RADIO STUDIES OF THE IONOSPHERIC D-REGION

By

T.J. STUBBS, B.Sc. (Hons)

A Thesis
presented for the degree of
DOCTOR OF PHILOSOPHY
at the
UNIVERSITY OF ADELAIDE
(Physics Department)

CONTENTS

PAGE NO.

PAGE NO.

PAGE NO.

PAGE NO

PREFACE

To the best of the author's knowledge, this thesis contains no material previously published or written by another person, except where due reference is made in the text. It contains no material which has been submitted or accepted for the award of any other degree or diploma in any University.

(T.J. Stubbs)

ACKNOWLEDGEMENTS

The work described in this thesis was carried out under the supervision of Dr B.H. Briggs using the facilities of the University of Adelaide. The author is greatly indebted to Dr Briggs for his very helpful and enthusiastic guidance throughout the project.

The continued interest and help of Dr R.A. Vincent in many aspects of the work has also been appreciated. In connection with the Woomera project, the technical expertise of Mr L.A. Hettner and Mr N. Wild is gratefully acknowledged, as is the willing co-operation of Mr C. Low of the Weapons Research Establishment.

Discussions with Dr W.G. Elford and Professor C.O. Hines have provided the author with valued insights into various facets of the project.

Finally, thanks must go to my Mother, to Mrs Judy Taylor and to Mrs Patricia Coe for invaluable assistance in producing the thesis in its final form.

For most of the time, the author was in receipt of a Studentship from the Commonwealth Scientific and Industrial Research Organisation.

INTRODUCTION

This thesis contains the results of a study of irregularities and drifts in the D-region of the ionosphere. The observations were carried out using the facilities of the Buckland Park radio field station near Adelaide, South Australia (35°S, 139°E). The results presented cover the years 1972 to 1974. The work was undertaken at a time when relatively little use had been made of D-region partial reflections to obtain ionospheric drift measurements. It was certainly not widely accepted that the drifts so obtained were necessarily related to the neutral air motions at the same heights.

The initial chapter contains a brief review of the ionospheric D-region with particular emphasis on its dynamical properties and their measurement. During the course of the current work a new partial reflection experiment was established at Woomera, South Australia (31°S, 136°E). In Chapter 2 the experimental arrangement, both there and at Buckland Park, is reviewed together with an outline of main data analysis procedures.

One of the main objects of the drift work carried out in 1972 was a comparison with neutral wind measurements at similar heights made by meteor techniques. The motivation for, and the results of, this comparison are presented in Chapter 3. In the two subsequent chapters the partial reflection drifts are considered in more detail. Chapter 4 deals with the more regular features, both mean and harmonic, of the drifts whereas some of the fluctuations with time scales of a few hours or less are considered in Chapter 5.

Some of the first results of the comparisons with rocket neutral wind measurements made at Woomera appear in Chapter 6, although for many of the rocket flights the Woomera site was not in operation and the only drift values available were those from Buckland Park. On several occasions late in 1973 both drift recording systems were operational and a comparison of the respective drift values forms the subject matter of Chapter 7.

Up to this stage the emphasis has been on the horizontal drift velocities of the ionospheric irregularities. In Chapter 8 the nature of the irregularities themselves is discussed. The drift analysis technique yields information about the ground diffraction pattern. The relationship between these ground patterns and the ionospheric irregularities is considered since the patterns exhibit certain striking variations in time and height.

Finally the work is summarised and suggestions are made as to which path future developments should follow. It is the author's belief that despite the questions raised or left unanswered, the contents of this thesis should considerably enhance the status of the partial reflection technique as an indicator of neutral air dynamics at D-region heights.

CHAPTER ONE

THE D-REGION AND ITS DYNAMICS

The region of the ionosphere known as the D-region extends from about 50 km to 100 km and is fairly inaccessible from the point of view of scientific study. Even as recently as 20 years ago so little was known about the region that the designation "ignorosphere" was an appropriate one. The recent increase in understanding, at least of the structure of the D-region, has come about largely as a result of the study of partially reflected radio waves. And whilst most of our present knowledge of its complex dynamics has come from other techniques, many of the questions still remaining may yet find their resolution in the same source.

1.1 Nature of the D-region

1.1.1 Formation.

The production of D-region ionisation cannot be explained by the influence of solar X-rays alone although the ionisation of neutral air components by X-rays with wavelengths less than 20Å is a major source of ions. This source explains the solar cycle variations in electron density, the X-ray flux changing by a factor of 2 over a cycle. A less variable source of D-region ions is provided by the ionisation of nitric oxide by Lyman-α radiation. Although nitric oxide is a very minor constituent of the region resulting largely from nitrogen diffusing down from the E-region and reacting with atomic oxygen, it so happens that the Lyman-α wavelength of 1216Å coincides with one of the atmosphere's spectral windows. Below about 70 km a third source of ions becomes important - bombardment by cosmic rays. These particles collide more frequently with atmospheric molecules as they penetrate further into the atmosphere to regions of higher density, but on the other hand the rate of loss of electrons by attachment increases at the same time. This production/loss balance results in a maximum concentration from this source at about 65 km. This region exhibits solar control only in the sense of a switch causing or cutting off photo-detachment of electrons. Consequently the electron density in this region shows little variability during the day and is almost zero at night.

1.1.2. Structure

The total reflection of radio waves from the E-region at heights of around 110 km has long been a useful tool for ionospheric study but DIEMINGER (1952) reported finding on ionograms evidence of weak echoes from below the E-region and extending down to 70 km, implying a region of inhomogeneous ionisation. Subsequently GARDNER and PAWSEY (1953) suggested a technique which utilised these reflections to deduce an electron density profile. This technique was based on the different behaviour of the two magneto-ionic components of a linearly polarised radio wave. It is found that the amplitude ratio of the extraordinary echo to the ordinary echo varies with height in a manner dependent on the electron density distribution. This so-called Differential Absorption Experiment (DAE) has been widely applied (for example FEJER and VICE, 1959; BELROSE and CETINER, 1962; GREGORY and MANSON, 1967). BELROSE and BURKE (1964) modified the resultant profiles to take account of the energy dependence of collision cross section and PIGGOTT and THRANE (1966) have pointed out that at low heights the DAE results may have to be corrected for gradients in collision frequency. An interesting variant of the experiment (TITHERIDGE, 1962a) is to use just the ordinary component but at two different frequencies.

The overall picture that emerges is of daytime profiles which increase from about 100 electrons/cc at 60 km to about 100,000 electrons/cc at 100 km, with a decrease of between 2 and 3 orders of magnitude at night. However this increase with height is neither smooth nor monotonic and there are variations with solar cycle, season and, on a smaller time scale, with sunspot number (TITHERIDGE, 1962a). GREGORY (1956) found from a study of partial reflections that the D-region has properties which vary widely over the full height range, with structure and behaviour at 90 km appearing quite different from that near 60 km. ELLYETT and WATTS (1959) comprehensively reviewed numerous early partial reflection studies for evidence of stratification or 'preferred heights' of echoes. They also included optical and rocket techniques in their survey and concluded that whilst there was evidence for considerable fine structure, the only consistent long-term stratum was at about 85 km. Since then, further evidence for stratification has been obtained by various workers. In particular, GREGORY (1961) found at Christchurch (43°S) that the thickness of preferred regions increased with height whilst TITHERIDGE (1962b) found very stable strata at intervals of about 10 km from 70 - 100 km.

The interpretation of results such as these is made difficult because of the lack of understanding of the reflection process itself. Two favoured models (GREGORY and VINCENT, 1970) are of reflection from discrete bounded irregularities on the one hand, and on the other, of scattering from a volume in which inhomogeneities are continuously distributed. GREGORY and MANSON (1969a) obtained results more consistent with scattering from sharply bounded horizontally stratified irregularities than with volume scattering but it may be that the relative importance and validity of the two models varies with height.

1.1.3. Absorption

An important property of the D-region is its absorption of radio waves. Absorption is dependent on both electron concentration and collision frequency. The former basically increases upward through the D-region whereas collision frequency decreases. This results in maximum absorption in the vicinity of 80 km. A maximum near midday reflects the importance of the solar ionisation but the seasonal variation exhibits puzzling deviations from the expected winter minimum (BEYNON and JONES, 1965). On some winter days the absorption is much higher than can be explained by normal day-to-day variability. This feature has become known as the 'winter anomaly'.

1.1.4. Inter-region coupling

The detailed study of the D-region has had an important consequence. Up till now the ionosphere has been studied largely in isolation but it has become apparent both from comparative studies of individual events (GOSSARD and PAULSEN, 1968) and from observations of wavelike variations in D-region properties (GREGORY, 1965) that the upper atmosphere is in fact coupled to the meteorological realm of the troposphere and stratosphere. The hope for a full understanding of this coupling lies in the development of better D-region observation techniques which would ideally be capable of tracing an event in the lower atmosphere through to its ultimate effect at ionospheric levels. Most of the connections reported so far must be considered suggestive rather than convincing. Two promising tacks appear to be comparison of 'ground level' pressure and mean ionospheric wind fluctuations (HAUBERT, 1972) and particularly studies of the ionospheric response to stratospheric warmings (MANSON, 1968).

1.1.5. Turbulence

One final characteristic of the D-region which distinguishes it from the higher regions of the atmosphere and perhaps belongs equally in the next section, is the existence of turbulence - complex small-scale dissipative and non-linear motion fields which seem to terminate at close to 100 km (BLAMONT, 1963). This turbulent mixing gives rise to the observed chemical composition of the region (COLGROVE, JOHNSON and HANSON, 1966) which cannot be explained on the basis of molecular diffusion alone. More important from the point of view of this thesis are the effects of turbulence on wave motions in the medium and the part that it plays in producing ionisation inhomogeneities or irregularities.

1.2. Dynamics of the D-region

The importance of waves in D-region dynamics arises from the exponential growth of wave amplitude with height. This growth is required to keep the energy density constant in regions of exponentially decreasing density. A factor which acts to counter this growth to some extent (HINES, 1972) is the increase of eddy and molecular viscosity with increasing height. The relative importance of these two effects, and the effect of solar radiation in general varies greatly with season and geographical location and the formulation of a definitive model of wind and temperature at D-region heights is a very complex problem. Some of the earliest attempts at such a model were those of PANT (1956) KELLOGG and SHILLING (1957) and MURGATROYD (1957). As further experimental observations became available these were in turn modified by other workers, for example BATTEN (1961), KOCHANSKI (1963) and GROVES (1969). A particular problem is still the lack of Southern Hemisphere data.

It has become customary to discuss the various contributions to D-region wind fields in terms of mechanisms of progressively decreasing temporal scales. This procedure will be adopted in the ensuing discussion, but it should be noted that the various components do not exist in isolation and their interaction may be of great importance.

1.2.1. Prevailing Wind

The term 'prevailing wind' is assigned to that component of the wind which remains relatively constant over periods of several days. This general circulation arises from a geostrophic balance between Coriolos forces due to the earth's rotation and pressure forces brought about by thermal gradients in the atmosphere. At middle latitudes the prevailing flow is predominantly zonal and is eastward between 70 and 100 km for most of the year. The meridional component appears to be generally directed away from the summer pole, a result which would be somewhat surprising on radiative arguments alone. The prevailing wind varies appreciably throughout the year with strong annual and semi-annual components (GROVES, 1972). There is some evidence for even longer period variations, notably 24-month (TEPTIN, 1971) and 26-month (SPRENGER and SCHMINDER, 1967).

1.2.2 Planetary Waves

It is an interesting observational fact that the mean wind on successive days can be quite variable, often exhibiting wavelike fluctuations. MULLER (1972), in spectrally analysing winds deduced from drifting meteor trails, found prominent peaks in the vicinity of periods of several days. Planetary waves with spatial scales of order 1000 km and periods of 3 - 8 days are well known from tropospheric and stratospheric studies, and there is a possibility that such waves could propagate up through the D-region and appear as perturbations on the general circulation. This possibility was investigated theoretically by CHARNEY and DRAZIN (1961) who concluded that such propagation would only be likely at times when the mean zonal winds at upper levels were a minimum. This would be in spring and perhaps briefly in autumn. More recently, GREGORY and MANSON (1969b) found that variations in D-region electron density profiles could be accounted for by the upward propagation of planetary and cyclonic waves.

1.2.3 Tides

The earth rotates in the gravitational fields of the sun and moon and in the thermal radiation field of the sun. This results in periodic variations in parameters describing the behaviour of its oceans and atmosphere. Whereas the lunar tide is of great importance in the oceans such is not the case for the atmosphere. The subject has been comprehensively reviewed by MATSUSHITA (1967) but suffice it to say, in the present context, that the lunar influence at D-region heights is negligible. The main driving force for atmospheric tides is in fact solar heating and the basic periods of the tides are thus submultiples of the solar day, principally 24 hr and 12 hr. The development of tidal theory has been a tortuous one (SIEBERT, 1961; LINDZEN, 1968; CHAPMAN and LINDZEN, 1970) and a brief outline only will be given here.

One of the earliest problems associated with tidal theory (CHAPMAN, 1924) came from the observation that the semidurnal tide was more prominent than the diurnal although the main source, solar heating, was a diurnal effect. A lot of prominence was given to a 'resonance' theory which was based on the atmosphere having a natural oscillation period close to 12 hours. One of the shortcomings of this theory was that it had a sensitivity to mesopheric temperature variations which was not borne out by observations. The real answer lies in the way in which the atmosphere responds to solar heating with the distribution of absorbing gases in height and latitude being of paramount importance. Tides are now understood in terms of modes, Hough functions with different vertical structures and latitudinal variations. The latter are in essence standing wave patterns between the poles.

The semidiurnal tide is largely excited by ozone absorption around 50 km (BUTLER and SMALL, 1963) and is associated with modes of positive equivalent depth. The fundamental semidiurnal mode, designated (2,2), has, except in the mesophere where it becomes evanescent, a vertical wavelength of about 100 km, much greater than those associated with diurnal modes. The Hough function has a fairly constant amplitude over most latitudes in keeping with the distribution of ozone. Modes with a shorter wavelength can become important higher in the ionosphere because, although they are less strongly excited, their amplitude tends to grow more quickly as they propagate upwards.

LINDZEN (1966) showed that the diurnal tide could not be explained purely in terms of modes of positive equivalent depth and derived from the tidal equations modes of negative equivalent depth, essentially trapped modes inhibited from propagating freely in the vertical direction. This had been suggested by GREEN (1965) as a qualitative explanation of the weak excitation of the diurnal tide. The diurnal tide is now accepted to have as its two main thermal drives absorption by ozone and water vapour (LINDZEN, 1967) but with the short wavelength propagating modes confined to low latitudes and high altitudes. Most of the energy for the diurnal oscillation is trapped at the level of absorption in evanescent modes.

What was once one of the most puzzling aspects of tidal observations can be readily explained in terms of this theory. Northern Hemisphere (55°N) results at 90 km (for example GREENHOW and NEUFELD, 1955) had consistently shown the dominance of the semidiurnal tide whereas results from Adelaide (35°S) often gave much greater prominence to the diurnal tide (ELFORD, 1959). This is because the Hough function associated with the main propagating diurnal mode (1,1) has a significant amplitude at 35° latitude whereas at

55^0 the only diurnal modes of importance are evanescent. However the agreement between tidal theory and observation is not always as satisfactory as this.

1.2.4 Gravity Waves

Gravity waves, introduced to ionospheric physics by HINES (1960), differ from tides in that they are only spasmodically generated and are of sufficiently small scale that Coriolos forces can be ignored. They have been proposed as the major contributor to the deviations from regular and periodic behaviour that ionospheric wind systems exhibit. Gravity waves are a particular class of internal waves confined to a limited frequency range, the upper limit of which is determined by atmospheric parameters and corresponds to a period of about 5 minutes. Unlike the more familiar acoustic waves they have wavelengths dependent both on frequency and direction of phase propagation. In addition the direction of oscillation of the medium in which they travel is almost transverse to the direction of phase propagation. Phase and energy do not propagate in the same direction and for short wavelengths the difference approaches 90^0.

Most suggested sources for gravity waves are tropospheric. One possible generating mechanism would be standing waves associated with topographic features such as mountain ranges whilst another could be the moving interface between air masses of differing density such as would be provided by a tropospheric weather front (GOSSARD, 1962). A mechanism for their generation in thunderstorms has been proposed by PIERCE and CORONITI (1966). This is based on updraughts of air overshooting their new equilibrium level and going into oscillation. CHIMONAS and HINES (1970) considered the possibility of gravity waves being induced by a solar eclipse whilst YUEN, WEAVER, SUZUKI and FURUMOTO (1969) presented evidence for earthquake-induced waves which travel up to heights of 350 km.

The question of propagation of gravity waves is complicated by consideration of factors such as background temperatures and winds (HINES and REDDY, 1967), and of strong shears in the wind (BRETHERTON, 1966). These factors imply restrictions on the spectrum of waves reaching the ionosphere, but despite the filtering there is little doubt that the waves do play an important part in the ionosphere. Both phase height fluctuations in the E-region (VINCENT, 1969, 1972) and wind perturbations in drifting meteor trails (REVAH, 1970) have properties consistent with the amplitude, phase and periods of gravity waves. The tracing of an ionospheric gravity wave effect back to its source near ground level is made more difficult by the fact that the direction of energy propagation of a gravity wave is dependent on its period, with long period waves propagating almost horizontally. Consequently the source and effect may be horizontally separated by thousands of kilometres. The place of gravity waves in an overall picture of ionospheric motions has been reviewed by HINES (1963). It is a matter of some contention as to where the short wavelength end of the gravity wave spectrum merges into turbulence. The two forms of motion are closely related (HODGES, 1967).

1.3. Techniques of Wind Measurement

Many techniques have been developed for the study of winds in the height region 50 - 100 km. The region is beyond even the highest balloons (CONOVER and WENTZIEN, 1955) and well below the height where satellites can be maintained in orbit. The study of winds must therefore depend largely on tracers, either natural or injected by man, which remain long enough in the region to be influenced by the complex local motions. Natural tracers are few, infrequent except in the case of meteors and are beyond the control of the experimenter, whereas man-made tracers are often expensive and necessarily short-lived.

1.3.1 Visual techniques

Visual and photographic tracking of long-enduring visible meteor trails provided the first knowledge of the nature of winds at D-region heights (OLIVIER, 1942; LILLER and WHIPPLE, 1954), giving early evidence of the large velocities and shears that can occur there. The rarity of such phenomena make the method little more than a curiosity. Similar techniques have been more recently applied to the spectacular auroral displays of higher latitudes (KIM and CURRIE, 1958; DAVIS, 1960). Perhaps the most intriguing of all the visual methods is the study of noctilucent clouds (WITT, 1962; FOGLE and HAURWITZ, 1966). These remarkable clouds appear at about 82 km and provide direct evidence of motion and wave activity. Unfortunately they occur rarely and require special viewing conditions: the 82 km region must be sunlit against a dark background and the sky must be free of tropospheric clouds.

The development of sophisticated visual techniques using Fabry-Perot interferometers affords a good opportunity of using airglow (light emission at specific wavelengths by molecules at ionospheric heights) to measure winds, again with the proviso of favourable tropospheric conditions. The author is aware (from work being done at the Mawson Institute, Adelaide) of two ways in which this can be achieved. The first is by measurement of the Doppler frequency shift in a particular direction from which the radial wind velocity may be deduced. The second is based on the fact that the intensity of light received at the ground fluctuates as though irregular conglomerations of light-emitting molecules were drifting overhead. The situation is thus analagous to the ionospheric spaced receiver techniques discussed in the next section if several detectors are used. A similar time-delay analysis can therefore be applied to derive velocities.

1.3.2 Meteor radar

Undoubtedly the technique which has yielded the most wind information is the radio tracking of drifting meteor trails. The method has the advantage of being continuous over the full day and over the height range 70 - 100 km in general. Hence it is able to resolve both periodic components and vertical gradients. It is based on the fact that a meteor travelling through the ionosphere leaves in its wake an ionized trail which, if sufficiently long lasting, will drift with the local wind and Doppler shift the frequency of a radio signal reflected from it. Thus radiometeor systems in general provide a radial velocity component and the point in space at which it was determined. The total wind is therefore incompletely determined from one trail and some spatial and temporal averaging is necessary.

Most meteor systems are one of two types - continuous wave or pulsed. A continuous wave Doppler system was initially developed in California by MANNING, VILLARD and PETERSON (1950) and subsequently SPIZZICHINO, DELCOURT, GIRAUD and REVAH (1965) reported on a similar system at Garchy in France. GREENHOW (1950, 1952) was responsible for the pioneering work using a phase-coherent pulse method at Manchester, England. This system has produced information on turbulence and irregularities (GREEHOW, 1959) as well as a comprehensive wind picture (GREENHOW and NEUFELD, 1955, 1961). The Adelaide radiometeor experiment (ROBERTSON, LIDDY and ELFORD, 1953) uses a combination of continuous and pulsed radio and will be discussed in more detail in Chapter 2. All the above systems use frequencies in the vicinity of 30 MHz and an upper limit of 110 km is set by the diffusive expansion of the meteor trail. The upper limit can be raised considerably by using a lower observing frequency (BROWN, 1973).

1.3.3 Rocket releases

One of the most important types of artificial tracer, namely sodium vapour, was proposed by MANRING et al (1959). When sodium vapour is injected into the atmosphere its interaction with sunlight causes the trail to become luminous and consequently its movement can be resolved by 3-site photography providing the background sky is dark. Despite the fact that this restricts its use to half-hour periods near sunrise and sunset the technique has been extensively used by the initial group (MANRING, BEDINGER and KNAFLICH, 1962) and others. BLAMONT and deJAGER (1961) found strong evidence for a turbopause from sodium vapour trails whilst KOCHANSKI (1964) reported on a series of 25 experiments, largely at Wallops Island, which had enabled estimates to be made of the relative importance of prevailing, tidal and gravity wave contributions over the height range 70 - 190 km.

In order to overcome the twilight restriction applying to sodium vapour trails, techniques were developed for generating lingering self-luminous trails that could be laid at any time during the night (ROSENBERG, GOLOMB and ALLEN, 1963a). It was found that such trails could be produced by the chemical reactions of atomic oxygen with nitric oxide or with vaporised aluminium and its oxides. The release of trimethyl aluminium - TMA - (ROSENBERG, GOLOMB and ALLEN, 1963b) has been particularly favoured.

The idea of gun-launched sensors (BULL, 1964) is an attractive one on economic grounds. It can be used to inject either radar chaff for subsequent tracking or visual vapour trails. The method was used over several nights by MURPHY, BULL and EDWARDS (1966) to lay TMA trails.

Two further rocket techniques that have proved valuable are the 'falling sphere' (BARTMAN, CHANEY, JONES and LIU, 1956) and grenade explosions (STROUD, NORDBERG and WALSH, 1956). The falling sphere can be either a light rigid sphere or an inflatable balloon. Its trajectory is tracked by radar as it falls through the atmosphere. Greatest accuracy is obtained with a slow rate of fall and the method is most useful below 80 km where the atmosphere is sufficiently dense. Winds and temperature between about 40 and 90 km (NORDBERG and SMITH, 1964) can be determined indirectly by the use of grenades. A sequence of grenades is released from an ascending rocket at intervals of several kilometres. Photographic and photoelectric observations are made of the time and location of each explosion. This information, together with the time and direction of arrival of the acoustic waves at the ground as recorded by an array of microphones, is sufficient to determine a wind and temperature profile. It is worth noting that the effect of winds on the refraction of sound waves is the basis of a further technique not connected with rockets- the average behaviour of winds even at D-region heights can sometimes be deduced from the distribution of zones of silence associated with an explosion at ground level. (MURGATROYD, 1957).

1.3.4 Scatter measurements

The frequency spectra of incoherently back-scattered radio signals often show small shifts caused by the macroscopic motion of the ionospheric plasma. Doppler methods can thus be used to arrive at ion velocities. The use of this procedure has been largely confined to E and F region motions and the results have been recently summarised by EVANS (1972). The velocities can be determined to within a few metres/sec over a sounding period of 10 minutes.

IYENGAR (1970) has used fluctuations in forward-scatter amplitudes over three paths to determine, from time shifts of maximum signal incidence, the bulk motion of auroral particles.

1.4. Ionospheric Drifts

The topic of ionospheric drifts is treated separately because there is as yet no consensus of opinion regarding the geophysical significance of the drifts. Some of the objections to a 'wind' interpretation will be dealt with in Chapter 3. The present discussion will be concerned with a description and discussion of the techniques, both experimental and analytical, which have been developed to derive the drift velocities.

The ionosphere is inhomogeneous and irregular and if, while being illuminated from below by radio waves, it varies in time due either to random internal motions or a systematic total drift, the resultant wave fields at the ground will vary. Possible causes of the amplitude fluctuations of reflected radio waves were first discussed by RATCLIFFE and PAWSEY (1953). Subsequently PAWSEY (1935) concluded that horizontal movements in the vicinity of the reflection region were a major cause of fading, after finding that fading was still severe when multiply - reflected waves and interference between magneto-ionic components had been eliminated by using a pulse system and circular polarisation. Various methods have been developed to sample the moving diffraction pattern on the ground, and to relate its parameters to those of the actual reflection region. It has generally been assumed that the diffraction pattern will move with twice the velocity of the reflecting irregularities, although it is possible to formulate models of the reflection process for which such an assumption would appear invalid. Tests by FELGATE (1970) and WRIGHT (1972) indicate that the assumption is justified.

1.4.1 Detailed spatial sampling

One approach to the problem has been to construct large aerial arrays capable of sampling the pattern at a sufficient number of points to render it visible. The direction and magnitude of its motion across the array can then be observed directly. BRIGGS (1968) has shown how the velocity can be derived rigorously from two such samples of the pattern taken a short time interval apart. A 4 x 4 matrix of aerials with spacing 100m was used by MACDOUGALL (1966) whilst HAUBERT and

DOYEN (1966) sampled less of the pattern but in more detail by using a 6 x 6 matrix of aerials separated by only 30m. The largest such arrangement is the Buckland Park array of 89 crossed dipoles filling an approximately circular area of 1 km diameter (BRIGGS et al, 1969). Such direct observations of the pattern have yielded valuable information on the statistical properties of the patterns (FELGATE and GOLLEY, 1971a) as well as indications of rapid velocity changes. However it is a very expensive way of routinely measuring ionospheric drift velocities except in the visual approximation. Furthermore, the receivers used in this experiment are insufficiently sensitive for observation of D-region partial reflections.

1.4.2 Three spaced receivers

The measurement of drift using the minimum number of three aerials was developed by KRAUTKRAMER (1943, 1950) and MITRA (1949). A triangular arrangement of receiving aerials is sufficient to determine the drift from the similar but time-displaced fading observed at each aerial. However the sparseness of the spatial information must be compensated for by a more detailed temporal sampling. The length of fading record chosen then represents a compromise between conflicting requirements: the ideal of a constant velocity over the interval sampled-minimum record length - and the desire for minimum sampling errors - maximum record length - (AWE, 1964a) and hence minimum errors in the estimation of diffraction pattern parameters (AWE, 1964b).

An extension of the spaced receiver experiment which attempts to include vertical movements (GUSEV and MIRKOTAN, 1956; MIRKOTAN, 1958) involves the use of two closely spaced frequencies. They are reflected from slightly different heights and spatial information is thus obtained in three dimensions. This technique has been developed in its most sophisticated form by WRIGHT and FEDOR (1969). A four-dimensional (time and space) least squares correlation analysis appropriate to the technique was reported by FEDOR (1967). The interpretation of the results so obtained in terms of horizontal and vertical motions is however open to criticism on several grounds (BRIGGS, 1972), and for the most part the assumption of the simpler technique that the wind field is purely horizontal is a valid one, at least for D-region heights. Higher in the thermosphere vertical velocities may become quite significant (REES, 1969) and the assumption is probably unjustified there.

1.4.3 Methods of analysis

Several methods of varying complexity have been used to derive velocities from the spaced receiver fading records.

The fluctuations in the received signal amplitude are due to a combination of two factors - an overall steady motion of the diffraction pattern past the receivers, and random changes in form of the pattern as it moves. The simplest approach (MITRA, 1949) is to ignore these internal changes and attribute all the fading to the average horizontal motion. The velocity can then be calculated from the time displacements for maximum similarity of the fading records and the respective spacings of the receiving locations. Velocities calculated by this 'similar fades' method are termed 'apparent'.

The technique of 'full correlation analysis' (FCA) developed by BRIGGS, PHILLIPS and SHINN (1950) and extended to anisometric patterns by PHILLIPS and SPENCER (1955) attempts to allow for both the contributions to the fading in deriving the velocity. An additional parameter, designated Vc, and known as the characteristic velocity, is derived from the spatial and temporal correlation functions and this, together with the geometric form of the diffraction pattern, is used to produce a modified velocity, known as the 'true' velocity. Many modifications of FCA have been proposed, some aimed at simplification (for example YERG, 1955) or improvement (CHAPPELL and HENDERSON, 1956) and others to introduce such refinements as statistical weighting (SALES and BOWHILL, 1962)

The assumptions of FCA have been widely questioned. The analysis assumes the temporal and spatial correlation functions to be identical whereas KELLEHER (1966) found that they are often different in form, particularly when Vc is large. BROWN and CHAPMAN (1972) concluded that the shapes of the functions were similar but the spatial correlation was generally underestimated, this effect being most marked for small aerial spacings. The dependence of drift parameters on the triangle size is well established (KELLEHER, 1966; BEYNON and WRIGHT, 1969; GOLLEY and ROSSITER, 1970, 1971; SASTRI and RAO, 1971) although the effect is not well understood. ROSSITER (1970b) showed that the effect could not be explained by aerial coupling alone but FEDOR and PLYWASKI (1972) have subsequently queried the sensitivity of his experiment and shown that a coupling of 20 db is sufficient to produce an error of 50% in the velocity. Whatever the cause of the effect, it has been shown that it can be eliminated by appropriate choice of triangle size, the optimum size being different for total (GOLLEY and ROSSITER, 1970) and partial (GOLLEY and ROSSITER, 1971) reflections.

The assumption of statistically stationary data may frequently be invalid (FELGATE and GOLLEY, 1971b). Large fluctuations in E-region velocities over a short period were observed by MACDOUGALL (1966). FELGATE and GOLLEY (1971a) have suggested that a varying velocity could introduce an apparent positive "dispersion" into the records. Records are said to exhibit dispersion if the drift velocity is dependent on fading frequency; the dispersion is positive if the velocity increases with increasing frequency. JONES and MAUDE (1965) found 94% of their records to be dispersive, a fact they attributed to wave motions at the reflection level. Hower MCGEE (1966) and GOLLEY and ROSSITER (1971) found the majority of Adelaide and Cambridge drift records failed to yield any evidence of dispersion.

Correlation analysis has at least been favourably served by computer modelling tests. PITTEWAY, WRIGHT and FEDOR (1971) used a model of random point scatterers whereas GUHA and GELLER (1973) assumed a one-dimensional perfectly reflecting wavy ionosphere. In each case correlation analysis was able to determine the overall motion of the ionosphere from an analysis of the ground pattern.

Due to its relative simplicity, and hence economy, the 'similar fades' method has been adopted by many of the groups observing drifts on a routine basis. One of the main purposes of a worldwide network of drift stations will be the study of variations with geographic and geomagnetic location and it is therefore unfortunate that no uniformity presently exists in the treatment of drift data. SPRENGER and SCHMINDER (1969) investigated the relationship between 'apparent' and 'true' velocities and found the quality of agreement depended on the form of 'similar fades' analysis used, the averaging of individual drift vectors being preferable to the averaging of time delays. In general it appears that 'apparent' velocities are consistently too high whilst the 'true' velocities of FCA underestimate the drift unless the receiving triangle is sufficiently large.

1.4.4 Partial reflection drifts

Most observations of ionospheric drifts have used total reflections from the E and F regions. The weak backscatter from D-region irregularities has only recently been utilised for drift work. (AWE, 1961; FRASER, 1965, 1968; FRASER and KOCHANSKI, 1970; ROSSITER, 1970a; MANSON, GREGORY and STEPHENSON, 1973). The use of partial reflections has the advantage that the height of observation can be largely controlled by the experimenter, whereas in the case of total reflections the observation height is dependent on the radio frequency used and on ionospheric conditions. This height can vary quite markedly and the study of periodic motions is made hazardous by the fact that records obtained early in the day, for example, may refer to a height some 20 km different from those recorded at midday. The control of observation height also makes partial reflection drifts eminently suitable for comparisons with accepted neutral wind measuring techniques. These two aspects of time variation and comparison in a selected height interval form the bulk of the work described in this thesis.

CHAPTER TWO

EQUIPMENT AND EXPERIMENTAL PROCEDURES

The basic requirements of a partial reflection experiment are a relatively noise-free site, a powerful transmitter and sensitive receiving equipment. Two such sites will be described in this chapter : the well-established Buckland Park station near Adelaide (BRIGGS et al, 1969) and a new experimental site near Woomera. The basic frequency of operation in each case is 1.98 MHz. The geographical positions of the two sites are shown in FIGURE 2.1.

2.1 BUCKLAND PARK

2.1.1 Transmission

The transmitting array consists of four centre-fed folded half-wave dipoles arranged in the form of a square. Opposite pairs of aerials are joined in parallel. The sense of polarisation can be changed between the linear, and left and right-handed circular modes by changing the relative phases of the currents in the two pairs of dipoles. The transmitter has a peak power of 25 kw and produces Gaussian-shaped pulses of between 25 and 200 μsec duration at a repetition frequency of 25 or 50 Hz.

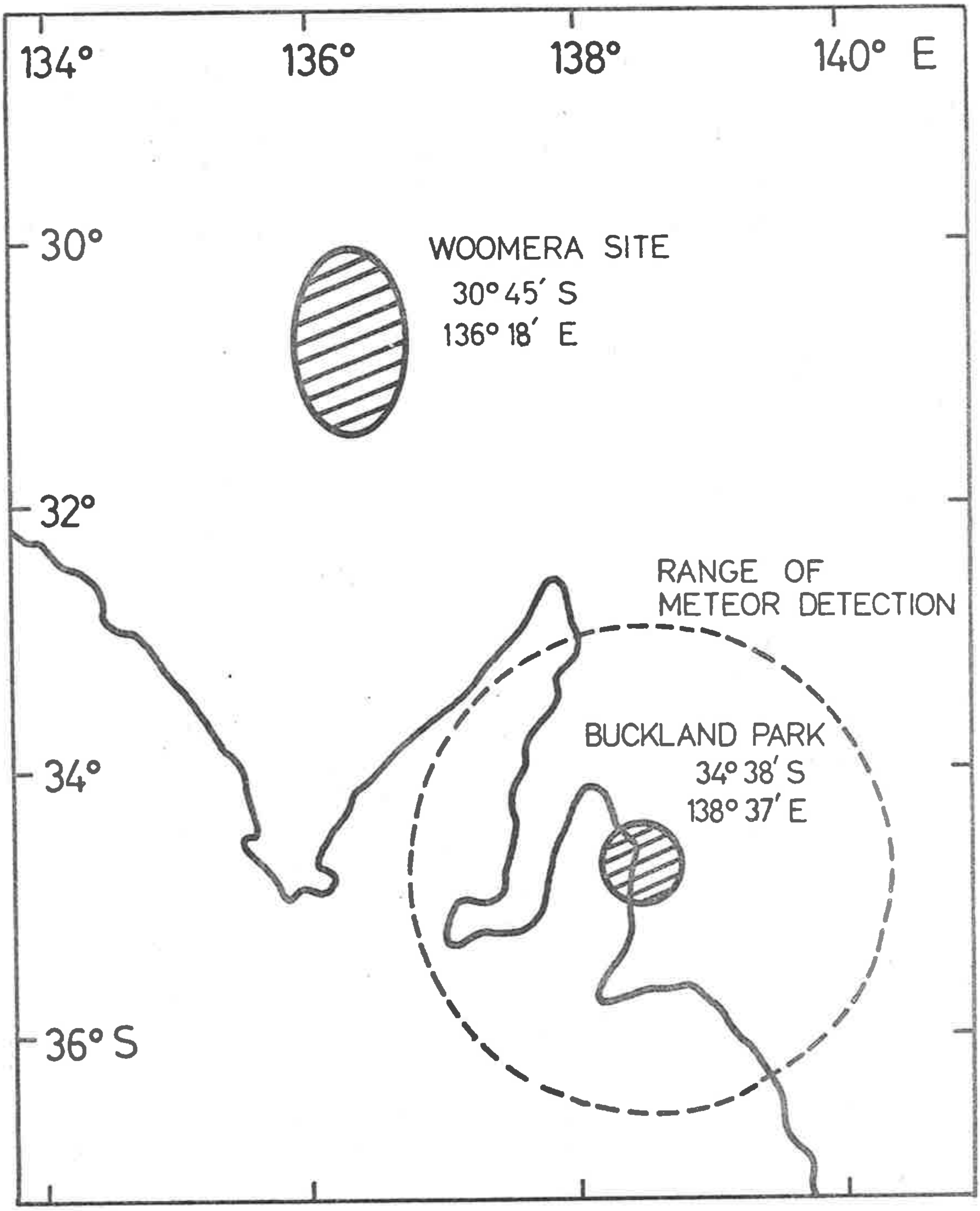


FIGURE 2.1 *The areas of sky at 85 km associated with the receiving systems for the Adelaide and Woomera drift experiments (shaded) and the Adelaide meteor-radar experiment (enclosed by the broken circle).*

The height resolution attainable using a pulse radar system is inversely proportional to the pulse width employed, a resolution of 4 km requiring a pulsewidth of about 30 μsec. A much narrower pulse would give improved resolution but would require a wider bandwidth receiver accepting noise over a wider frequency range. In general, a pulse width of between 25 and 35 μsec was employed. Occasionally at times of low absorption or intense ionospheric focussing, a repetition rate of 50 Hz is impractical because the echoes from a particular pulse are interfered with by multiply-reflected E or F-region echoes from the previous pulse. In such cases the lower pulse rate of 25 Hz must be accepted to allow time for the attenuation of the totally reflected radiation. Although the extraordinary wave is preferentially reflected from discontinuities in electron concentration, it is also highly absorbed during the day. Thus ordinary polarisation was generally used during the day and extraordinary at night. The use of linear polarisation is undesirable because of mutual interference between the two magnetoionic components (PHILLIPS, 1951).

