant to the preceding section, since there we have assumed the vector to be latitude independent.

A decrease in amplitude of the vector at high latitudes might be expected since at least the diurnal variation amplitude of the surface temperature decreases with increasing latitude. However, the vector used in Chapter V, and deduced from London results at sunspot minimum indicates little difference in

amplitude or phase with the Huancayo vector, so that to a first approximation it might be assumed to be constant in value with latitude.

Glokova (1960) has employed a rather different approach on the data used by Quenby and Thambyahpillai i.e. the neutron and meson data for Huancayo at sunspot minimum. Using a sliding mean technique to eliminate seasonal variations, she obtains the monthly values of the daily variation of the meson data during sunspot minimum, and shows that they approach a minimum of about 0.11% at 0530 hours during 1954. By comparison with the nucleonic data which goes to zero at this time, she deduces that the minimum value is an atmospheric vector. She repeats the meson analysis for Christchurch and Cheltenham and obtains similar vectors, but smaller in amplitude i.e. 0.05% and 0.08% respectively. From this Glokova deduces a latitude effect.

We need to consider two points in relation to this effect. Both Christchurch and Cheltenham are at very similar latitudes, only differing in longitude. According to Duperier (1948), stations at similar latitudes show little difference in atmospheric temperature variations despite different climates. Thus one would expect similar values for their correction vectors. Also Glokova's method does not specifically select those months when the nucleonic daily variation is zero, so that there is no certainty that there is not a residual daily

variation in the vectors obtained for Christchurch and Cheltenham. Such a residual daily variation would tend to cause a reduction in the apparent amplitude of the correction vectors.

To sum up:- The evidence for and against a latitude variation of the atmospheric temperature correction vector is confused and indefinite. Hence we shall use the vector deduced by Quenby and Thambyahpillai (1960) for all stations considered in Chapter 6, discussing the effect of small variations in amplitude and phase of the vector when summing up.

CHAPTER V - THE LONDON EAST-WEST EXPERIMENT

5.1. Introduction

Using an analogue computer, Brunberg and Dattner (1953) were able to determine the initial and final directions of cosmic ray particles in a simple magnetic dipole field. This was done for a whole series of latitudes and directions of arrival on the surface of a sphere surrounding such a dipole. Since to a first approximation the magnetic field of the earth is a simple dipole field they were able to immediately apply their results to cosmic rays arriving at the surface of the earth. These results were valid in the rigidity range 10 to 700 GV and supplemented earlier calculations by Stormer (e.g. 1955) and analogue calculations by Malmfors (1945), which were done for rather lower rigidities.

From these results and the directional measurements of Malmfors (1949) and Elliot and Dolbear (1951). Brunberg and Dattner (1954) were able to show that the main portion of the diurnal variation rigidity spectrum was probably contained in the range 20 to 40 GV.

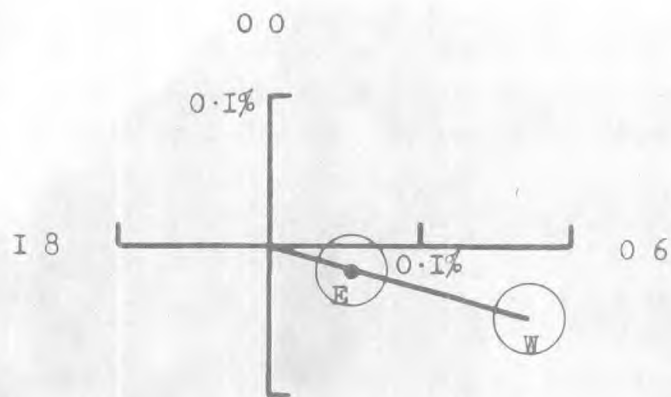
For latitude 50° , an examination of Brunberg and Dattner's curves in the range 20 to 40 GV shows that a telescope pointing at 45° to the vertical in a westerly direction will receive radiation originating from a direction roughly parallel to the earth's magnetic axis. A similarly situated telescope, but pointing at 45° to the vertical in an easterly direction, will receive radiation in this rigidity range originating from a direction

roughly in the equatorial plane. If we assume an anisotropy occurring somewhere near the equatorial plane, the westerly pointing telescope which receives radiation in the 20 to 40 GV range from roughly the same direction in space all day, should record only daily atmospheric variations, while the east pointing recorder should record similar daily atmospheric variations plus any cosmic ray daily variation, since the recorder scans a strip of the celestial sphere every 24 hours.

Elliot and Rothwell (1956) performed such an experiment during the period May 1954 to April 1955. The results of this experiment are summarized by figure 5.1 (1), taken from Elliot and Rothwell's paper. The figure shows the 24-hour and 12-hour components of the daily variation recorded by each telescope. The 12-hour components are not significant.

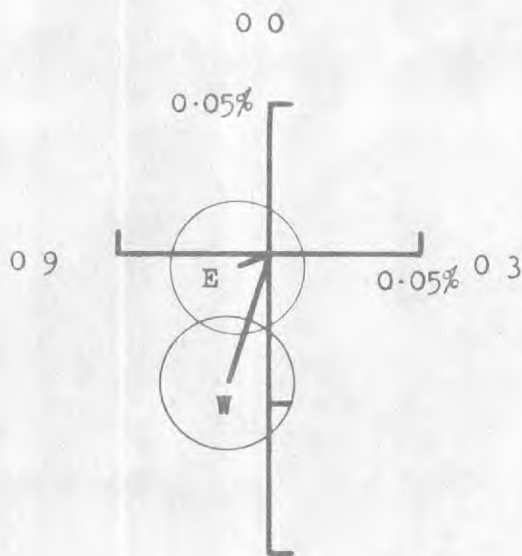
During the period of this experiment Possner and Van Heerden (1956) had observed apparent large changes in the phase of the vertically measured daily variation, these changes occurring over relatively short periods of time. The effect of such changes was to cause the yearly average of the daily variation to have an unusually small amplitude.

From figure 5.1 (1) we see that the 24-hour component recorded by the east-pointing recorder was considerably smaller than that registered by the west-pointing recorder. To see whether this was due to short term changes in time of maximum,



1.2

24-HOUR WAVES



0.6

12-HOUR WAVES

DIURNAL AND SEMI-DIURNAL VARIATIONS

OBSERVED IN LONDON 'EAST-WEST' EXPERIMENT

FIGURE 5.1 (i)

the monthly data for both telescopes was analysed. This analysis showed the variations to be consistent from month to month, and not to contain wide fluctuations in the time of maximum.

The results for the 24-hour waves shown in figure 5.1 (i) are difficult to interpret. From the initial hypothesis we should expect the west-pointing recorder to register the smallest amplitude diurnal variation and the east-pointing recorder to register the largest. The reverse was observed.

For the rigidities initially specified, 20 to 40 GV, Brunberg and Dattner's curves also indicate that the phase of the diurnal variation recorded by the east-pointing telescope should be one to two hours earlier than that recorded by the west-pointing recorder. This is consistent with the errors indicated.

A possible explanation of the rather small amplitude of the diurnal variation recorded by the east-pointing telescope is that the atmospheric temperature effects were closely in anti-phase with the diurnal variation. If this were so however, then under the initial assumptions, the westerly pointing diurnal variation should be in anti-phase to the time of maximum actually recorded for it. At the time of the experiment (1955), the only experimental data on the atmospheric temperature correction to the diurnal variation was the soft-component experiment of Dawton and Elliot (1953) (see Chapter IV). This indicated a correction vector of about 0.2% with a maximum close to mid-

night. Such a correction makes the amplitudes more nearly equal, but tends to make the phase of the easterly-recorded diurnal variation later than the westerly-recorded diurnal variation, although not significantly so. The basic anomaly is not removed by the correction.

It might be possible to account for the observed amplitude by decreasing the mean rigidity of the anisotropy, or by postulating a direction of anisotropy at an angle of 70° to the equatorial plane. Both of these suggestions are untenable when considered against the diurnal variation measured vertically at the equator and at very high latitudes.

In chapter I we discussed experiments that indicated the daily variation to be a true primary cosmic ray effect and not a purely atmospheric effect. Because of this Elliot and Rothwell (1956) suggested that the E-W results they had obtained might be explicable in terms of a local modulation mechanism, occurring close to the earth, in its magnetic field.

Brunberg and Dattner's curves cannot be used to investigate this idea, because their analogue computer did not give the details of the trajectories in the magnetic field but only the direction of the particle before it entered the field, and its final direction when it arrived at the surface of the sphere surrounding the dipole. For particles in the rigidity range 20 to 40 GV the earth's effective magnetic field stretches

out to about five earth radii, and it is in this region we require to investigate the trajectories of the cosmic ray particles.

We shall consider the possibilities of the 'local modulation' hypothesis in the next section.

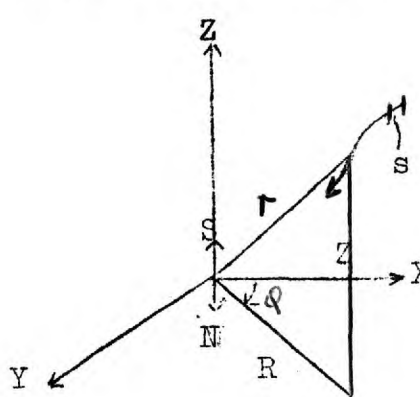
5.2. Investigation of the 'Local Modulation' Hypothesis

In order to investigate the trajectories of cosmic ray particles in the earth's magnetic field, it is necessary to integrate the equations of motion of the particles. Since this is rather laborious it was decided to do a pilot calculation on the trajectories of 20 GV, 40 GV and 60 GV particles arriving at 45° to the vertical in the east and west directions, to see whether more detailed calculations were justified.

There is a point of interest to be noted here about the method of calculation. Observers on the earth measure the final directions of arrival of cosmic ray particles and wish to know the initial directions. Fortunately the equations of motion of a charged particle in a dipole magnetic field are such that a positive particle injected into the field and arriving in a given direction is equivalent to a negative particle being ejected in the given direction at the point of arrival. Thus it is possible to integrate back along the path of the particle observed to arrive at the surface of the earth. This was how Brunberg and Dattner's analogue computer worked. They ejected electrons from the sur-

face of a magnetised sphere, called a terrella, and measured the final directions of the electrons when they were no longer influenced by the magnetic field.

The equations of motion of a charged particle in the field of a magnetic dipole were set-up many years ago (c.f. Störmer 1953). After a good deal of rearrangement, they may be written



(for a negative particle) as

$$\frac{d^2 R}{ds^2} = \left[\frac{2\gamma}{R} + \frac{R}{r^3} \right] \left[\frac{2\gamma}{R^2} + \frac{3R^5}{r^5} - \frac{1}{r^3} \right]$$

$$\frac{d^2 Z}{ds^2} = \left[\frac{2\gamma}{R} + \frac{R}{r^3} \right] \frac{3RZ}{r^5}$$

.....5.2.(1)

$$\left(\frac{dR}{ds} \right)^2 + \left(\frac{dZ}{ds} \right)^2 = 1 - \left(\frac{2\gamma}{R} + \frac{R}{r^3} \right)^2$$

$$R^2 \frac{dQ}{ds} = 2\gamma + \frac{R^2}{r^3}$$

where s = arc length of trajectory defined by $s = vt$

v = velocity of particle, t = time.

The unit of length in the equations is the so-called Störmer unit, defined by

$$C_{St} = \sqrt{\frac{Mc}{mv}} \text{ cms. where } m = \text{mass of particle}$$

e = charge of the particle

M = magnetic moment of dipole.

γ is a constant of integration introduced during the setting up of the equations, and lies in the range $-\infty < \gamma < \infty$. An

alternative method of considering it is to see that

$- 2m\mathbf{v}\gamma$ = Angular momentum (at infinity) of the particle about the Z-axis.