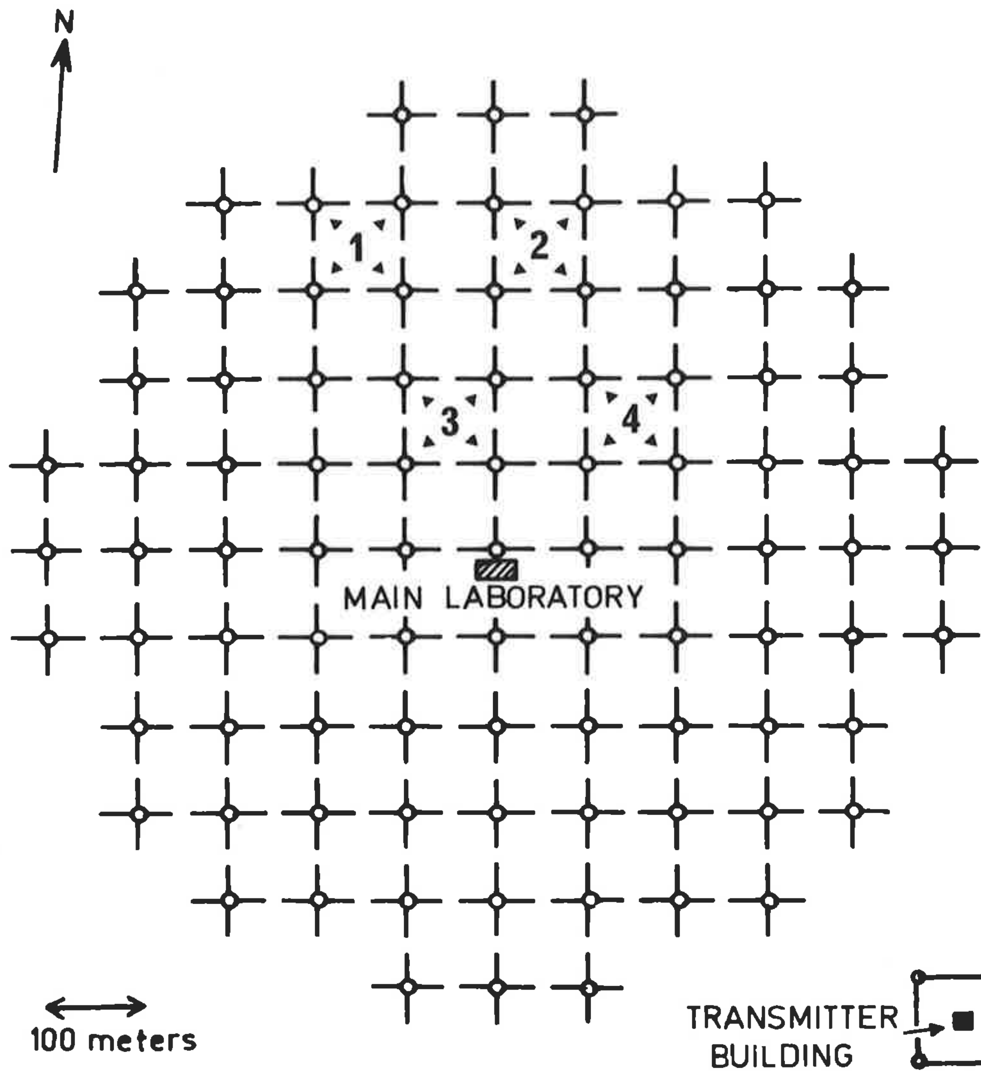
2.1.2 Reception

The Buckland Park array consists of 89 orthogonally crossed half-wave dipoles (spacing 91m) covering an area approximately 1 km in diameter. Each dipole is suspended some 11m above the ground and linked to the central hut by cable cut to an integral number of half wavelengths. The receiving aerials were selected from the array as indicated in FIGURE 2.2. Two features worthy of note are the grouping of four aerials connected in parallel to form each receiving "location", and the choice of two approximately equilateral receiving triangles. The use of four dipoles in parallel, and consequent narrowing of beamwidth, has the twofold advantage of increasing the signal-to-noise ratio and decreasing the response to off-vertical echoes. FIGURE 2.3 compares the computed polar response for four dipoles in parallel above an assumed perfectly conducting ground with the response for a single dipole. The most striking feature is the reduction in beamwidth in the plane perpendicular to the plane of the dipole. The advantages of an equilateral configuration have been indicated both theoretically (BARBER, 1957) and experimentally (GOLLEY and ROSSITER, 1970). The effective spacings between the group 'centres' in the two nearly equilateral receiving triangles (123 and 234 in FIGURE 2.2) are 204, 204 and 182 metres. This is sufficient to minimise the 'triangle size effect' without being so large that the fading at adjacent aerials is poorly correlated.

Valve receivers with a gain of 120 db and a bandwidth of 60 KHz were used. The height of observation was selected by

FIGURE 2.2

The aerial layout at Buckland Park showing the groupings used for the spaced receiver drift experiment.


N
1
2
3
4
MAIN LABORATORY
100 meters
TRANSMITTER
BUILDING

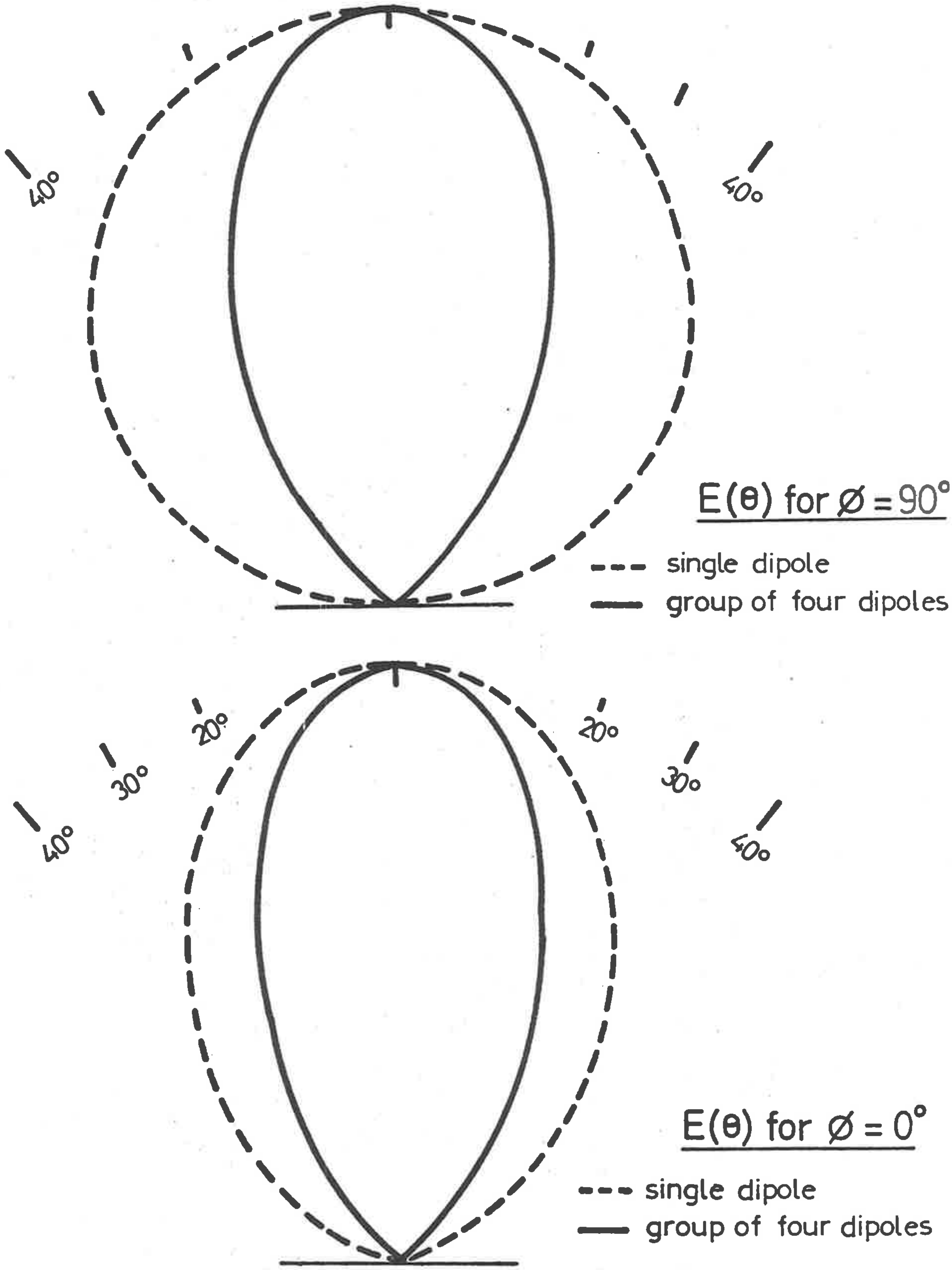


FIGURE 2.3 *The polar response of a single dipole and a group of four dipoles in the planes perpendicular to (ϕ = 90°) and in the direction of (ϕ = 0°) the dipoles.*

a range-scanning gate; the gating pulse was automatically stepped to a new height interval with each transmitter pulse, restarting at the lowest height when the chosen cycle was complete. This system enabled virtually simultaneous sounding of a number of heights, the number being limited primarily by the dynamic range of the receivers. During 1972 a 5 km gate width was used and cycle of 5 heights, which meant that for a transmitter repetition rate of 50 Hz the fading at each height was sampled at a frequency of 10 Hz. Generally in 1973 a cycle of 10 heights was sampled with a 2 km gate width for better vertical resolution in the study of waves and for rocket comparisons.

The receiver outputs were digitized to 64 levels and recorded on 7-channel magnetic tape for computer analysis. The length of fading record chosen for each velocity determination was 3 minutes - a compromise between the need for an adequate number of fading cycles to reduce statistical sampling errors and the desire for effectively instantaneous velocity determination.

2.2 WOOMERA

2.2.1 Transmission

The transmitting array constructed at Woomera is similar to that used at Buckland Park, with four centre-fed half-wave folded dipoles supported by steel towers 30m high

FIGURE 2.4

The upper section of one of the transmitting aerial towers showing the ends of two of the spaced dipoles.

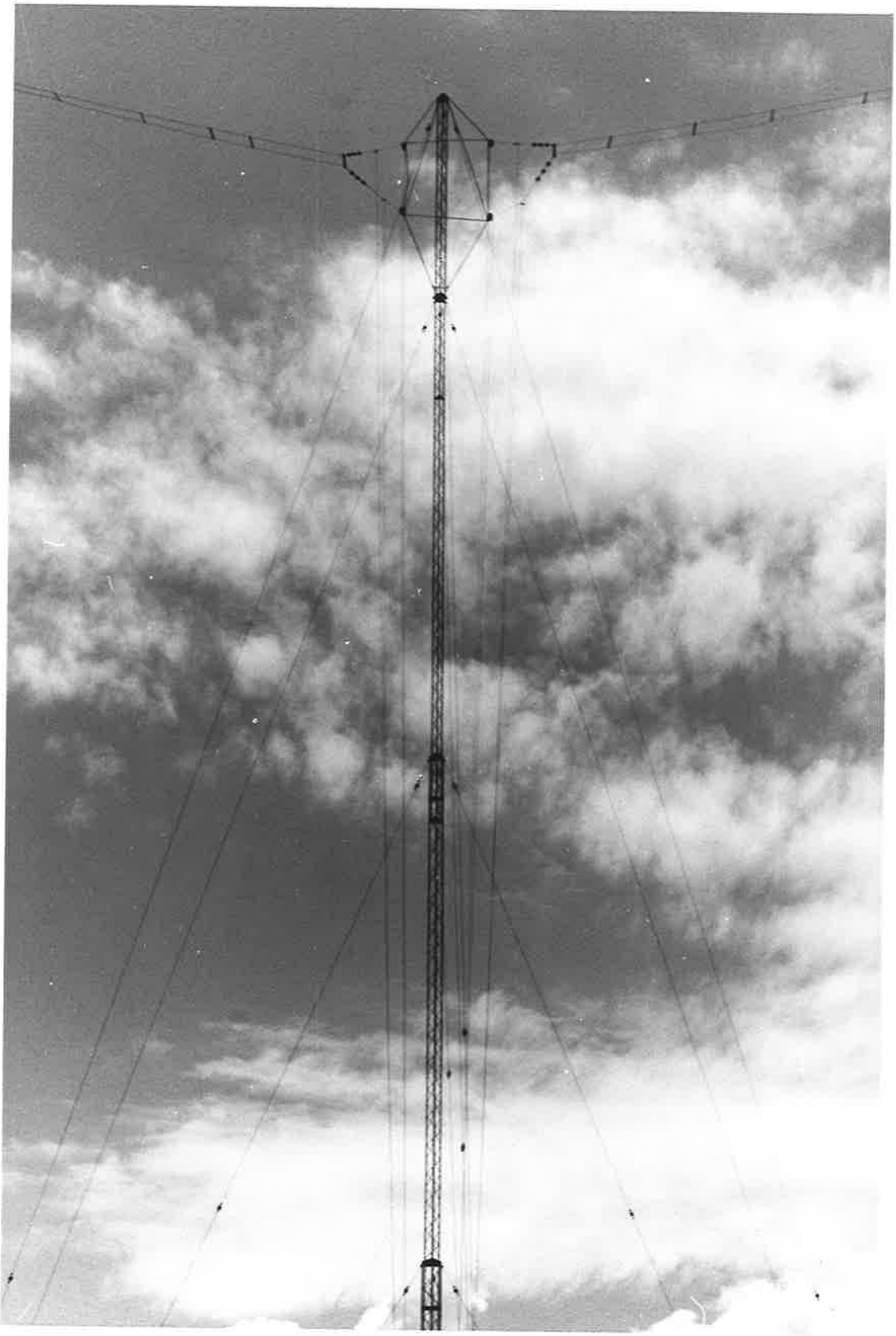


FIGURE 2.5

The four towers supporting the four transmitting dipoles.

The feeds to the centre of each dipole are just visible.

(FIGURE 2.4) and forming the sides of a square some 75m in scale (FIGURE 2.5). Each aerial is connected to a central control box by half-wavelengths of spaced open wire transmission line. The impedance of each feeder/dipole combination is resistive and close to 400Ω. A long spaced line, supported 1.4m above the ground by wooden posts, connects the control box to the transmitter at the central laboratory. This line is matched, at their respective ends, to the transmitter and to the aerial combination to avoid the possibility of standing waves and ensure maximum efficiency. The control box houses vacuum relays for rapid polarisation change. Circular polarisation is achieved by introducing an extra quarter-wavelength of cable in the feed to one pair of opposite dipoles.

The present Woomera transmitter has a peak power of 10 kw and is generally used to transmit pulses of width 30μsec at a frequency of 50 Hz. Again the pulses are Gaussian in shape to minimize interference to the adjacent commercial radio band whilst maintaining an acceptable degree of narrowness.

2.2.2 Reception

The nature of the receiving site and economic constraints dictated that the Woomera receiving arrangement be, at least initially, the simplest possible. Four half-wave dipoles are situated at the corners of a square of side 180m where they are supported 11m above the ground (FIGURE 2.6). Each dipole is connected to the central laboratory by polythene sheathed coaxial cable with a characteristic impedance of 75Ω. To ensure maximum power transfer, the cable is matched to the dipoles via a transformer and series tuned circuit housed in a small box at the top of each central support. Trenches were dug to protect the cable.

The receiving and digitizing equipment is pictured in FIGURE 2.7. It is similar to that described for Buckland Park, but more versatile and compact. The four receivers (lowest unit) use semiconductors rather than valves and the gains can be adjusted at both the IF and RF stages. The local oscillator can be used in a continuous or pulsed mode. The position of the pulse and the output of each receiver can be monitored. The width and duration of the pulse can then be adjusted to suppress all but the echoes of interest. The range scanning gate (third unit down) can be triggered from the 50 Hz mains, externally as required, or internally at 1, 5, 25 or 50 Hz. The starting height for the scan can be adjusted continuously using the 2 km range markers as a guide or by using the fixed delays calibrated for 40, 50, 60, 70, 80, 90 and 110 km. The number of height steps in a scan can be 4,5,10,20,40 or 100 and the width of the gate can be 2,5 or 10 km. The second unit from the top contains the crystal

FIGURE 2.6

One of the four receiving aerials.

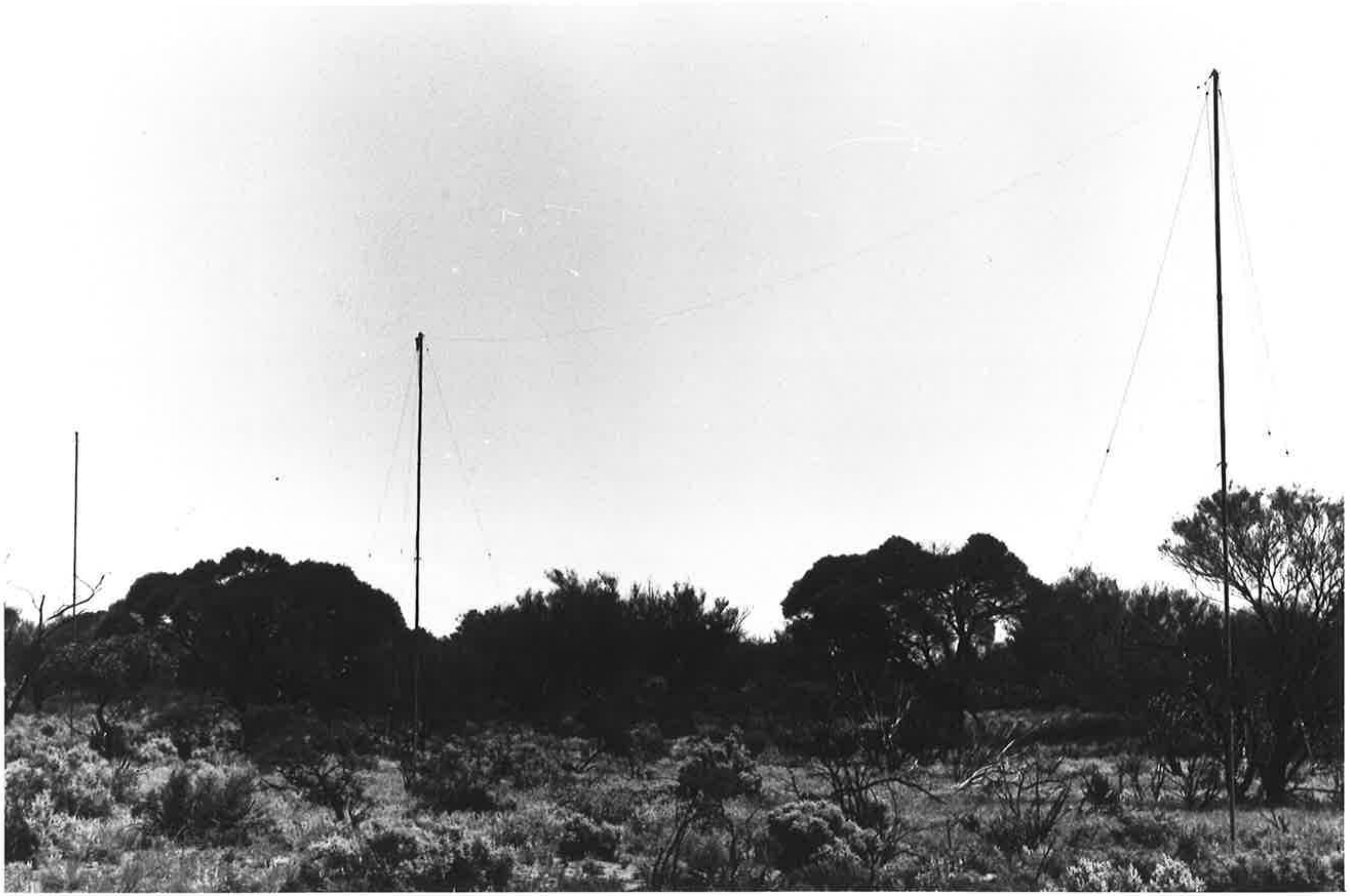
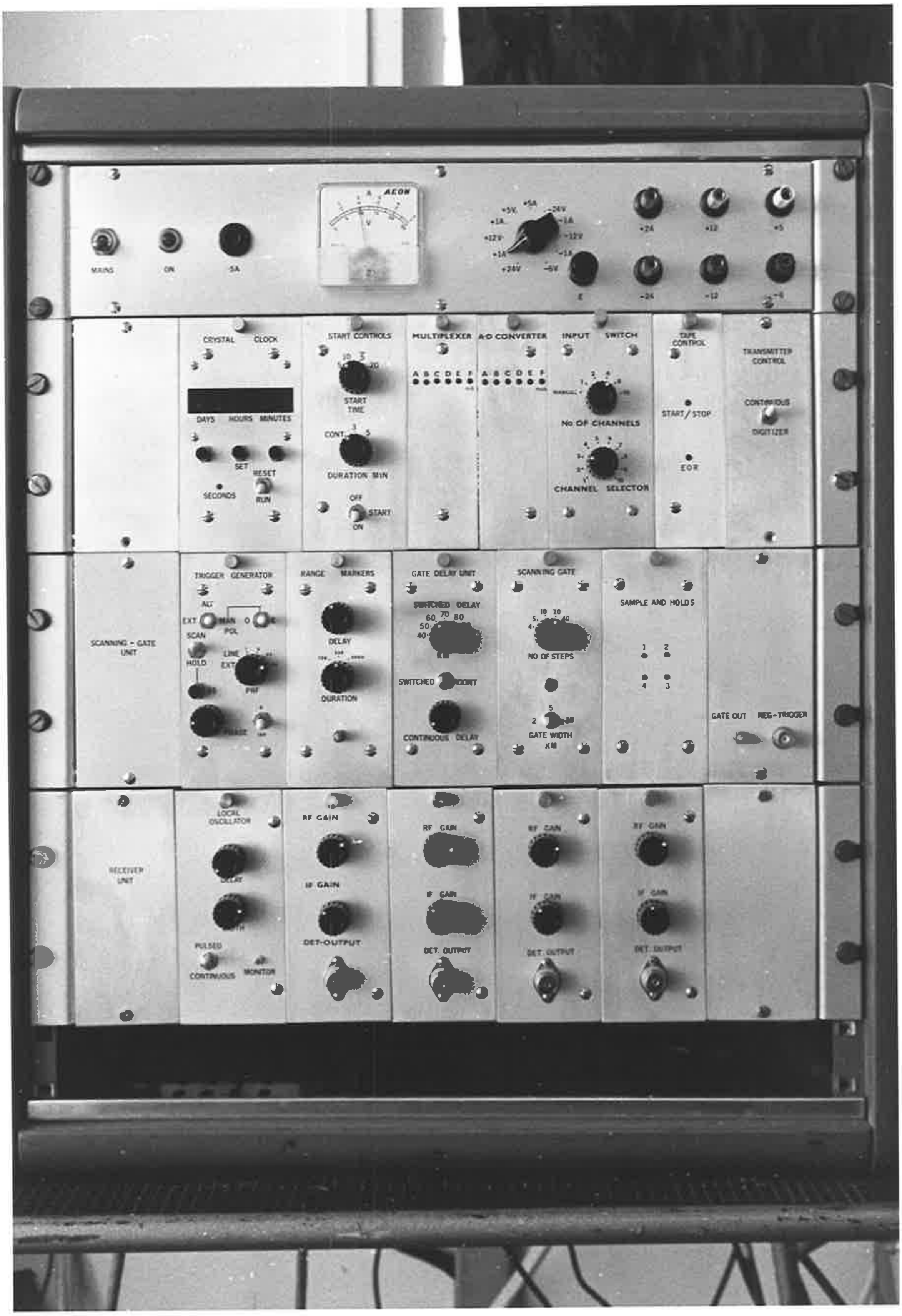


FIGURE 2.7

The receiving and digitizing equipment.


AEON
MAINS
ON
5A
CRYSTAL CLOCK
DAYS HOURS MINUTES
SET
RESET
RUN
SECONDS
START CONTROLS
START TIME
DURATION MIN
OFF
START
ON
MULTIPLEXER
A-D CONVERTER
INPUT SWITCH
No OF CHANNELS
CHANNEL SELECTOR
TAPE CONTROL
START/STOP
EOR
TRANSMITTER CONTROL
CONTINUOUS
DIGITIZER
SCANNING - GATE UNIT
TRIGGER GENERATOR
RANGE MARKERS
DELAY
DURATION
GATE DELAY UNIT
SWITCHED DELAY
CONTINUOUS DELAY
SCANNING GATE
NO OF STEPS
GATE WIDTH KM
SAMPLE AND HOLDS
GATE OUT
REG-TRIGGER
RECEIVER UNIT
LOCAL OSCILLATOR
PULSED
CONTINUOUS
MONITOR
RF GAIN
IF GAIN
DET-OUTPUT
DET. OUTPUT

clock, digitizing and tape control equipment. The clock displays days, hours and minutes and this information is written on the tape each 10 seconds. 1, 2, 4, 8 or 10 channels of data can be digitized and recorded. Only 4 are required for the echo amplitude information but the additional channels may be used in future for such additional parameters as phase. The tape can be run continuously or for intervals of 3 or 5 minutes and recording can be arranged to start on multiples of 5, 10 or 20 minutes. It is possible to start on the odd 5 minutes (for example 1215, 1225, 1255 ...). This is to enable both Buckland Park and Woomera to transmit for 3 minutes every 10 minutes but without mutual interference, which can sometimes occur when multiple-hop echoes are present at night.

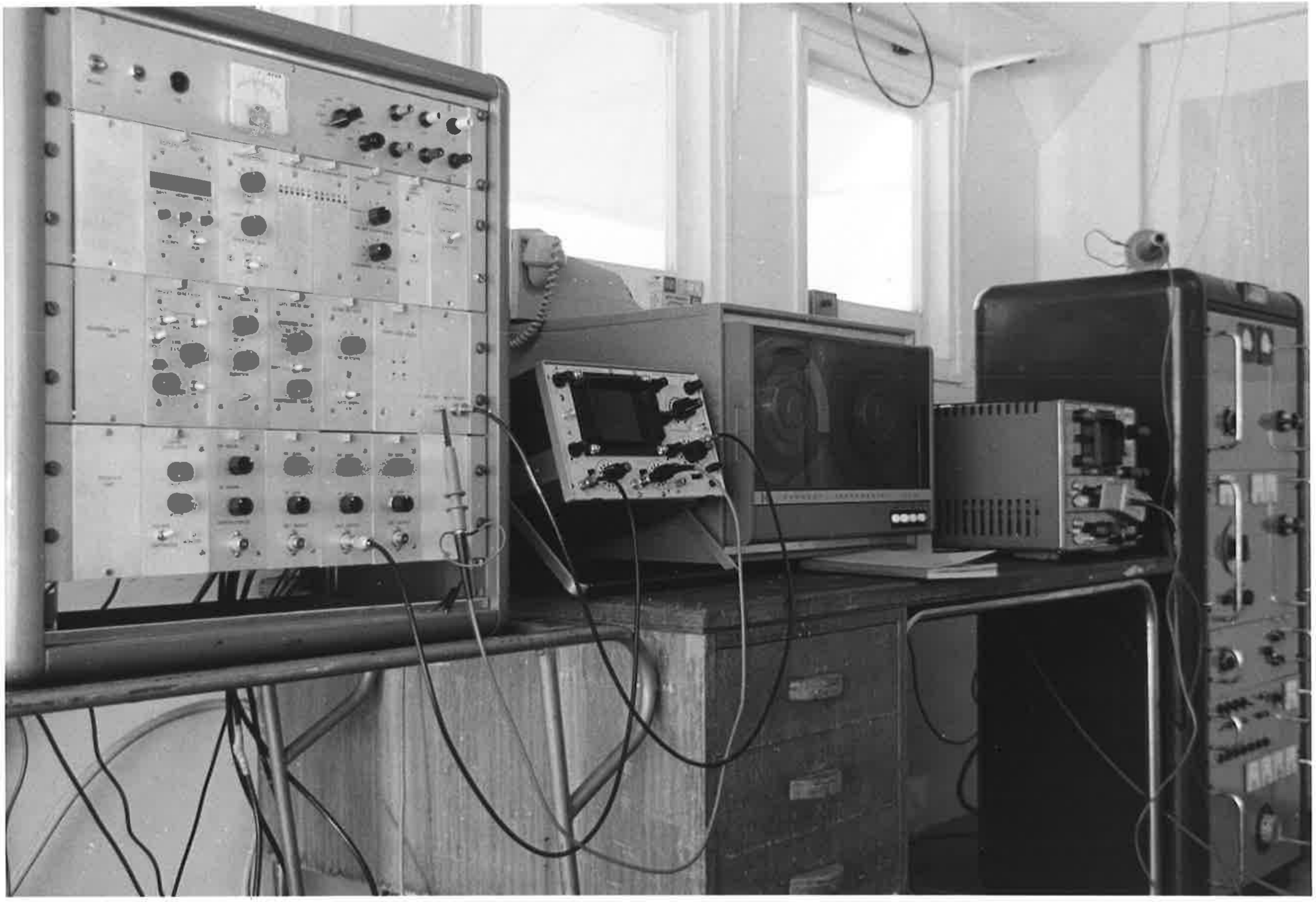
FIGURE 2.8 shows the arrangement within the central laboratory. From the left are the receiving equipment just described, an oscilloscope to monitor the gate position and receiver outputs, the tape recorder, a further oscilloscope to display the transmitter pulse and, on the extreme right, the transmitter itself. The tape recorder is a Kennedy incremental recorder used at a density of 556 bits per inch.

2.3 Analysis

Each tape recorded at Buckland Park or Woomera consisted

FIGURE 2.8

The arrangement in the central laboratory at the Woomera site.



of a series of 3-minute records. Each record in turn contained information from a number of heights, the sampling frequency for each height being dependent on the number of heights scanned and the transmitter repetition rate. For consistency, the data at each height was averaged, if necessary, so that there was one data point every 2/5 sec. Each record for analysis then consisted of 450 data points with digital levels between 0 and 63. The records were then analysed using the form of full correlation analysis outlined by BRIGGS (1967) and adapted for computer use by FOOKS (1965).

The following information was included in the computer print-out for each record analysed:

- the day, time and height interval
- the mean and standard deviation of the amplitude for each fading record
- the six auto - and cross-correlation functions computed at intervals of 2/5 sec for 25 positive and negative shifts
- the maximum value of each cross-correlation function and the time shift associated with it
- the magnitude and direction of the apparent velocity
- the parameters of the correlation eclipse; namely the major axis, axial ratio and the

ellipse orientation

- the magnitude and direction of the true velocity

the value of the random change parameter Vc

This information was scanned before accepting any velocity determination. Records were rejected if they failed to satisfy certain criteria:

(i) The mean of the fading had to lie between 4 and 60 for each receiver. A mean signal amplitude outside these limits was taken to indicate unduly truncated fading with the receiver output either zero or saturating for too much of the record.

(ii) The maximum value of cross-correlation for all three pairs was required to be in excess of 0.2

(iii) Records had to be free from significant high frequency noise. This proved relatively easy to detect, because when it was present the auto-correlation functions failed to extrapolate smoothly to unity at zero time shift.
The cross-correlation functions also tended to peak spuriously at zero lag, usually resulting in negligible time shifts and unreasonably large apparent velocities.

(iv) Oscillatory correlation functions were not accepted because of the impossibility of unambiguously determining the time shifts between receiving sites.

(v) Records were rejected if any of the cross-correlation functions had not reached a maximum or any of the auto-correlation functions had not fallen to .5 within the number of shifts chosen for the standard analysis i.e. a period of 10 seconds.

(vi) Occasionally the square of the random velocity, Vc, was negative and in such cases the true velocity was regarded as unreliable.

2.4 NATURE OF RESULTS

Few results have yet been obtained using the Woomera drift equipment which was only completed late in 1973. On the other hand, this thesis contains the results of many thousands of drift determinations from Buckland Park. Throughout 1972 records were taken on an average of 7 or 8 days a month. Where possible, they were taken every 10 or 20 minutes throughout the full 24 hours, with the aim of looking for evidence of tidal influence in the drift results. The receiving equipment was continuously monitored and the receiver gains adjusted to give maximum continuity of data from the 85 - 95 km height range. This choice was largely governed by the prevailing echo structure. During the day the total reflection echo from the E-region is usually centred close to 100 km, with D-region partial reflections returned from between 60 and 95 km. At night 2 MHz radio waves are totally reflected from the F-region near 230 km, and quite strong partial reflections nearly always occur between 80 and 100 km.

For routine analysis the records for only one of the possible receiving triangles were analysed. However comparison of simultaneous values from two triangles gives a useful indication of the uncertainty inherent in each drift result, even though the two triangles are not completely independent. Throughout the thesis the drift magnitudes quoted are half the magnitude of the diffraction pattern velocity across the ground. Except where otherwise stated, directions are measured clockwise from geographic North and represent the direction to which the ionosphere appears to be drifting. When components are considered, the ambiguous words 'easterly' and 'westerly' will not appear. Instead the suffix '-ward' will be used. TABLE 2-1 lists for both triangles the magnitude and direction of apparent and true

TABLE 2.1

APPARENT VELOCITY								TRUE VELOCITY							
Triangle 1		Triangle 2		Triangle 1		Triangle 2		Triangle 1		Triangle 2		Triangle 1		Triangle 2	
Mag	Dir	Mag	Dir	Mag	Dir	Mag	Dir	Mag	Dir	Mag	Dir	Mag	Dir	Mag	Dir
65	130	60	127	78	238	79	237	46	144	39	133	55	258	36	231
63	100	58	102	118	215	116	226	53	123	45	128	69	207	69	216
51	282	69	275	86	149	96	159	45	260	47	272	79	144	65	171
74	142	79	145	99	246	119	215	57	151	60	156	60	234	60	226
83	286	71	298	85	159	90	163	60	275	61	309	67	153	67	166
114	300	80	285	120	230	107	254	38	272	61	289	45	226	51	259
64	103	78	121	93	232	78	253	38	115	40	113	54	247	61	273
52	91	31	100	98	123	91	139	47	98	24	107	77	126	58	136
63	118	65	126	91	92	79	98	43	123	45	130	58	95	43	113
100	280	96	298	87	141	99	142	59	261	66	291	69	133	75	134
60	133	54	138	121	237	80	202	35	122	43	111	43	243	65	215
66	163	64	161	90	128	90	129	22	154	20	132	72	126	74	133
84	298	95	280	95	121	92	116	75	276	78	282	70	119	70	133
102	281	98	286	62	35	63	55	65	270	76	287	49	70	62	88
45	153	46	156	95	111	81	110	42	131	36	138	65	104	65	119
158	265	110	281	91	73	82	80	49	252	79	294	75	86	71	89
61	145	65	149	68	225	56	239	32	147	38	146	50	254	24	251
101	296	116	289	71	143	67	141	68	271	65	305	55	150	57	157
71	169	61	163	86	130	64	122	42	162	45	164	51	116	47	144
52	161	59	154	65	260	66	238	32	178	32	154	42	251	41	234
57	190	64	186	72	231	86	233	36	182	43	168	23	272	15	267
58	151	73	158	63	138	61	137	40	149	53	152	61	133	51	149
112	201	100	207	58	231	50	251	49	196	57	185	42	244	47	278
82	214	65	264	71	154	74	124	55	213	31	235	51	149	55	144
109	163	97	157	50	81	40	86	74	167	65	167	44	84	27	73
57	259	74	254	61	197	71	207	37	278	43	283	37	215	28	215
76	158	83	170	59	202	59	220	68	146	66	152	29	212	25	226
56	191	56	195	79	123	79	124	37	189	41	177	44	135	45	123
72	164	77	170	68	202	70	202	68	146	76	150	44	200	48	216
94	160	79	151	55	157	63	169	51	165	52	157	54	133	52	153
87	160	92	159	72	208	71	218	59	163	59	150	33	222	29	221
86	146	64	134	59	125	61	107	63	133	58	142	39	123	40	107
82	164	82	163	43	72	46	68	72	163	73	160	41	87	45	80
54	257	56	253	46	125	49	124	50	274	44	279	43	128	46	138
110	161	112	165	47	200	47	197	88	168	78	159	41	204	39	198
69	197	117	210	39	177	45	210	55	213	34	221	32	201	30	193
82	198	83	211	43	114	48	115	56	200	60	195	38	105	34	117

velocity obtained on January 23, 1972 for heights between 80 and 95 km. Deviations were calculated in the following way: the value of 'true' velocity magnitude obtained using one triangle was subtracted from the corresponding value for the other triangle and this difference was squared. The 74 squared differences were then averaged to obtain the root mean square velocity difference. The other three parameters were treated similarly. The resultant r.m.s. variations are:

apparent velocity magnitude	-	±7 m/sec
apparent velocity direction	-	$\pm 7^0$
true velocity magnitude	-	±5 m/sec
true velocity direction	-	$\pm 8^0$

Uncertainties of the same order have been reported by HARNISHMACHER (1968) for total reflections from the E-region and by HAUG and HOLT (1968) for D-region partial reflections.

An even better indication of the reliability of individual drift values is provided by a time sequence of values at a specific height. The greatest number of drift readings obtained for one height interval in a day was on December 12, 1972 at 85 - 90 km when 89 records satisfied the acceptance criteria. These results are plotted in FIGURE 2.9. It must be remembered that no values are rejected provided the records satisfy the criteria previously mentioned. The computer printout for each record is scanned independently and is accepted irrespective of whether the computed drift agrees with surrounding determinations. The small residual scatter in FIGURE 2.9 is therefore indicative of the reliability of the drift technique rather than the selection process.

2.5 The Adelaide Meteor System

Observations of drifting meteor trails have been made at Adelaide since 1952. The system has undergone many changes in the intervening period but remains the same in concept - a combined continuous wave and pulse technique. The most recent descriptions have been given by ROPER (1965) and ELFORD (1968) and only the main features will be mentioned here. The equipment basically consists of two transmitters, continuous wave and pulse, at Adelaide and a main receiving station at St. Kilda, 25 km to the north of the transmitters.

The first parameter fundamental to meteor wind measurement is the line-of-sight drift velocity of individual meteor trails. This is obtained by measuring the rate of

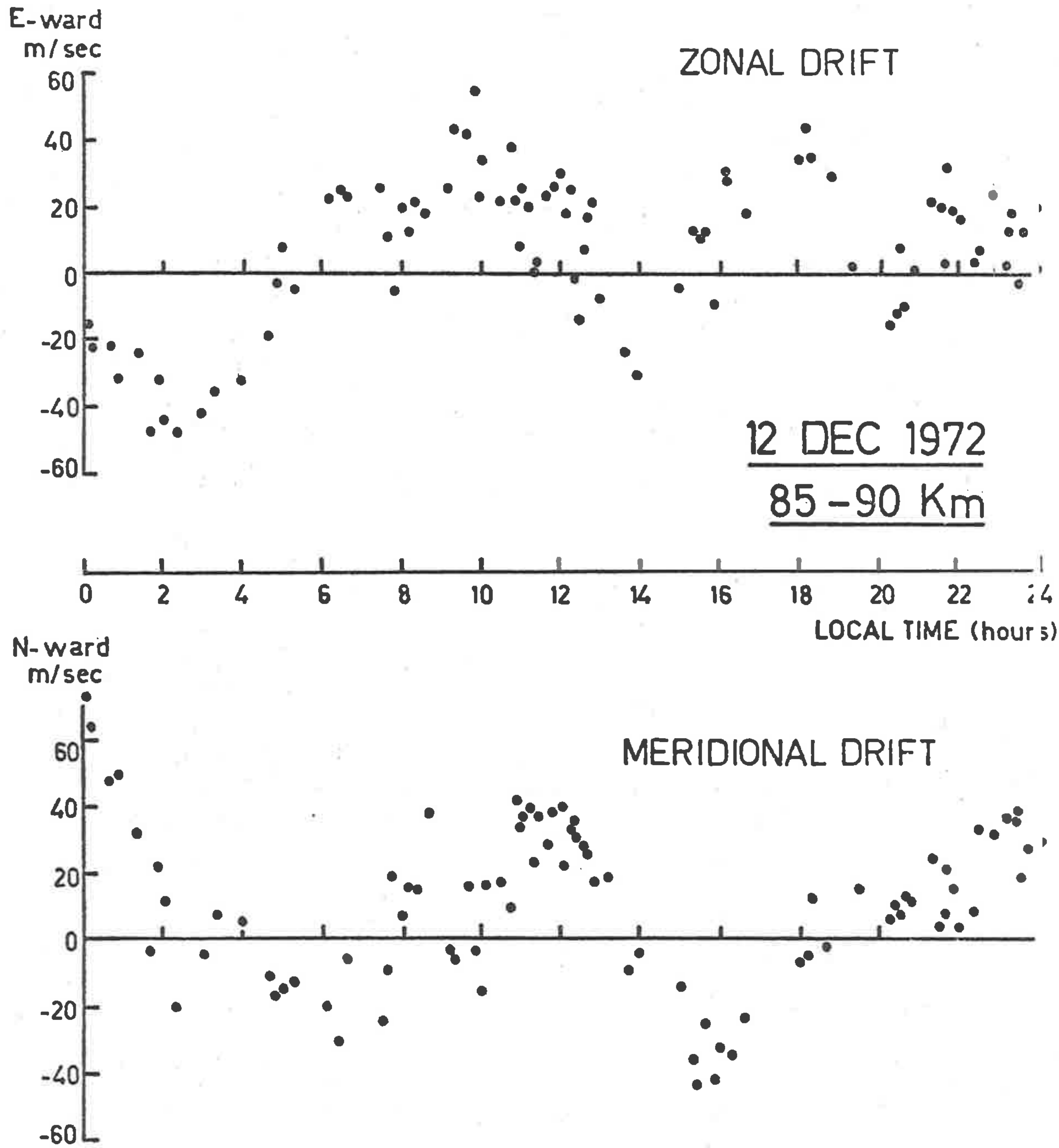


FIGURE 2.9 *The individual drift results at 85 - 90 km on December 12, 1972.*

increase of the phase path of the sky wave, its phase being compared with the reference phase provided by the ground wave. The sense of the Doppler frequency shift is determined from a sawtooth modulation of the continuous wave signal. The second parameter measured is the location of the reflection point. Its direction is found by a comparison of the phases of the sky waves at five spaced receiving aerials at St. Kilda. A pulsed transmission is used to determine the slant range of the echoing point. The difference between the arrival times of a ground pulse and the associated sky wave pulse is measured and an appropriate correction made for the separation of transmitting and receiving sites. The system accepts meteor echoes from all azimuths and from elevations within 60^0 of the zenith so echoes are received from an area some 200 km in radius centred on the receiving site.

Because only the line-of-sight component of the velocity of any measured trail is known, it would be necessary to use at least two separated reflection points occurring simultaneously at the same height to fully specify the instantaneous wind. This is rarely feasible and in practice a model is fitted to the data using a least-squares analysis developed by GROVES (1959). The model incorporates a periodic variation in time, with periods of 24, 12 and 8 hours, and a polynomial variation in height of the amplitude and phase of the tidal components and the amplitude and direction of the prevailing component. Generally second, third and fourth order polynomials are used, which necessarily results in a considerable smoothing of the wind profiles and their temporal variation.

CHAPTER THREE

A COMPARISON OF IONOSPHERIC AND METEOR TRAIL DRIFTS

3.1 Interpretation of Ionospheric Drifts

One problem which must be considered when discussing ionospheric drifts is the extent to which layers or clouds of charged particles follow the motion of the neutral air. Where the gyrofrequency of the electrons and ions greatly exceeds the collision frequency with neutrals, electromagnetic effects become dominant and the ionisation can move quite independently of the neutral wind. This is true of the upper E-region and particularly of the F-region. However at D-region heights electromagnetic effects are expected to be of little importance (RATCLIFFE, 1959) and it might be deduced that irregularities in the D-region would be good sensors of neutral air motion. This view has been widely questioned.

The mechanism by which the irregularities are formed is of great importance. If the irregularities are turbulence-induced then they will probably move with the background wind as it is generally expected that turbulence is carried along by the ambient wind (HINES, 1972). This is a feasible mechanism in the turbulent realm of the D-region but higher in the ionosphere the most likely source of irregularities is the passage of propagating waves. In this case the irregularities might move with the phase velocity of the wave rather than the local wind. A more realistic picture would possibly include a broad spectrum of such waves (HINES and RAO, 1968) in which case it is extremely doubtful whether any meaning could be extracted from drift results. In fact HINES and REDDY (1967) have shown that a wide range of internal gravity waves can propagate upwards without great attenuation but with possible directional filtering effects. They suggested that in summer the waves reaching ionospheric heights would be mainly eastward-propagating and that this bias would therefore be reflected in drift results. Some results consistent with this suggestion have been reported (RAO and RAO, 1963). If waves appear to be the probable cause of E- and F-region irregularities there is no reason why they could not also be the major producer of D-region irregularities (ROSSITER, 1970a).

There is, however, one way in which a wind interpretation of drifts may be retained even in the face of wave-induced and wave-transported corrugations. This is because of the existence of 'critical' layers where the frequency of a propagating wave is Doppler-shifted to zero by the background wind. It has been proposed (HINES, 1968) that these waves would receive enhanced observational emphasis. At a given height the enhanced irregularities associated with 'critical' waves would move as if with the neutral wind and so, if they were preferentially observed, the drift obtained would indeed represent the true background wind. The extent to which such a critical layer criterion is in fact operative in the real ionosphere must remain a matter for conjecture.

It has been assumed so far that the amplitude fading observed at the ground is entirely imposed on the radio wave at the reflection level. However it is by no means obvious that all the fading is imposed at this level (PHILLIPS, 1952). The process of partial reflection is poorly understood and the transmitted pulse passes twice through the atmosphere between the observer and the selected reflection height. It is conceivable that moving patches of ionisation in this intervening layer could modify the pulse amplitude. In view of this possibility, the validity of the heights assigned to drifts must be somewhat in question.

There are therefore quite a number of legitimate objections to the assumptions made in both the determination (§1.4) and interpretation (§3.1) of ionospheric drifts. This is why comparisons of drifts with accepted measurements of the neutral wind have assumed great importance. Because the applicability of many of the objections varies with height, it is imperative that such comparisons be carried out at all levels.

3.2 Comparison with other Techniques

In view of the large wind shears that can occur in the ionosphere it is desirable that the two techniques compared be characterised by good height resolution and similar spatial and temporal averaging. These criteria have not been fulfilled in many of the comparisons made previously.

GREENHOW and NEUFELD (1955) compared the daily mean prevailing wind measured by the Jodrell Bank meteor system with the mean wind based on a survey of Cambridge E-region drift results made by BRIGGS and SPENCER (1954). The same results were used by JONES (1958) to compare the phases of the semidiurnal tide. However the two sets of data were obtained in different years. BEYNON and GOODWIN (1967) compared spaced receiver results from oblique c.w. transmissions with seasonal averages of Cambridge E-region and Manchester meteor results. KOCHANSKI (1964) found good agreement at about 110 km between radio measurements of the plasma cloud associated with a luminous trail and the neutral gas motions as determined optically. SPRENGER and SCHMINDER (1968) have investigated in some detail both the average and periodic behaviour of winds at comparable heights obtained from meteors at Jodrell Bank and from low frequency totally reflected radio waves in North Germany. The station separation was 900 km although in recent years (LYSENKO et al, 1972) they have been able to use a local meteor radar for essentially common volume comparisons. MULLER (1968a) compared Sheffield meteor data with ionospheric drifts from Aberystwyth but with an average height difference of some 10 km. KENT and WRIGHT (1968) have reviewed in detail all these early comparisons. The general result from them is that, considering the lack of overlap, the main features of the total reflection E-region drifts are in good agreement with those from more accepted wind measuring techniques. More recent comparisons with phase path (VINCENT, 1972) and meteor (FELGATE et al, 1974) techniques tend to confirm this impression.

In contrast to this large body of comparison data for total reflections, comparatively little use has been made of partial reflections, and what has been carried out can be briefly summarised as follows. FRASER and KOCHANSKI (1970) found encouraging agreement for New Zealand partial reflection drift data with circulation models and with very distant rocket measurements of the winds at similar heights. GREGORY and REES (1971) compared noon drift profiles with rocket data but since the rocket results were for below 60 km there was no height overlap. Additionally there was a horizontal separation of several hundred kilometres. A rocket comparison with some height overlap (WRIGHT and FEDOR, 1967; WRIGHT, 1968) indicated, for partial reflections from sporadic E, a relatively poor agreement in magnitude and a fair agreement of directions. ROSSITER (1970a) made a common-volume comparison between ionospheric drifts (total and partial reflections) and the Adelaide meteor radar measurements. Results were taken for only a few hours of the day near noon and the scatter of individual comparisons indicated a possible unreliability in individual drift determinations. There was also a tendency for the drifts to be systematically too small, although this tendency and some of the directional scatter may well have resulted from the use of a rather small right-angled triangle of side 91m. It was this work that prompted the current experiment and particularly the improvements in experimental method and analysis described in Chapter 2.

3.3. Compatability of the Adelaide partial reflection and Meteor Systems.

3.3.1 Height coverage.

From the point of view of height distribution, the two systems are well suited for comparison. As mentioned earlier, the best height range for continuous drift results using partial reflections is 80-95 km. The meteor technique is most reliable at the heights where the number of detectable ionisation trails is largest. FIGURE 3.1 shows a smoothed distribution of the occurrence of meteor echoes versus height, based on the flux detected at 27 MHz in 1972. The distribution peaks at around 89 km which is almost the centre of the optimum interval for drift measurements. It was decided that the interval 85-90 km should be the subject of the most intensive comparison. The choice of a 5 km height interval was based on the need for an adequate number of meteors in a time interval short enough to preserve tidal information. It was also limited more fundamentally by the width of the transmitted radio pulse used in the drift experiment. Both factors precluded any attempt at obtaining a better height resolution.

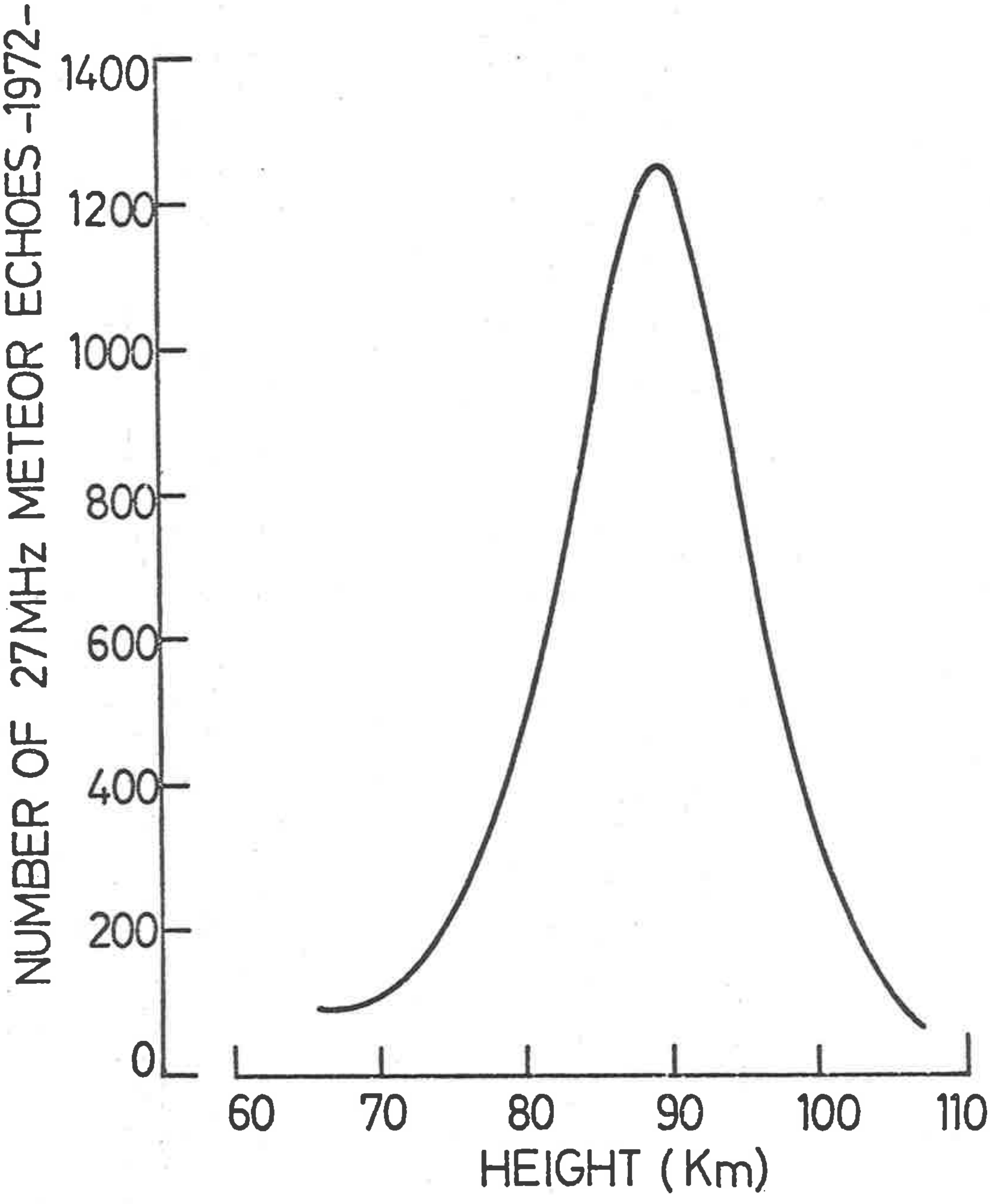


FIGURE 3.1 *The height distribution of 27 MHz meteor trail echoes based on the 1972 data.*

3.3.2 Temporal averaging

With the 5 km height interval it was found that any 3-hour period generally included an adequate number of meteors (i.e. 3 or preferably more) for a wind determination. In order to have similar time averaging for the two systems the Groves model analysis was not used. Instead all the meteor line-of-sight velocities in each 3 hour/5 km interval were subjected to a simple averaging of horizontal components to produce a single horizontal vector velocity for that interval. The drift results were treated in the same way, all the values in a given data interval being averaged to produce a mean.

The distribution of drift determinations during the day was generally much more even than the distribution of meteor echoes. The main features of these diurnal variations are displayed in FIGURE 3.2. The factor-of-3 variation in the meteor echoes results from the observer's changing orientation with respect to the directional meteor flux. In the case of the drift determinations, the variation is due in part to the changing ionospheric structure, and also to local impulsive radio noise which was often particularly strong between 1700 and 2000 local time.

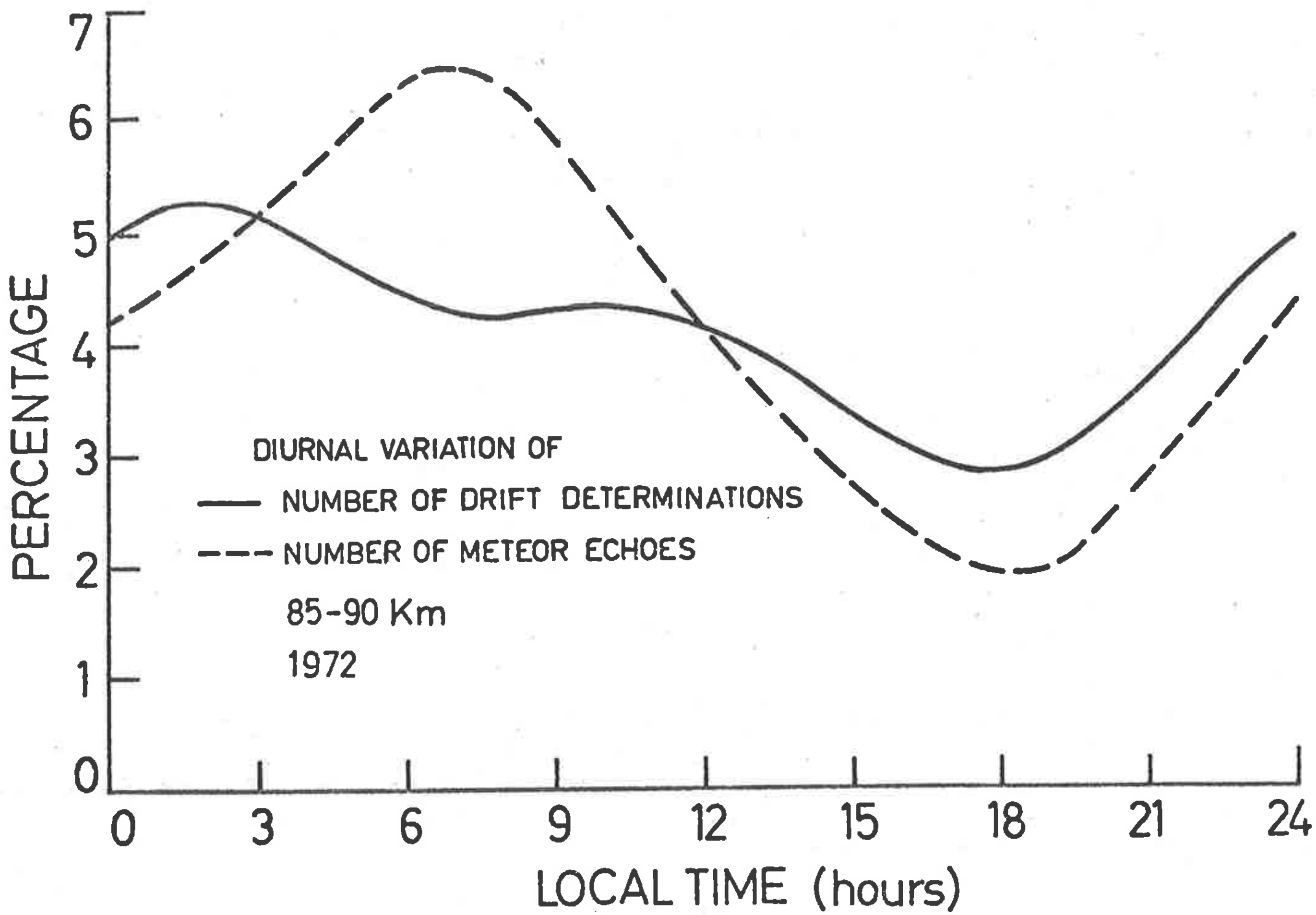


FIGURE 3.2 *A comparison of the most typical diurnal variations in meteor numbers and successful drift determinations.*

3.3.3 Spatial Averaging

It is in the extent of their horizontal sensitivity that the two systems are least compatible, although they include a common volume. Their respective collecting areas (and that for the Woomera drift system) at a height of 85 km are shown in FIGURE 2.1. Whereas meteors are observed over a circle or radius 200 km, the Buckland Park ionospheric reflections (as calculated from the projected half-power response of the receiving aerials) are confined to a roughly circular region only 30 km in radius. In fact it is probably much smaller because of the narrow range of angles over which the ionosphere scatters. Whilst this is a considerable improvement on most previous comparisons involving partial reflection drifts, it should still be noted that successive meteor echoes may be up to 400 km apart and the meteor echoes will often be in excess of 100 km distant from the region for which ionospheric drifts are determined.

3.4 Results and Discussion

During 1972, the meteor system was run routinely for three days a month and for longer periods in March, June, September and December. Where possible the drift equipment was operated over the same periods, with the receiver gains being continuously monitored and set to obtain good fading records from the 85-90 km range. Due to the large dynamic range of the echoes and the linear nature of the receivers, reasonable data continuity over the full 24 hours was usually only obtained for this height range. However, during February, March and May, special attention was paid to also obtaining continuous drift information from one of the other height ranges covered by the two techniques - 95-100, 90-95 and 80-85 km respectively.

The results for all months at 85-90 km are shown in FIGURES 3.3 - 3.17. Each velocity component is plotted at the centre of the three-hour period for which it represents the average. Points corresponding to the ionospheric drifts are joined by continuous lines, and the meteor winds by dashed lines.

The degree of correlation between the two sets of results varies. Some general and specific features are worthy of comment. In view of the uncertainties inherent in the drift technique, the overall agreement is very encouraging. If the meteor technique measures the neutral wind (and it is widely agreed that it does) then it must be concluded that for much of the time so does the drift technique. Nor are the various discrepancies necessarily

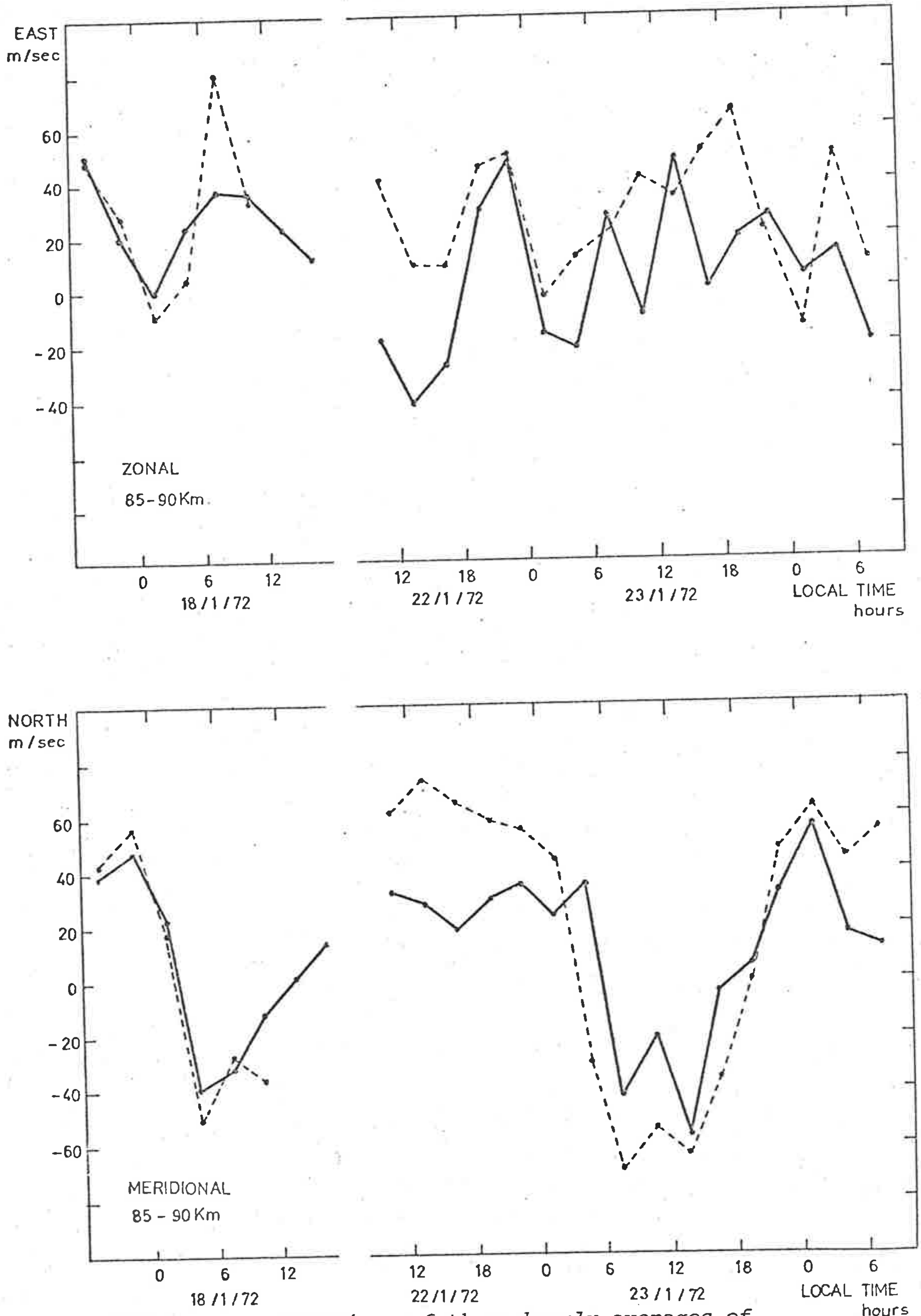


FIGURE 3.3 *A comparison of three-hourly averages of partial reflection drifts (dots joined by full lines) and meteor trail wind determinations (broken lines).*

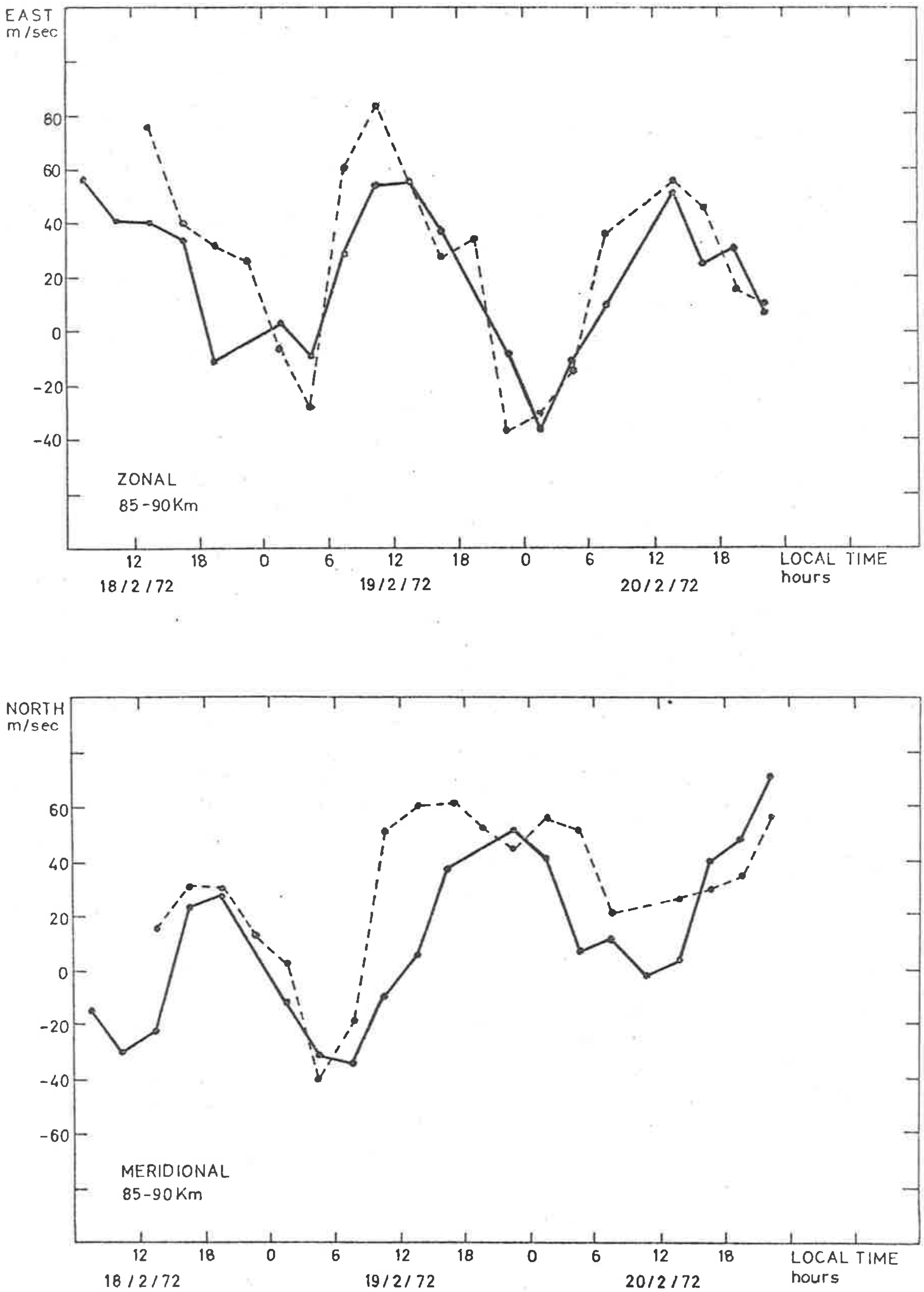
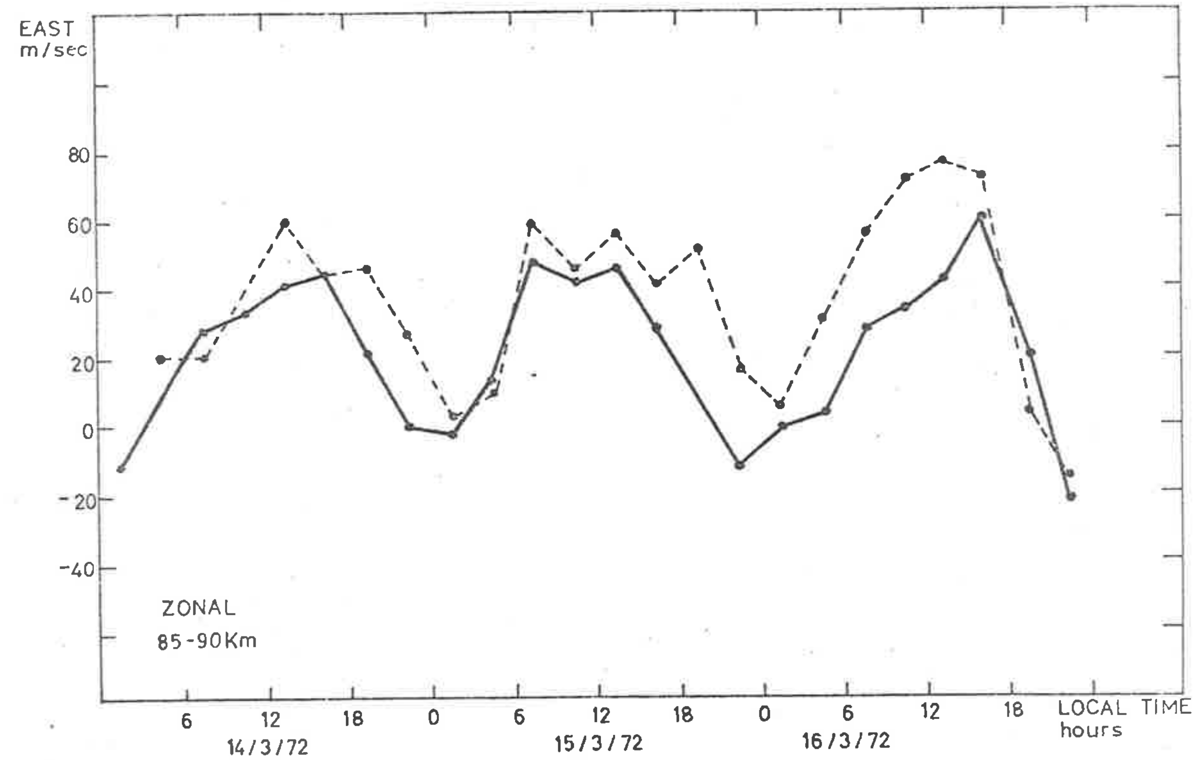


FIGURE 3.4 February 1972, 85 - 90 km.



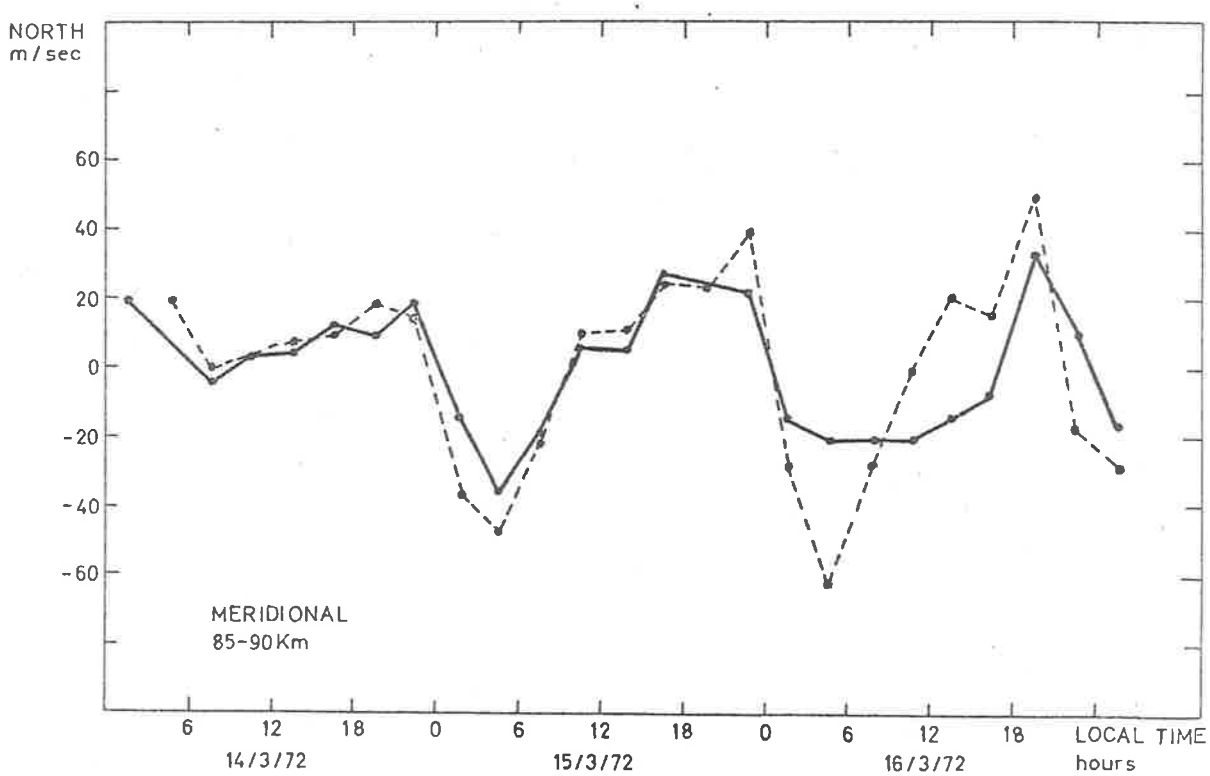
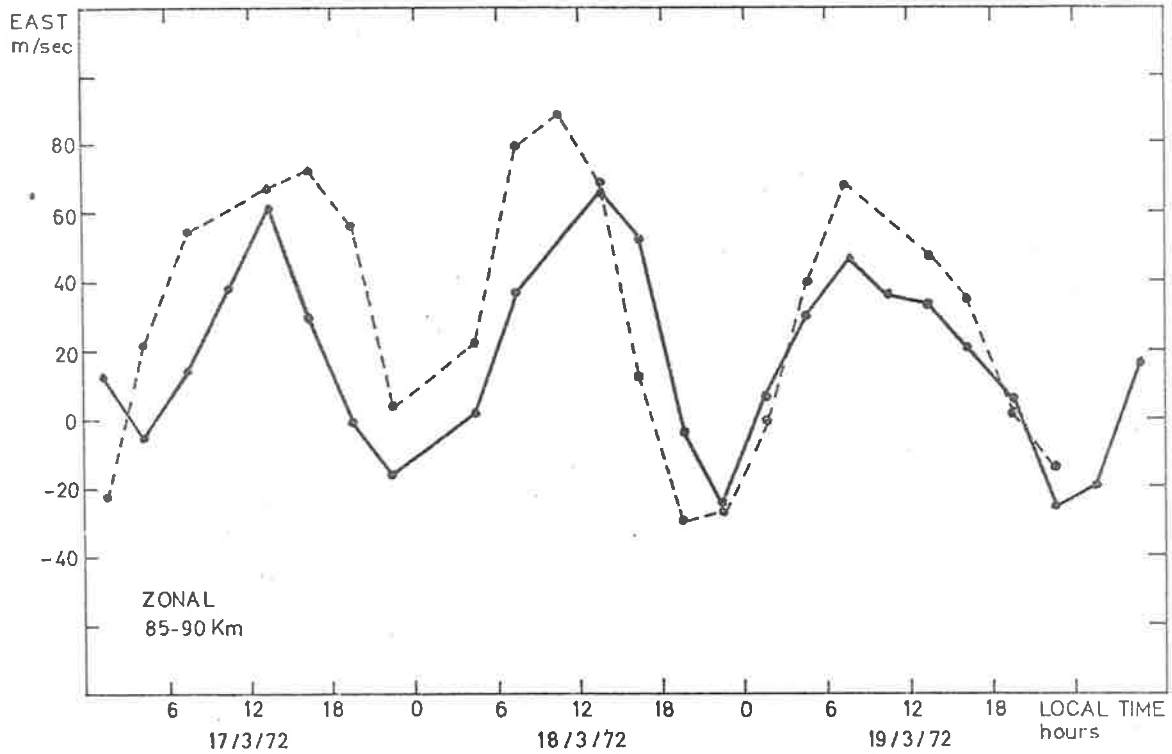


FIGURE 3.5 March 1972 (I) 85 - 90 km.



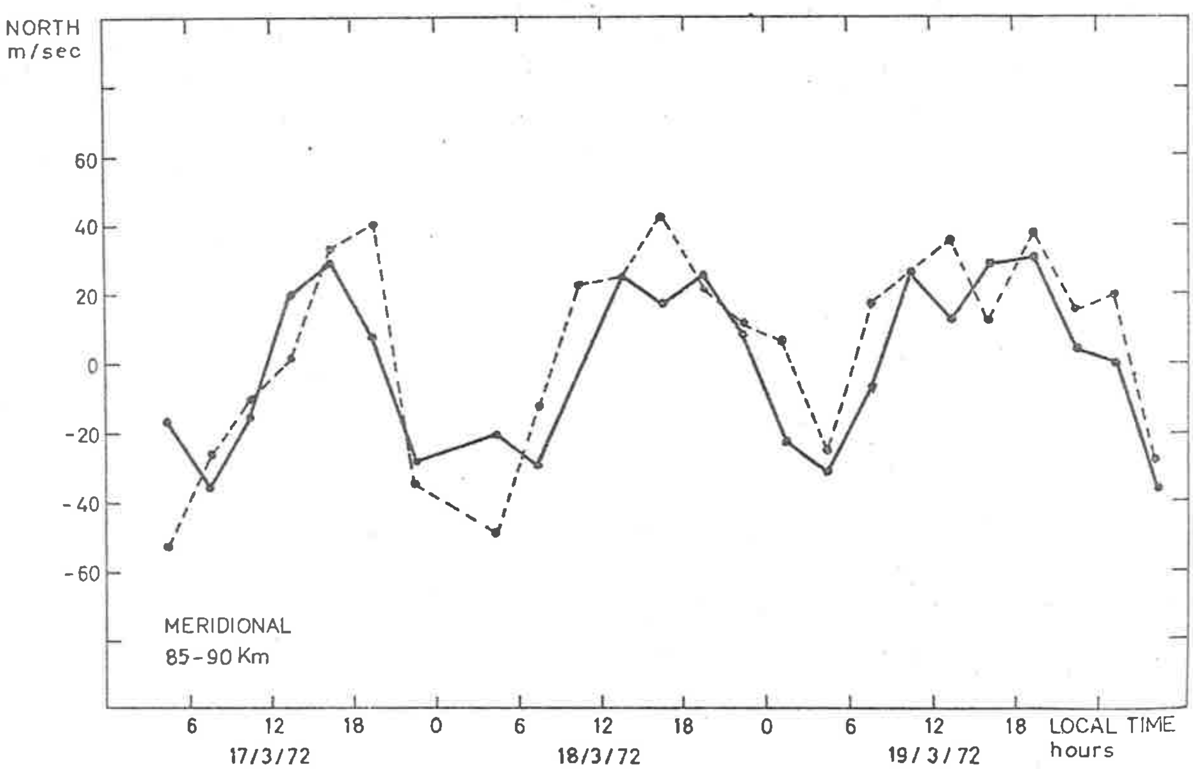
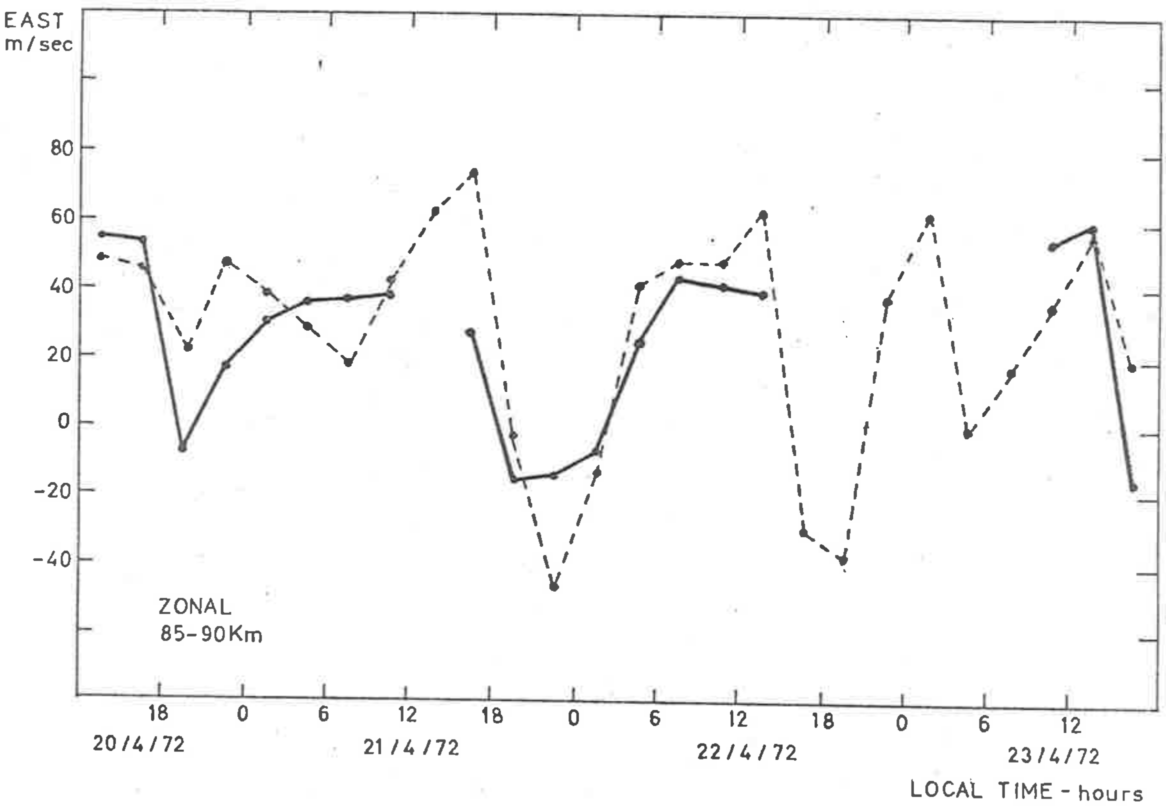


FIGURE 3.6 March 1972 (II) 85 - 90 km.