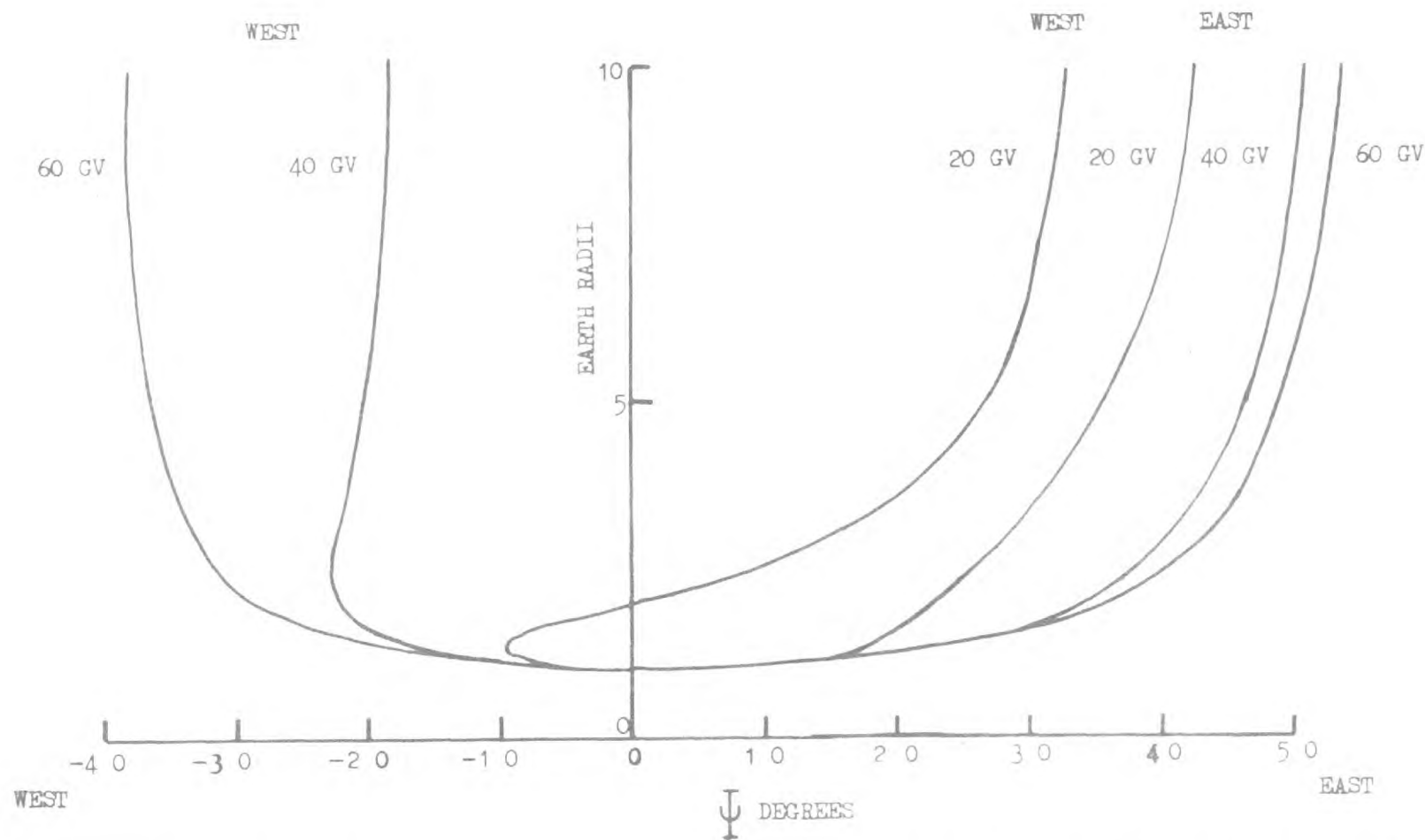
In the general case numerical methods are necessary for the solution of equations 5.2.(1). We selected the Runge-Kutta fourth order method as being the most suitable for the facilities available i.e. ²/desk electrical calculating machine. The main advantages of this method are that no special procedure is required to start the solution, and the interval of integration may be altered at any stage of the computation without disrupting the entire set-up. The price to be paid for these advantages is that each integration step is split into four parts, which with a complex equation makes substitution back rather involved, and also it is not possible to ~~directly~~ compute ^{directly} the error inherent in the calculation. The first point was not important in this case, since the equations are comparatively straightforward. Also we were able to get round the problem of the errors. All that we required of our trajectories was that they should not be more than one to two degrees in error in the 'directions at infinity' i.e. their directions outside the influence of the magnetic field. Since Brunberg and Dattner's curves have been satisfactorily checked against the very accurate calculations of Störmer (e.g. 1955), which were of known error, all we needed to do was check that the ends of our trajectories fitted smoothly to the curves of Brunberg and Dattner. This ensured that no serious error

had been propagated.

Initially it was necessary to choose an integration interval small enough to account for the detailed orbit but sufficiently large to avoid unnecessary computational labour. Later as the trajectory became less curved we introduced a wider integration interval to take advantage of this.

One of the inherent problems of numerical solutions is that if a computational error is made, then due to the nature of the process involved, it very rapidly propagates. Unfortunately it is often difficult to detect such errors by visual inspection of the results, so that the error may not be detected until a great deal of re-computation is required. To guard against this it is normal computational practice to build error checks into the calculation. In this case we used Milne's correction formula to check each interval.

The results of our pilot calculations are shown in figures 5.2 (i) and 5.2 (ii). They are expressed as plots of distances from the earth, in units of earth radii, against two angles Φ and Ψ . These are the angles used by Brunberg and Dattner to illustrate the final directions of the cosmic ray particles, and are illustrated in the sketch. The dotted line



'PILOT' TRAJECTORIES IN EQUATORIAL PLANE

FIGURE 5.2(1)

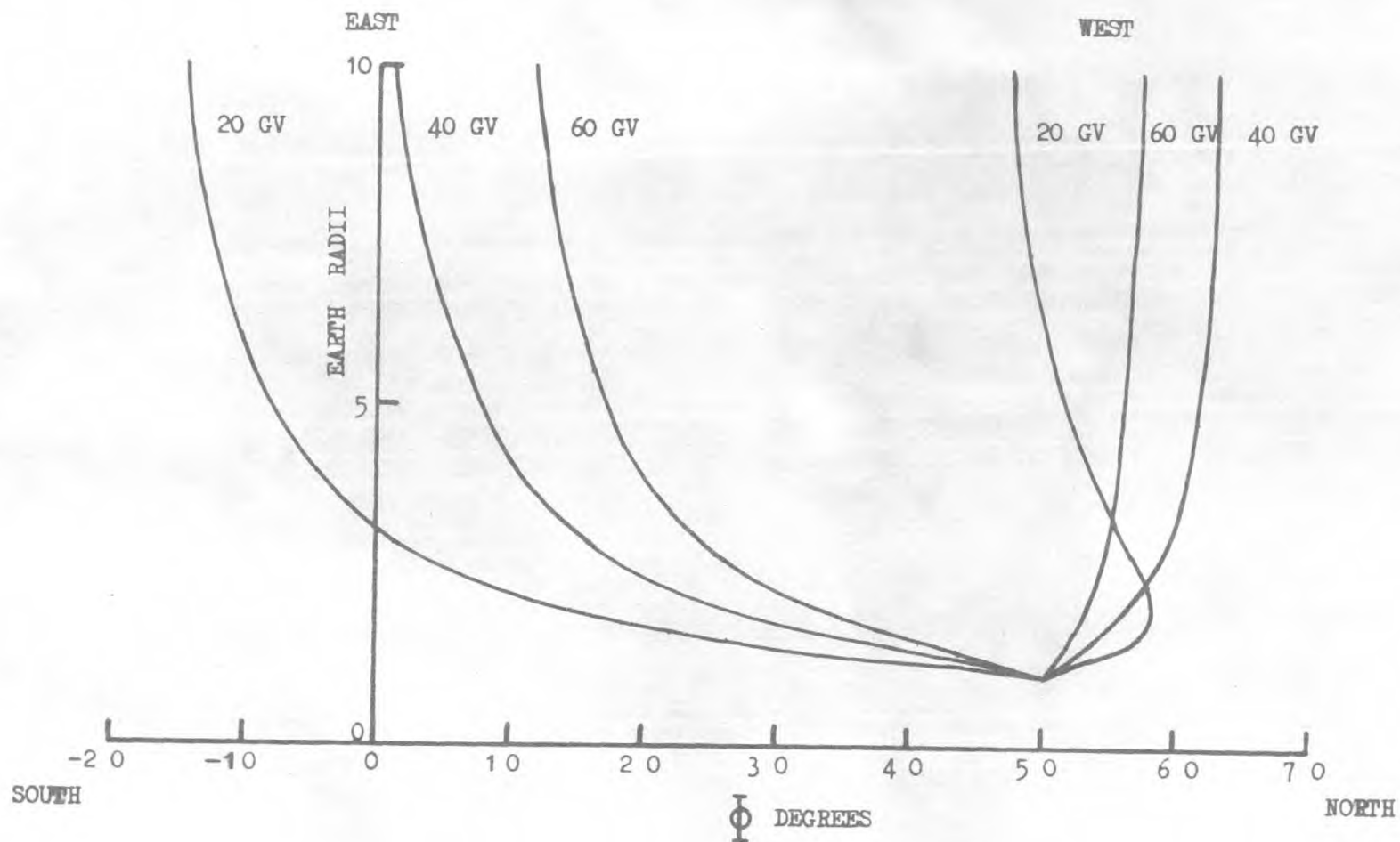
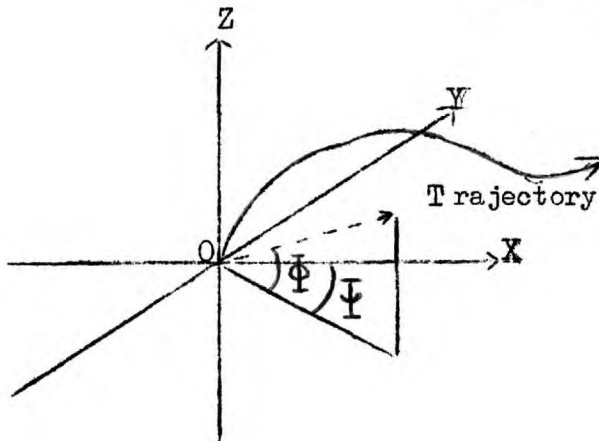


FIGURE 5.2(11)

'PILOT' TRAJECTORIES IN MERIDIANAL PLANE



line with the arrow head on is parallel

to the final direction of the trajectory, and the direction is defined by Φ and Ψ as indicated. Ψ is positive or negative according to whether it is east or west of the earth-sun

line. For each rigidity

there are two trajectories, one corresponding to particles arriving at 45° to the vertical from the west, and at a similar angle from the east.

An examination of figures 5.2. (i) and (ii) does not indicate any solution other than that already suggested by Elliot and Rothwell i.e. that below 20 GV the Φ values are tending to converge, and there is little difference between the Ψ values. Thus if we postulate an anisotropy at about 20° to the equatorial plane with a mean rigidity of rather less than 20 GV, then it seems possible that a daily variation of equal amplitude and equal time of maximum might be observed by two recorders operated as in the London E-W experiment. If we accept the large, and as we now know, (Chapter IV), incorrect, atmospheric temperature correction implied by Dawton and Elliot's experiment (1953),

this is the situation we get.

However the explanation runs into considerable difficulties. The main difficulty is the daily variation measured at the equator and at high latitudes. Measurements in Europe and in Tasmania and New Zealand indicate a rough symmetry about the equator for the amplitude of the daily variation. As shown in Chapter VI, the latitude effect for the 'hard' component is about 1:1.3 for the equator and stations at about 50°N . Since the vertical magnetic cut-off rigidity at the equator is approximately 15 GV, a daily variation with a mean rigidity of this order would ~~probably~~ show a much greater latitude effect. Thus the possibilities of a 'local' modulation seems unlikely to offer a solution to the anomaly observed by Elliot and Rothwell (1956).

As we have used only a simple dipole field in our calculations, there is a possibility that the non-dipole parts of the earth's magnetic field could cause serious changes in the trajectories computed. This is certainly so near geomagnetic cut-off rigidities, but the corrections are small at rigidities of 20 GV and above.

5.3. An Atmospheric Temperature Correction Explanation

The first point to be decided is the choice of atmospheric temperature correction vector. This is effectively decided for us. During the approximate period of Elliot and Rothwell's experiment, (May 1954 to April 1955), the correction vector

mentioned in the previous chapter can be deduced from measurements made at about this time. This vector, $(0.13 \pm 0.03)\%$ at (0400 ± 1) hours, was deduced from measurements that included those from two inclined telescopes, although they were inclined in a North-South direction. With this in mind we shall use the vector to correct the East-West results.

Figure 5.3. (i) shows the effects of such a correction. The east vector now has an amplitude of $(0.10 \pm 0.04)\%$ with a time of maximum (1424 ± 0136) hours, and the west vector an amplitude $(0.13 \pm 0.04)\%$, time of maximum (1008 ± 0126) hours. With the errors indicated neither amplitudes or phases can be considered significantly different. Let us consider in detail the basis of the experiment to see if we can reconcile experiment and theory

What we need to do is to obtain the effective $\bar{\Phi}$ and $\bar{\Psi}$ angles for the two recorders using Brunberg and Dattner's curves, assuming an extra-terrestrial anisotropy. We shall use Dorman's definition of the effective angles as

$$\bar{\Phi} = \frac{\int \Phi \frac{dN}{dp} E(p) dp}{\int \frac{dN}{dp} E(p) dp}; \quad \bar{\Psi} = \frac{\int \Psi \frac{dN}{dp} E(p) dp}{\int \frac{dN}{dp} E(p) dp}$$

where $E(p)$ = rigidity spectrum of the diurnal variation

$\frac{dN}{dp}$ = differential rigidity response of the detector

The integrals are taken over the appropriate rigidity range of the spectrum $E(p)$.

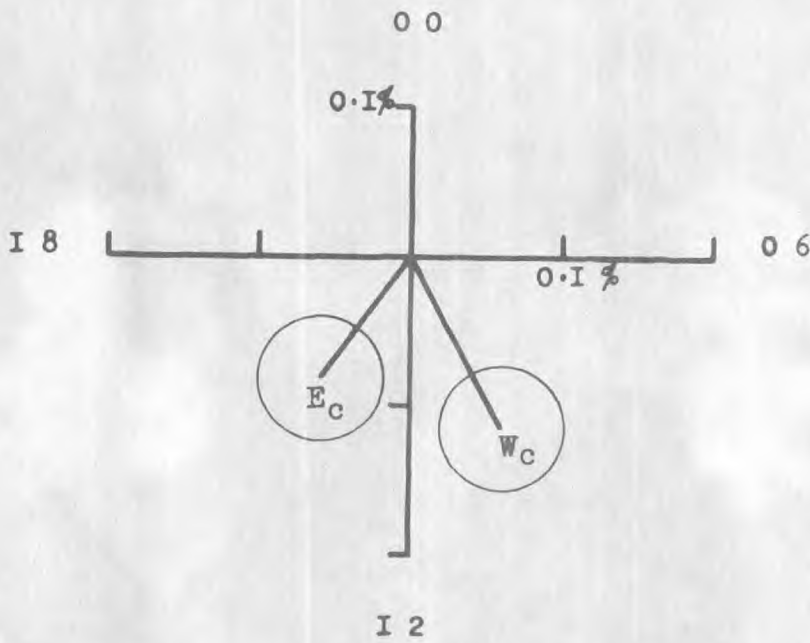


FIGURE 5.3 (1)

ATMOSPHERIC TEMPERATURE CORRECTED DIURNAL VARIATIONS

OF LONDON 'EAST-WEST' EXPERIMENT

To calculate these effective angles we need to make certain assumptions of which the most doubtful is about the rigidity spectrum of the diurnal variation. For this we shall adopt the general shape of the spectrum due to Elliot (1960) i.e. a flat spectrum over the range 15 to 50 GV, dropping off fairly rapidly at each end. This spectrum appears to be in general accord with previous diurnal variation directional and latitude measurements (e.g. Brunberg and Dattner 1954). We shall also assume that we can use the sea-level μ -meson detector differential response curve obtained by Webber and Quenby. This was deduced from vertical measurements but there seems no obvious reason why it should vary seriously for the directional measurements done in Elliot and Rothwell's experiment.