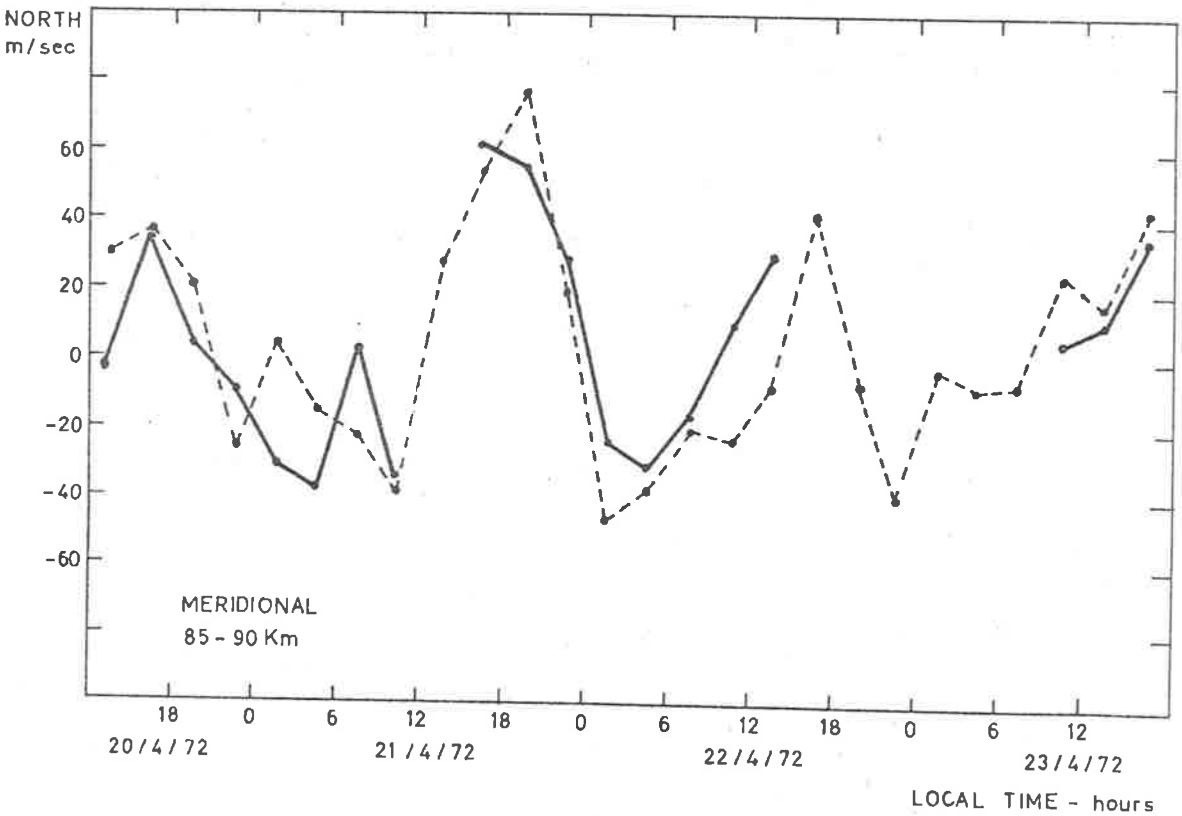


FIGURE 3.7 April 1972, 85 - 90 km.

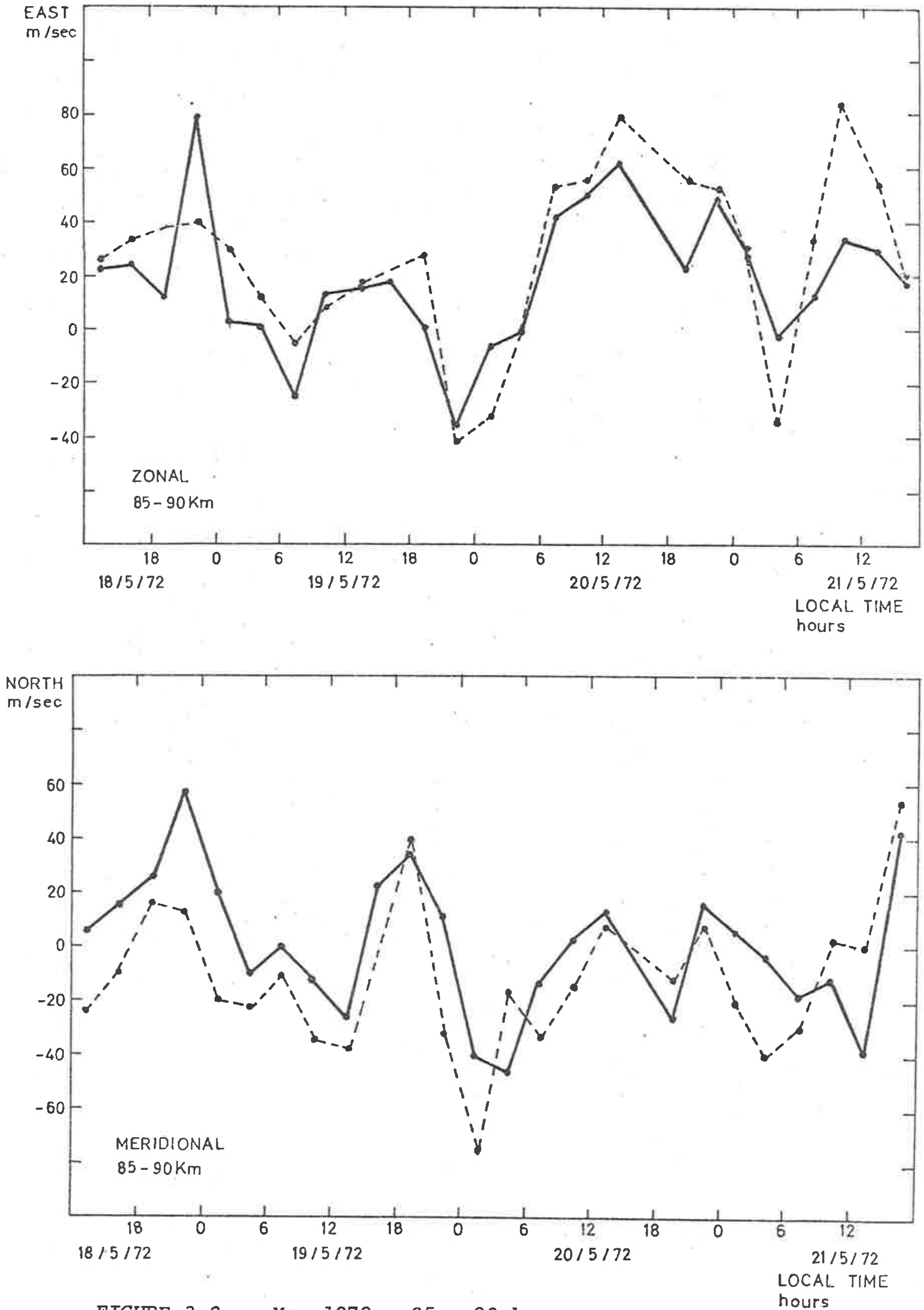
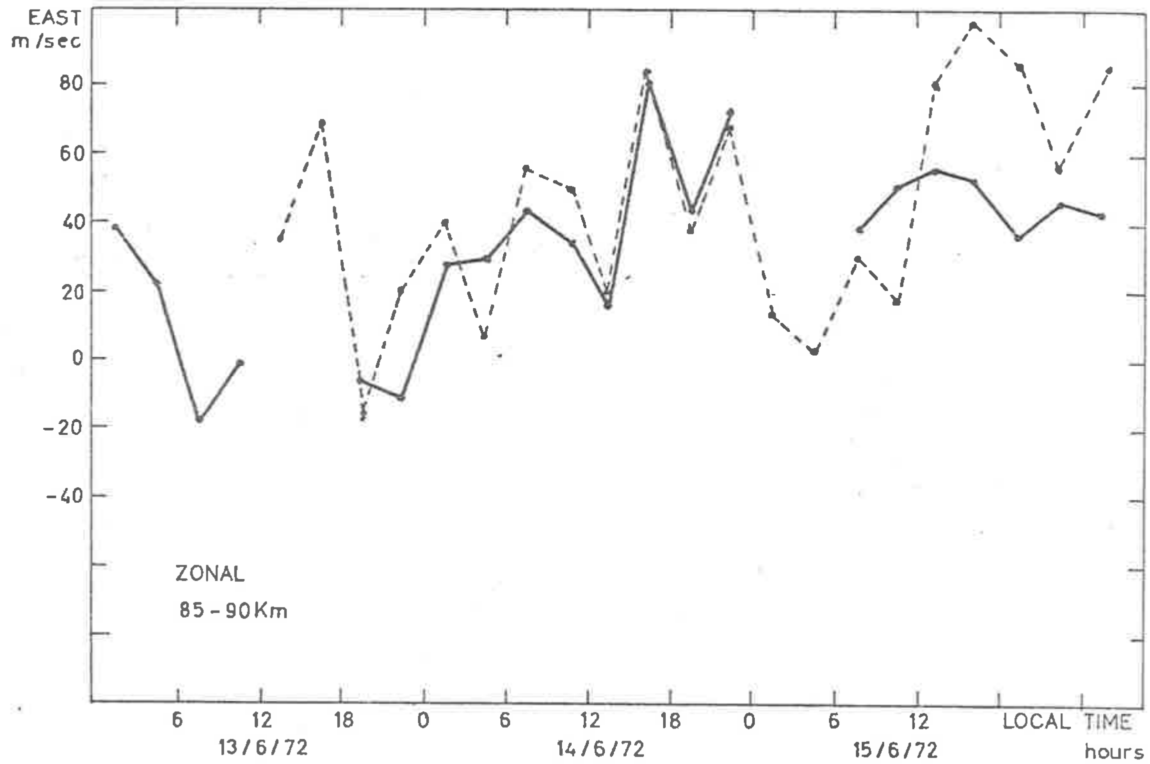


FIGURE 3.8 May 1972, 85 - 90 km.



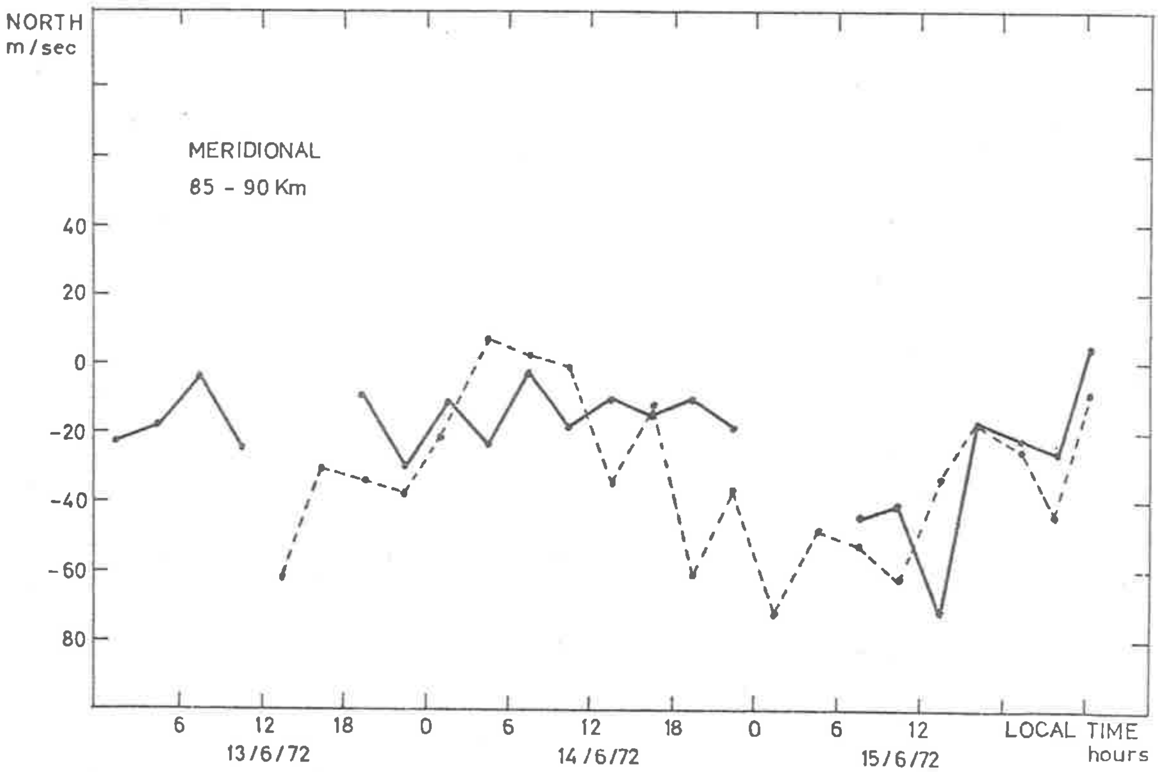


FIGURE 3.9 June 1972 (I) 85 - 90 km.

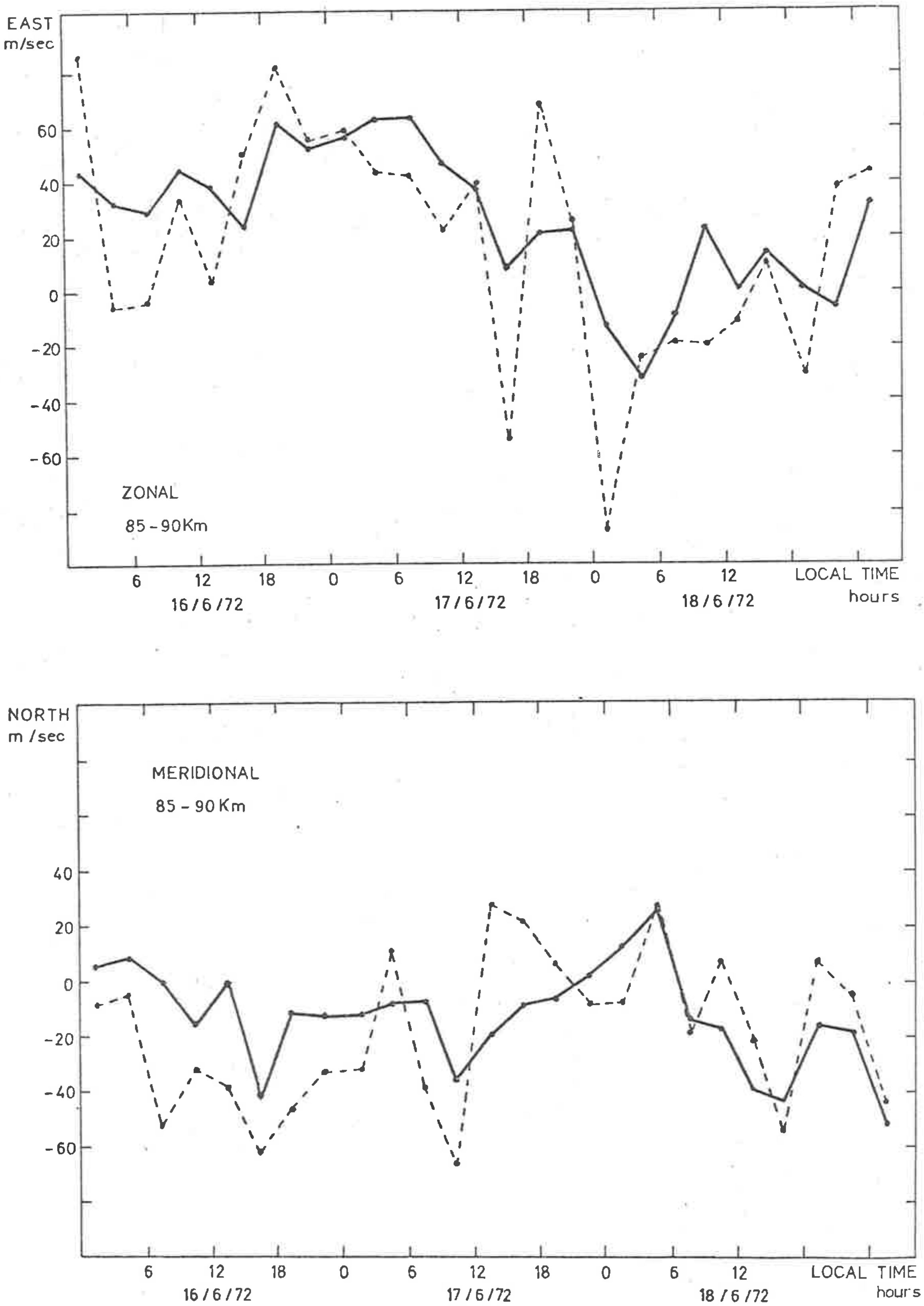


FIGURE 3.10 June 1972 (II) 85 - 90 km.

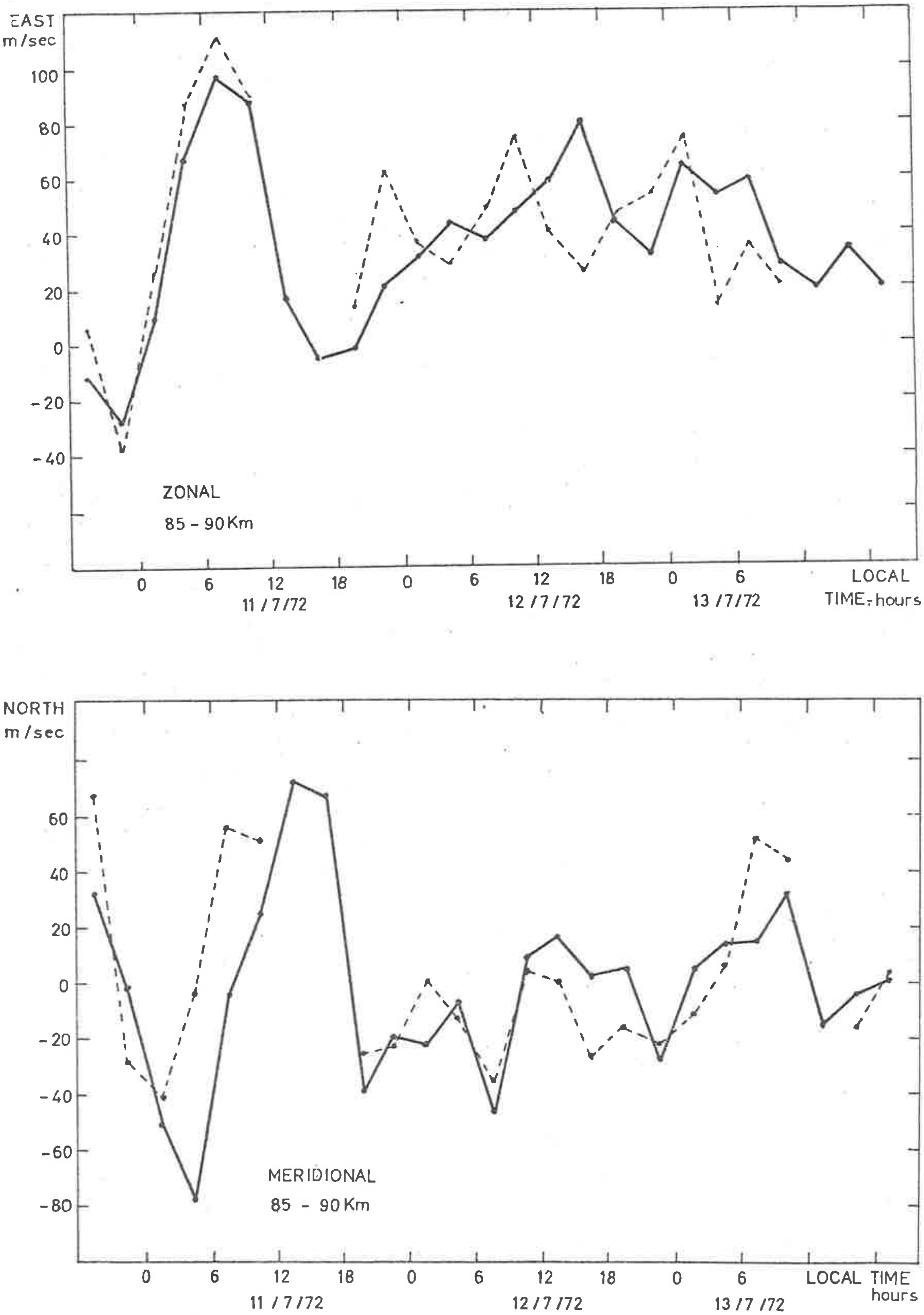
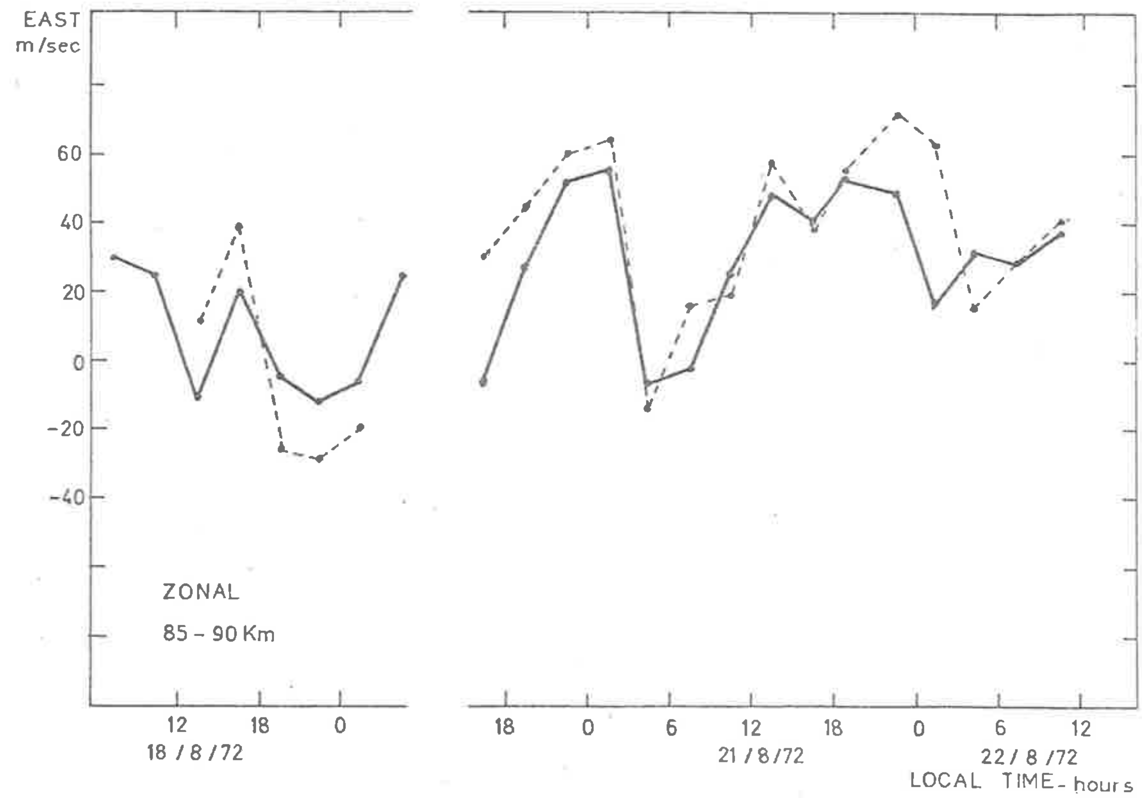


FIGURE 3.11 *July 1972, 85 - 90 km.*



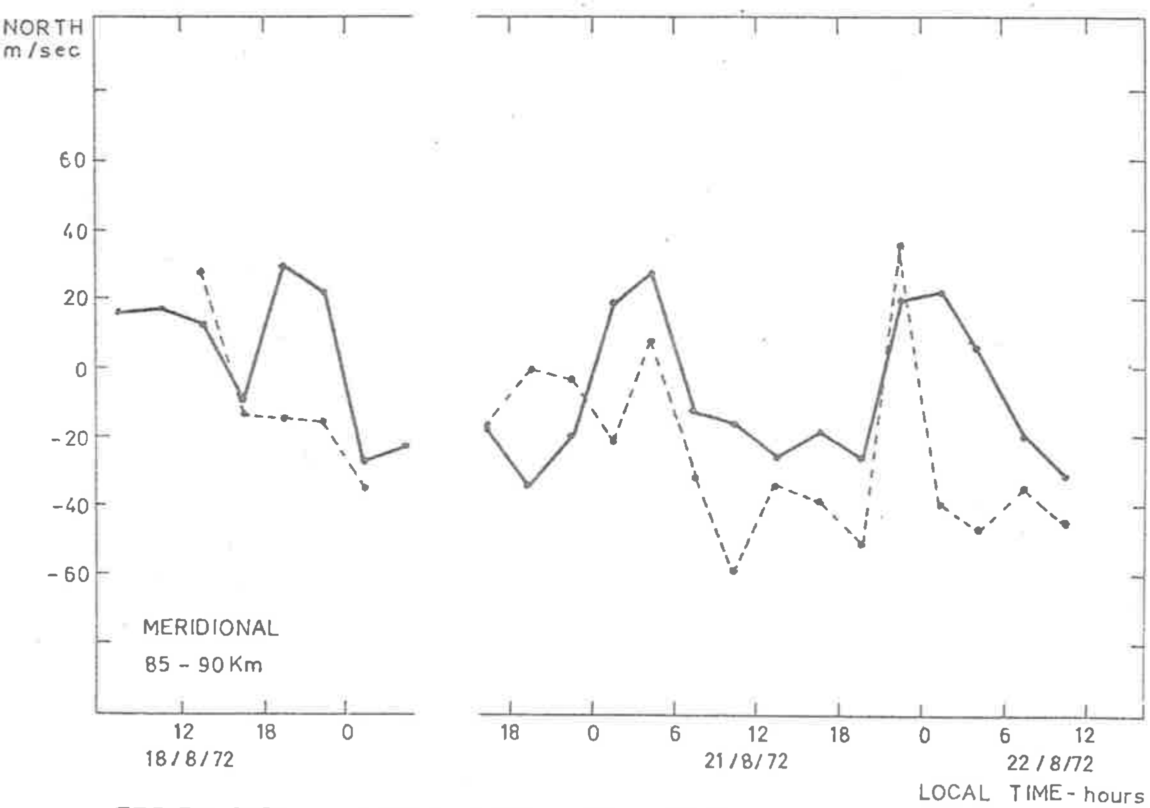


FIGURE 3.12 August 1972, 85 - 90 km.

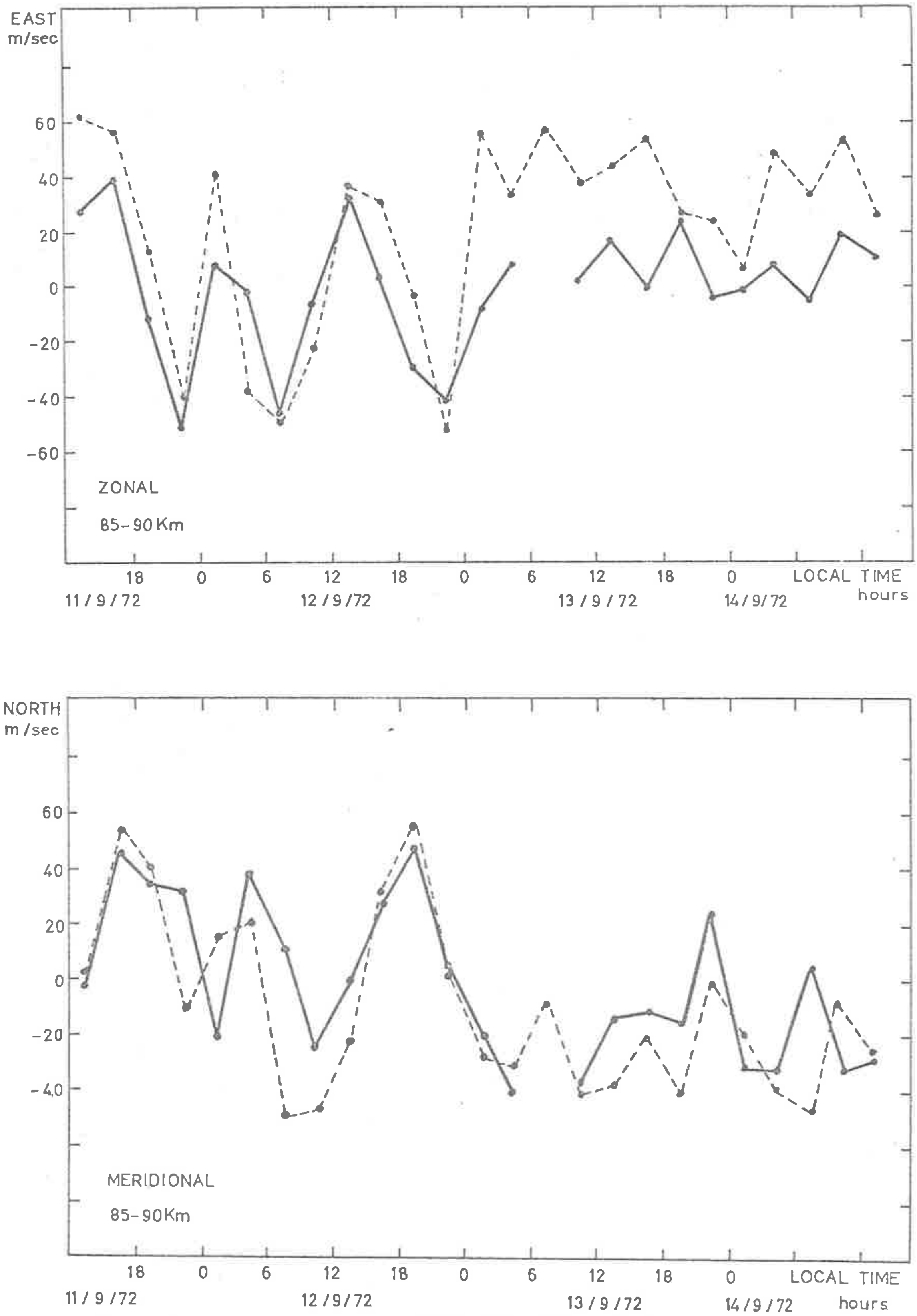


FIGURE 3.13 September 1972 (I) 85 - 90 km.

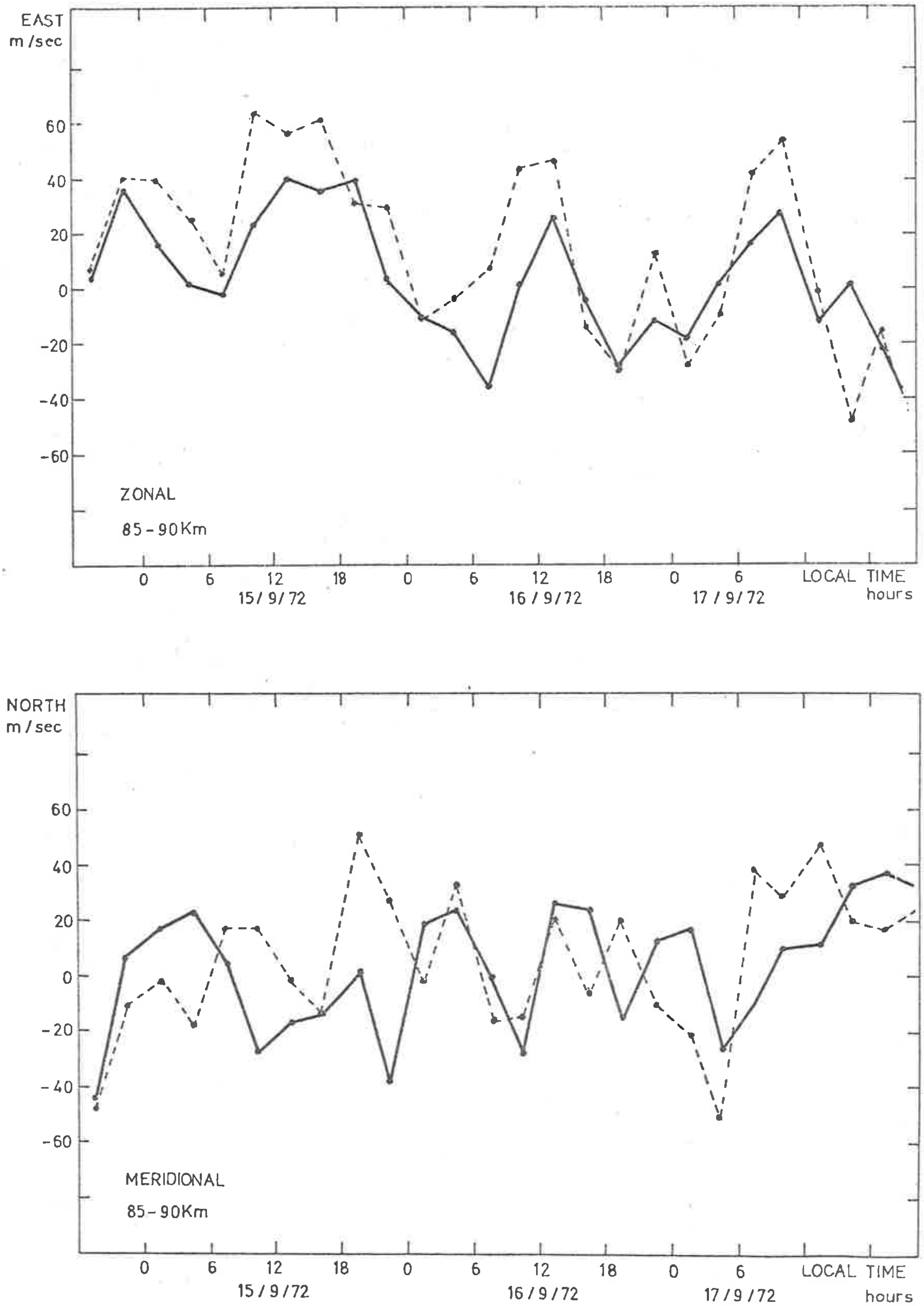


FIGURE 3.14 September 1972 (II) 85 - 90 km.

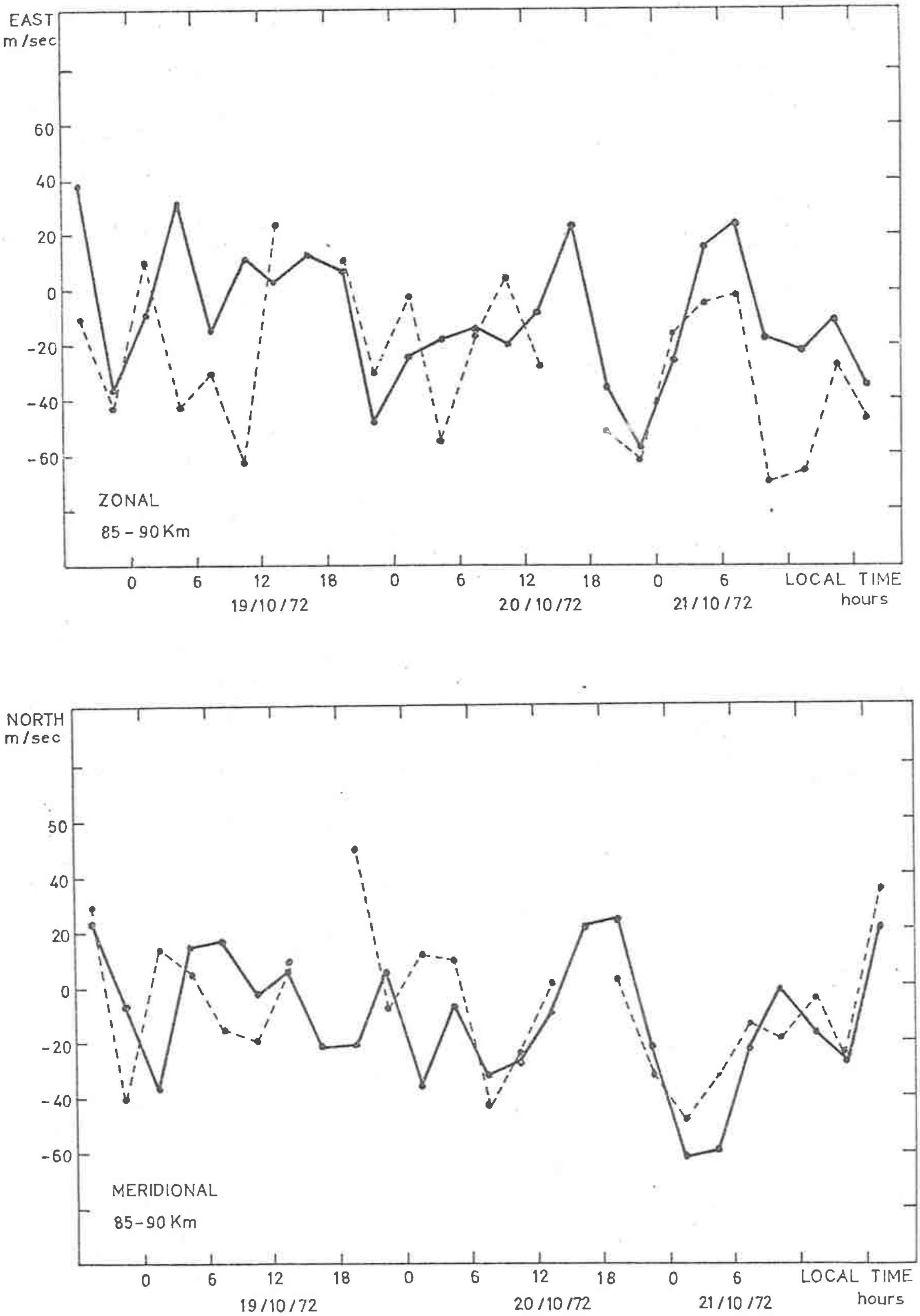
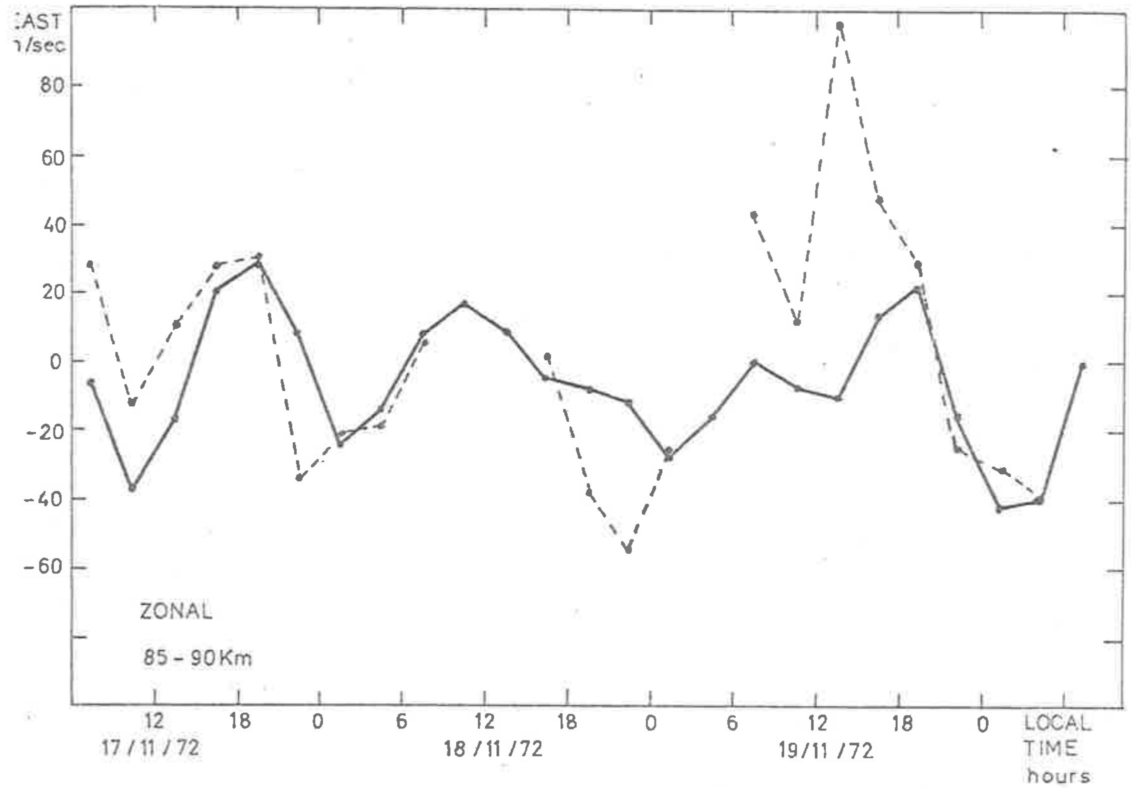


FIGURE 3.15 *October 1972, 85 - 90 km.*



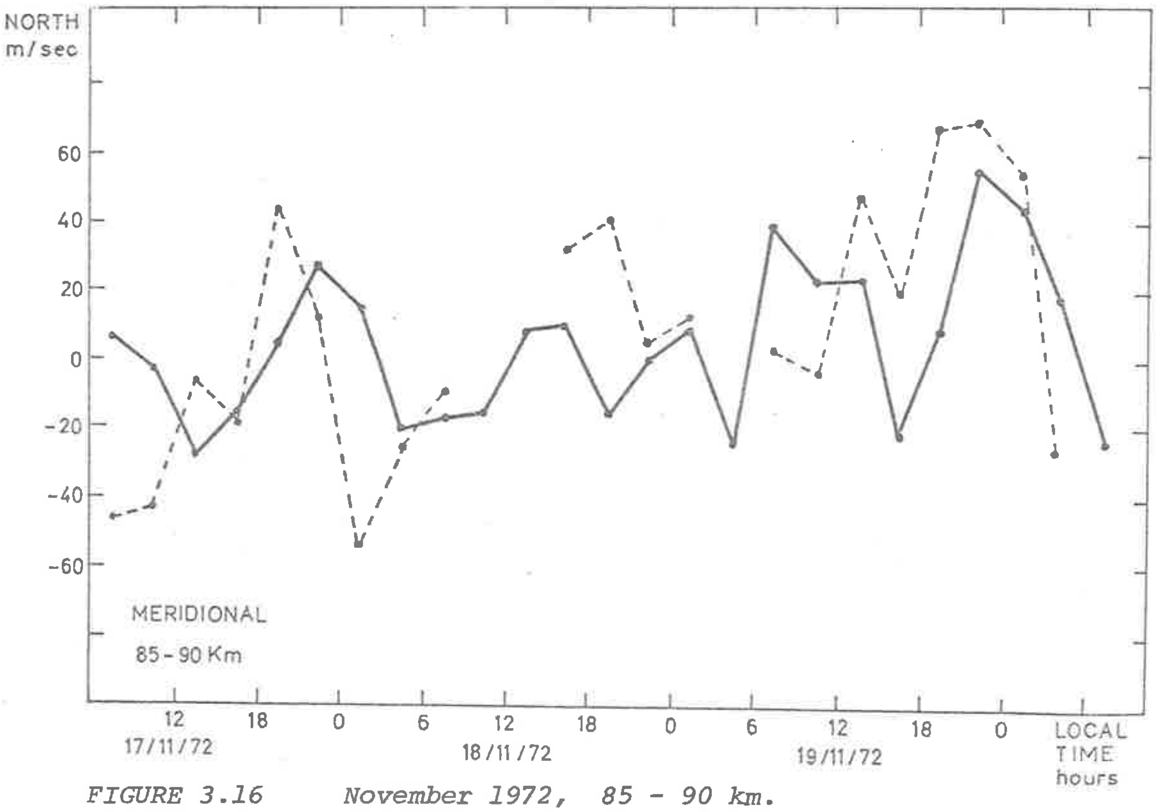


FIGURE 3.16 November 1972, 85 - 90 km.

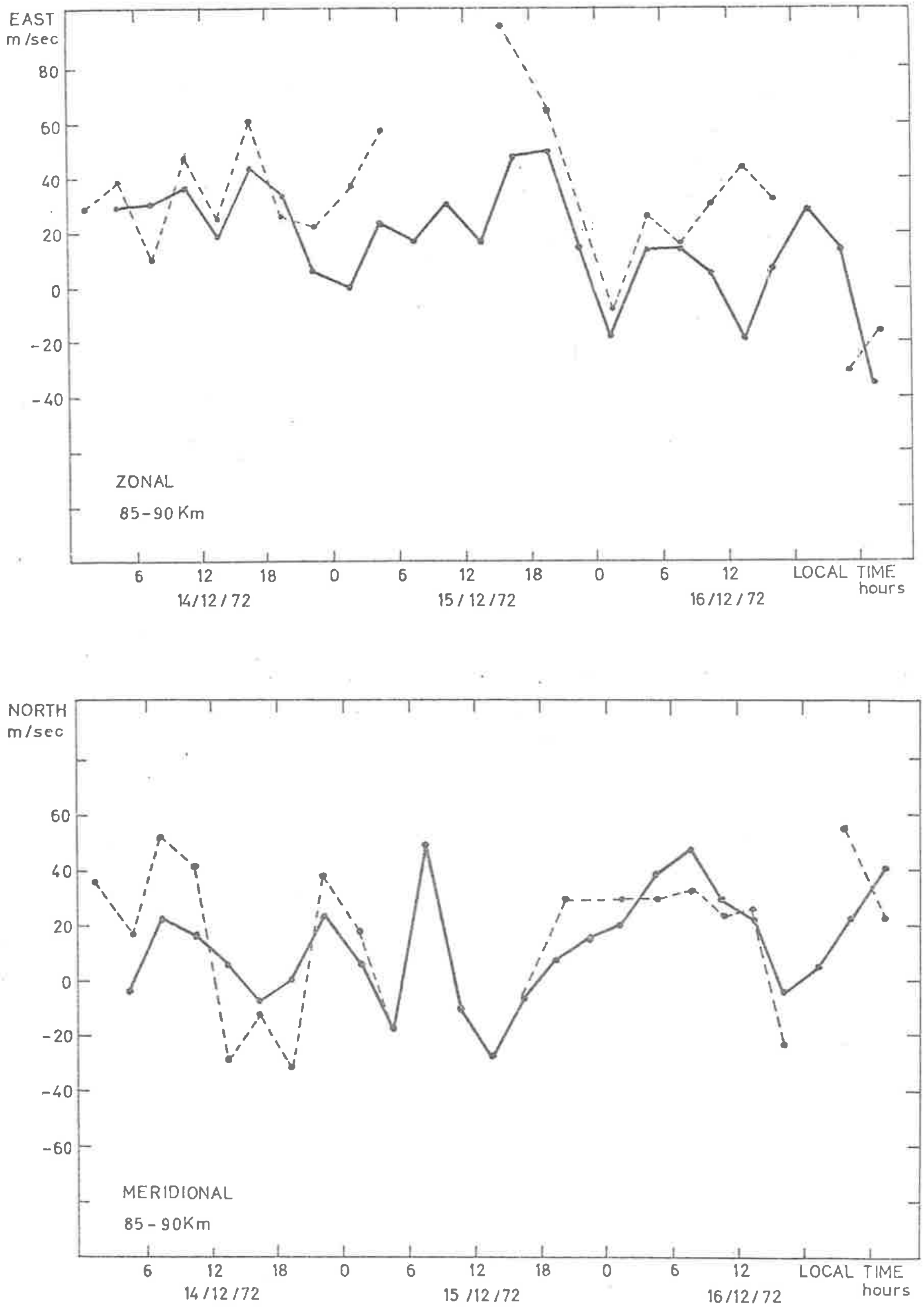


FIGURE 3.17 December 1972, 85 - 90 km.

indicative of shortcomings in the drift values. Some may be due to the various forms of averaging used. Apparent phase shifts in the results are to be expected from both the time and height widths of the basic data interval chosen. Within a particular 3-hour interval, the times at which meteor echoes occur may be quite different from the times at which drifts can be successfully determined. Whilst this difference only approaches 3 hours on occasions of low data density, it could still be significant when the wind is changing rapidly. Within the 5 km height interval, differences between the mean height of the meteors and the height from which the bulk of the radio power is returned may be even more important, especially at times of high wind shear.

One systematic feature of the results is the tendency of the drifts not to reproduce the extreme values of the meteor winds. This may indicate that the triangle size effect has not been completely eliminated. This is not surprising even though the receiving station spacing is greater than the average ground pattern size. On some occasions when the pattern size exceeds the spacing, the drifts will be too small but there will be no compensating tendency for them to be large when the pattern size is small. This underestimation of the velocity is however only some 10% on the average. The drift technique is quite able to detect velocities in excess of 100 m/sec.

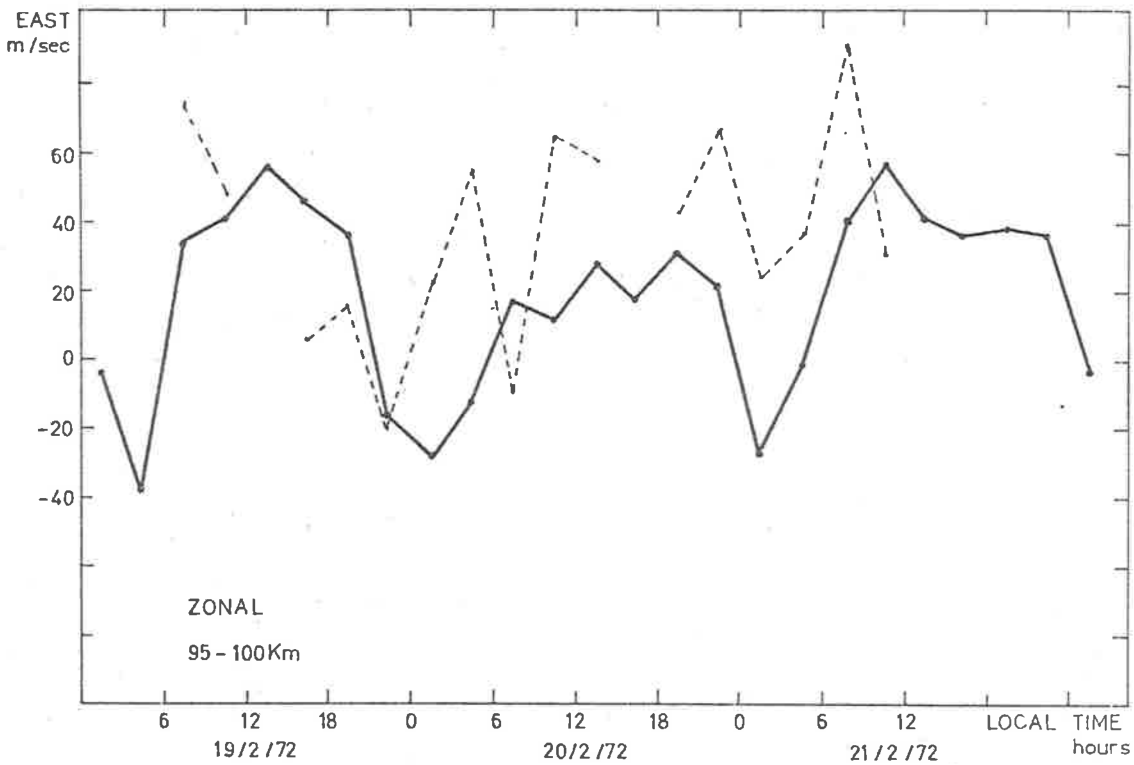
The quasi-sinusoidal nature of the results is quite obvious and there is little evidence for the drifts failing to show tidal influence on occasions when the meteor winds do. This feature of Rossiter's results (ROSSITER, 1970a) resulted largely from comparison with model-fitted meteor data. Whilst the model-fitting procedure (GROVES, 1959) enables the maximum amount of information to be extracted from the data, it also means that because of the influence of other times and heights, tidal components will automatically be introduced into a short segment of results at a given height.

In fact, in the present results some discrepancies may perhaps be better explained in terms of the greater variability of the meteor results. For example, late on June 17 (FIGURE 3.10) a sharp variation in the zonal component of the meteor results is hard to reconcile with tides as it does not appear later in the meridional component. There were only 5 meteors between 1500 and 1800, 3 between 1800 and 2100 and 5 between 2100 and 2400. The horizontal scale of the region observed by the meteor system is 400 km. Internal gravity waves with vertical wavelengths of the order of 10 km and periods of an hour or so will have large horizontal wavelengths, typically of the order of hundreds of kilometres (HINES, 1960) so that the gravity wave wind systems are slightly tilted. It is therefore possible that the line-of-sight velocities measured from successive meteor trails occurring at the same height but separated by large horizontal distances may be inconsistent. Thus the meteor wind determinations would be suspect when only a few widely spaced meteors are detected. By contrast, the wind fields over the much narrower partial reflection region are probably more homogeneous.

The quality of the agreement seems more variable in the winter and spring months. The winter divergence is consistent with the fact that wave activity is more prevalent at ionospheric heights in winter than in other seasons (for example NORDBERG et al, 1965). This may be due to either an increase in the strength of the generating source or a lower attenuation by the wind structure of the underlying regions as the waves propagate upwards. The spring behaviour raises an interesting possibility. The semi-diurnal tide is then at its most prominent in relation to the other dynamical contributions, and it may be that it is a more local phenomenon than the prevailing and diurnal components. If the importance of the higher order modes changed rapidly with latitude or longitude then the difference in the collecting areas of the two systems would become more crucial.

It is of interest to note that at no stage during the entire year are the D-region drifts consistently more eastward than the meteor winds. This would appear to rule out the possibility (discussed earlier) that the drifts may be strongly influenced by eastward-propagating gravity waves generated by weather systems in the troposphere.

The results for the other three height ranges are presented in FIGURES 3.18 - 3.20. At 80-85 km the agreement is uniformily good, as it is at 90-95 km with the exception of part of one cycle of the zonal component where the 6-hour phase shift is suggestive of a height difference between the two sets of data. At 95-100 km the agreement is only fair, but the drifts exhibit a clear periodic behaviour in both components. The meridional component consistently lags the zonal component by about 6 hours, as would be expected for a basically diurnal tide. The variability of the meteor component is in keeping with the combined effects of the



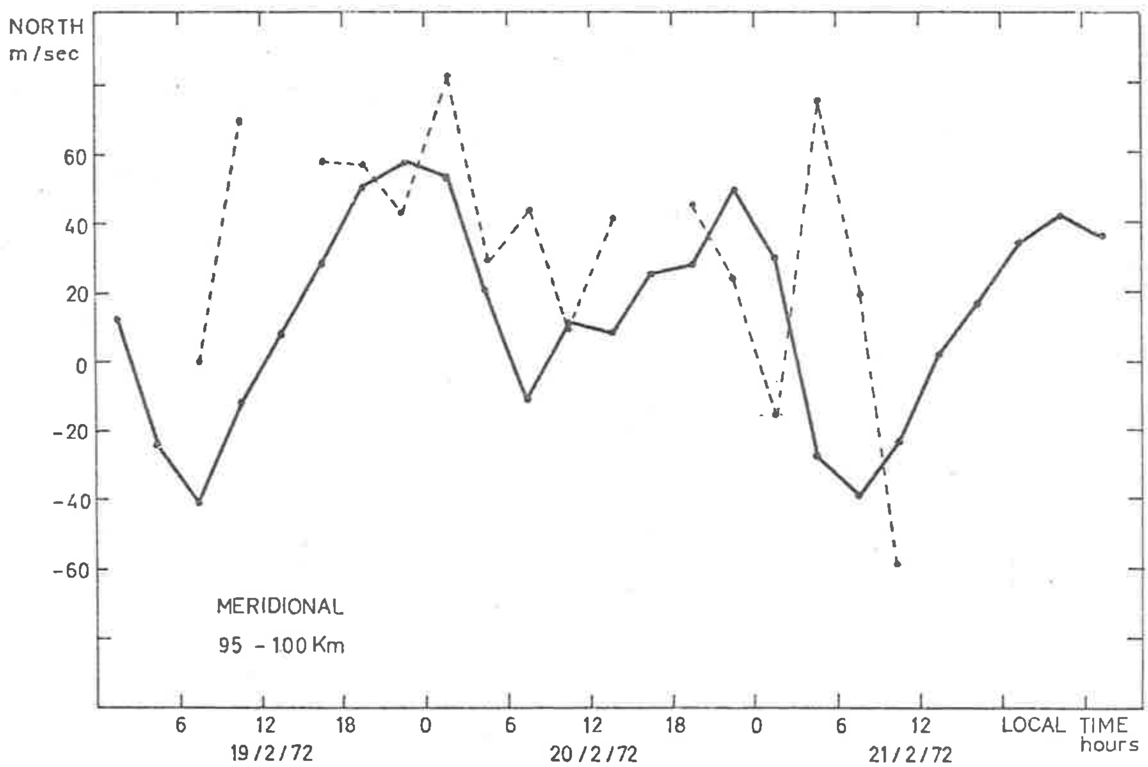
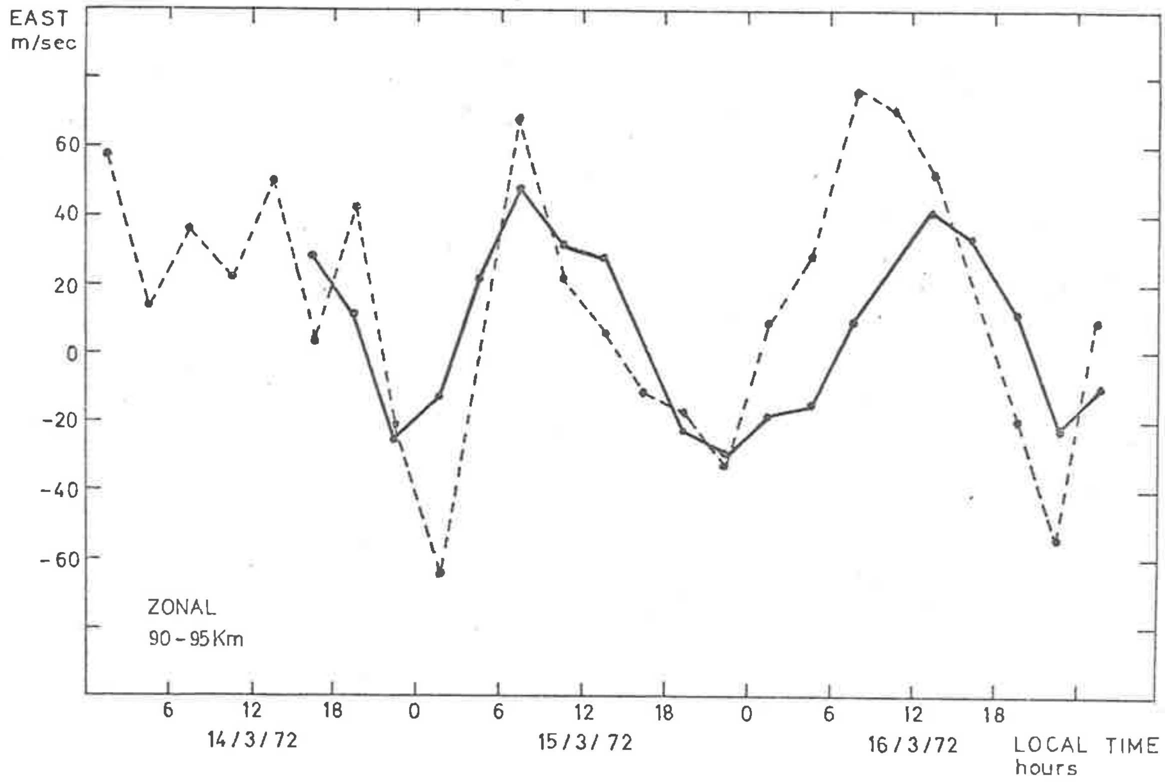


FIGURE 3.18 February 1972, 95 - 100 km.



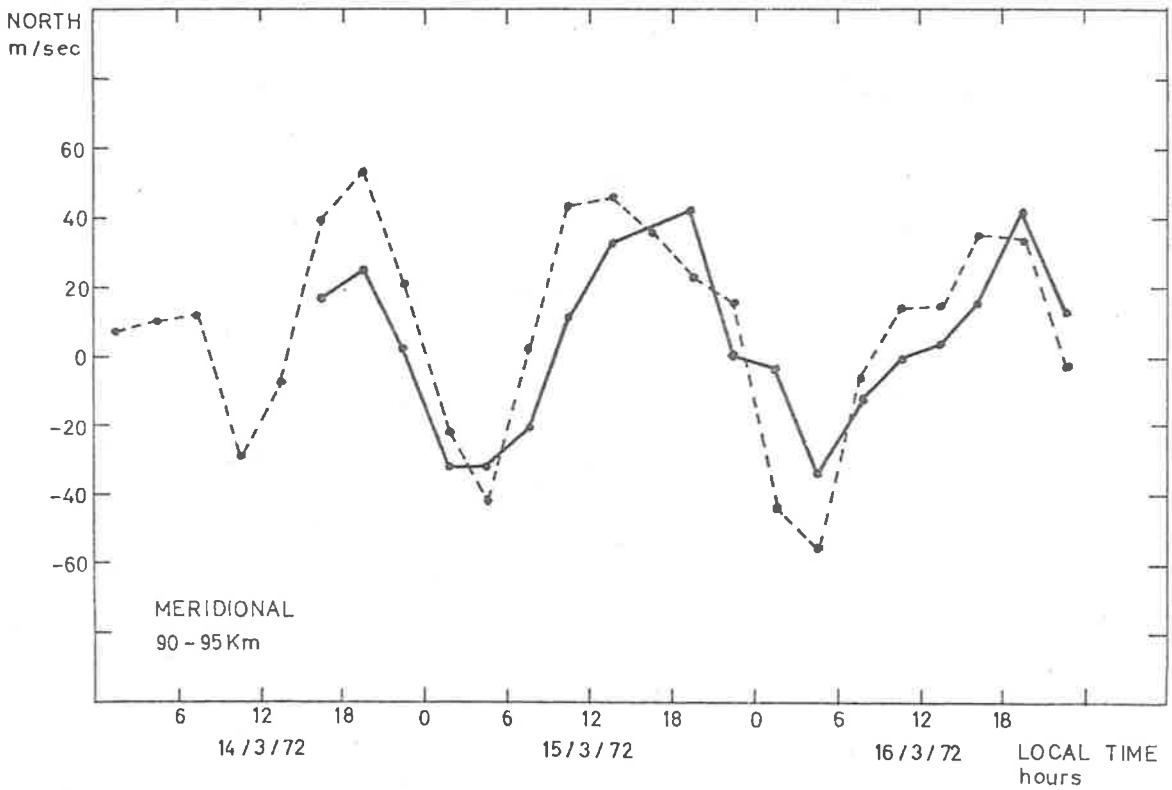


FIGURE 3.19 March 1972, 90 - 95 km.

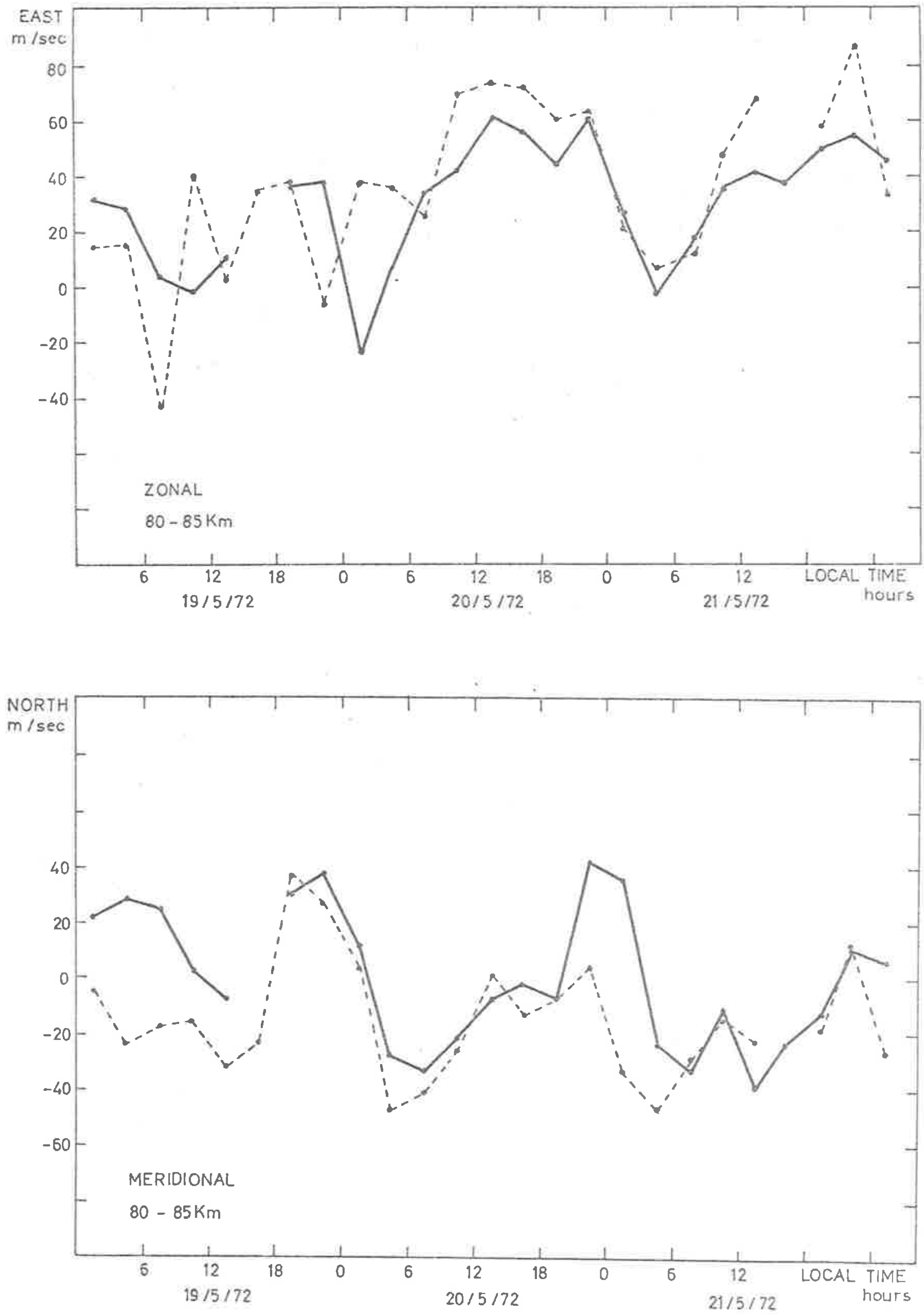


FIGURE 3.2o *May 1972, 80 - 85 km.*

greater gravity wave amplitude predicted for that height and the reduced meteor flux.

In summary, then, it would appear that in general the D-region drifts and winds derived from distorted meteor trails represent the same physical quantity. For much of the time the agreement is so good as to preclude any doubts. On other occasions there are differences which could be due to the differing parameters of the observing systems or to inadequacies in one or both of the techniques. It will however become apparent in the next chapter that even in some of the months where the correlation is relatively low, both the mean and periodic components of the D-region drifts behave in a manner entirely consistent with a neutral wind interpretation.

CHAPTER FOUR

THE DRIFT RESULTS - PREVAILING AND PERIODIC COMPONENTS

In this chapter, the more regular features of the drifts will be considered - notably the mean and periodic aspects of the motion. Each day's data was subjected to harmonic analysis, a least squares fit being made to a mean plus 24 hour, 12 hour and 8 hour periodic components. The choice of only these three periods was based largely on the experience of the analysis of many years of meteor winds. Harmonic analysis could not be applied to days when there were gaps of several hours in the data, even though on many such days there were still drifts distributed over the complete 24 hour period. Therefore in order to retain as much meaningful information as possible, the prevailing component was determined from a simple averaging of all the results for a particular day, providing the coverage was sufficiently uniform to smooth out the tidal variations. This approach has of course been widely used but usually when results were only available over a set period of the day (for example FRASER, 1968; ROSSITER, 1970a). Prevailing components can then be misleading (STUBBS and VINCENT, 1973) to an extent largely dependent on the phase stability and amplitude of the local tides.

4.1 Prevailing Component

4.1.1 Annual variation

The zonal and meridional components of the 1972 daily mean drifts are plotted for four heights in FIGURES 4.1 and 4.2 respectively. A positive drift is defined as eastward in the case of the zonal component and northward in the case of the meridional component. One striking feature of the graphs is the great variation within the data for each month. Even so, there are clear trends over the year. The curves represent 11-term power series that have been fitted to each set of data on a least squares basis. They serve to highlight the main features of the seasonal variation and also the gradual change in character with height. This is most marked in the zonal component (FIGURE 4.1) where the basically annual variation at 80-85 km has changed to a semiannual variation by 95-100 km. It is also clear that, on the average, the meridional mean is much less than the zonal mean. There are still individual days with prevailing N/S components of between 30 and 40 m/sec but generally the meridional curves deviate little from zero. Whilst there is no strongly preferred meridional direction, the basic zonal motion is clearly eastward.

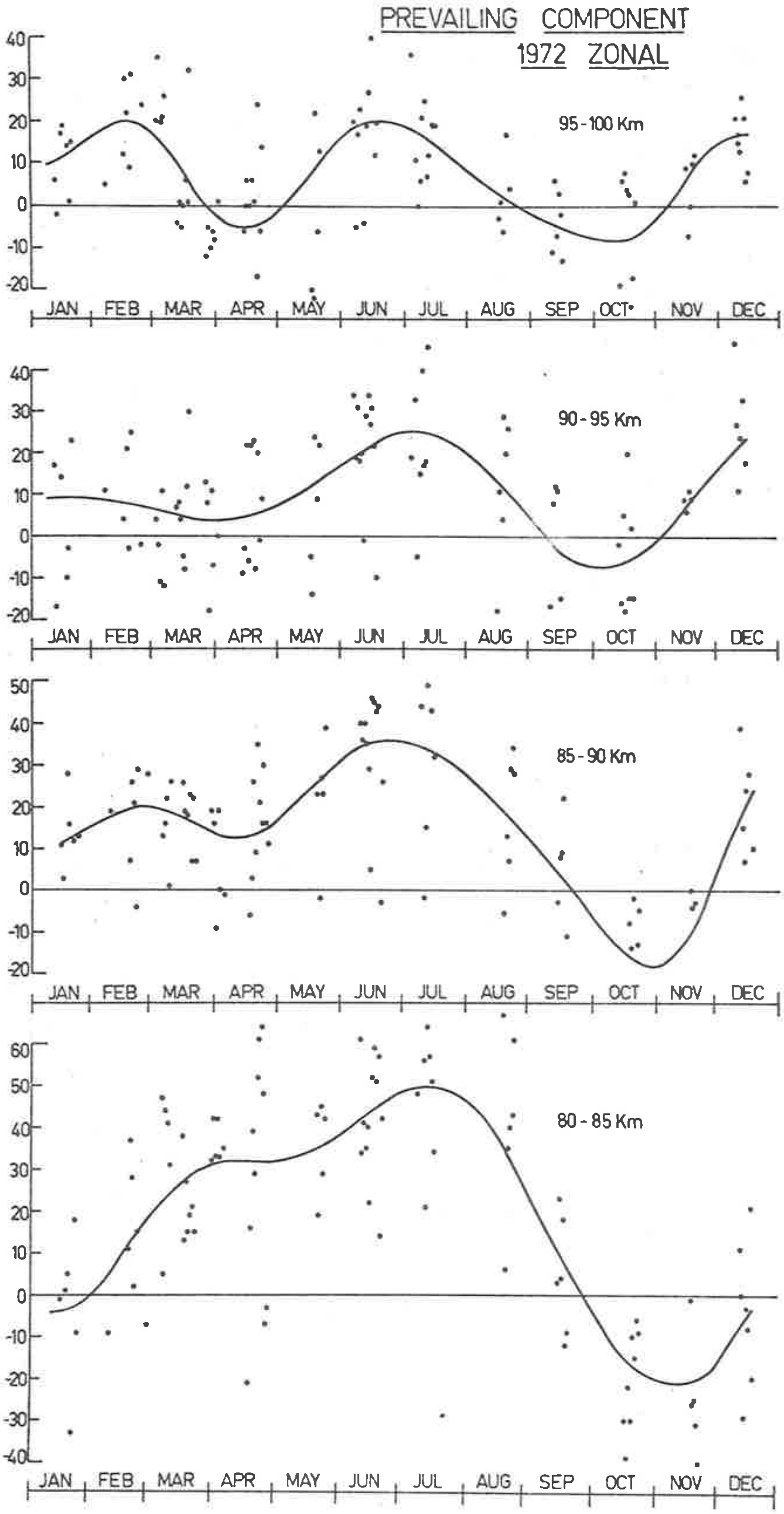


FIGURE 4.1 *The zonal component of the mean drift for individual days of 1972. The smooth curves are power series fitted by a least squares criterion.*

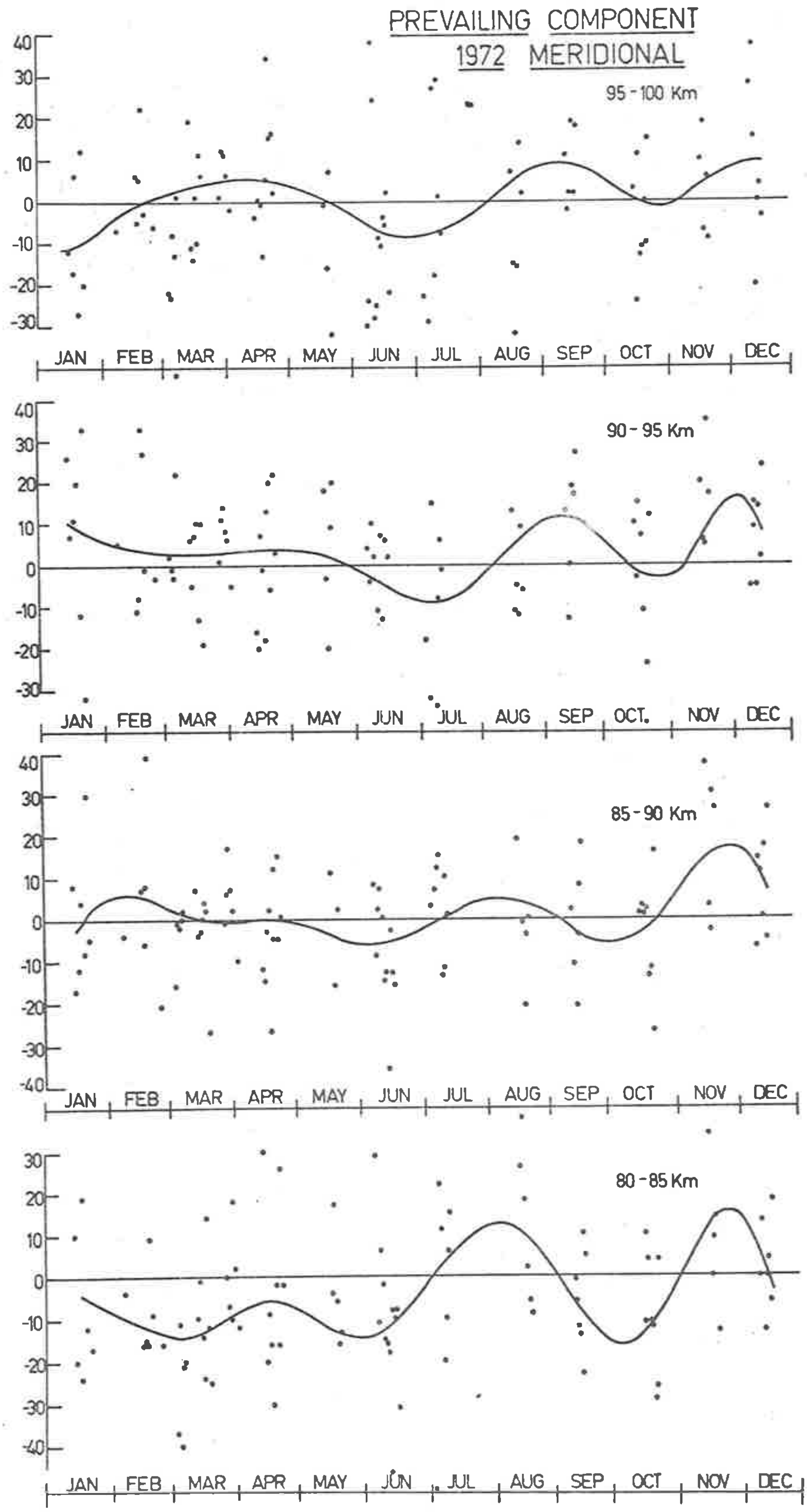


FIGURE 4.2 The meridional component of the mean drift for individual days of 1972. The smooth curves are power series fitted by a least squares criterion.

The physical content of the curves is perhaps clearer when they are reproduced in the form of contour maps. FIGURE 4.3 is the contour plot for the zonal component with the shaded sections representing westward drift. Below 95 km the drifts display the characteristics of the mesopheric monsoonal circulation - strongly eastward in winter with a westward motion in the late spring and early summer months. The reversal is most marked at the lower heights. Above 95 km, the autumn drifts and the sense of shear in the winter drifts indicate the intrusion of winter westward motion. In the meridional contour plot (FIGURE 4.4) the shaded areas represent southward drift. The meridional trends are less easy to delineate. The strongest southward belt occurs in winter and the strongest northward belt in summer. This corresponds to a poleward motion in winter and motion away from the pole in summer - again a fairly consistent feature of neutral wind results. There are other interesting aspects of the meridional contours but the velocities are generally too low to allow generalisations. One further feature which may be significant is the reversal in the sense of shear between the autumn months of March and April and the spring months of September and October.