Taking into account the opening angles of the telescopes and the difference between the geographic and the simple geomagnetic dipole East-West plane, we computed the two asymptotic angles to be expected for each recorder. The results are shown in Table 5.3A. The angles have been computed for two rigidity ranges, 10 to 30 GV and 10 to 50 GV. We have assumed that at 10 GV the amplitude of the diurnal variation has fallen to one half of its value in the range 15 to 50 GV. The errors quoted are estimated from the errors given by Brunberg and Dattner for their curves and from the deviations from the curves due to the non-dipole terms in the earth's magnetic field, estimated by Quenby and

Wenk (1961). These deviations are 4° to 5° at this latitude and longitude.

p GV	WEST	EAST
10 $\rightarrow$ 30	$\bar{\Phi} = 23 \pm 5^\circ$	$\bar{\Phi} = -6 \pm 10^\circ$
	$\bar{\Psi} = 41 \pm 7^\circ$	$\bar{\Psi} = 41 \pm 6^\circ$
10 $\rightarrow$ 50	$\bar{\Phi} = 32 \pm 5^\circ$	$\bar{\Phi} = -5 \pm 8^\circ$
	$\bar{\Psi} = 32 \pm 6^\circ$	$\bar{\Psi} = 42 \pm 6^\circ$

TABLE 5.3.A

An examination of Table 5.3.A shows that varying the high rigidity end of the diurnal spectrum has little effect on the values of $\bar{\Phi}$ and $\bar{\Psi}$. The most interesting point about this table is that while the east-pointing recorder certainly sees primary particles coming largely from the equatorial plane, the west-pointing recorder sees such particles coming from an effective direction about 25° to 35° to this plane, an angle rather less than a normal inspection of Brunberg and Dattner's curves suggests.

Two rigidity ranges have been used since there is some evidence that at sunspot minima i.e. 1953-54, that the amplitude of the daily variation becomes rather small relative to its value at sunspot maximum especially at low latitude stations (see Chapter 6). This can be interpreted as a decrease in the mean rigidity of the variation from sunspot maxima to minima.

Taking the errors quoted in Table 5.3.A and the errors shown in Figure 5.3. (i) then there no longer appears to be any

anomaly since the results of table 5.3.A are consistent with diurnal variations of equal amplitude and phase for each recorder. Obviously with such large errors this cannot be considered as completely satisfactory, but it is reassuring to find that the London East-West experiment is not seriously inconsistent with our present picture of the daily variation.

CHAPTER VI - THE 22-YEAR VARIATION

6.1. Introduction

Elliot and Thambyahpillai (1953) and others (Sarabhai and Kane 1953) have shown that there is an apparent long term change in the time of maximum of the diurnal variation of the hard component. In Chapter I we have discussed the technique of measuring the daily variation with telescopes at angles inclined to the vertical, as a means of establishing the existence of a primary variation. Considerations of two such sets of measurements, made in Stockholm during 1940 and in Manchester during 1949-50 (Malmfors 1949, Elliot and Delb  ar 1951) caused Elliot and Thambyahpillai to believe that a change in the time of maximum had occurred, independent of atmospheric effects. To check this they examined the phase of the diurnal variation, as measured at the Carnegie Institute stations and others, over an extended period from 1932 to 1952. This analysis revealed several important points.

A distinct world-wide variation was found. This seemed a good indication that the variation was real and not due to atmospheric effects. The variation in the times of maxima was also found to follow an apparent twenty-two year cycle, although the data was insufficient for this point to be certain. Such a variation seemed highly significant since the true periodicity

of the sunspot cycle was known to be twenty-two years. Elliot and Thambyahpillai pointed out that while it was known that the phase of the diurnal variation was affected by geomagnetic activity (Sekido and Kodama 1952) this seemed improbable as the cause of the variation observed, since geomagnetic activity is known to follow a largely eleven-year cycle (Chapman and Bartells 1940).

Changes in the amplitude of the diurnal variation were not discussed by Elliot and Thambyahpillai, but were considered by Sarabhai and Kane (1953). The changes in the amplitudes were not as marked as those in the times of maxima, and appeared to follow something closer to an eleven year, rather than a twenty-two year, periodicity.

The main weakness of the above analysis was a lack of knowledge of the effects of atmospheric temperature variations, changes in atmospheric pressure having been corrected for. All that was known at the time was that such temperature effects were serious (Chapter I).

In the following sections we shall bring Elliot and Thambyahpillai's results up-to-date, make a correction for the effects of atmospheric temperature changes, and compare these corrected results with comparable measurements of the nucleonic component, which is essentially independent of atmospheric temperature changes (Chapter I).

For the purposes of continuity we shall use only the Carnegie Institute ion-chamber data in our analysis. The temperature correction used will be that deduced by Quenby and Thambyahpillai (1960), for reasons discussed in section 4. of Chapter IV. The Carnegie Institute stations are

	GEOMAGNETIC LATITUDE	LONGITUDE	VERTICAL CUT-OFF G.V.
HUANCAYO	-0.6°	353.8°	14.3
CHELTHENHAM	50.1 °	352.3°	2.19
CHRISTCHURCH	-48.6 °	45.2°	2.53

6.2. Results of Atmospheric Temperature Correction

For ease of expression and handling we shall initially divide these results into two sub-sections, 'The Phase and 'The Amplitude'.

The Phase of the Diurnal Variation

Figure 6.2.(i)a shows a plot of the atmospheric temperature corrected time of maximum of the diurnal variation, against time in years. Results are shown for each of the three stations. Due to the latitude effect 2.3 hours have been added to each of the Huancayo values to make them directly comparable to the values for Cheltenham and Christchurch. This compares with the 3.3 hour addition used by Elliot and Thambyahpillai for the uncorrected results.

There are several points about the data that require

• = HUANCAYO
 ○ = CHELTENHAM
 x = CHRISTCHURCH

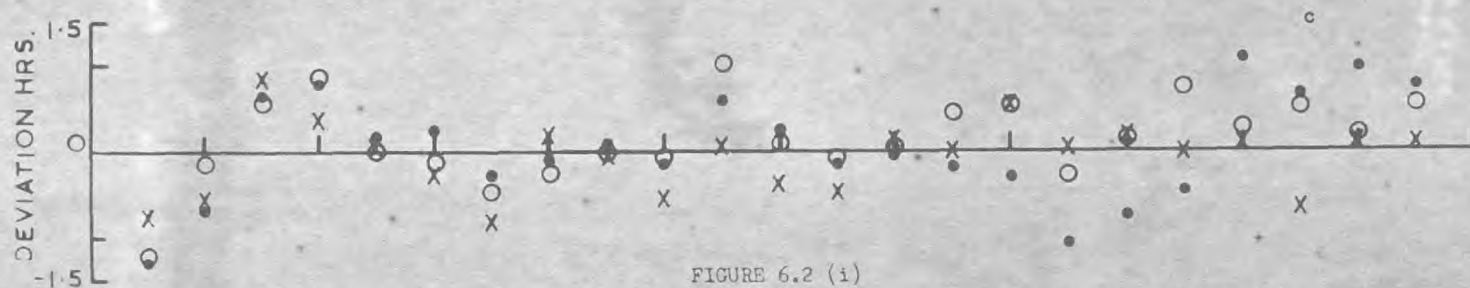
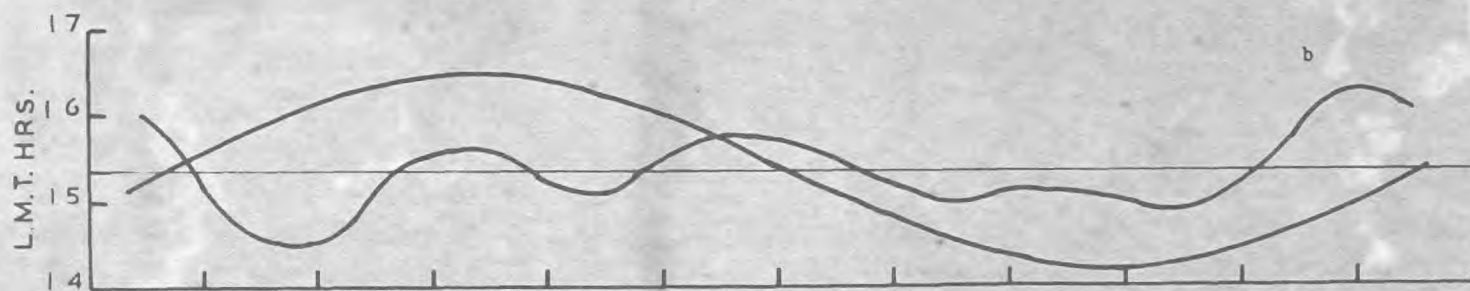
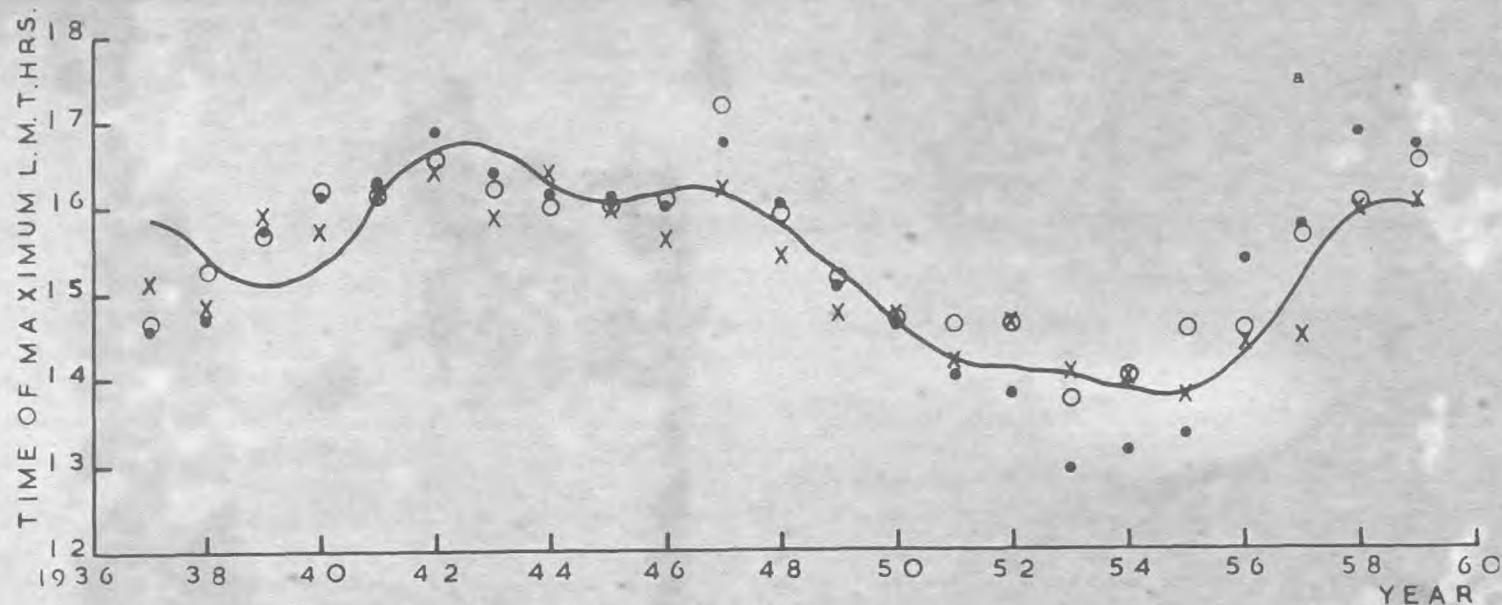


FIGURE 6.2 (1)

LONG TERM CHANGES IN PHASE OF DIURNAL VARIATION

explanation. Quenby and Thambyahpillai (1960) showed that during 1953 certain of the months at Huancayo had zero amplitude when corrected for the effects of atmospheric temperature. These months have been excluded from the corresponding Cheltenham and Christchurch data. They also showed that the corrected amplitude was effectively zero throughout 1954 at Huancayo, so we have interpolated a phase value for the station. Data for all stations during the period 1956-1959 are taken from the paper by Forbush and Venkatesan (1960). From this paper we have also taken the re-pressure corrected values for Huancayo from 1946 onwards.

For any Fourier analyses we have used the period 1957-1958, but 1959 values have been included to see how well they agree with the predicted values. No attempt at a secular correction to the data was made before analysis, since we could only have applied a linear correction, and this is rather questionable.

An inspection of figure 6.2.(i)a shows a close similarity in the changes observed at each of the stations. To put this similarity on a quantitative basis we correlated the changes occurring at pairs of stations. The results of the correlation are shown in table 6.2.A.

CORRELATION COEFFICIENT		REGRESSION COEFFICIENTS	
$r_{H.CH}$	0.93	$\beta_{H.CH}$	0.69 ± 0.09
		$\beta_{CH.H}$	1.2 ± 0.07
		$\beta_{H.CR}$	0.63 ± 0.12
$r_{H.CR}$	0.88	$\beta_{CR.H}$	1.2 ± 0.09
$r_{CH.CR}$	0.89	$\beta_{CH.CR}$	0.85 ± 0.10
		$\beta_{CR.CH}$	0.92 ± 0.10

TABLE 6.2.A

The 0.001 significance level (Fisher 1946) for a set of twenty observations is 0.68. The values of the correlation coefficients shown in Table 6.2.A thus establish the significance of the changes observed, and since we are dealing with three independent and widely separated stations, imply that the variation is genuine primary cosmic ray effect.

It is interesting to note that Forbush and Venkatesan (1960) using a more elaborate method of analysis on the non-atmospheric temperature corrected case, for which the equivalent correlation coefficients are not significantly different from those above, have shown that the correlation between stations is probably as close to 1:1 as the 'noise-level' of the ion-chambers permit.

We have also shown in Table 6.2.A the equivalent regression coefficients with their standard errors. The internal agreement between pairs of coefficients for Huancayo is not good, but initially we shall make our Fourier analysis on the mean values of the three sets of data and consider individual stations later.

In our preliminary Fourier analysis we found the first six harmonic components of the mean values of the data shown in figure 6.2.(i) (a). These harmonic coefficients are shown in Table 6.2.B.

HARMONIC COEFFICIENTS	AMPLITUDE	TIME OF FIRST MAX.	PERIOD YEARS
$a_1 = -1.2$ $b_1 = 17.0$	17.0	1942.7	22
$a_2 = 4.2$ $b_2 = -3.0$	5.1	1946.9	11
$a_3 = 1.1$ $b_3 = -3.8$	4.1	1942.8	7.3
$a_4 = 4.4$ $b_4 = -2.6$	5.1	1942.0	5.5
$a_5 = -0.4$ $b_5 = -2.2$	2.2	1940.2	4.4
$a_6 = 2.6$ $b_6 = -0.3$	2.6	1940.6	3.7

TABLE 6.2.B

The units of the harmonic coefficients are degrees of longitude, each degree being equivalent to four minutes of time.

To decide the significance of the above coefficients we need to assess their errors. Since we are dealing with the phase of the diurnal variation, we need to ~~first~~ ^{first} determine the

mean error of the amplitude of the diurnal variation at a given station, then use this mean error and the mean value of the amplitude to determine the mean phase error.

In order to clarify the rather complicated procedure we enumerate the following steps. Starting with the raw cosmic ray data we obtain the error of the bihourly values. A Fourier analysis of the data to obtain the diurnal variation will reduce the error by a factor of approximately two. This yields the error of the diurnal variation data. This diurnal variation is now compounded with the temperature correction vector, with its associated error. A combination of these two gives the error of the atmospheric temperature corrected diurnal variation. If we now Fourier analyse twenty-two of these values we reduce their error by a factor of approximately three. We now have the errors of the twenty-two year, Fourier analysis, coefficients for individual stations. Since the values in figure 6.2.(1) (a) agree rather well it seems reasonable to take the error of the mean of the three sets of data assuming this to be $\frac{1}{\sqrt{3}}$ of the individual values. This is now the error of the harmonic coefficients in Table 6.2.B.

Following the procedure we find^a/standard deviation error of $\pm 0.023\%$ for the temperature corrected diurnal variation amplitudes. This corresponds to a phase error of $\pm 6_2^{10}$ at Cheltenham and Christchurch, and to $\pm 7_2^{10}$ at Huancayo. A

twenty-two point Fourier analysis reduces these errors to $\pm 0.007\%$ for the amplitude, and in the phase to $\pm 2^\circ$ at Cheltenham and Christchurch and to $\pm 2\frac{1}{2}^\circ$ at Huancayo. For the error of the mean values of the phases of the three stations we obtain a mean standard deviation of

$$\sigma_{\text{phase}} = \pm 1.2^\circ$$

This is very much a minimum error, so we have set our significance level for the harmonic coefficients in Table 6.2.B at 3σ i.e. 3.6° .

As a check on the complicated procedure used above in determining σ_{phase} , we found the standard deviation of the values for each station about the mean value. This gave a value of

$$\sigma = \pm 1.4^\circ$$

in reasonable agreement with the value above, and is an indication that we have not grossly underestimated the errors involved.

Returning to Table 6.2.B, we see that the first four harmonics are significant by our criteria. It is difficult to give any physical significance to the third and fourth harmonics and they may just indicate that the phase and amplitude of the eleven year cycle vary from solar cycle to solar cycle.

In figure 6.2.(i) (b) we have plotted the twenty-two year wave, and the sum of the other three harmonics. These help to give a physical picture of the situation. We have an

apparent twenty-two year wave with minima at times of sunspot maxima, combined with a small and erratic eleven year wave that approximately follows the eleven year sunspot cycle.

The thick, full line in figure 6.2.(i) (a) is the sum of the two curves drawn in (b) and is a reasonably good fit. This is emphasised in 6.2.(i) (c) where the deviations from the fitted curve are plotted. Since 3σ for each of the individual points is approximately 1.3 hours, none of the deviations are significant.

As we pointed out, the internal agreement of the pairs of regression coefficients shown in Table 6.2.A was not good for Huancayo. To investigate this we Fourier analysed each set of data for the three stations. The results are shown in Table 6.2.C.

For each station the significance level is indicated. Thus we see that each station has a significant twenty-two year wave present, but only Huancayo has any significant higher harmonics i.e. the second and third.

Figure 6.2.(ii) shows the twenty-two and eleven year waves on harmonic dials, both for the individual stations and the mean values. We see emphasised the point that only Huancayo has a significant eleven-year wave, and then only just so. We shall consider this point later.

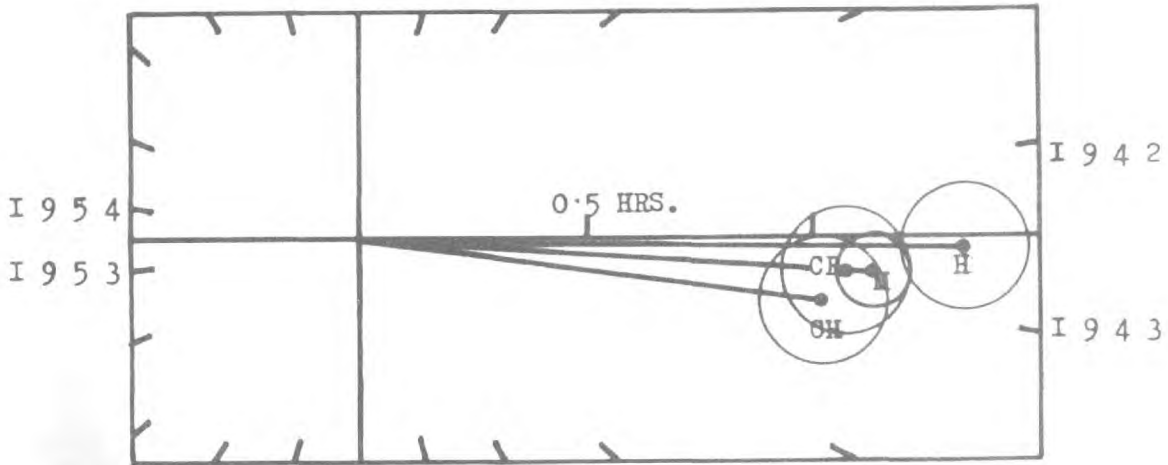
Summing up; when corrected for the effects of atmosphere temperature variation, the time of maximum of the diurnal

HARMONIC COMPONENTS OF THE PHASE VARIATION

HUANCAYO $3\sigma = 6.9$			CHELTENHAM $3\sigma = 5.7$			CHRISTCHURCH $3\sigma = 5.7$		
HARMONIC COEFFICIENTS	AMPLI- TUDE	TIME OF FIRST MAX.	HARMONIC COEFFICIENTS	AMPLI- TUDE	TIME OF FIRST MAX.	HARMONIC COEFFICIENTS	AMPLI- TUDE	TIME OF FIRST MAX.
$a_1 = -0.2$ $b_1 = 19.9$	19.9	1942.5	$a_1 = -2.1$ $b_1 = 15.2$	15.3	1943.0	$a_1 = -1.1$ $b_1 = 15.9$	15.9	1942.7
$a_2 = 6.9$ $b_2 = -5.2$	8.6	1946.9	$a_2 = 3.9$ $b_2 = -3.1$	5.0	1946.8	$a_2 = 1.7$ $b_2 = -0.9$	1.9	1947.2
$a_3 = 3.2$ $b_3 = -7.9$	8.5	1942.9	$a_3 = -0.7$ $b_3 = -3.6$	3.7	1942.3	$a_3 = 1.1$ $b_3 = -0.1$	1.1	1944.2
$a_4 = 5.1$ $b_4 = -4.4$	6.7	1941.9	$a_4 = 4.5$ $b_4 = -2.6$	5.2	1942.0	$a_4 = 3.8$ $b_4 = -0.8$	3.9	1942.3
$a_5 = -0.5$ $b_5 = -3.5$	3.9	1940.2	$a_5 = -1.7$ $b_5 = -1.8$	2.5	1940.9	$a_5 = 0.9$ $b_5 = -1.4$	1.7	1940.7
$a_6 = 2.6$ $b_6 = -1.5$	3.0	1940.4	$a_6 = 3.4$ $b_6 = -0.3$	3.4	1940.7	$a_6 = 2.0$ $b_6 = 0.8$	2.2	1940.9

TABLE 6.2 C

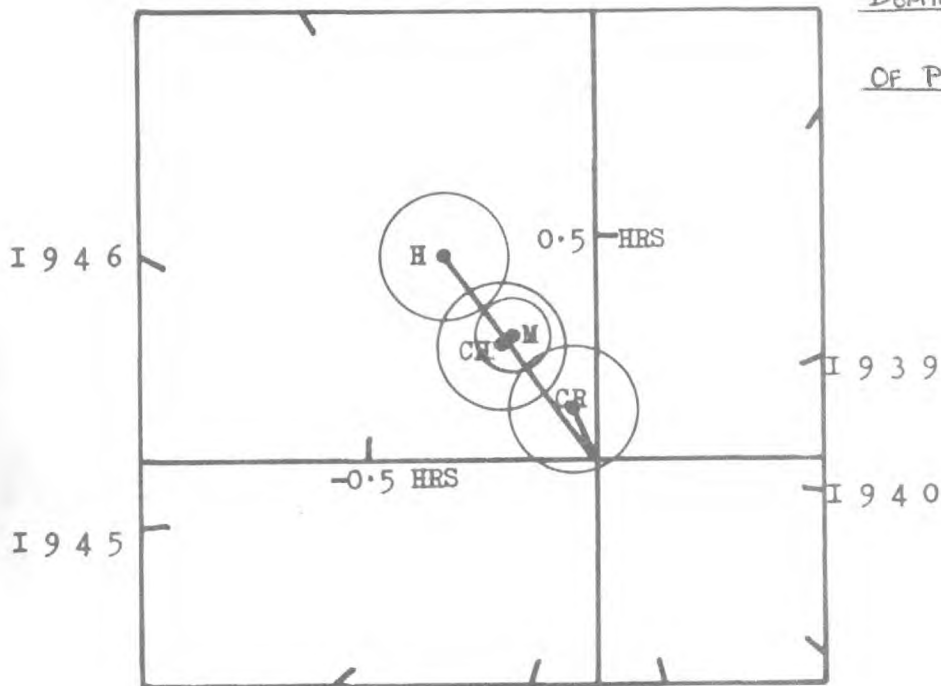
1937



1948

22 - YEAR PERIOD

1937



DOMINANT HARMONICS

OF PHASE VARIATION

1943 1942

11 - YEAR PERIOD

FIGURE 6.2 (11)

variation still contains a significant twenty-two year variation. Higher harmonics may exist, physically they must be represented as an eleven year wave, but possibly only occurring close to the equator.

The Amplitude of the Diurnal Variation

The entire analysis used above was repeated on the long term changes observed in the amplitude of the atmospheric temperature corrected diurnal variation. In Table 6.2.D. are shown the correlation and regression coefficients of the changes between pairs of stations.

CORRELATION COEFFICIENT		REGRESSION COEFFICIENTS	
$r_{H.CH}$	0.70	$\beta_{H.CH}$ $\beta_{CH.H}$	0.83 ± 0.16 0.84 ± 0.16
$r_{H.CR}$	0.74	$\beta_{H.CR}$ $\beta_{CR.H}$	1.14 ± 0.15 1.68 ± 0.15
$r_{CH.CR}$	0.78	$\beta_{CH.CR}$ $\beta_{CR.CH}$	1.01 ± 0.14 0.60 ± 0.14

TABLE 6.2.D.

The correlation coefficients are not as high as in the case of the phase, but all exceed the 0.001 significance level, establishing the significance of the changes observed and again implying a genuine primary cosmic ray effect. Internal agreement

between the pairs of regression coefficients is better than for those obtained in the phase analysis.

Figure 6.2.(iii) shows the data to which the above correlation analysis was applied. a shows the three sets of values with the mean fitted curve, b the dominant harmonics i.e. the first harmonic and the sum of the significant higher harmonics, and c the deviations of the data in a about the mean fitted curve. There is one experimental point to be noted. After the temperature vector was applied, the mean amplitude at Huancayo was less than that at Cheltenham and Christchurch, i.e. 0.17% to 0.205%. To enable us to use a common fitted curve we have added 0.035% to each of the Huancayo values. Since we are only considering the long term changes, this is quite valid.

In Table 6.2.E. are shown the harmonic coefficients of the first six harmonics of the mean of the amplitude data of the three stations.