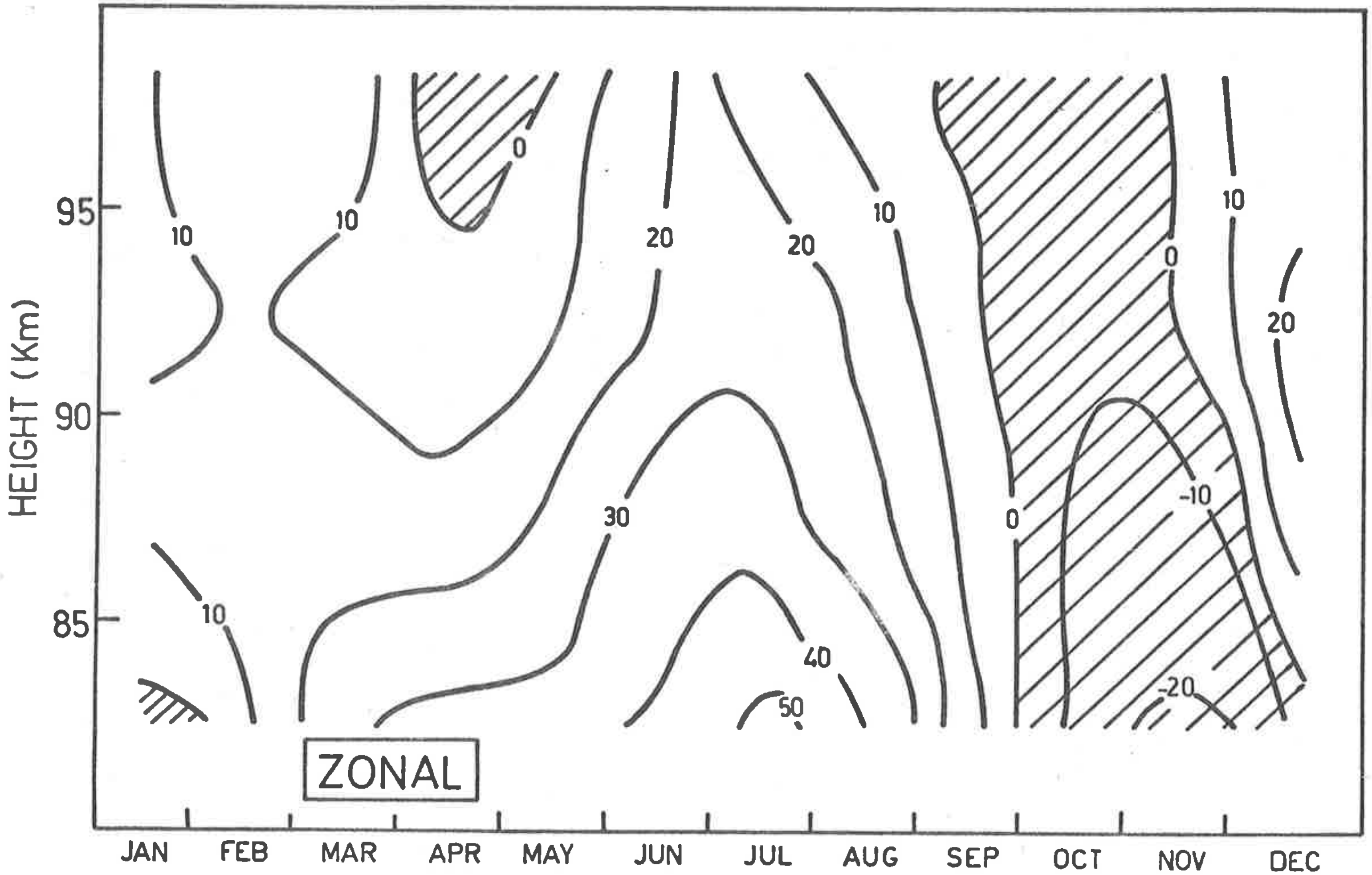


FIGURE 4.3 The prevailing drifts for 1972 - zonal component

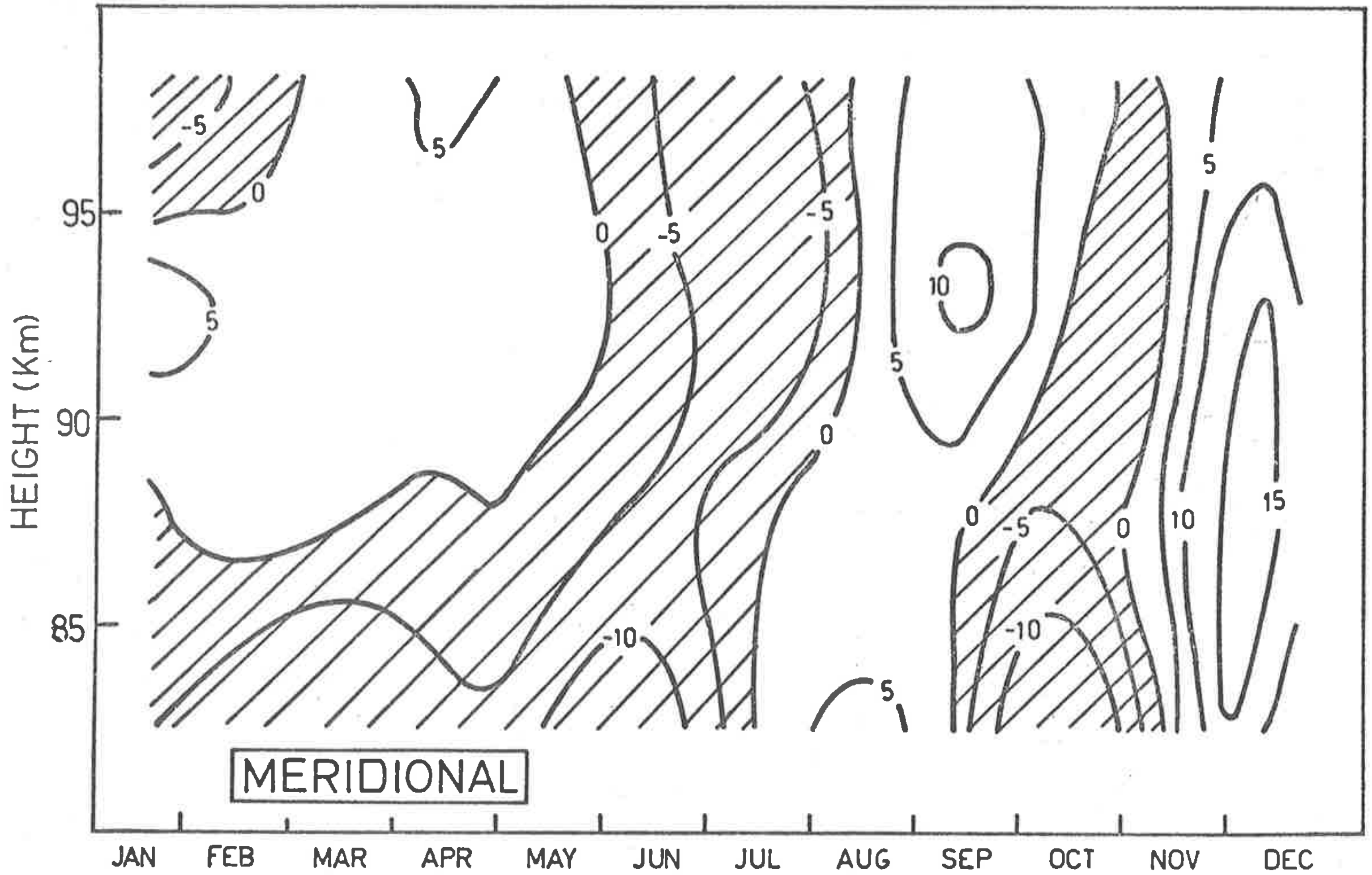


FIGURE 4.4 The prevailing drifts for 1972 - meridional component

4.1.2 Monthly averages

The monthly mean height profiles of the drifts provided a convenient form for comparison with a neutral wind model. One of the most recent and comprehensive of such models is that of GROVES (1969). It is based largely on rocket and meteor data from Northern Hemisphere stations. Values are given at monthly intervals for each 10 degrees of latitude over the height range 60-130 km. By interpolation, values for the middle of each month and for 35^0 latitude were obtained from the model for heights between 70 and 110 km. For each month of 1972 the drifts were then averaged for all 5 km intervals with sufficient data. All values below 80 km were necessarily obtained from daytime results only.

The drift and model height profiles are compared in FIGURES 4.5 - 4.8. The zonal profiles agree well, bearing in mind that there is undoubtedly considerable year-to-year variability in the prevailing wind (ELFORD and McAVANEY, 1971) and that the model includes data from many years of observation. The summer profiles in particular indicate more westward motion than the model would predict. In the meridional profiles, the model fairly consistently over-estimates the magnitudes of the drifts, particularly for

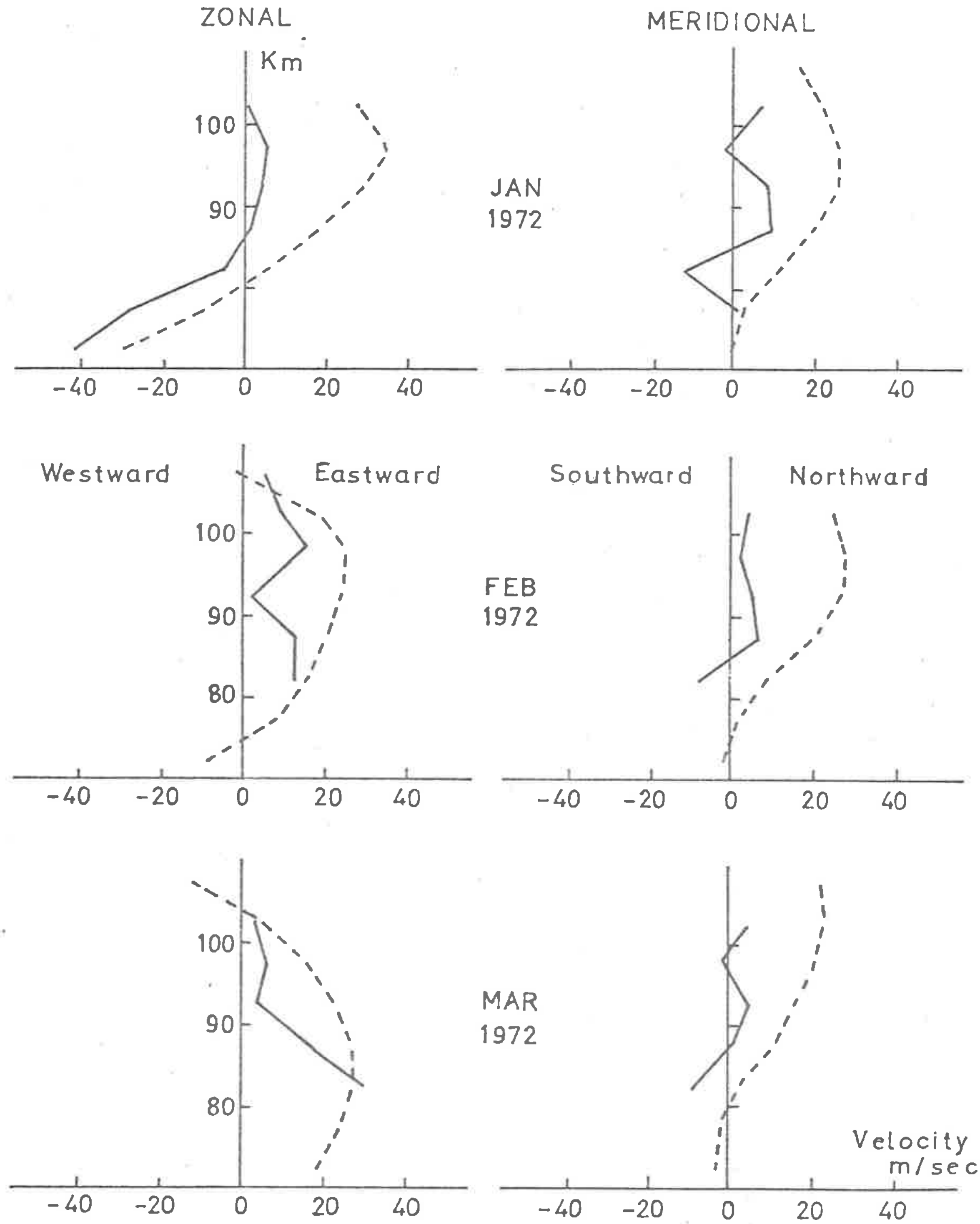


FIGURE 4.5 A comparison of monthly mean drift profiles (full lines) with the 1969 Groves wind model (broken lines)

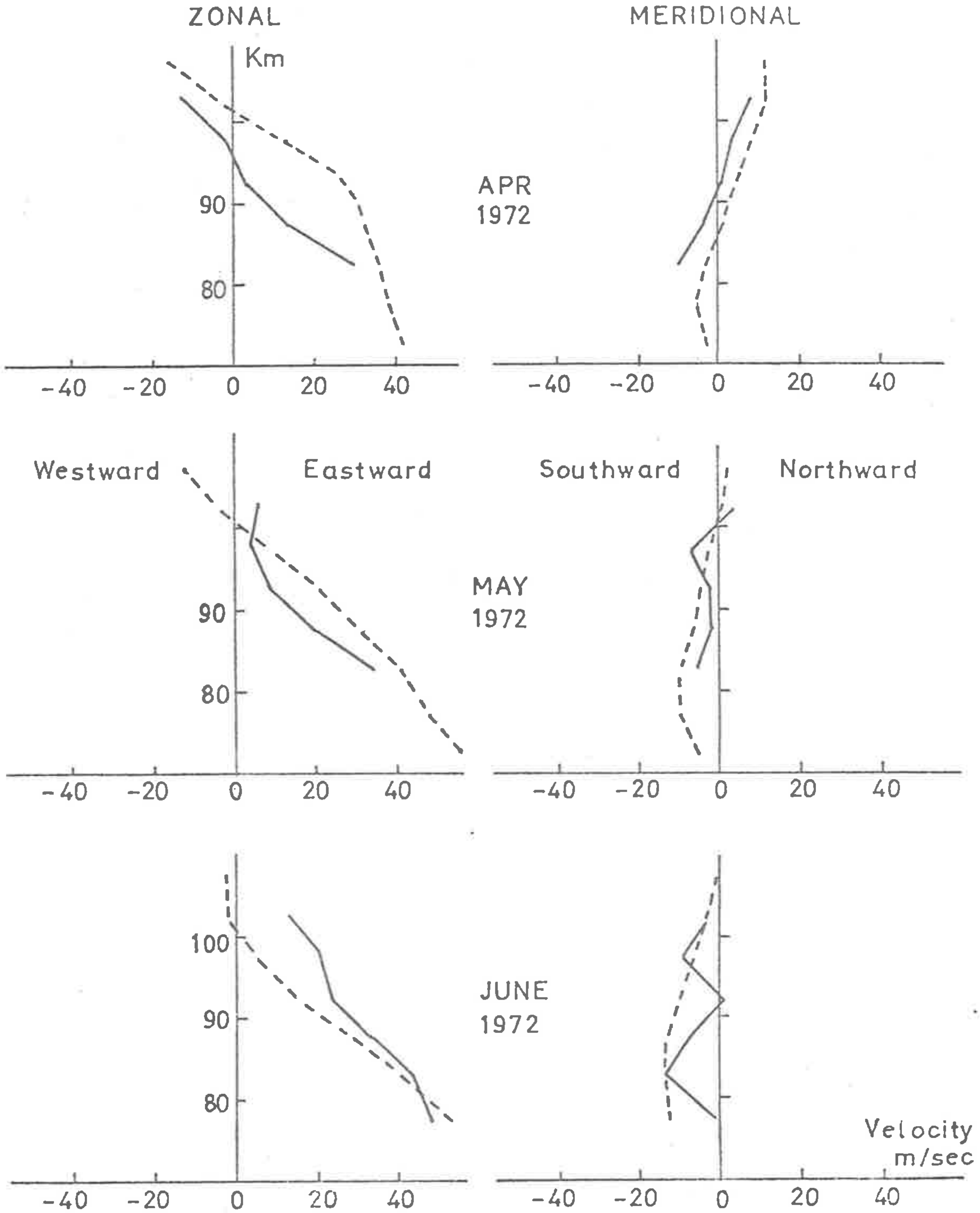


FIGURE 4.6 A comparison of monthly mean drift profiles (full lines) with the 1969 Groves wind model (broken lines).

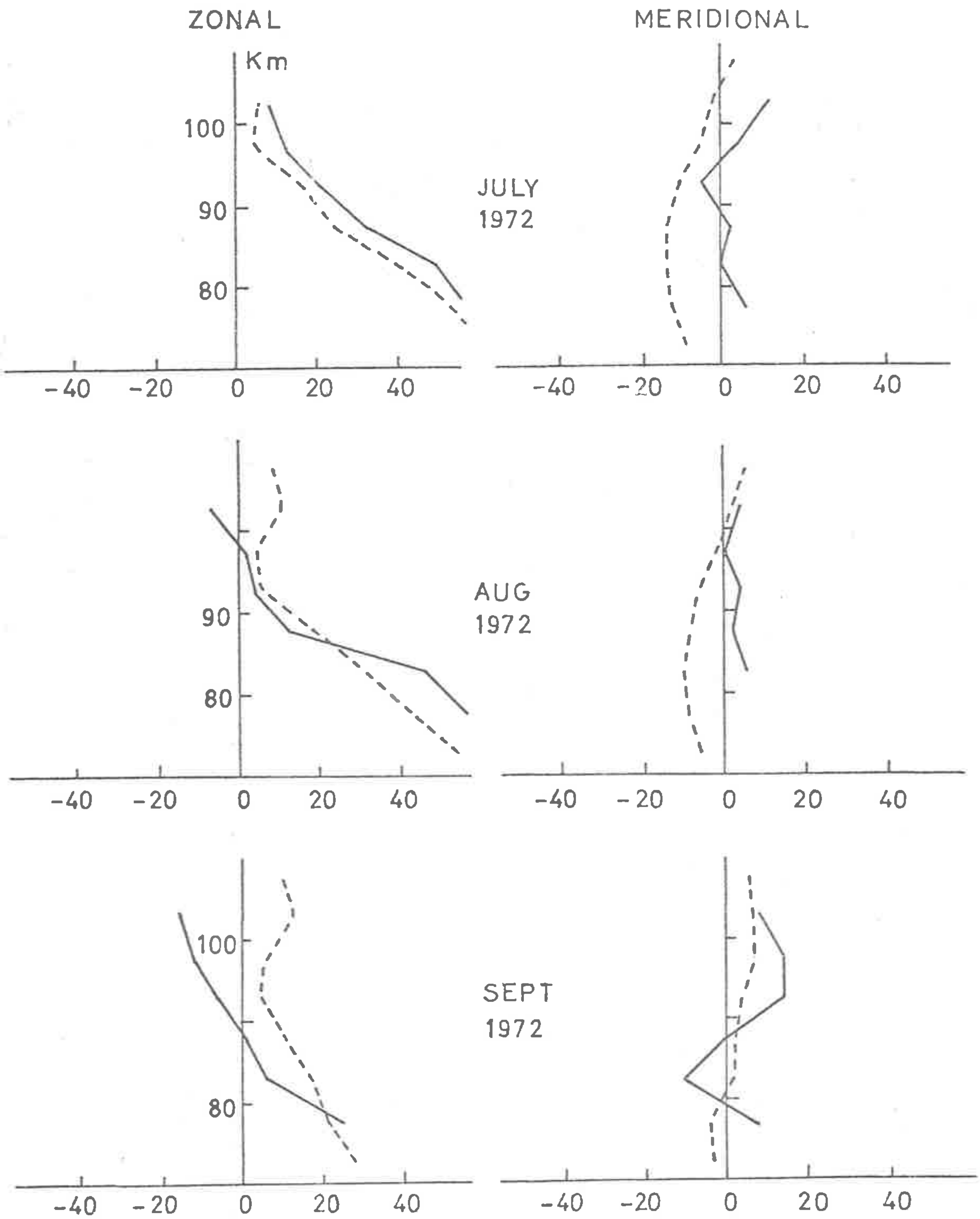


FIGURE 4.7 A comparison of monthly mean drift profiles (full lines) with the 1969 Groves wind model (broken lines).

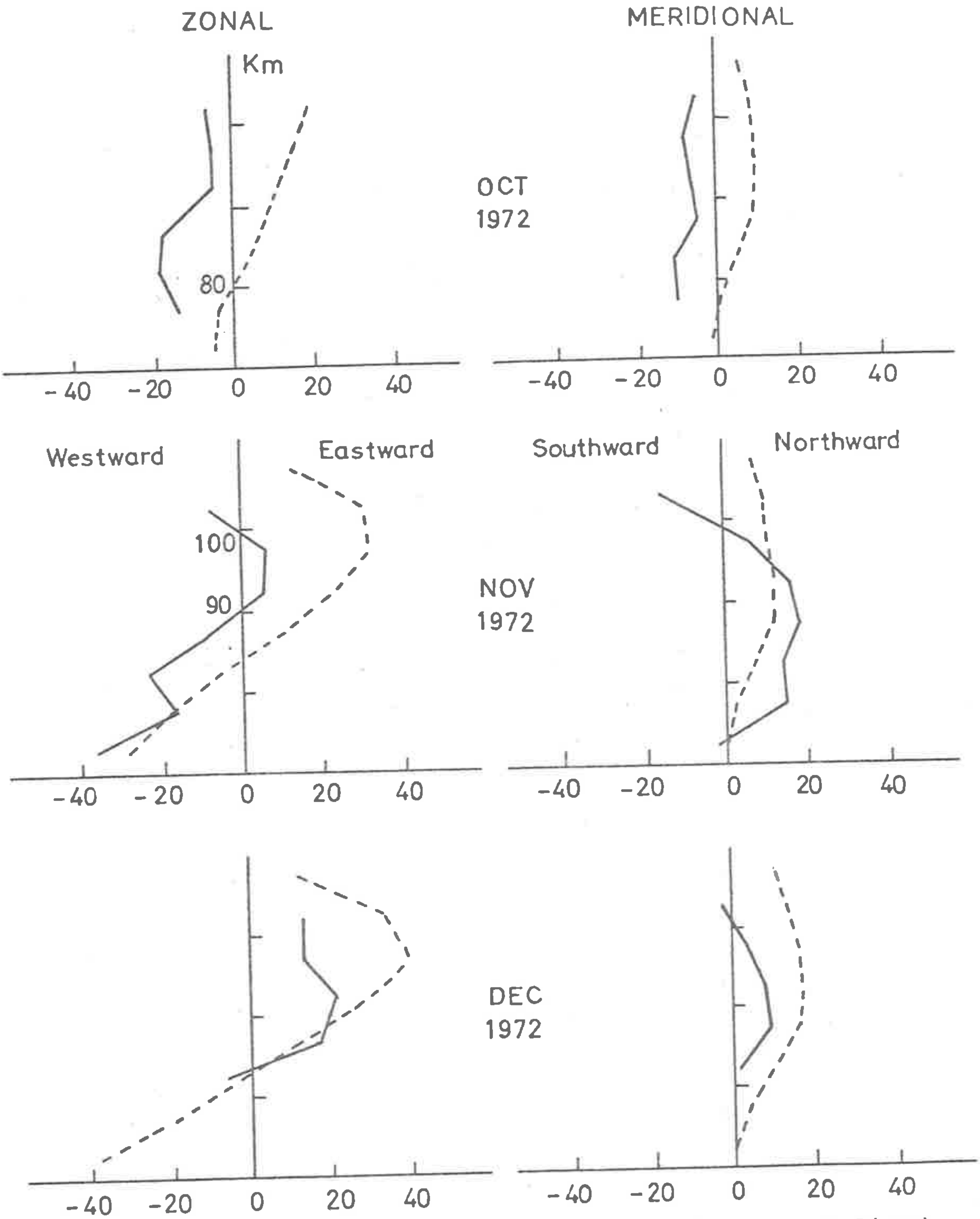


FIGURE 4.8 *A comparison of monthly mean drift profiles (full lines) with the 1969 Groves wind model (broken lines).*

the summer months of December, January and February. Overall, the model provides a much better picture of the winter behaviour than it does of the summer behaviour. This underlines the need for more Southern Hemisphere observing stations and for long-term observations aimed at a better understanding of how the mean motion changes over a number of years.

4.1.3 Planetary waves

In the discussion of FIGURES 4.1 and 4.2, mention was made of the large variations that the prevailing component of the drifts often exhibits within a short period. Although it is not clear from those plots because of the cramped time scale, these variations are sometimes quite systematic, following a wavelike pattern over several successive days.

The longest consistent example of this quasi-oscillatory behaviour in the mean drift is depicted in FIGURE 4.9 where the mean is plotted for six successive days of June 1973. At the time, 2 km gate widths were being used, and the drift results were averaged in 4 km blocks centred on 82, 86 and 90 km, the data continuity being poorer above 92 km. For both components at the three heights, there is a clear wavelike variation of the mean with a period in the vicinity of 4 or 5 days. It seems likely that the fluctuations

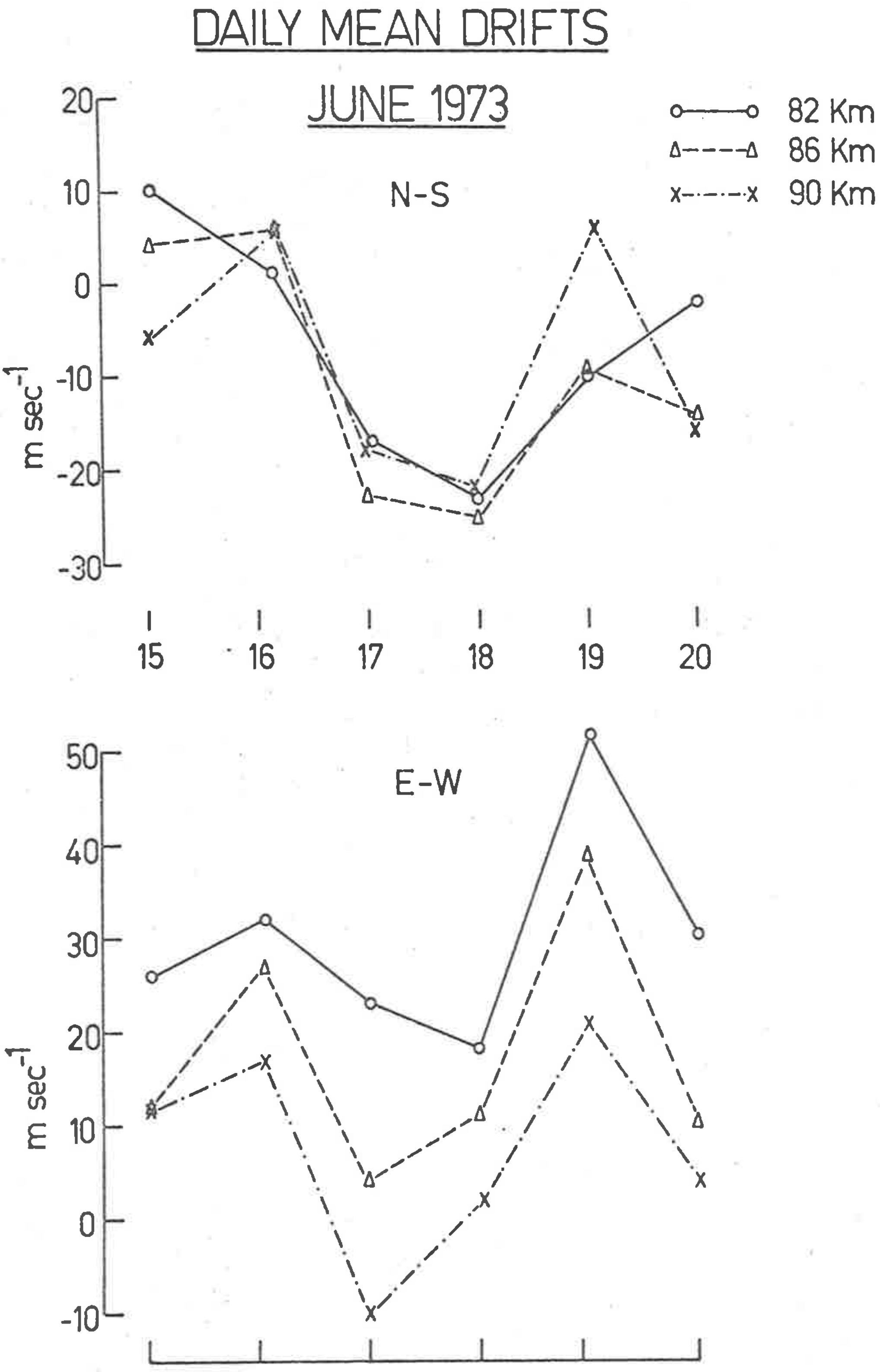


FIGURE 4.9 *The mean drift for six successive days of June, 1973.*

resulted from the passage through the region of a planetary wave.

4.2. Periodic Components

The importance of periodic components in the D-region drifts can be most easily seen from polar plots of the hourly wind vector. In many cases, the drift consists almost entirely of oscillations with 24- and 12- hour periods. FIGURE 4.10 shows four examples with a predominantly diurnal variation. Each numbered point represents the appropriate hourly drift average. The position of the point gives the direction towards which the ionosphere is drifting, and its distance from the origin gives the magnitude of the drift. In the four examples of FIGURE 4.11 the variation is basically semidiurnal.

The anticlockwise sense of rotation of the drift vector throughout the day, evident in all diagrams, is consistent with the typical Southern Hemisphere behaviour of the solar tides. It has been pointed out that the direction of rotation is more dependent on the tidal modes of importance than the hemisphere under consideration (BLAMONT and TEITELBAUM, 1968), but at 35^{0}S the expected rotation is anticlockwise. In analysing the drifts, the zonal and meridional components have been independently subjected to harmonic analysis. The phase of the zonal component is

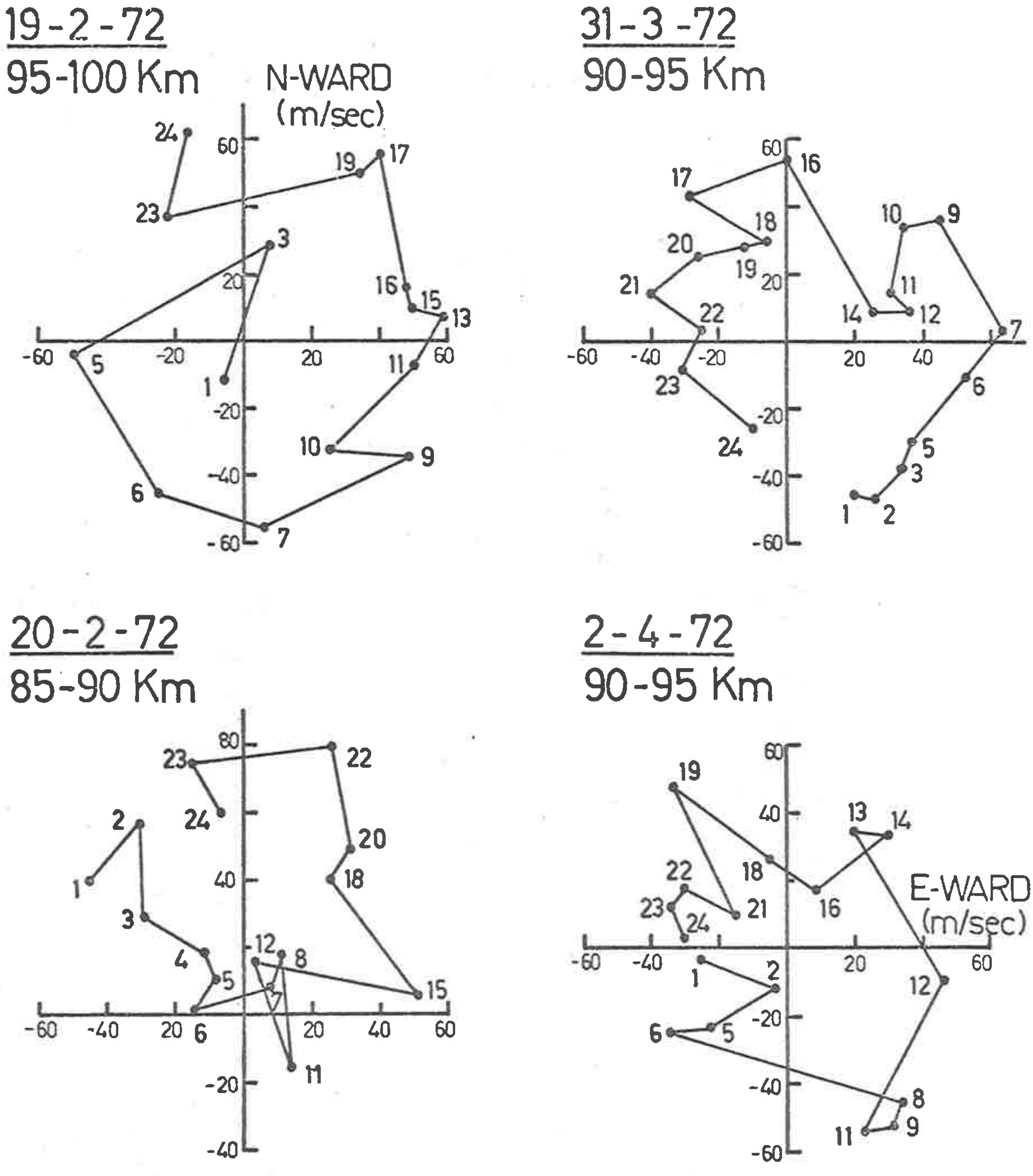


FIGURE 4.10 *Polar plots of the diurnal variation in the hourly drift averages.*

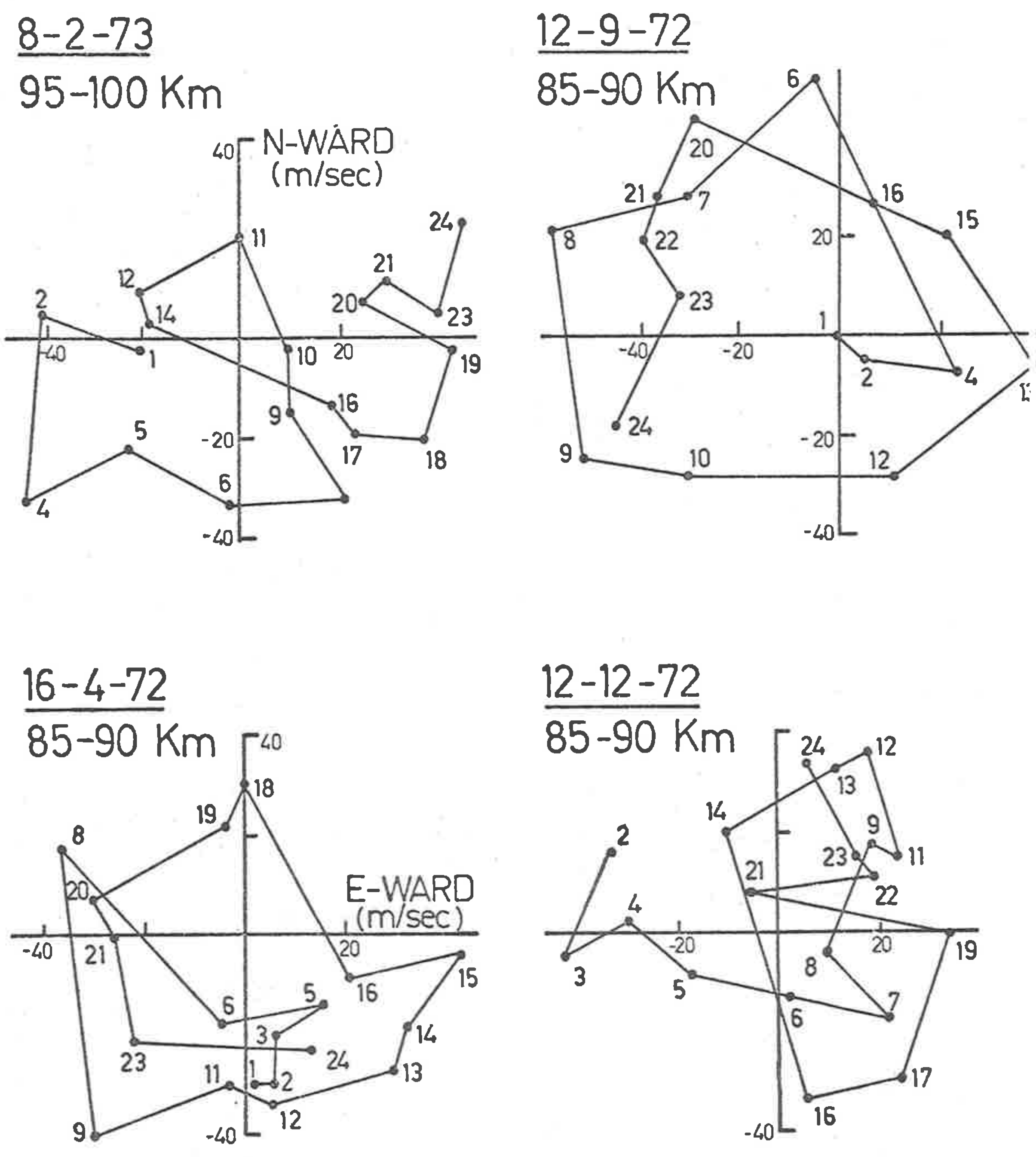


FIGURE 4.11 *Polar plots of the diurnal variation in the hourly drift averages.*

defined as the local time in hours of the maximum eastward drift and the phase of the meridional component is defined as the time of northward maximum. For an anticlockwise, circular, tidal rotation, the northward maximum will lag the eastward maximum by 6 hours in the case of a diurnal oscillation and by 3 hours for a semidiurnal oscillation.

The word 'tide' is used in its most general sense in this thesis because of its convenience. The term should properly be associated with particular modes. However the unambiguous identification of tidal modes requires observations over a range of latitudes (TESTUD, 1973). It might also be argued that a tide should be deduced from repeated oscillations over a series of days rather than from a single day. This would be feasible if the atmospheric oscillations were always stable, as for example in the long diurnal sequence in March, 1972 (FIGURES 3.5, 3.6). However one of the remarkable features of the results is the variability of the tides, and particularly the amplitude of the diurnal component. Two examples of dramatic decay in the tidal amplitude are given in FIGURES 4.12 and 4.13. The origin of this variability is as yet unknown. It may be due to interference between various tidal modes, to changes of composition in the source region or to changes in the damping effects of molecular viscosity.

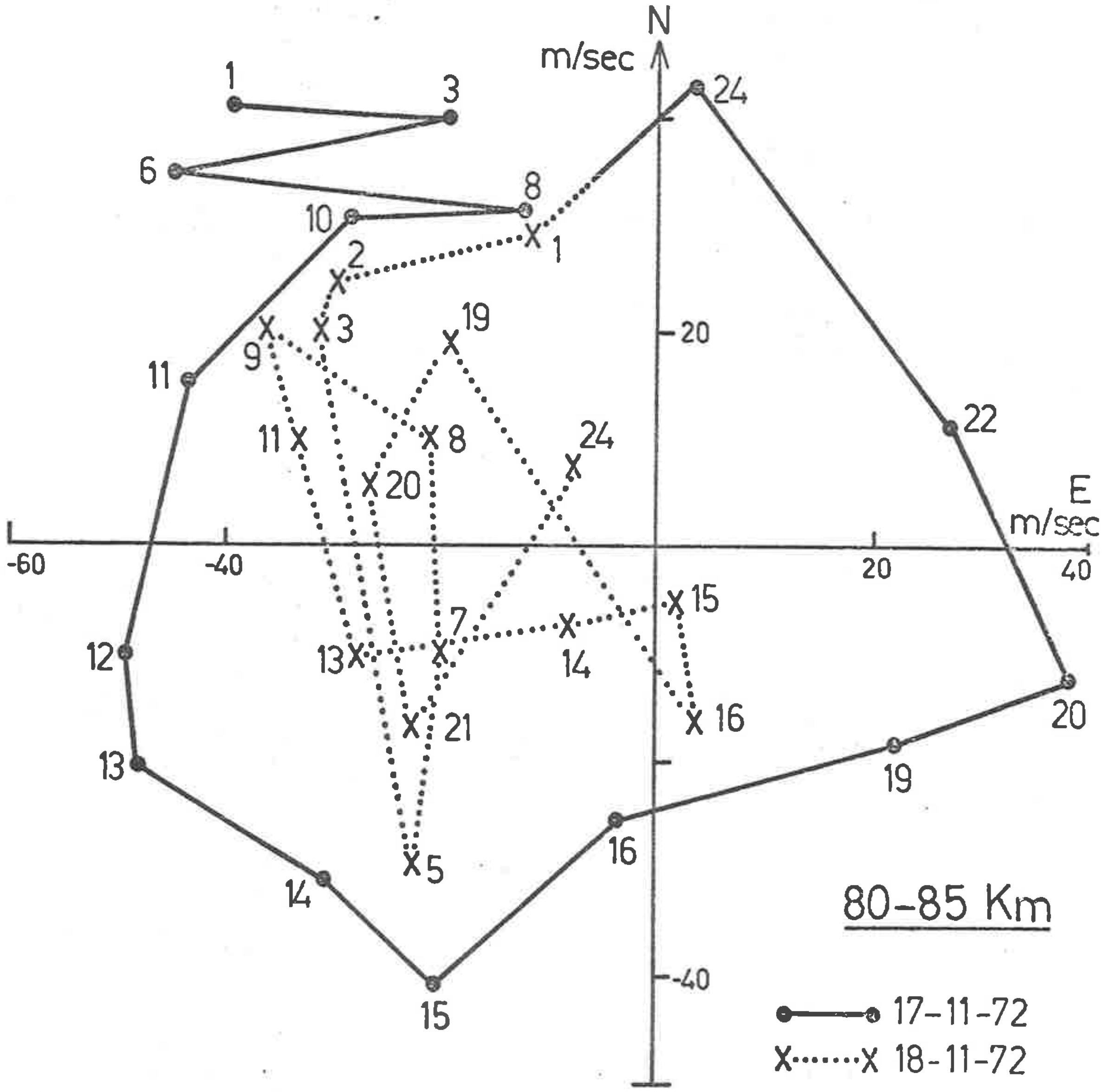


FIGURE 4.12 Polar plot of hourly drift averages on the 17th (full lines) and 18th (dotted lines) of November, 1972.