<u>HARMONIC COEFFICIENTS</u>	<u>TABLE 6.2.E</u> <u>AMPLITUDE %</u>	<u>TIME OF FIRST MAX.</u>
$a_1 = 0.001$ $b_1 = 0.040$	0.040	1942.4
$a_2 = 0.028$ $b_2 = 0.009$	0.029	1937.5
$a_3 = 0.006$ $b_3 = -0.019$	0.020	1942.9
$a_4 = 0.014$ $b_4 = -0.023$	0.027	1941.6
$a_5 = -0.015$ $b_5 = -0.011$	0.019	1939.6
$a_6 = 0.005$ $b_6 = -0.009$	0.010	1940.0

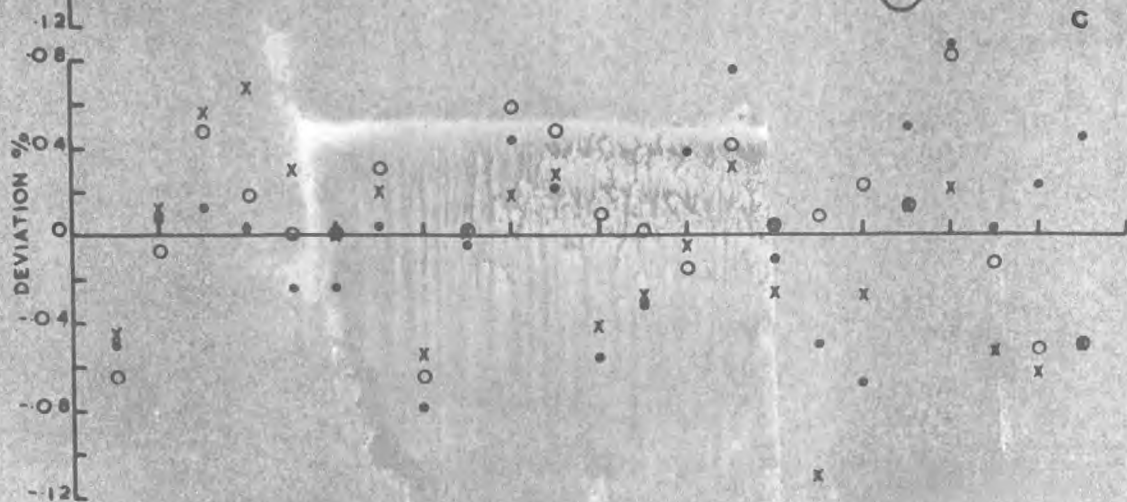
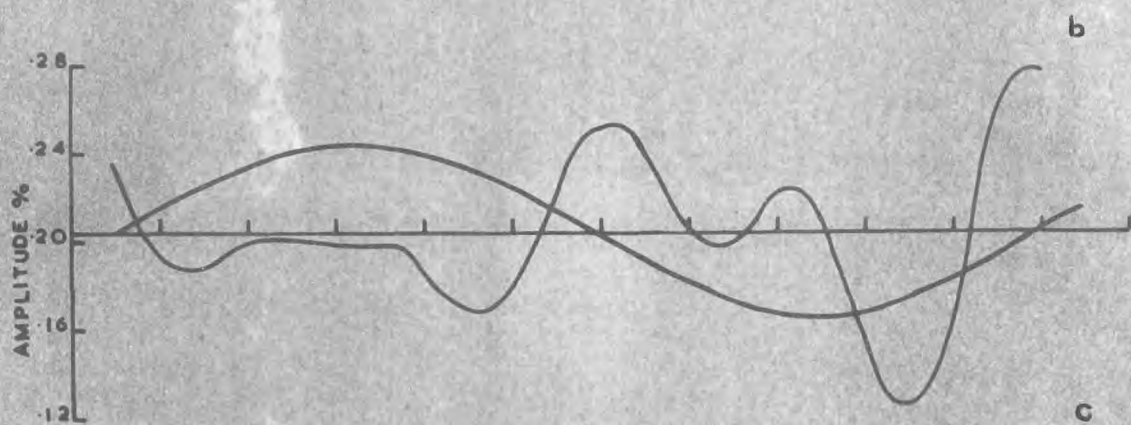
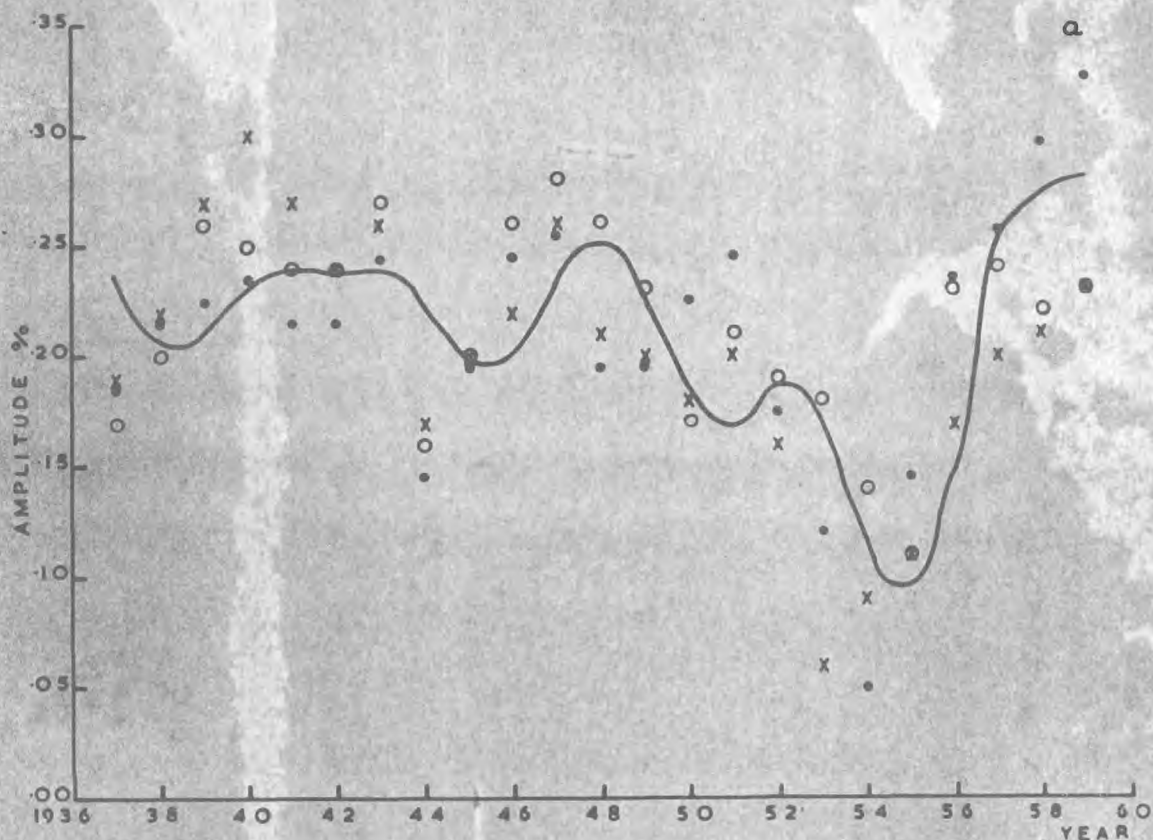


FIGURE 6.2(111)

LONG TERM CHANGES IN
AMPLITUDE OF DIURNAL VARIATION

The significance level is taken as $3\sigma = 0.012\%$. Examining the Table we see that the first five harmonics are significant. Once again we can attribute no physical significance to the third, fourth and fifth harmonics, and shall assume that they indicate variations in phase and amplitude of the eleven year wave from period to period. An interesting point is that the magnitude of the amplitudes of the second to the fifth harmonics relative to the amplitude of the first harmonic are a good deal larger than in the case of the phase.

This point is emphasised in figure 6.2.(iii)b where the twenty-two year wave and the sum of the other four harmonics are plotted. In general the twenty-two year wave is considerably modified by the irregular eleven year wave, much more so than in the time of maximum analysis. In particular the modification is quite small during roughly the first solar cycle, but during the second, approximately 1948 to 1958, it is considerable, especially at the minimum 1954-55 and at the maximum 1958.

Figure 6.2.(iii)c shows the deviations of the three sets of data about the mean fitted curve in a. The scatter is rather wide. Three standard deviations for the individual points is approximately 0.069% so that several of the deviations could be significant. Since there is some uncertainty in the total error it does not seem fruitful to ~~further~~ ^{further} consider these deviations.

Finally we considered the individual stations and did a separate Fourier analysis on each one. The results of this analysis are given in Table 6.2.F. Significant levels for each station are indicated.

An examination of Table 6.2.F. shows that at Huancayo the first five harmonics are significant, at Cheltenham the first and the fourth and at Christchurch the first, second and fourth. To illustrate this more clearly we have plotted the first, second and fourth harmonics on a series of harmonic dials in figure 6.2.(iv).

Two interesting points are shown on these dials. The amplitude of the twenty-two year for Cheltenham is nearly twice that at the other two stations, although statistically this difference is not significant. Also we see that the magnitude of the eleven-year wave is greater than that of the twenty-two year wave at Huancayo, although again this difference is not statistically significant.

In summing up we can say that there exists a twenty-two year wave in the diurnal amplitude data, that occurs at each station considered. It has the same phase as the wave found in the time of maximum data. There also exists a slightly smaller and rather irregular eleven year wave, which occurs at each station except Cheltenham, and there it only just misses being significant. This eleven year wave shows pronounced

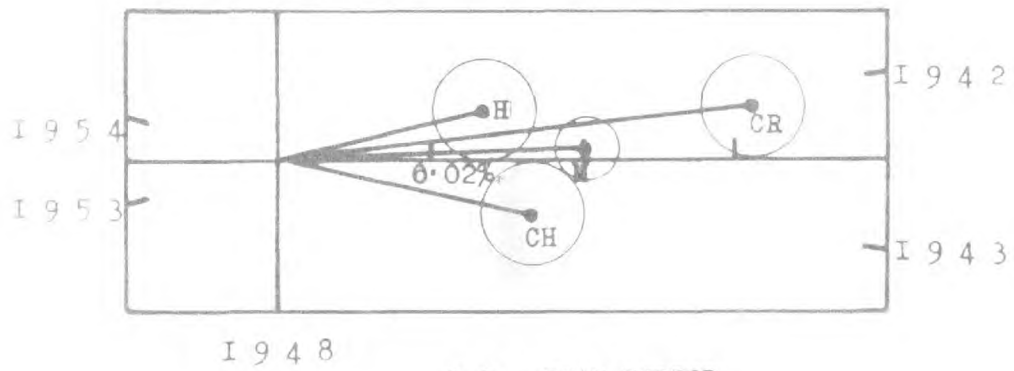
HARMONIC COMPONENTS OF THE AMPLITUDE VARIATION

HUANCAYO $3\sigma = 0.021\%$			CHELTENHAM $3\sigma = 0.021\%$			CHRISTCHURCH $3\sigma = 0.021\%$		
HARMONIC COEFFICIENTS	AMPLI- TUDE	TIME OF FIRST MAX.	HARMONIC COEFFICIENTS	AMPLI- TUDE	TIME OF FIRST MAX.	HARMONIC COEFFICIENTS	AMPLI- TUDE	TIME OF FIRST MAX.
$a_1 = 0.007$ $b_1 = 0.026$	0.027	1941.6	$a_1 = -0.007$ $b_1 = 0.032$	0.033	1943.2	$a_1 = 0.007$ $b_1 = 0.062$	0.062	1942.9
$a_2 = 0.037$ $b_2 = 0.003$	0.037	1937.1	$a_2 = 0.020$ $b_2 = 0.005$	0.021	1937.4	$a_2 = 0.027$ $b_2 = 0.021$	0.034	1938.2
$a_3 = 0.023$ $b_3 = -0.020$	0.030	1943.5	$a_3 = -0.007$ $b_3 = -0.017$	0.018	1942.0	$a_3 = 0.000$ $b_3 = -0.019$	0.019	1942.5
$a_4 = 0.014$ $b_4 = -0.035$	0.037	1941.5	$a_4 = 0.022$ $b_4 = -0.015$	0.027	1942.0	$a_4 = 0.008$ $b_4 = -0.022$	0.023	1941.4
$a_5 = -0.021$ $b_5 = -0.002$	0.021	1939.4	$a_5 = -0.016$ $b_5 = -0.020$	0.026	1940.9	$a_5 = -0.015$ $b_5 = -0.004$	0.016	1940.5
$a_6 = 0.009$ $b_6 = -0.010$	0.013	1940.2	$a_6 = 0.001$ $b_6 = -0.016$	0.016	1939.8	$a_6 = 0.001$ $b_6 = -0.001$	0.012	1940.6

TABLE 6.2 F

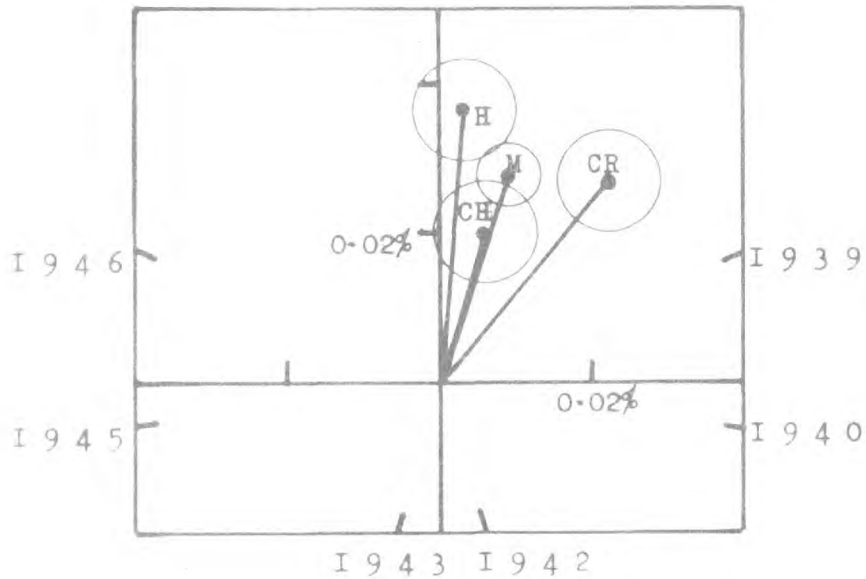
1937

- 107 -



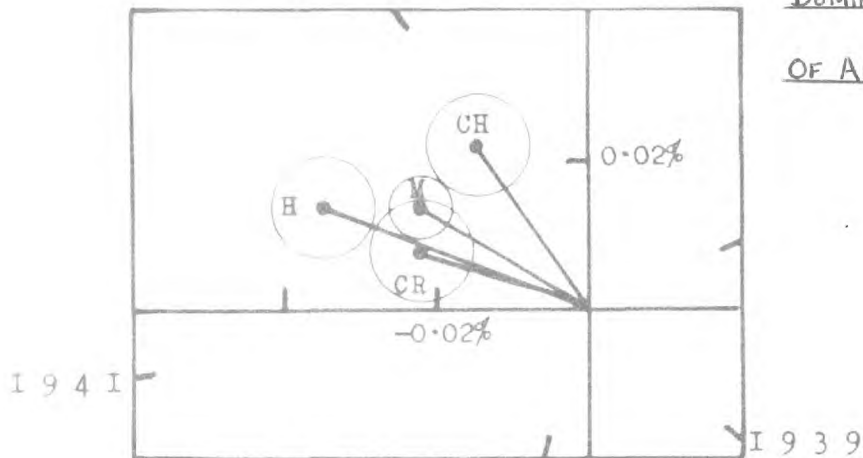
22 - YEAR PERIOD

1937



11 - YEAR PERIOD

1937



DOMINANT HARMONICS
OF AMPLITUDE VARIATION

5.5 - YEAR PERIOD

FIGURE 6.2 (iv)

maxima and minima at sunspot maxima and minima.

Since there appears to be some evidence that the amplitude of the diurnal variation is sensitive to solar activity, it seems to be of interest to compare magnetic activity with the changes observed in the amplitude. This type of analysis has been done previously (Sarabhai and Kane 1953, Sarabhai et al. 1954) using K_p figures as a measure of magnetic activity, but no very definite result has been obtained. However, these attempts have used non-atmospheric temperature corrected results.

The choice of magnetic index to indicate magnetic activity is not entirely obvious. What we require is some indication of the electromagnetic condition of the space surrounding the earth. Many workers have used K_p for this, but this index does not seem very sensitive to the cosmic ray changes associated with disturbed solar conditions. A more promising choice seems to be the planetary magnetic index A_p , which is a measure of the variability of K_p , and hence it is assumed, of changes in electromagnetic conditions in interplanetary space close to the earth.

Figure 6.2.(v) shows a plot of atmospheric temperature corrected amplitude of the diurnal variation (left-hand scale) and of magnetic index A_p (right hand scale) over the period 1937 to 1959. A_p is indicated by the thick line. The figure shows an apparent correlation between the two sets of data, particularly for the troughs during 1944-5, 1948-9 and 1954-5,

COMPARISON OF LONG TERM CHANGES IN A_p AND AMPLITUDE OF DIURNAL VARIATION

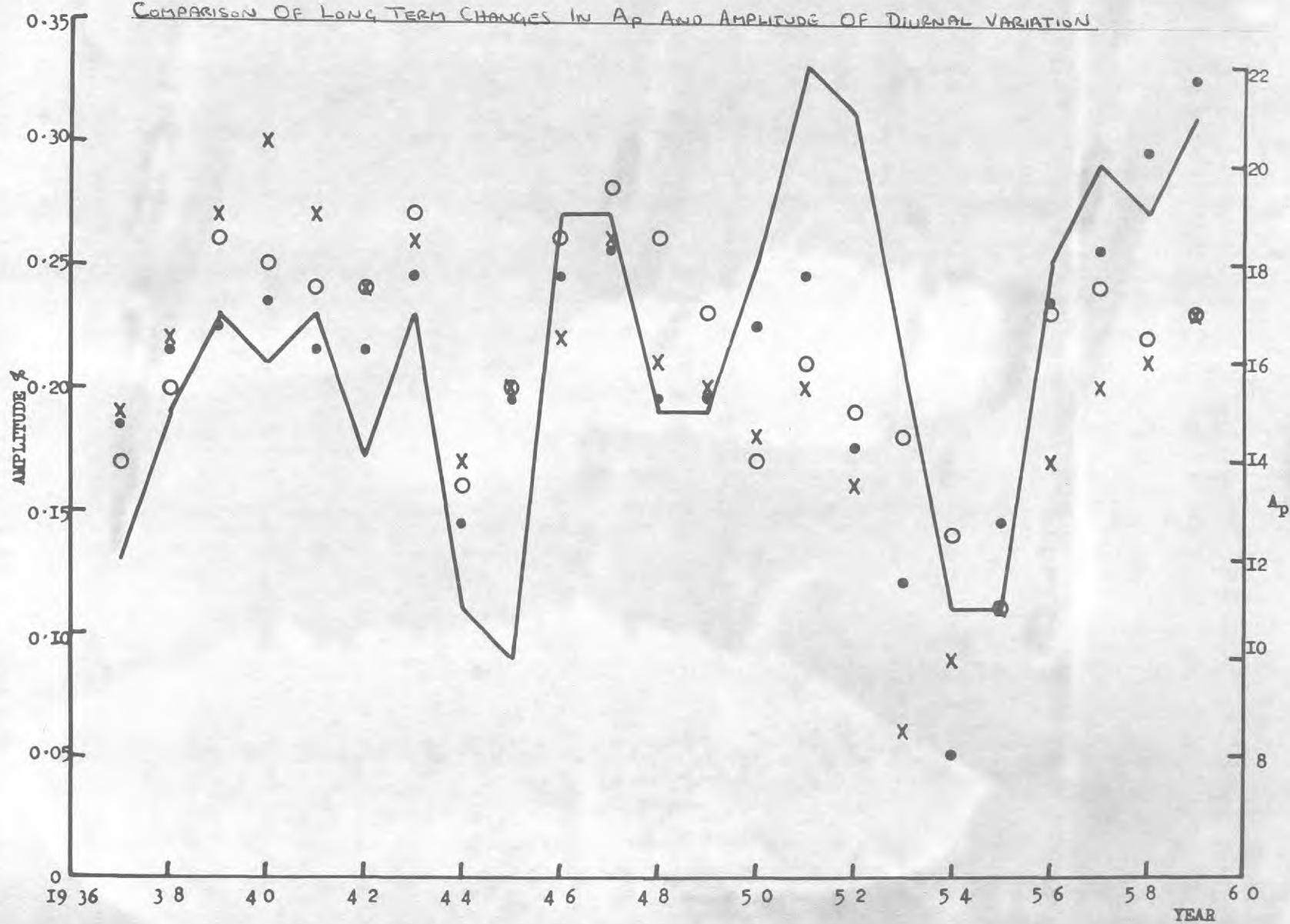


FIGURE 6.2 (v)

and for the peaks of 1946-7, 1957,8,9. The apparent peak in the amplitude data during 1951 is not statistically significant.

To investigate the relationship more quantitatively we correlated the changes in amplitude with the changes in A_p . A correlation coefficient of + 0.51 was obtained, which exceeds the 0.05 significance level. However this does not tell us a great deal since magnetic activity tends to follow solar activity (Chapman and Bartells 1940) and it could be the eleven year waves in the two sets of data that are correlating. To test this, we Fourier analysed the A_p values for the first six harmonics. The largest amplitudes obtained were for the five point five and four point four year period waves, so we extracted the five point five year waves from each set of data and re-correlated. It made negligible difference.

Thus, even using atmospheric temperature corrected amplitudes we find no firm relationship between the diurnal variation amplitude and magnetic activity as measured at the surface of the earth.

6.3. Comparison with the Diurnal Variation of the Nucleonic Component

In Chapter I it was indicated that the nucleonic component was effectively free from atmospheric temperature effects. Since both the ion-chamber and the neutron monitor are omnidirectional devices, limited only by atmospheric absorption, then,

initially neglecting the rigidity dependence of the two components, we should expect reasonable agreement between the diurnal variation of the nucleonic component and the atmospheric temperature corrected diurnal variation of the ionization component.

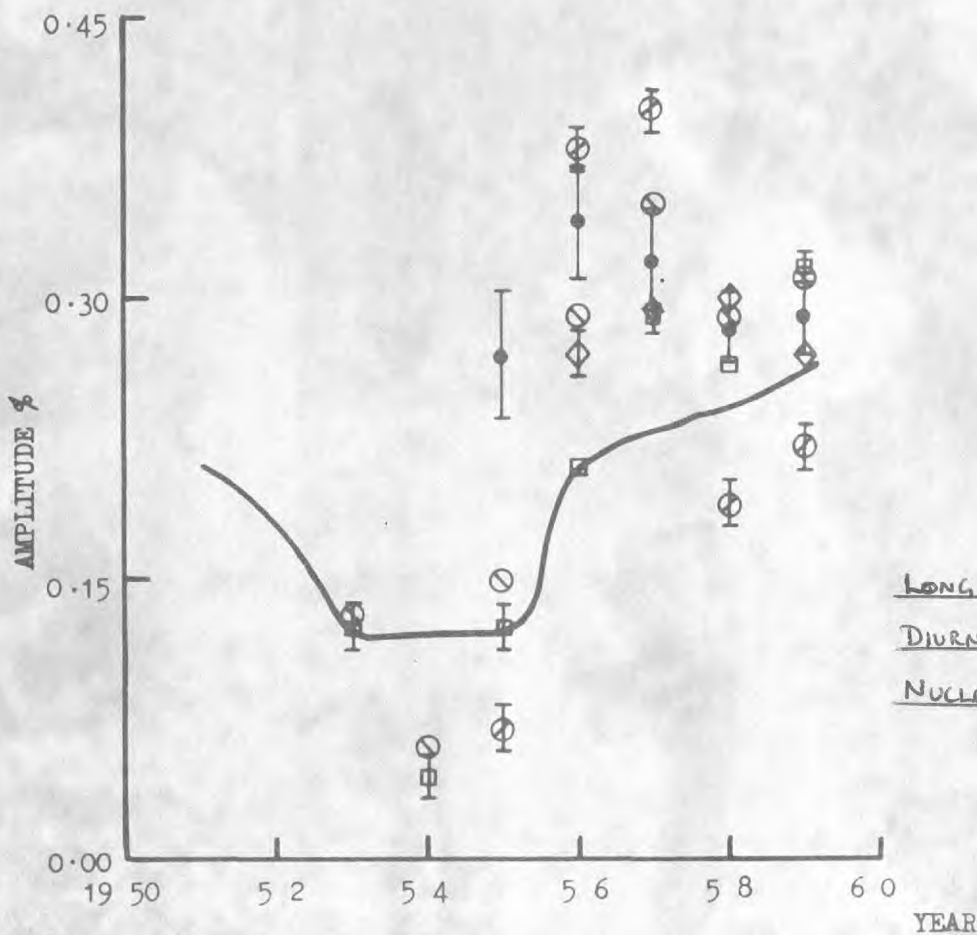
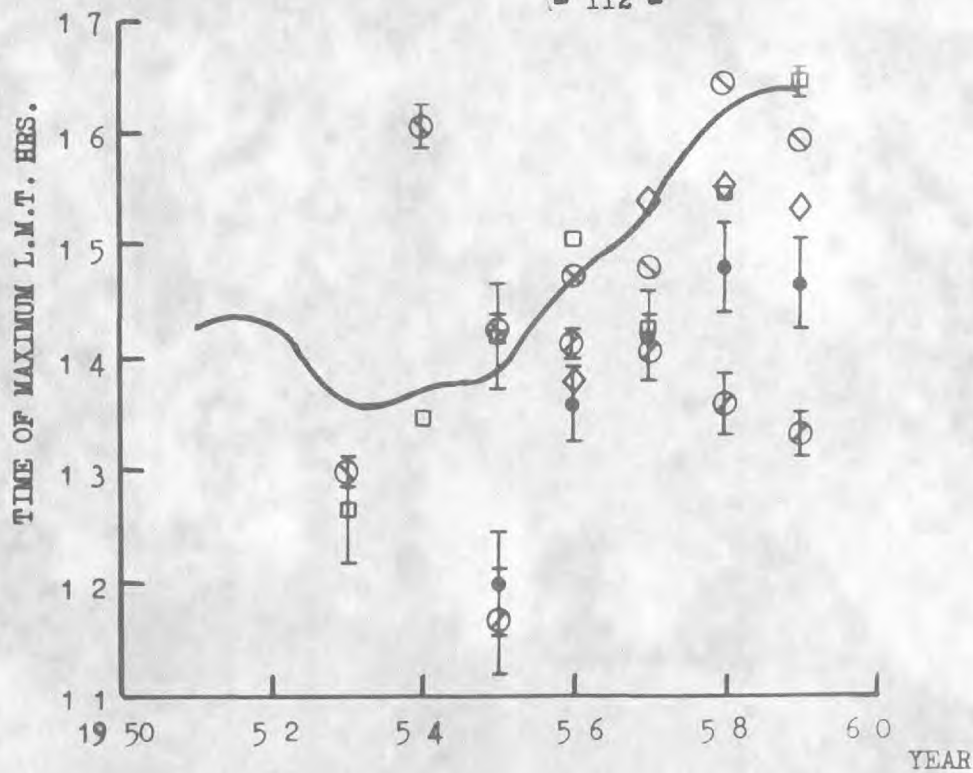
To check this, and hence check the consistency of the results obtained in the previous section, we collected nucleonic data over as long a period as possible. Neutron monitors were not in use much before 1950-51 and we could not obtain reliable data on the diurnal variation before 1953. The data actually used is that from Climax, Huancayo, Mt. Norikura, Ottawa and Leeds.

Figure 6.3.(i) shows a comparison between the diurnal variations of the nucleonic component and of the μ -meson component. The upper diagram shows the time of maximum changes, and the lower diagram, the amplitude changes. The thick line on each diagram represents the mean value of the μ -meson data, so that the diagrams represent the way in which the nucleonic data is scattered about the mean μ -meson results.

There are a number of experimental points to be noted. 2.3 hours has been added to each of the Huancayo times of maximum for the reasons discussed in section 6.2. Similarly 1.2 hours has been added to each of the Mt. Norikura times of maximum. It was necessary to interpolate the 1954 Huancayo phase value, since the amplitude was negligible for this period. The

- II2 -

NUCLEONIC COMPONENT



LONG TERM CHANGES IN
DIURNAL VARIATION OF
NUCLEONIC COMPONENT

FIGURE 6.3(i)

Huancayo and Climax data for 1956 was also interpolated since it was not available to us. In the amplitude diagram 0.035% has been added to each of the Huancayo values (section 6.2) and 0.018% to each of the Mt. Norikura values. We are assuming that the rigidity dependences of the two components are the same for the diurnal variation. We shall discuss this point later. Finally, the errors shown were computed from counting rates and represent the minimum errors. Where no errors are indicated they are of the same order as the size of the plotting dots except for Huancayo 1954, 56, Climax 1956.