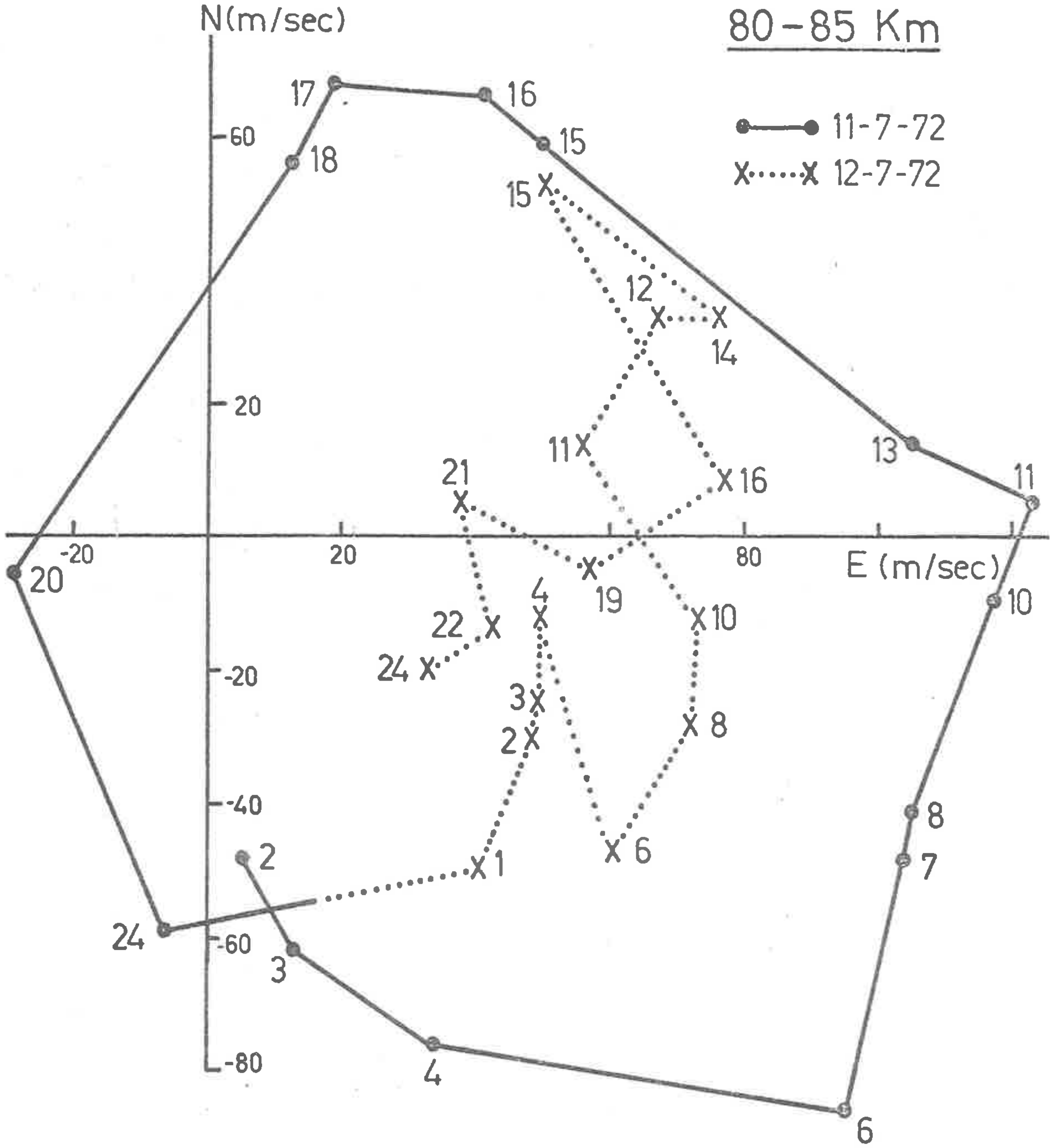


FIGURE 4.13 *Polar plot of hourly drift averages on the 11th (full lines) and 12th (dotted lines) of July, 1972.*

It has also been suggested (HINES, private communication) that the changes may result from interaction between the tides and the meridional component of the background wind.

4.2.1 24 - hour period

The way in which the scalar amplitude of the diurnal tide at 85 - 90 km varied throughout 1972 is shown in FIGURE 4.14. Each point represents the average of the zonal and meridional amplitudes for a particular day and the solid line joins the mean amplitudes for each month. The July peak is probably misleading because of the influence of the one very large value. The main feature of the plot is the consistently large amplitude during the late summer and the autumn. There are also indications of minima in early winter and early summer.

The tidal data can be more meaningfully displayed by making use of harmonic dials. Both phase and amplitude information can then be used to obtain a vector average for a particular period. The results are plotted in this way in FIGURES 4.15 - 4.17, being grouped into months or pairs of months, depending on the number of data points. The small open circles represent the amplitude and phase of the zonal (left half of diagrams) and meridional components (right half) for the individual days adjacent to them. The

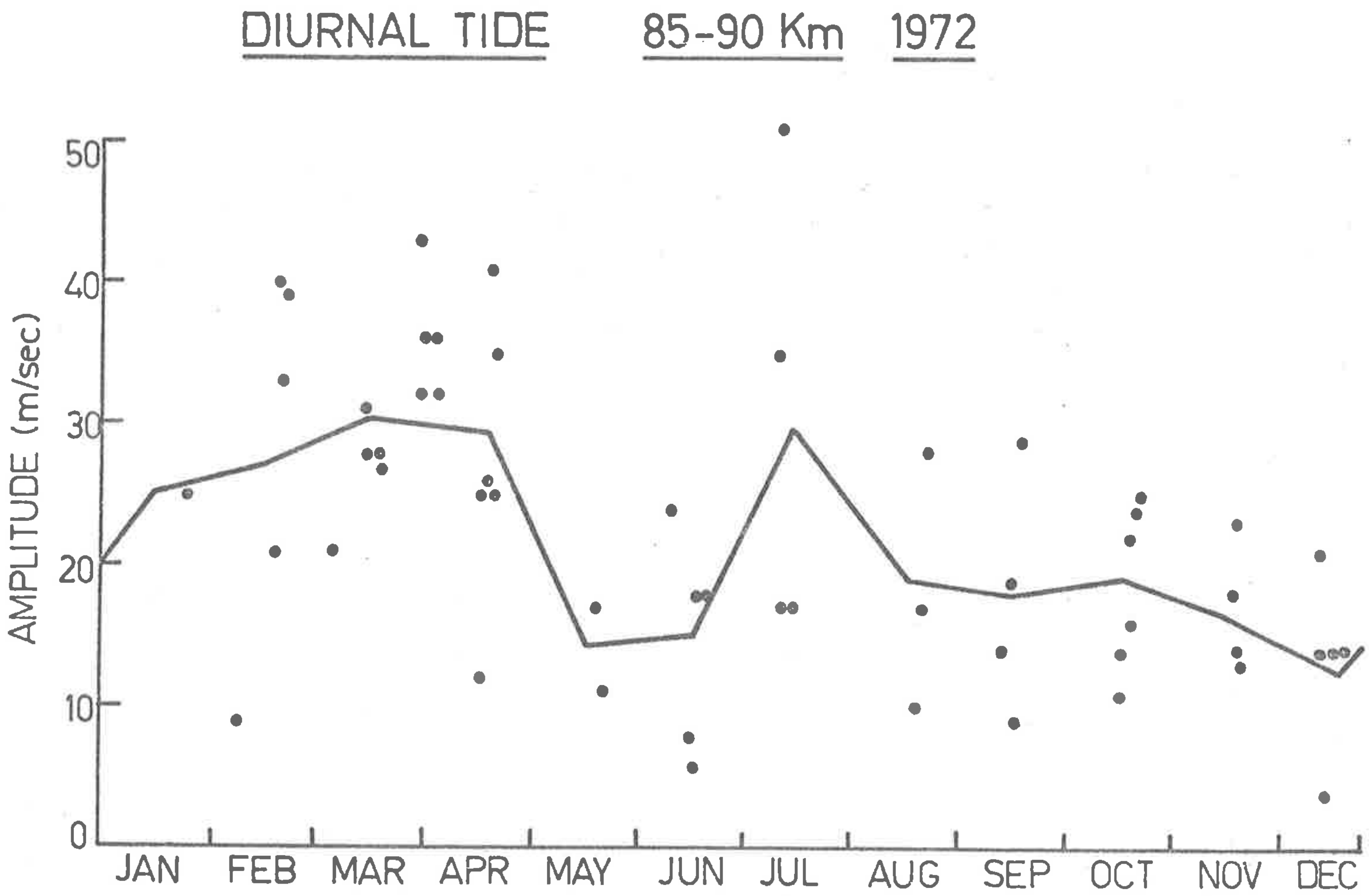


FIGURE 4.14 *The scalar amplitude of the diurnal tide for individual days of 1972. The full line joins the respective monthly averages.*

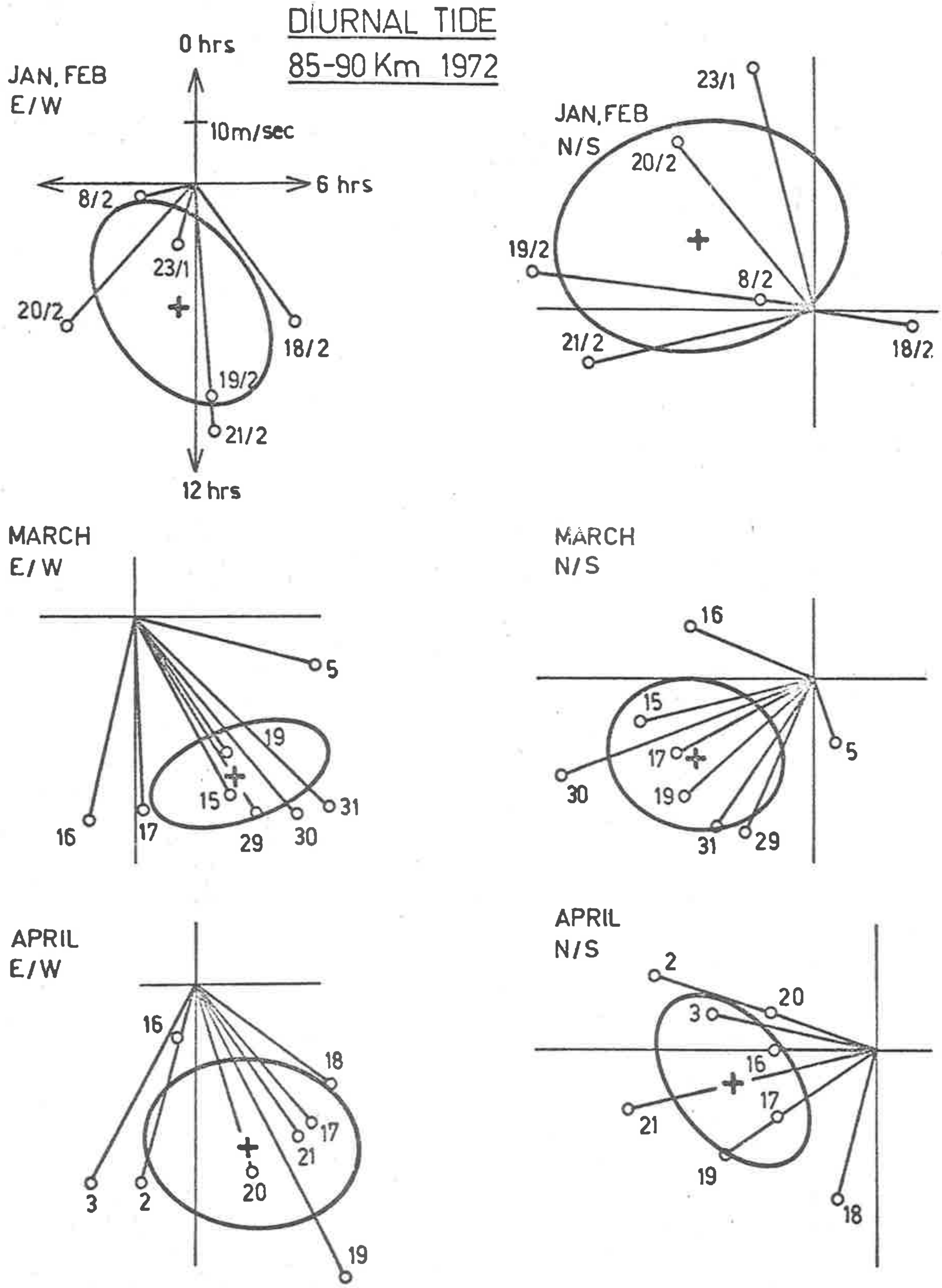


FIGURE 4.15 *Harmonic dials of the diurnal tide as deduced from the 1972 drift results.*

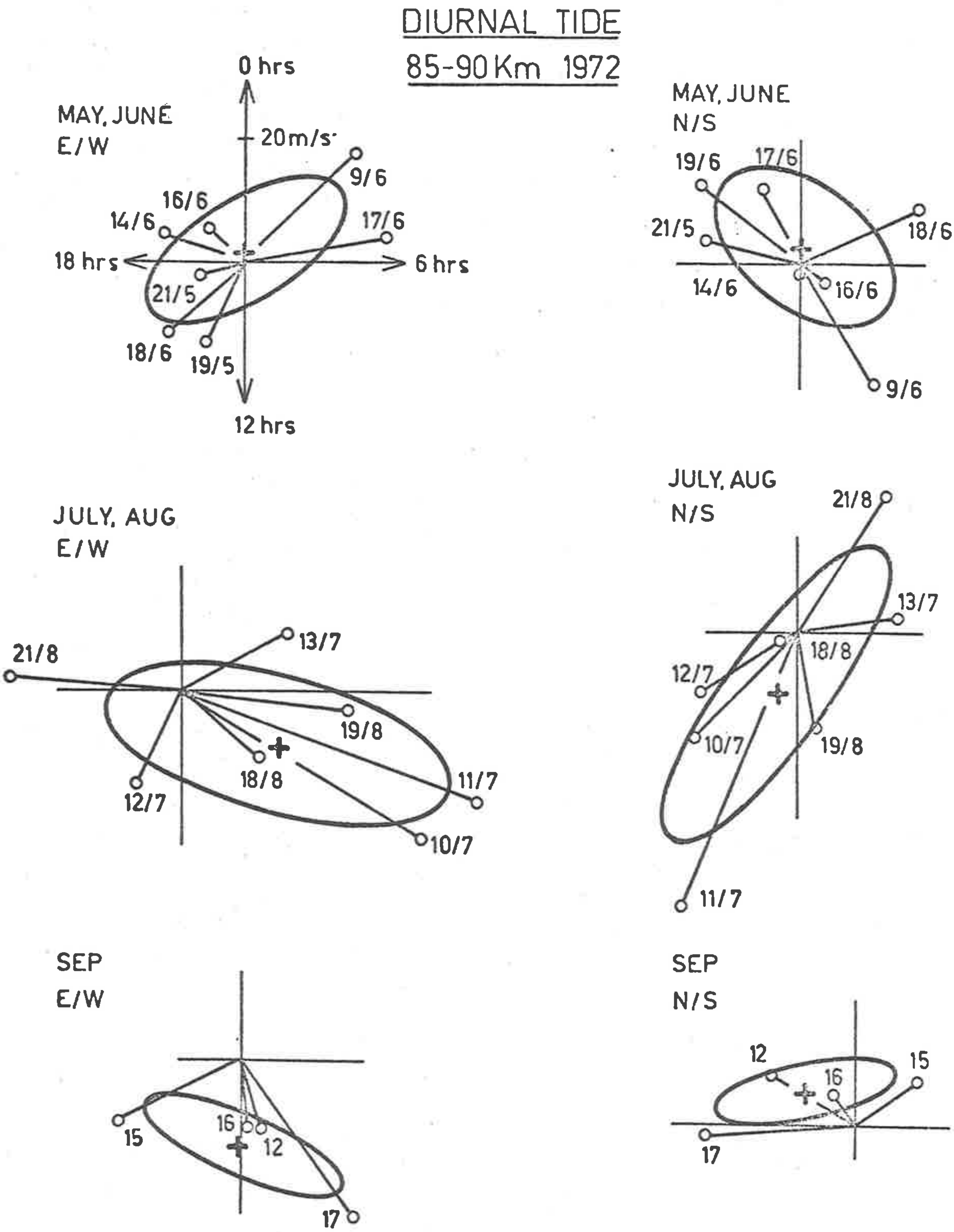


FIGURE 4.16 *Harmonic dials of the diurnal tide as deduced from the 1972 drift results.*

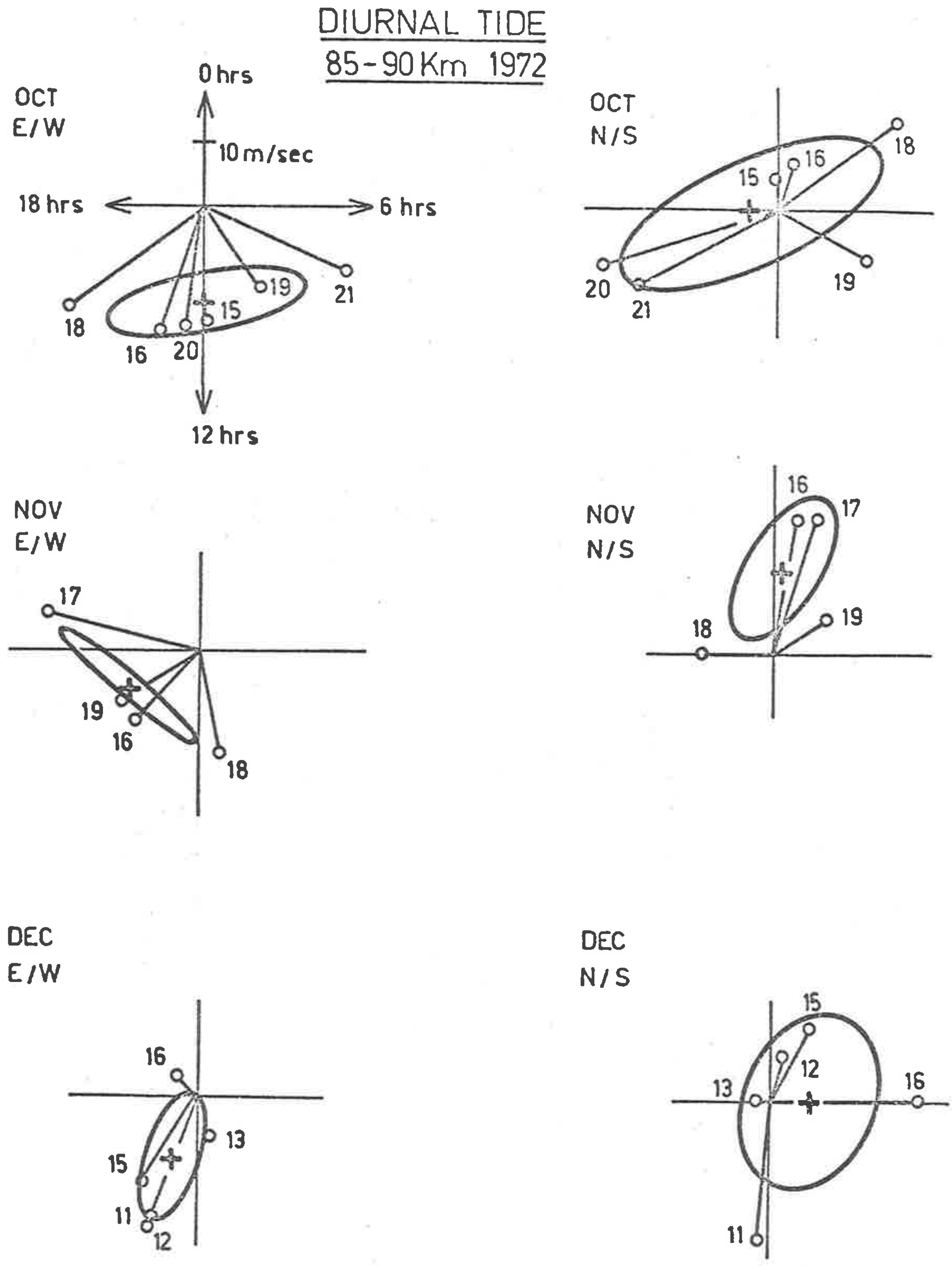


FIGURE 4.17 *Harmonic dials of the diurnal tide as deduced from the 1972 drift results.*

cross and probable error ellipse are the result of a vector averaging. There are several interesting features of these averages:

(i) December is the only month when the relative phase lag of the meridional component is not close to six hours. In general, the amplitudes determined independently from the E/W and N/S components agree well, although there is sometimes a suggestion of ellipticity.

(ii) In autumn the amplitude is greatest and successive days are not usually very different in phase.

(iii) In winter, the diurnal component varies unpredictably in phase from day to day with the average contribution being small, particularly in June.

(iv) In fact, with the exception of winter, the phase of the diurnal tide appears to be fairly stable over the year, with the eastward maximum generally occurring within a few hours of local noon.

Another way in which wave effects can be studied is in terms of vertical structure. Rocket firings, for example, usually produce a vertical wavelength, which characterises the wind profile. However the only way that this can be associated unambiguously with a tide is if information is also available about how the profile changes in time. The drift technique is capable of providing both height profiles, albeit over a somewhat limited range, and detailed time sequences.

During several days of March, 1973, drifts were measured at 2 km intervals between 80 and 100 km, the height range often being scanned at several different receiver gain settings to obtain data from as many intervals as possible. There was clear evidence of a dominant 24 hour oscillation at all heights. In FIGURE 4.18, the results have been averaged in three-hour blocks and sequential zonal profiles plotted over a 24-hour period. The resultant picture is very much like looking, at 3-hourly intervals, at a basically monochromatic wave moving down through the observed region. A similar pattern is evident in the meridional profiles (FIGURE 4.19) with the same section of the wave being seen 2 frames (6 hours) later. In general, the impression is of being able to see about half the wave which appears to have an amplitude of around 60 m/sec and a vertical wavelength in the vicinity of 35 km. A similar sequence, this time over 12 hours, is shown in FIGURE 4.20.

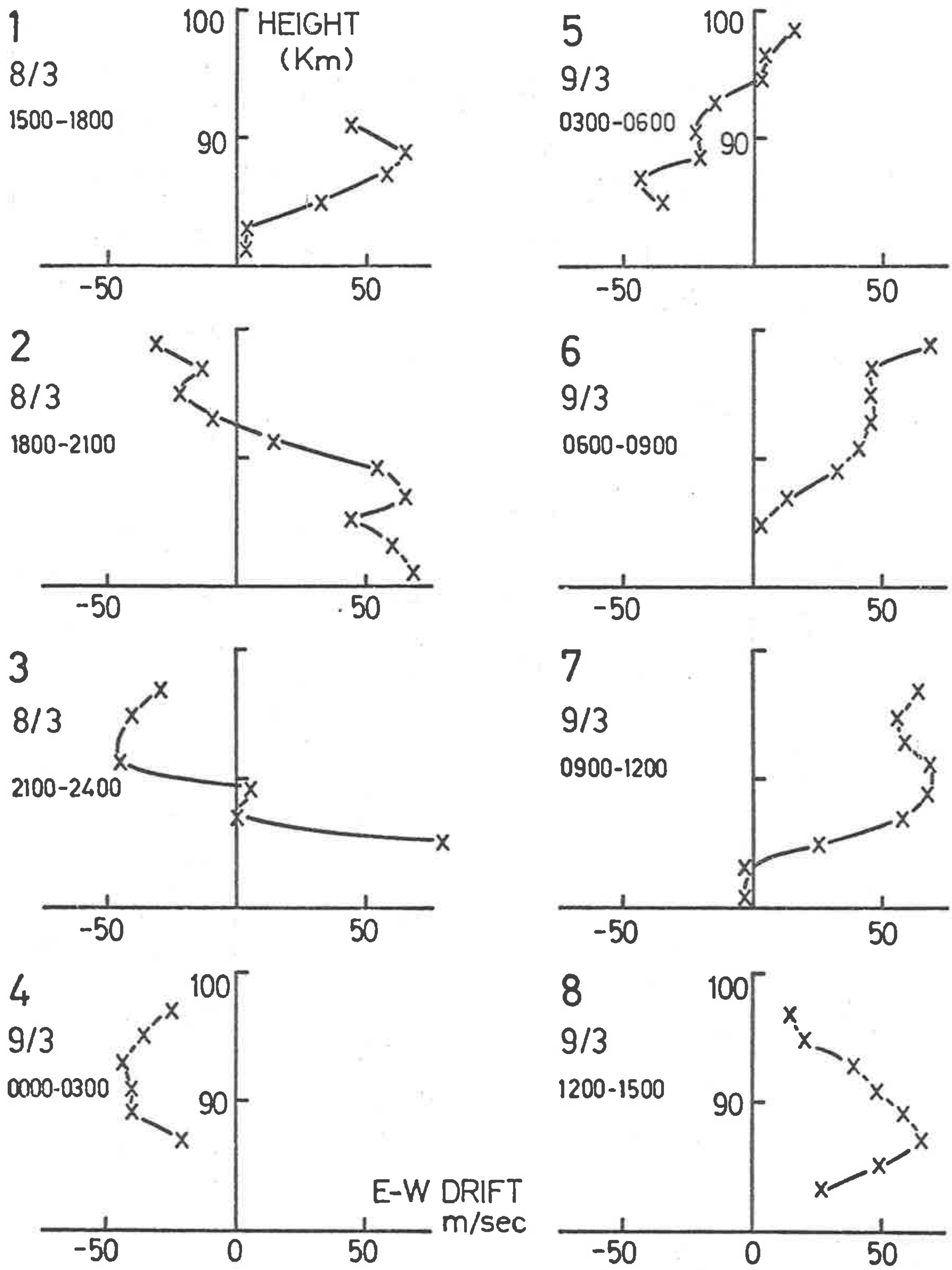


FIGURE 4.18 *Sequential three-hourly average drift profiles - zonal component.*

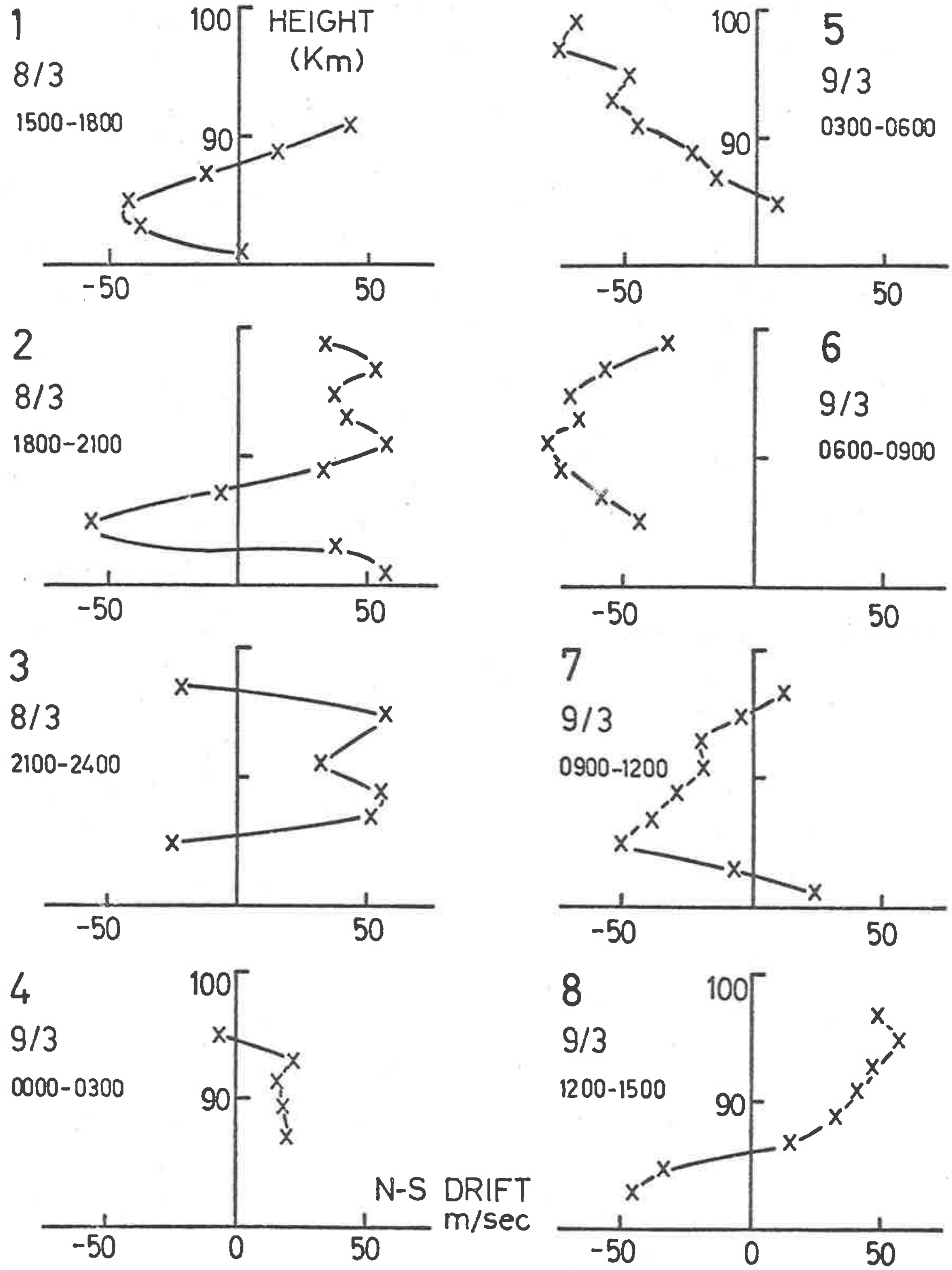


FIGURE 4.19 *Sequential three-hourly average drift profiles - meridional components.*

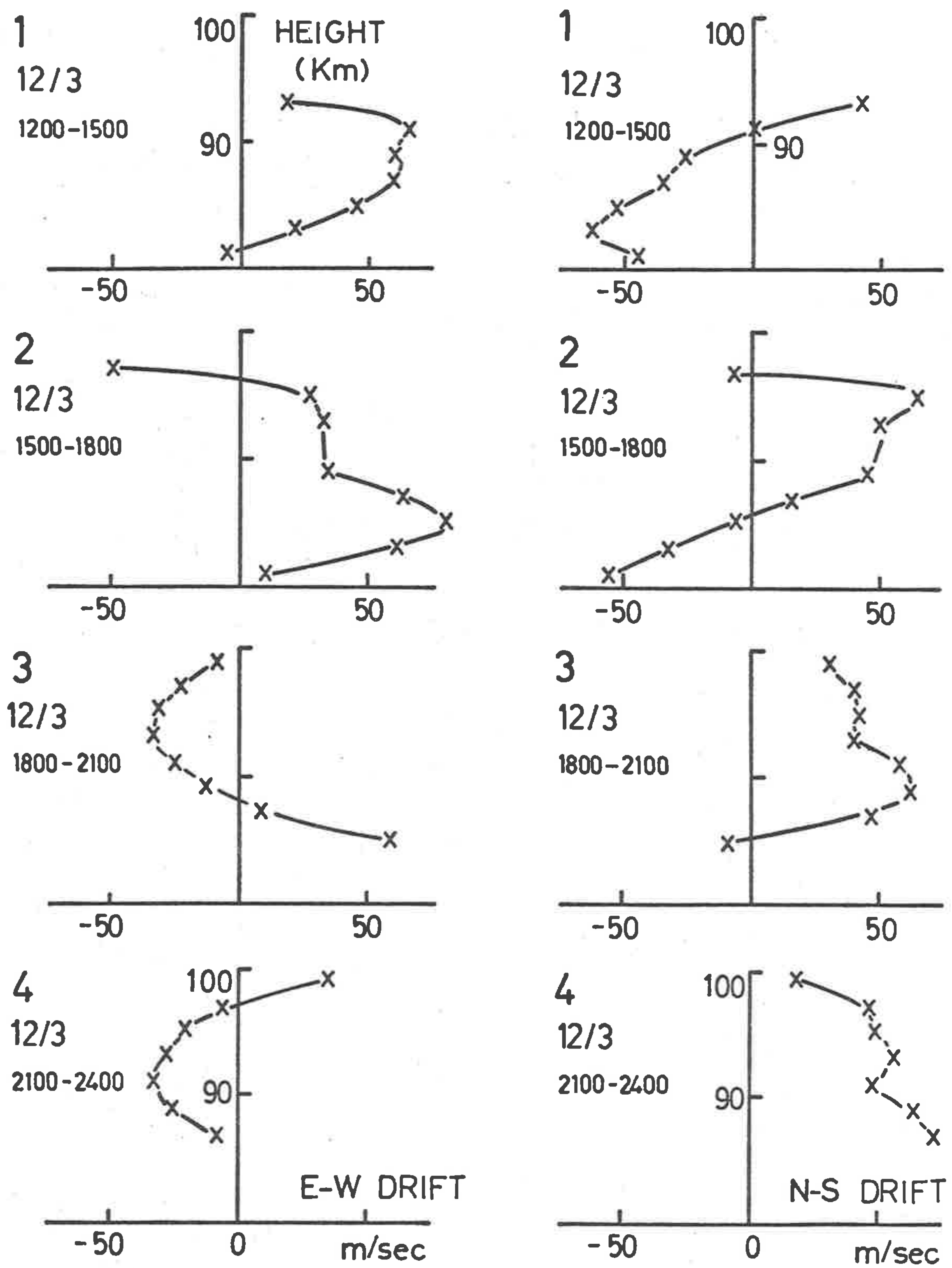


FIGURE 4.20 *Sequential three-hourly average drift profiles - zonal (left) and meridional components.*

4.2.2 12-hour period

The scalar amplitudes of the semidiurnal oscillation at 85-90 km are plotted in FIGURE 4.21. There seems to be a fairly clear semiannual effect with maxima near the autumn and spring equinoxes. A comparison with the corresponding graph of the diurnal amplitudes (FIGURE 4.14, plotted on a smaller scale because of the large values) shows that in spring the semidiurnal component attains an equal importance to the diurnal component but for most of the year, and particularly in autumn, it plays the lesser role.

Harmonic dials for the 12-hour oscillation appear in FIGURES 4.22 - 4.24. Again they are quite illuminating:

(i) Even where the error ellipse is large compared to the average value, the tide seems well determined with a three hour phase shift, in the sense of an anticlockwise rotation, being evident in all months except during the winter.

(ii) Early in winter (May, June) the amplitudes are small as was the case for the diurnal tide, but the phase seems much more consistent.

(iii) The phase appears to be most variable in the months just before the equinoxes and most stable

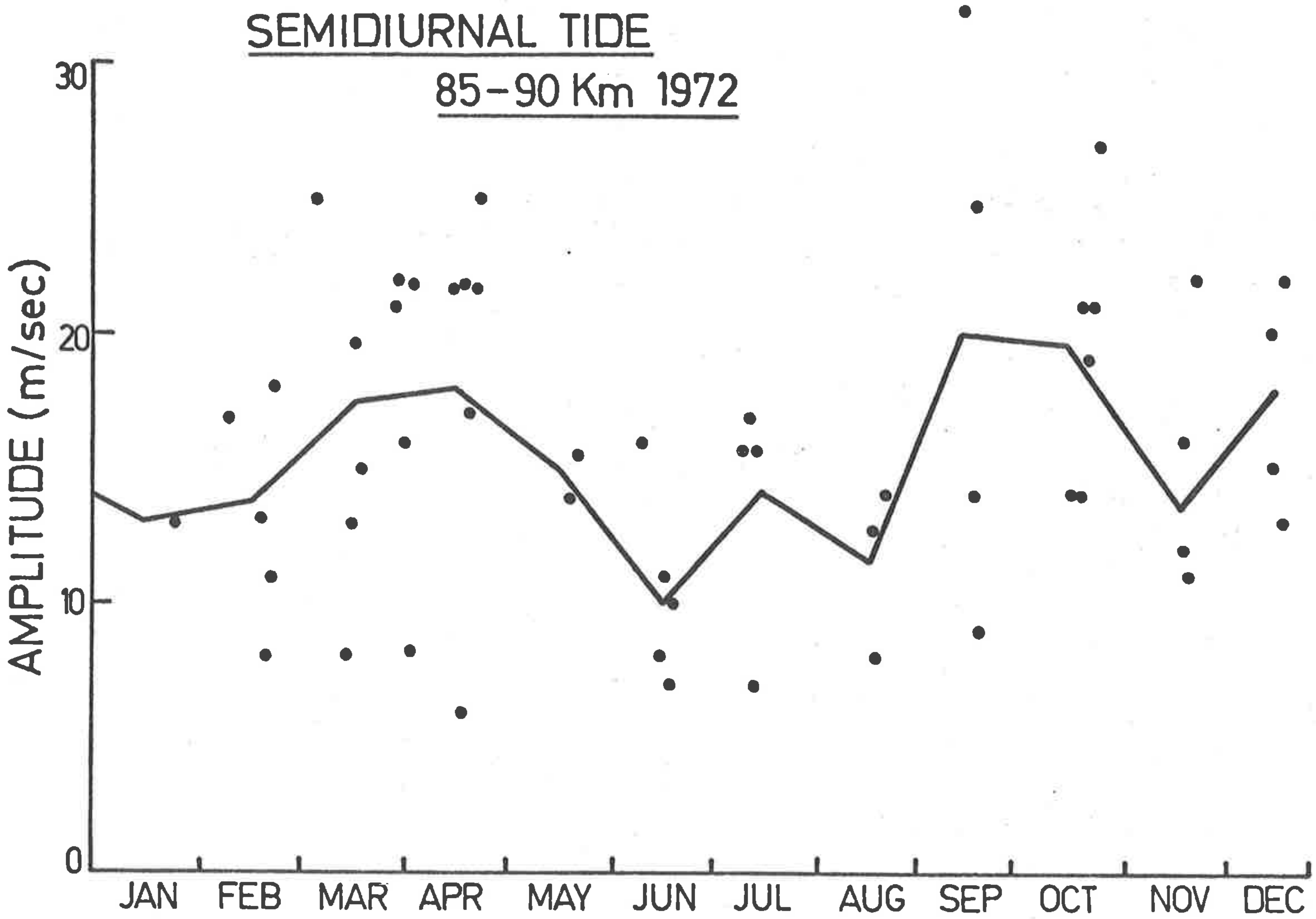


FIGURE 4.21 *The scalar amplitude of the semidiurnal tide for individual days of 1972. The full line joins the respective monthly averages.*

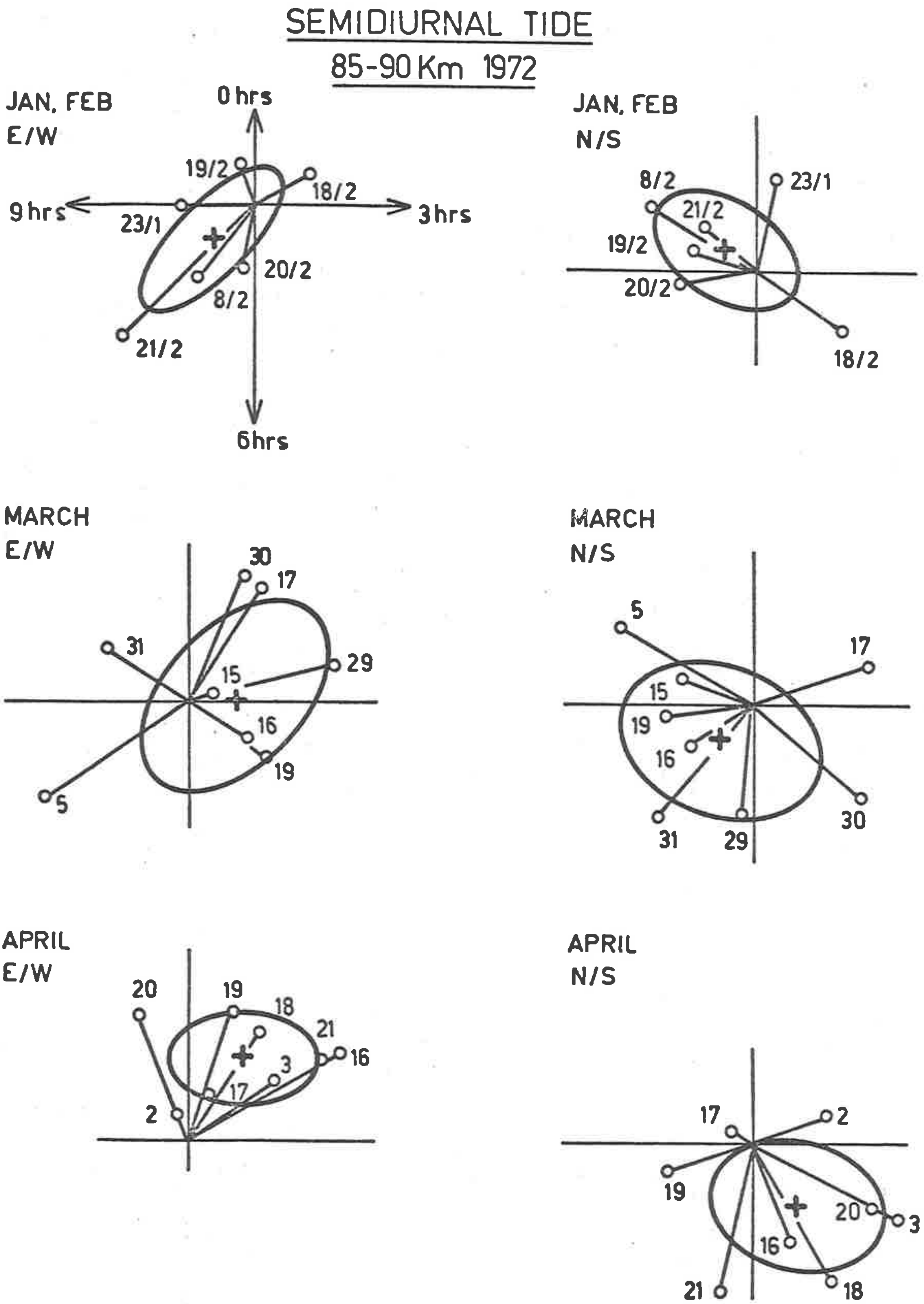


FIGURE 4.22 *Harmonic dials of the semidiurnal tide as deduced from the 1972 drift results.*