If we consider the phase diagram first, then apart from the Climax value during 1954, only the Leeds values deviate seriously and significantly from the mean curve. The very high value for Climax during 1954 is due to a period of several months when the time of maximum became very late i.e. 2300 hours L.M.T. (Conforto and Simpson 1957). If these months are removed, then the phase agrees with the value interpolated for Huancayo. Since this very late time of maximum is anomalous and does not represent the average behaviour, the wide deviation is not serious.

All of the Leeds values shown are internally consistent i.e. they do not contain a small number of months during which the phase was widely different from the average phase (Marsden and Begum 1960). Thus with the exception of 1956, '57 the Leeds values must be considered as significantly different from the

general trend.

Considering the lower, amplitude, diagram we see that the agreement with the mean curve is not very good during 1956, 1957. The high Climax and Huancayo values during 1957 are possibly due to the fact that only half of the year's data was available to us i.e. July to December. However, once again the most serious and significant deviations are for the Leeds values. Since there appears no physical reason why the Leeds data should differ so considerably from the general trend, and considering the large jump in amplitude from 1955 (0.07%) to 1956 (0.38%), there seems some reason for considering the Leeds data as not characteristic of the general world-wide trend. It is interesting to note that the general level of the nucleonic component intensity at Leeds shows a number of anomalous and discontinuous changes during this period (Webber 1960).

In general there seems a fair measure of agreement between the long-term changes in the ionization, and nucleonic, component diurnal variations, except that the comparison is over a rather short period of time.

In comparing the diurnal variations of the nucleonic and ionization components we have ^{far}so neglected the rigidity dependence of these two components. Webber and Quenby's (1959) response curves for the neutron monitor and the ion chamber show that there is no doubt that in general the nucleonic

component is related to the lower rigidity end of the primary cosmic ray spectrum. However the significant quantity in the present discussion is the rigidity spectrum of the diurnal variation. This is not known with any certainty and a discussion is complicated by the possibility that the spectrum varies over the solar cycle. However we shall try to indicate some facts that seem significant.

Using the years for which the neutron data is most reliable i.e. 1953, '55, '58 and '59, the ratio of the amplitudes between Climax and Huancayo is 1.3:1, while the total areas under the differential response curves (Webber and Quenby 1959), assuming the vertical cut-off at Climax to be 2.7 GV, (Quenby and Webber 1959) are roughly in the ratio 2.5:1. For the same four years the ionization data gives the ratio of amplitude between Huancayo and Christchurch and between Huancayo and Cheltenham as 1.0:0.8 and 1.0:1.0 respectively. The respective areas under the differential response curves are as 1.1:1.0 for both Christchurch and Huancayo and Cheltenham and Huancayo.

The close correlation between changes observed at Huancayo and at high latitude stations i.e. Cheltenham and Christchurch, and the fair agreement observed between the diurnal variation as measured by neutron monitors and ion-chambers over the period 1953 to 1959 also seems worthy of note.

The above data points to the fact that the diurnal variation

is a relatively high rigidity phenomena and that particles with rigidities below, say 10 GV, contribute only a small fraction to the daily variation. This hypothesis is strengthened by measurements made with directional telescopes. Brunberg and Dattner (1954) have shown that, using their diagrams for the curvature of cosmic rays in the earth's magnetic field in conjunction with directional telescope measurements made at Manchester (Elliot and Dolben 1951), and at Stockholm (Malmfors 1949) they could estimate the mean rigidity of the diurnal variation. This they found to be in the range 20 to 40 GV. These results were not corrected for atmospheric temperature effects. The Manchester data when roughly corrected for atmospheric temperature variations showed the mean rigidity to be in the range 15 to 30 GV. This value is supported by similar calculations made on atmospheric temperature corrected results of Mawson, Antarctica (Quenby and Thambyahpillai 1960) which also indicate a mean rigidity in the range 15 to 30 GV.

Finally, in a recent publication (Elliot 1960) Elliot has proposed a solar system modulation model of the cosmic radiation that seems capable of adequately explaining the main, observed cosmic ray variations, e.g. the eleven year cycle and the Forbush decrease. From this model he has constructed a diurnal variation rigidity spectrum which gives a diurnal variation at Huancayo similar to that observed. This spectrum indicates that particles in the rigidity range 1 to 10 GV contribute

relatively little to the diurnal variation.

If this is so, and the evidence cited above is in agreement with it, then the agreement observed between measurements of the diurnal variation of the nucleonic component and of the ionization component is explained, since the differential response curves of Webber and Quenby (1959) indicate that above 10 GV the rigidity responses for the two components are becoming rather similar.

6.4. Discussion

Let us first briefly review the experimental facts; the time of maximum of the diurnal variation of the hard component shows a significant twenty-two year period cycle. There is also a possible eleven year period cycle, perhaps only occurring at low latitudes. The maximum and minimum values of the twenty-two year cycle correspond to times of sunspot minima. The nodes of the cycle correspond to times of sunspot maxima. The eleven year cycle roughly follows the sunspot cycle, while the amplitude of the twenty-two year wave is three or more times that of the eleven-year wave. The amplitude is approximately one solar hour in magnitude.

For the changes in the amplitude of the diurnal variation the same situation exists, with small differences. A significant eleven year wave occurs at all three stations considered, while the ratio of the amplitudes of the twenty-two year and the eleven

year waves is about 1.3:1.0, the amplitude of the twenty-two year wave being about 0.04%.

It is convenient at this point to consider the effects of a possible difference between the correction vector used for Christchurch and Cheltenham, and that used for Huancayo. The discussion in section 4.5 indicates that the worst possible case is a reduction of slightly more than 50% in the amplitude of the correction vector we have already used. If we apply such a correction vector to the Christchurch and Cheltenham data, then we observe two differences in the results obtained compared with those that we have already discussed.

The first change is in the mean value of the phase and amplitude at each station. We find that there is no longer a difference in mean amplitude between the two high latitude stations and Huancayo. Also that the difference in phase is reduced to about 1.5 hours.

The other change is on the amplitudes of the variations observed, no shift in positions of maxima or minima occurring. The amplitude of the twenty-two year phase variation is increased by approximately 50%. The amplitude of the twenty-two year amplitude variation has the magnitude of its first harmonic reduced by approximately 40% and its higher harmonics by approximately 20%. This makes the harmonics at Cheltenham of doubtful significance.

Thus the net effect of the largest possible difference between correction vectors used for the high latitude stations and for Huancayo is to leave the twenty-two year variation basically unaffected, merely emphasizing the variation in the phase and reducing the magnitude of the variation in the amplitude.

An interesting point concerns the possible eleven year wave in the phase and why it perhaps only occurs close to the equator. As pointed out in the previous section, the Brunberg-Dattner plot of Ψ_E against rigidity for stations at low latitudes is a very steep function, small changes of rigidity causing relatively large changes in Ψ_E . Thus if the diurnal variation rigidity spectrum 'hardens' or 'softens' slightly over the solar cycle it would cause a change in Ψ_E near the equator that would not be apparent at higher latitudes.

Two main attempts have been made to calculate the rigidity spectrum of the diurnal variation. That due to Dorman (1956) is largely determined by some early, and doubtful, nucleonic component data, and by inaccurate differential response curves. It is thus unreliable and will not be considered further.

This leaves the spectrum due to Elliot (1960), which has been mentioned previously (Chapter V and section 6.3). The simplest means of calculating the diurnal variation from this spectrum is to choose a low latitude station, since this avoids

difficulties about corrections for the deflection of particles in the earth's magnetic field that occur in the discussion for high latitude stations. The most convenient component of the cosmic radiation to choose is the nucleonic component, since there is some doubt about the differential response curve for the ionization component at high rigidities (Webber and Quenby 1959).

Elliot (1960) has indicated values for the three parameters (M_s = the 'solar dipole' moment, α_K = A measure of the probability of absorption of cosmic rays by the sun; and p_0 = A measure of the strength of the scattering centres that cause deflection of the cosmic rays) that define his rigidity spectrum, for the years 1955, '57 and '58. Using these values we have computed the phase and amplitude of the diurnal variation of the nucleonic component at Huancayo (latitude -0.6° S geomagnetic). The results depend somewhat upon the estimates made of the effects of geomagnetic, instrumental and geometric, smoothing, but there is no serious discrepancy between the changes predicted over these four years with those actually observed for both the phase and the amplitude of the nucleonic component diurnal variation. The shortness of the period over which the comparison may be made is unsatisfactory. However, due to atmospheric temperature effects and the magnitude of the changes involved, the long term changes in the nucleonic

component, as opposed to the ionization component, are the most satisfactory means of determining the values of M_s , αK and p_0 . Since the nucleonic component has been recorded over a much shorter period of time than the μ -meson component, this makes the problem of obtaining the long-term changes in M_s , αK and p_0 rather difficult. However from the available data we may make a number of points.

If, as seems probable, Elliot's rigidity spectrum is a good approximation to the true diurnal variation spectrum, then since the factors determining its structure i.e. M_s , αK and p_0 , are determined by the general level of intensity, so that they must follow an approximate eleven year cycle, it offers the solution to the observed eleven year cycle in the diurnal variation data. Since the values of M_s , αK and p_0 were determined experimentally and are interdependent to a certain extent we are not justified in trying to extrapolate backwards in time. Hence we are not able to calculate the changes in phase and amplitude of the diurnal variation to be expected over a complete sunspot cycle. Until such a calculation is possible, and in view of the reasonable agreement over the period 1955-58, there seem some grounds for regarding the eleven year variation in the diurnal variation as provisionally explained.

This model of the diurnal variation also suggests a possible explanation of the twenty two year cycle. The simplest

solution of such a cycle is that the mechanism producing the diurnal variation caused it to be larger in amplitude and later in phase during one solar cycle, than in the next. Such a situation would occur in Elliot's model if M_s , αK and p_0 were greater in magnitude during the 1943-44 sunspot minimum than during the 1953-54 minimum. However, if this were so, then from the changes calculated for the period 1955-58, we should expect a difference in mean intensity $\sim 3\%$ between 1943-44 and 1953-54, while data from the Carnegie Institute stations (Forbush, 1958) shows that there is negligible difference between the mean intensities of these times.

However, the mean intensity is dependent on the overall value of the three constants M_s , αK and p_0 i.e. they can be adjusted relative to each other without altering the level of the mean intensity. However, such manipulation can cause a considerable variation in the diurnal variation, so that lacking experimental knowledge of M_s , αK and p_0 it is not at all possible to decide on the validity of the hypothesis outlined in the previous paragraph.

Summary

The diurnal variation of the ionizing component has been corrected for the effects of atmospheric temperature variations, using the vector obtained by Quenby and Thambyahpillai (1960). The corrected results confirm the twenty two year

variation in the phase of the diurnal variation noted by Elliot and Thambyahpillai (1953). There are indications that a similar wave exists in the changes with time of the amplitude of the diurnal variation. Variations with periods of eleven years are also present.

A comparison is made with the diurnal variation of the nucleonic component. Reasonable agreement is obtained, but over a rather limited period of time. Finally Elliot's theory (1960) of the diurnal variation is considered. It is found consistent with the changes observed over the period 1955 to 1958.

ACKNOWLEDGEMENTS

I am indebted to Professor P.M.S. Blackett for the privilege of working in his Laboratories. My thanks are due to Professor H. Elliot for suggesting the Soft Component experiment and for his constant help and encouragement.

Generally I should like to thank my colleagues of the Cosmic Ray Group for their interest. In particular I should like to thank Dr. J.J. Quenby for a number of useful discussions and Dr. T. Thambyahpillai for a great many stimulating and helpful discussions.

I am indebted to Mr. G. Cescotti for a number of technical discussions and for the design of the anti-coincidence circuit. Also to Miss A. Benton for assistance in the reduction of the Soft Component data.

Finally I am grateful to the Department of Scientific and Industrial Research for the provision of a Research Studentship.

APPENDIX

The Rigidity Dependence of the Soft Component

There are two methods of comparing the relative rigidity dependence of the soft and hard components. One is to use the earth's magnetic field as a spectrometer, i.e. make a latitude survey. The other is to select large Forbush decreases and to regress the changes in intensity of the two components during a decrease.

Several latitude surveys making simultaneous measurements of the hard and soft components, using geiger tube telescopes, have been made (e.g. Morris et al. 1948, Law et al. 1949). All suffer from poor statistical accuracy. That due to Law et al. is the most satisfactory. They observed a latitude decrease in the hard component of $(18 \pm 2)\%$, and in the soft component of $(20 \pm 2)\%$. Thus within the rather large errors there is no significant difference of rigidity response.

When comparing rigidity responses during a Forbush decrease it is better to regress the components separately against the changes in the nucleonic component, since the latter decreases approximately twice as much as the μ -meson component (e.g. Webber 1961). Also, comparative measurements giving the appropriate hard-nucleonic ratio exist (McCracken 1959). We need also to consider the question of atmospheric temperature. The Forbush decreases selected were those during which the 100 mb. layer height was approximately constant. The data was limited to six hours before, and eighteen hours after, the decrease. This was the shortest time consistent with obtaining definite results

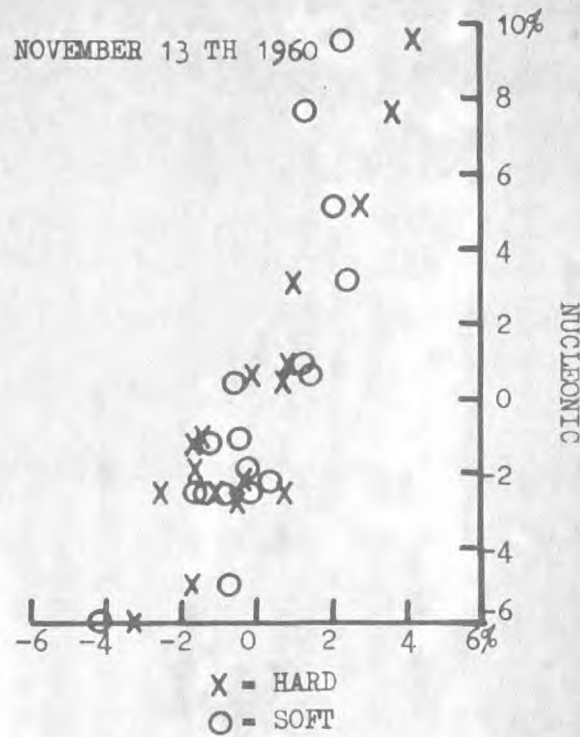
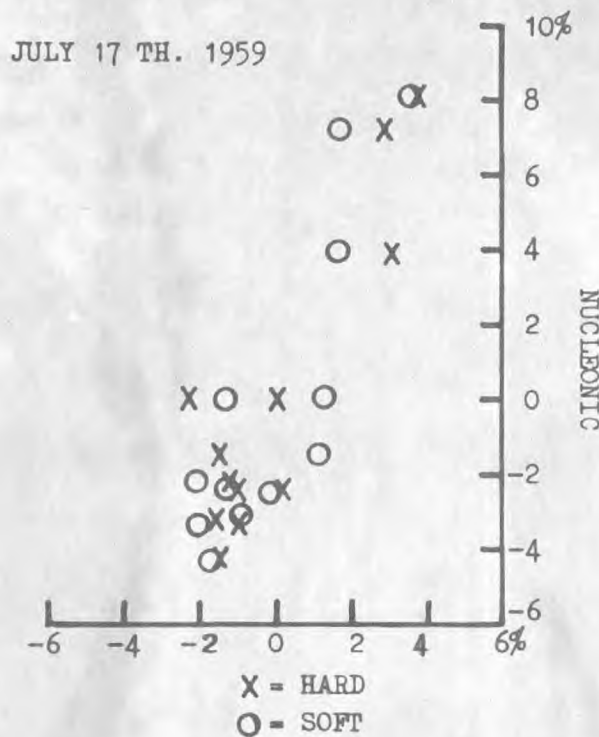
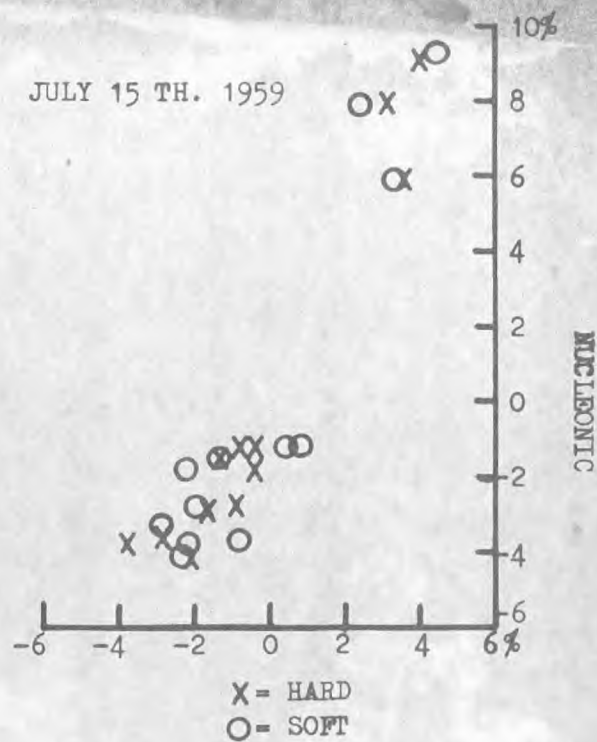
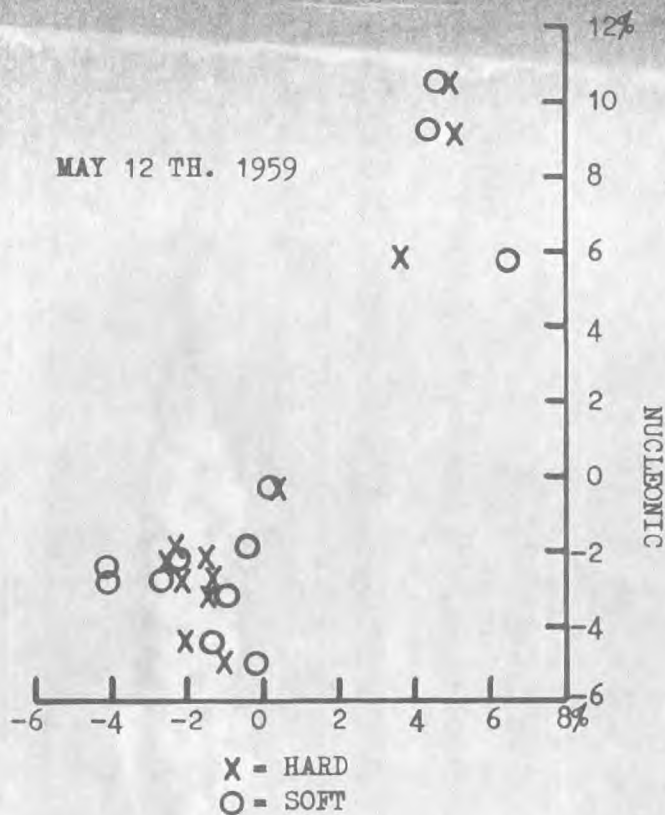
and reducing the possibilities of significant atmospheric temperature change. The use of the nucleonic component helps reduce this source of error too, since the nucleonic component is unaffected by these temperature changes.

Figure A.1 shows the combined scatter diagrams between the hard component and the nucleonic component, and the soft component and the nucleonic component, for each of the four decreases chosen. Bihourly values of the data have been used, since this was the shortest time period available to us in the nucleonic data, the standard errors being 0.44%, 0.47% and 0.80% for the nucleonic, the hard, and the soft components respectively. Thirty-four hours data, instead of twenty-four hours, has been used in the November 13th event to take advantage of a further decrease that occurred shortly after the main event. Each of the four events was a decrease of about 10% in the nucleonic component.

A visual inspection of the diagrams show that there is little difference between the two sets of points. This effect is put on a quantitative basis by the results of the regression analysis shown in Table A.a

<u>EVENT</u>	β_{NIH}	<u>HARD</u> $1/\beta_{IHN}$	β_{NIS}	<u>SOFT</u> $1/\beta_{ISN}$
MAY 12TH 1959	1.74 ± 0.16	1.90 ± 0.18	1.28 ± 0.38	1.80 ± 0.25
JULY 15TH 1959	1.78 ± 0.19	1.99 ± 0.25	1.87 ± 0.19	2.30 ± 0.37
JULY 17TH 1959	1.73 ± 0.29	2.26 ± 0.40	1.91 ± 0.39	2.70 ± 0.60
NOV. 13TH 1960	1.07 ± 0.21	1.26 ± 0.08	2.00 ± 0.40	3.07 ± 0.60
MEAN	1.58 ± 0.30	1.85 ± 0.37	1.76 ± 0.28	2.47 ± 0.47

TABLE A.a



REGRESSION DIAGRAMS BETWEEN HARD AND
SOFT AND NUCLEONIC COMPONENTS DURING
FIGURE A.1 FORBUSH DECREASES

The errors quoted in the Table are standard regression errors, while the errors of the means are the standard deviations within the sets of four results.

An examination of this Table indicates no difference in the rigidity responses of the hard and soft components except perhaps from the November 13th 1960 event, and here the differences are not significant. The hard-nucleonic component ratios are in agreement with those obtained by McCracken (1959).

As a further check we did a direct regression analysis between the soft and hard components. The results of the analysis are shown in Table A.β

EVENT	β_{IHS}	β_{ISH}
MAY 12TH 1959	0.76 ± 0.11	1.1 ± 0.15
JULY 15TH 1959	0.88 ± 0.16	0.86 ± 0.15
JULY 17TH 1959	0.72 ± 0.27	0.57 ± 0.21
NOV. 13TH 1960	1.01 ± 0.18	0.67 ± 0.12
MEAN	0.84 ± 0.11	0.80 ± 0.20

TABLE A.β

The errors were obtained in the same way as in Table A.α. We see once again that there is no evidence for a difference in rigidity response between the hard and soft components. Thus in general we find that during Forbush decreases the soft and hard components exhibit the same rigidity dependence.

This is in agreement with the results of Dawton (1952), when the mean decrease over a period of days is calculated for the two Forbush decreases he observed ($\sim$ 3-4% in the hard component).

Since the rigidity dependence of the hard and soft components is the same for latitude decreases and Forbush decreases, and the latter dependence extends to the order of 200 GV, we shall make the assumption that the rigidity dependence of the daily ^{variation} ~~variation~~ for both components is similar over the same rigidity range.

REFERENCES

- Barker, P. Phys. Rev. 100, 860, (1955).
- Blackett P. Phys. Rev. 54, 973, (1938).
- Brunberg, B. and Dattner, A. Tellus, 5, 135, (1953).
- Brunberg, B. and Dattner, A. Tellus, 6, 73, (1954)
- Brunt, D. Combination of Observations. C.U.P. 1933.
- Chapman, S. and Bartells, J. Geomagnetism. O.U.P. 1940.
- Conforto, A. and Simpson, J. Nuov. Cim. 5, 1052, (1957).
- Dawton, D. Ph.D. Thesis, Manchester Univ. (1952).
- Dawton, D. and Elliot, H. J. Atmos. Terr. Phys. 3, 295, (1953).
- Dolbear, D. and Elliot, H. J. Atmos. Terr. Phys. 1, 215, (1951).
- Dorman, L. Cosmic Ray Variations, Moscow, (1956).
- Duperier, A. Proc. Phys. Soc. 57, 464, (1945).
- Duperier, A. Proc. Phys. Soc. 61, 34, (1948).
- Duperier, A. Proc. Phys. Soc. A62, 684, (1949).
- Elliot, H. Proc. Phys. Soc. A62, 369, (1949).
- Elliot, H. Phil. Mag. 5, 601, (1960).
- Elliot, H. and Dolbear, D. J. Atmos. Terr. Phys. 1, 205, (1951).
- Elliot, H. and Rothwell, P. Phil. Mag. 1, 669, (1956).
- Elliot, H. and Thambyahpillai, T. Nature, 171, 918, (1953).
- Fisher, R. Statistical Methods for Research Workers - Oliver and Boyd, (1946).
- Forbush, S. Terr. Mag. 42, 1, (1937).
- Forbush, S. J. Geophys. Res. 59, 525, (1954).
- Forbush, S. Moscow CSAGI Conf. (1958).

- Forbush, S. and Venkatesan, D. J. Geophys. Res. 65, 2213, (1960).
- Franklin, E. and Loosemore, W. A.E.R.E. Report, (1950).
- French, W. and Chasson, R. J. Atmos. Terr. Phys. 14, 1, (1959).
- Ginsberg, V. Prog. Cosmic Ray Phys. Vol. IV - North Holland, (1958).
- Glokova, E. Proc. Moscow Conf. 4, 243, (1960).
- Hodson, A. Proc. Phys. Soc. A64, 1061.
- Hogg, A. Proc. Phys. Soc. A192, 128, (1947).
- Law, P. et al. Aust. J. Sci. Res. A2, 493, (1949).
- Lindgren, S. and Lindholm, F. Uppsala Univ. Report, (1960).
- Malmfors, K. Tellus, 1, 2, (1949).
- Marsden, P. and Begum, Q. Phil. Mag. 4, 1247, (1959).
- McCracken, K. Phys. Rev. 113, 343, (1959).
- Morris, P. et al. Phys. Rev. 74, 1102, (1948).
- Possner, M. and Van Heerden, I. Phil. Mag. 1, 253, (1956).
- Puppi, G. Prog. Cosmic Ray Phys. Vol. III - North Holland, (1956).
- Quenby, J. and Thambyahpillai, T. Phil. Mag. 5, 585, (1960).
- Quenby, J. and Webber, W. Phil. Mag. 4, 90, (1959).
- Quenby, J. and Wenk, G. To be published, (1961).
- Rossi, B. Phys. Rev. 57, 469, (1940).
- Rossi, B. Rev. Mod. Phys. 20, 537, (1948).
- Rossi, B. High Energy Phys. - Prentice Hall, (1952).
- Rossi, B. and Klapman, S. Phys. Rev. 61, 414, (1942).
- Sarabhai, V. et al. Phys. Rev. 96, 469, (1954).

- Sarabhai, V. and Kane, R. Phys. Rev. 90, 204, (1953).
- Schonland, B. et al. Terr. Mag. 42, 137, (1937).
- Sekido, Y. and Kodama, M. Report of Ionosphere Research in Japan, 7, No. 2, (1952).
- Sherman, H. New York Univ. AFOSR Report No. TR-56-32, (1956).
- Simpson, J. et al. Phys. Rev. 90, 934, (1953).
- Simpson, J. and Fagot, W. Phys. Rev. 90, 1068, (1953).
- Störmer, C. The Polar Aurora - O.U.P. (1955).
- Thambyahpillai, T. Ph.D. Thesis - Manchester Univ. (1953).
- Trefall, H. Proc. Guanajuato Conf. 1955.
- Trefall, H. Proc. Moscow Conf. 4, 258, (1960).
- Trumpy, B. Univ. Bergen Arbok, 2, 19, (1955).
- Trumpy, B. and Trefall, H. Physica, XIX, 636, (1953).
- Van Duuren, K. et al. Nucleonics, June 1959, 86.
- Van Zoonen, D. App. Sci. Research, B3, 377, (1953).
- Webber, W. Private Communication, (1960).
- Webber, W. Prog. in Cosmic Ray Phys. Vol. V (To be published) (1961).
- Webber, W. and Quenby, J. Phil. Mag. 4, 654.
- Wilkes, M. Oscillations of the Earth's Atmosphere - C.U.P. (1949).