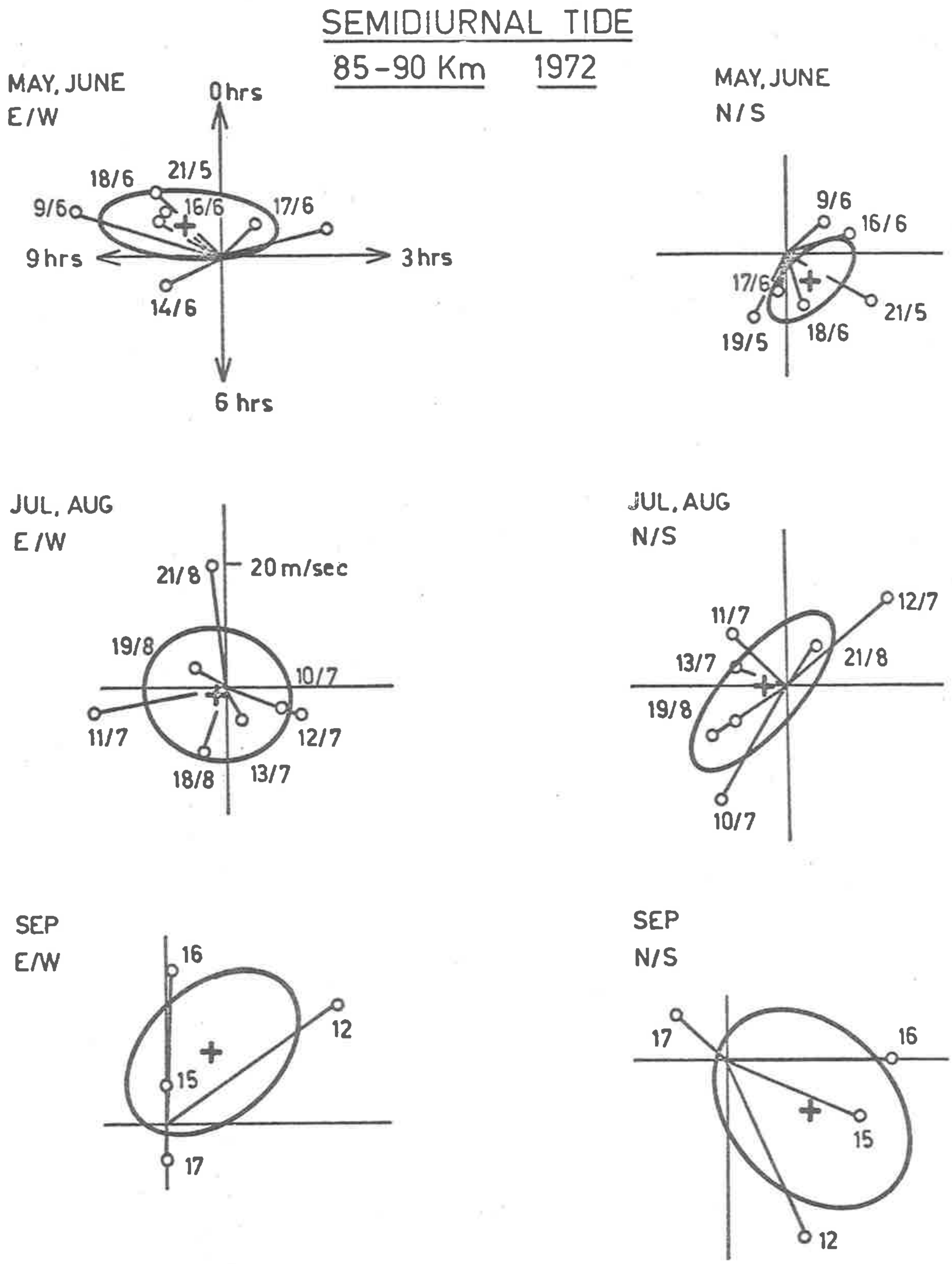


FIGURE 4.23 *Harmonic dials of the semidiurnal tide as deduced from the 1972 drift results.*

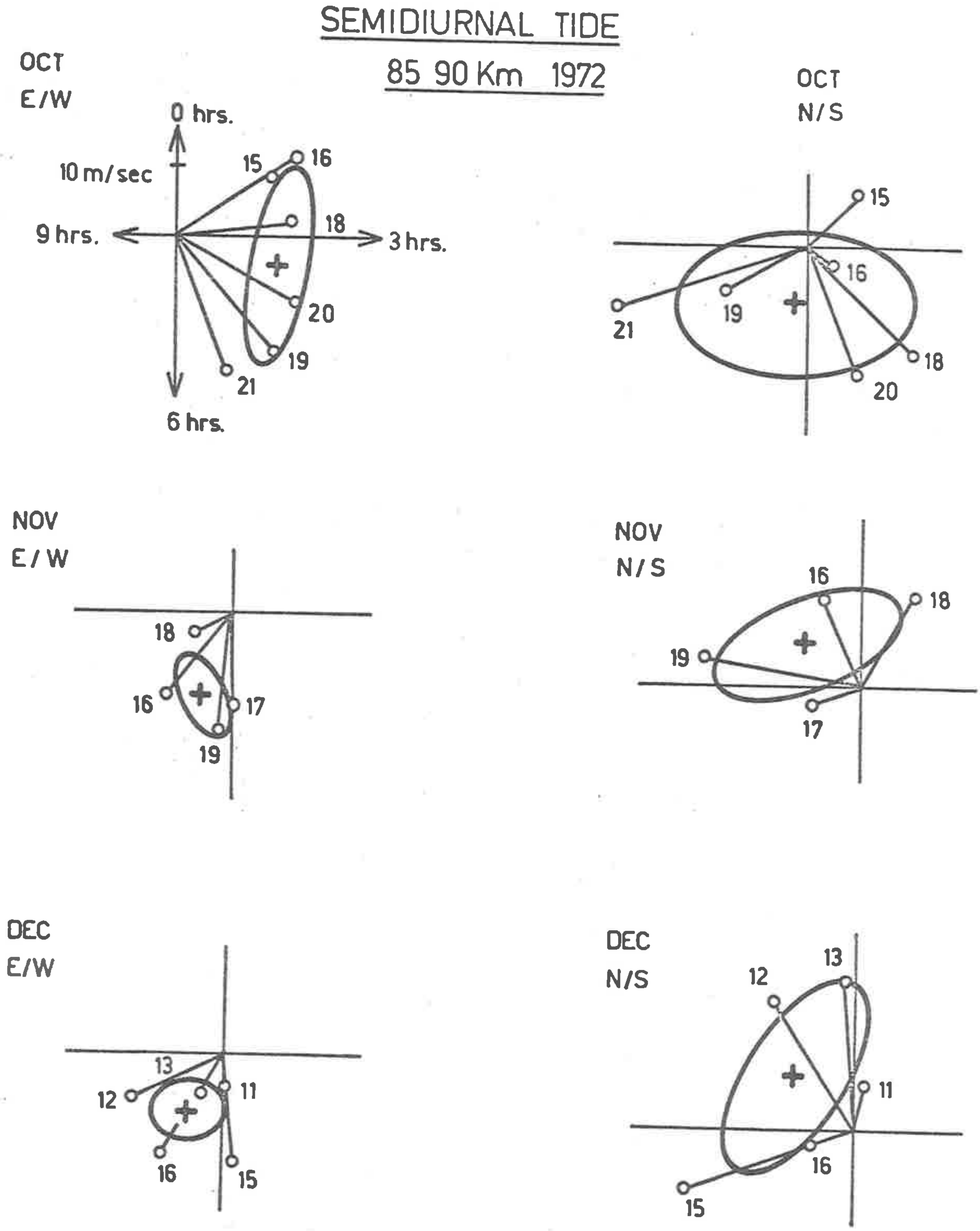


FIGURE 4.24 *Harmonic dials of the semidiurnal tide as deduced from the 1972 drift results.*

just before the solstices.

(iv) In contrast to the diurnal oscillation, the yearly picture is of a very variable phase with the whole range of phases being preferred at various times of the year. This implies that in averages of results taken at a particular time of the day over an extended period, the diurnal tide will almost certainly not be smoothed out although the semidiurnal tide may be.

4.2.3 8-hour period

Little can be said of the 8-hour oscillation because it was found to be small (usually little more than 5 m/sec) in amplitude and virtually random in phase, achieving some sort of significance only when the 24 and 12 hour amplitudes were small. A great deal of data would be needed to isolate any systematic features of its behaviour. The importance of the terdiurnal component undoubtedly varies geographically as it does for the other tidal components. In spectrally analysing Kazan meteor results for two months of 1964, POKROVSKIY and TEPTIN (1972) found quite a significant peak at 8 hours. From an analysis of Sheffield meteor data, MULLER (1966) reported that the amplitude of the 8-hour component was only 2 or 3 m/sec.

The present results and results from the Adelaide meteor system seem to indicate a slightly larger amplitude for 35^0 S.

4.3. Discussion

As mentioned in the opening chapter, the relative dominance of the diurnal tide found at many locations is no longer a surprising feature. It has been reported for both high (HOOK, 1970) and low (ALLEYNE et al, 1974) latitudes and is well in keeping with current tidal theory, being associated with evanescent and propagating modes respectively. In fact, it is instructive to compare the harmonic components derived from the present drift results with the predictions of theory.

The current state of the art has been well summarised by CHAPMAN and LINDZEN (1970) and the quantitative estimates quoted by them will be used. The values are based on a mixture of modes. Remembering that the theory is based on typical temperature profiles and neglects such factors as mean winds, dissipation and nonlinearity, the agreement is quite good. For the diurnal tide at 35^0 latitude, the amplitude given by theory (CHAPMAN & LINDZEN,Pg 150) is about 21 m/sec in each component which is not far from the average value of FIGURE 4.14. The predicted phases (CHAPMAN and LINDZEN Pg 151) of about 1200 for maximum eastward wind and 1800 for maximum northward wind are certainly the most typical of the results. Where data was obtained at several heights, a somewhat longer wavelength than the 28 km of theory generally seemed appropriate. It is not surprising that there should be strong seasonal variations in view of the changes that can occur in the local temperature and mean wind structure, changes which will affect the various refractive and filtering processes to which the tides are subjected. For the semidiurnal component (CHAPMAN and LINDZEN Pg 137), the predicted amplitude of around 16 m/sec is again fairly typical of the results. Only for the phase of the semidiurnal component do theory and experiment differ significantly, with theory predicting a northward maximum shortly after 1100 and 2300. This does not seem to be representative of the results.

On the average, the experimental results reveal three components of comparable magnitude - the diurnal and semidiurnal tides and the mean zonal flow. Minor contributions to the overall motion come from the terdiurnal tide and the mean meridional flow. Two further points of interest will now be considered briefly - namely the tidal momentum flux and the correlation between tides and the Sq current system.

The northward transport of momentum due to tidal motions is given by the expression ½ uv cos (ϕ_u - ϕ_v) where u and v are the amplitudes, and ϕ_u and ϕ_v the phases, of the E/W and N/S components respectively (NEWELL and DICKINSON, 1967). The values of this expression are plotted in FIGURE 4.25. Despite the considerable spread, particularly for the diurnal tide, there is evidence, at times, of substantial poleward momentum transport. This trend is most marked for the 24-hour tide from September to December. In the case of the 12-hour tide, the April - June values are all poleward. Similar systematic features have been found from more than six years of meteor observations (ELFORD, private communication). Near 90 km, the main feature of the meteor calculations for the 24-hour tide is a strong poleward transport from September to December. For the 12-hour tide at 90 km there is a small but consistent poleward bias. Above 95 km, very large poleward values are obtained during March and April. However the main feature of both sets of results is the consistent poleward transport by the diurnal tide in late spring and early summer - a contribution which appears sufficient to maintain the summer circumpolar vortex.

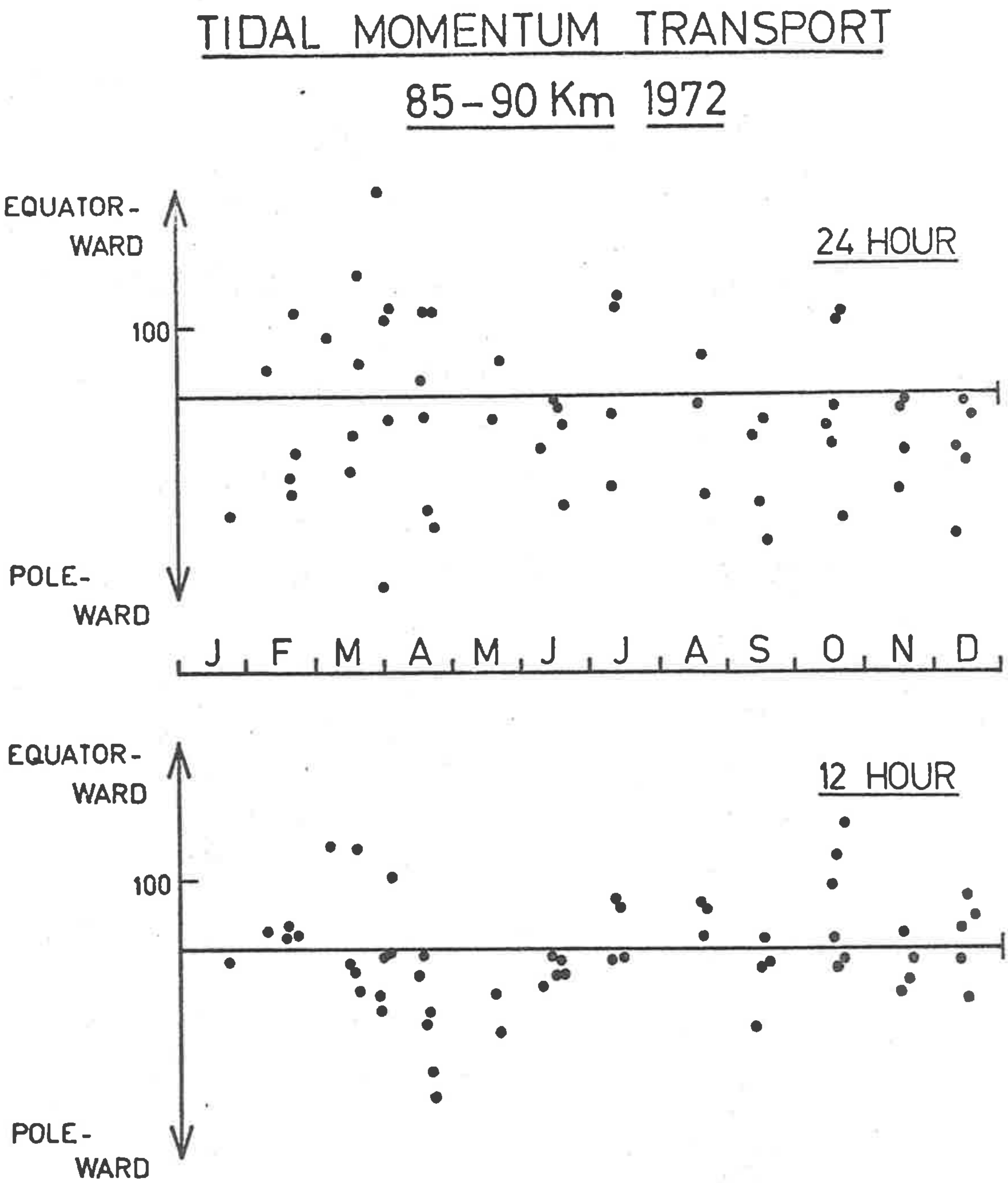


FIGURE 4.25 *The meridional momentum transport associated with the 24 hour and 12 hour periodic components at 85 - 90 km, 1972.*

The Sq current systems, associated with the movement of ionisation across the Earth's magnetic field, occur near 120 km. They have a day to day variability which may be due to conductivity changes in the E-region or variations in the Sq dynamo driving force.

The fluctuations in H seen on magnetograms are also due to magnetic disturbances of more distant origin. External magnetic effects can be largely removed, and a more accurate estimate of the Sq current variation obtained, by differencing the magnetograms of two stations of similar longitude but with latitudes above and below the Sq focus (OSBORNE, 1966). This procedure has been adopted in the calculation of Sq fluctuations in H for comparison with tidal amplitudes (FIGURE 4.26), using magnetograms from Port Moresby (13.8^{0}S) and Toolangi (37.5^{0}S). There is no correlation with the amplitude of either the diurnal or semidiurnal tide. This null result is not altogether surprising in view of the height difference between the drift data and the Sq currents. It may in fact imply that the Sq current system is driven by an evanescent, non-propagating tide, excited in situ at about 120 km. The amplitude of such a tide would decrease quickly with decreasing height and be insignificant at 90 km. The present null result is therefore compatible with the proposal

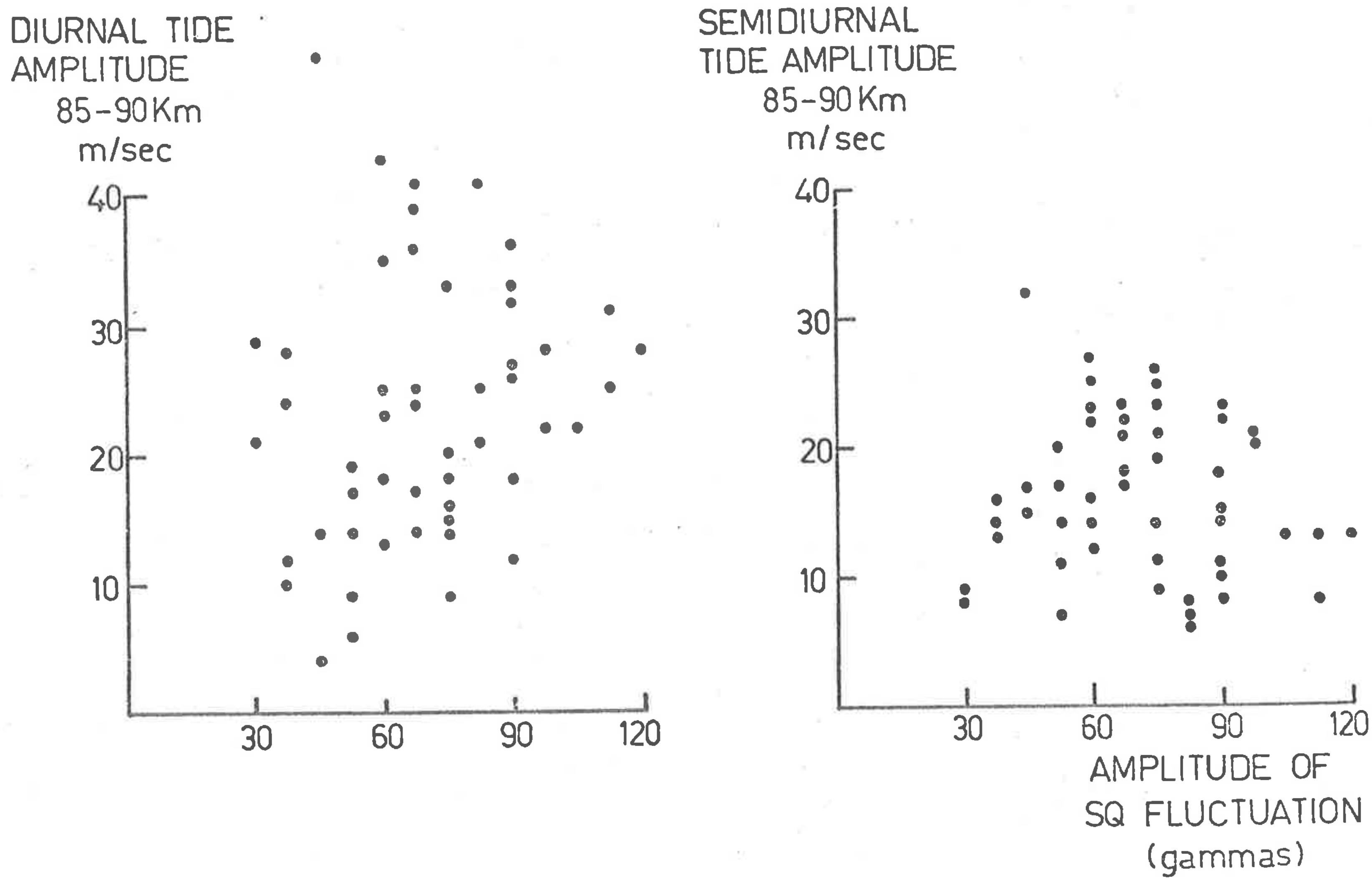


FIGURE 4.26 *The relationship between tidal amplitude and the amplitude of the Sq current system fluctuations, based on the 1972 data.*

by STENING (1970) that the evanescent diurnal mode (1,-1) is likely to be the dominant driving force for the Sq system.

CHAPTER FIVE

THE DRIFT RESULTS - SHORT-TERM VARIATIONS

In Chapter 4, the mean and tidal components of the drift results were considered. In this chapter, drifts will be considered from the point of view of the shorter-period fluctuations. In order to isolate the more rapid fluctuations, data must be available for a sufficiently long time sequence to enable the longer-period variations to be removed. The period 15 - 20 June, 1973 has been chosen for detailed analysis because of the continuity of the observations.

5.1 Residuals - a General Treatment

The data, smoothed to emphasise the general nature of the drifts over the period of interest, is displayed in FIGURES 5.1 and 5.2. FIGURE 5.1 is the 3-hourly running mean of the E/W component for the six days and FIGURE 5.2 the N/S component for the same period. The height ranges represent a combination of drifts from two 2 km intervals - for example the 86 km graph, which includes the most data, is based on the drifts for 84 - 86 km and 86 - 88 km.

The basic features of the more rapid fluctuations are

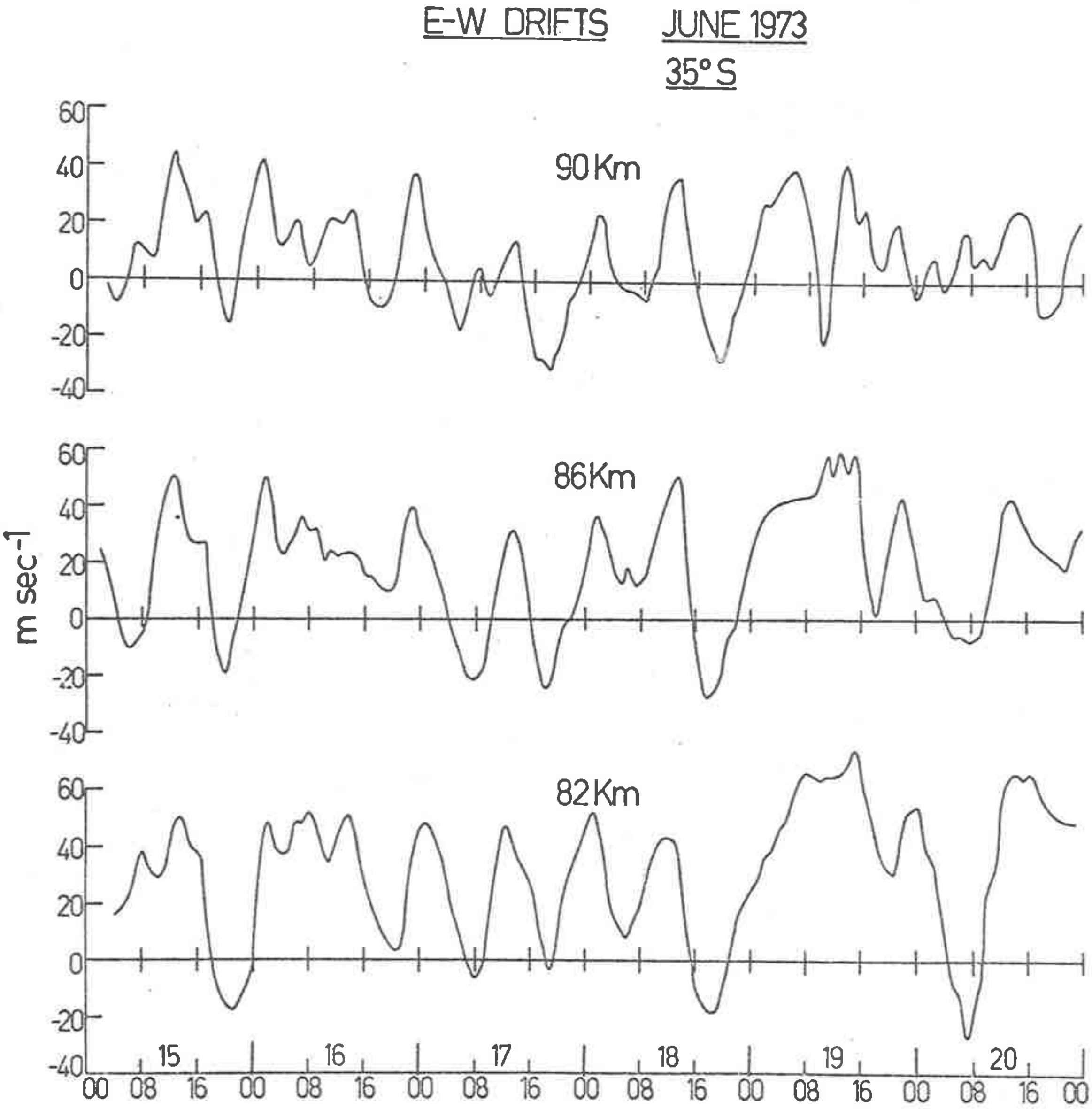


FIGURE 5.1 *The zonal drift behaviour for a sequence of six days in June, 1973. A three-hour running mean has been used to smooth the data.*

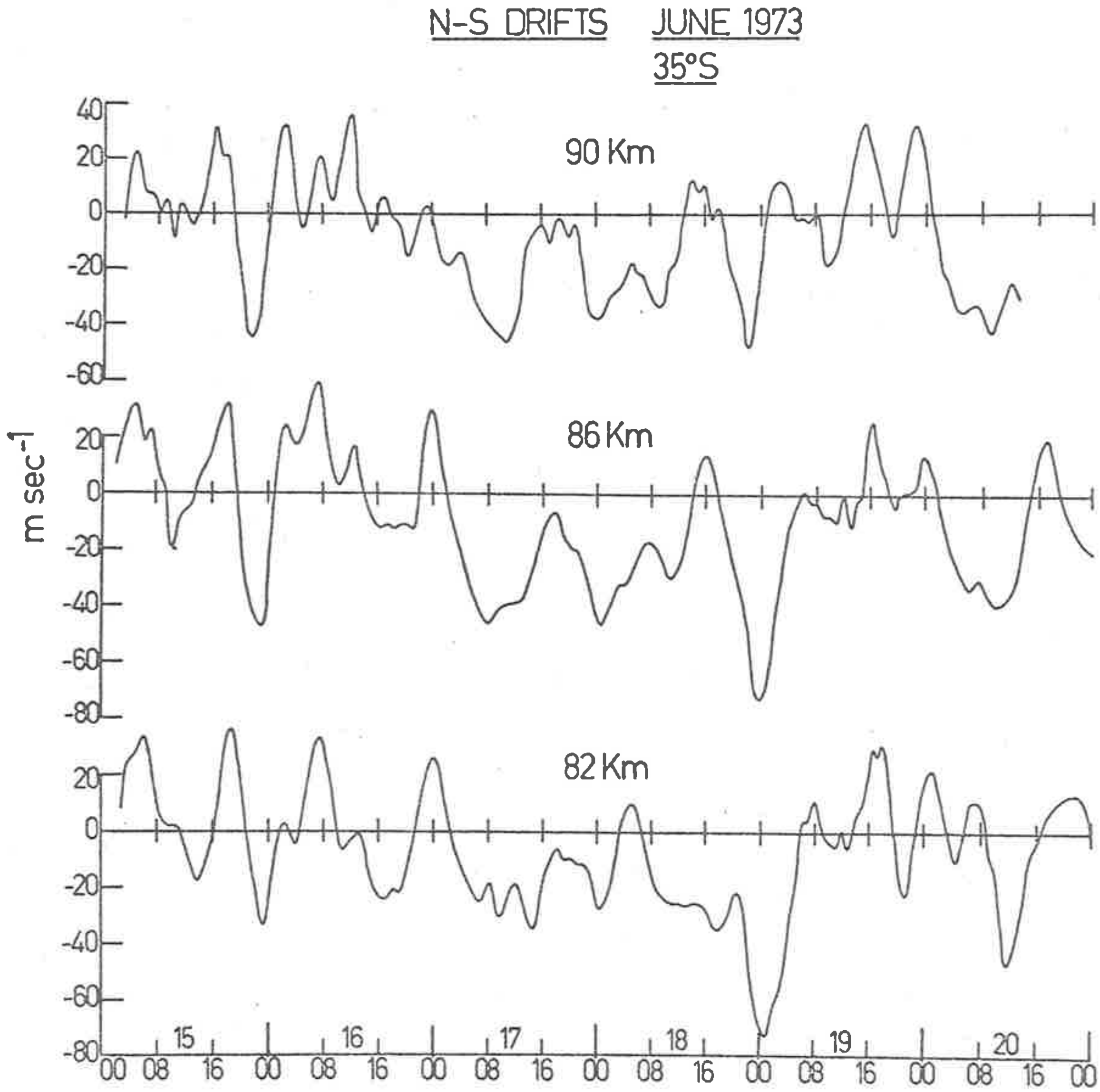


FIGURE 5.2 *The meridional drift behaviour for a sequence of six days in June, 1973. A three-hour running mean has been used to smooth the data..*

evident from FIGURE 5.3. Here the original, unsmoothed data for the six days has been filtered to remove periods less than one hour and greater than six hours. What remains is therefore a picture of the changing influence of periodicities between one and six hours. The filtering was done by taking a six-hour running mean of the hourly averages. Whilst this filter may not entirely remove the semidiurnal oscillation, later more rigorous filters have produced very similar results (VINCENT, private communication).

The r.m.s. amplitude for both the E/W and N/S components is about 15 m/sec but there is quite a lot of variability. For example, considering the N/S component, the short-term fluctuations are clearly more important on the 16th than they are on the 18th. Furthermore, the fluctuations on the 17th appear to have a longer period than do those on the 16th or 19th. Another feature of interest is the period from 1900 on the 18th to 1100 on the 19th. For this period the amplitude of the E/W residuals is very small compared to the N/S residuals, suggesting a 'polarisation' of the irregular wind fluctuations.

5.2 Residuals - a Specific Example

In this section the time resolution potential of the

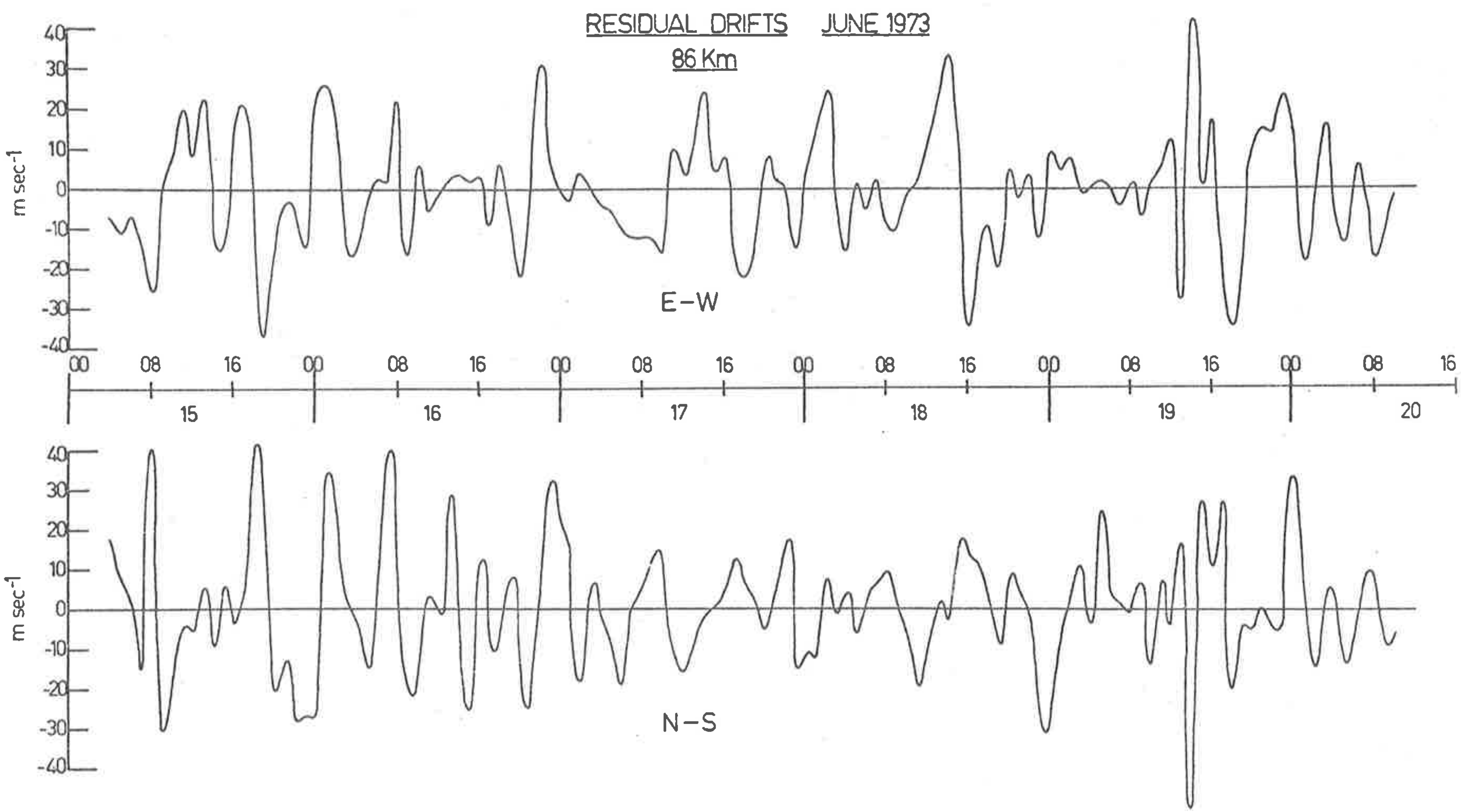


FIGURE 5.3 The drift fluctuations after filtering to remove periodicities less than one hour and greater than six hours.

drift technique is demonstrated by examining more closely a period of high data density between 1700 on June 17 and 0300 on June 18. The results for 4 height intervals are displayed in FIGURES 5.4 to 5.7. The individual drift values are represented by the dots and the continuous curve is the result of a harmonic fit of a mean and 24-, 12- and 8- hour components to the data from 1200 on the 17th to 1200 on the 18th. The period examined is fairly well centred on this interval, which is desirable because of the tidal variability. The deviations from the smooth curves are therefore due to random errors in the drift determinations and to fluctuations with periods less than 8 hours. It can be seen that they are often quite systematic and wavelike in form.

It is often difficult to assign a specific period or amplitude to the fluctuations and they are probably due in general to the superposition of several waves. Some of the variations are quite long lasting. For example at 82 - 84 km there appears to be a general 3-hour periodicity in the E/W component with maximum positive deviations near 1700, 2000, 2300 and 0200. This is more apparent if an envelope enclosing the individual points is imagined. The data density at 88 - 90 km is insufficient to see any definite short-term features. This highlights

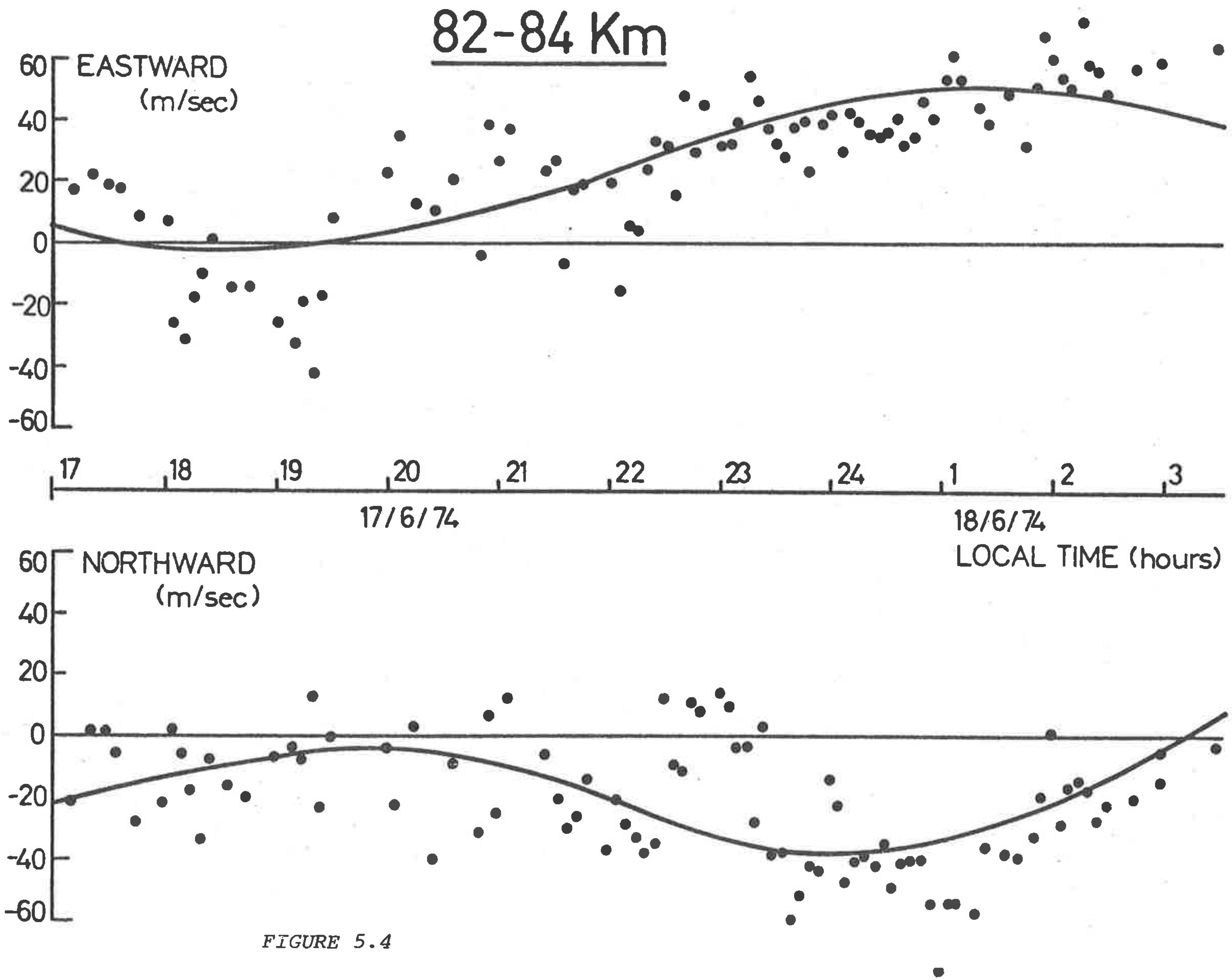
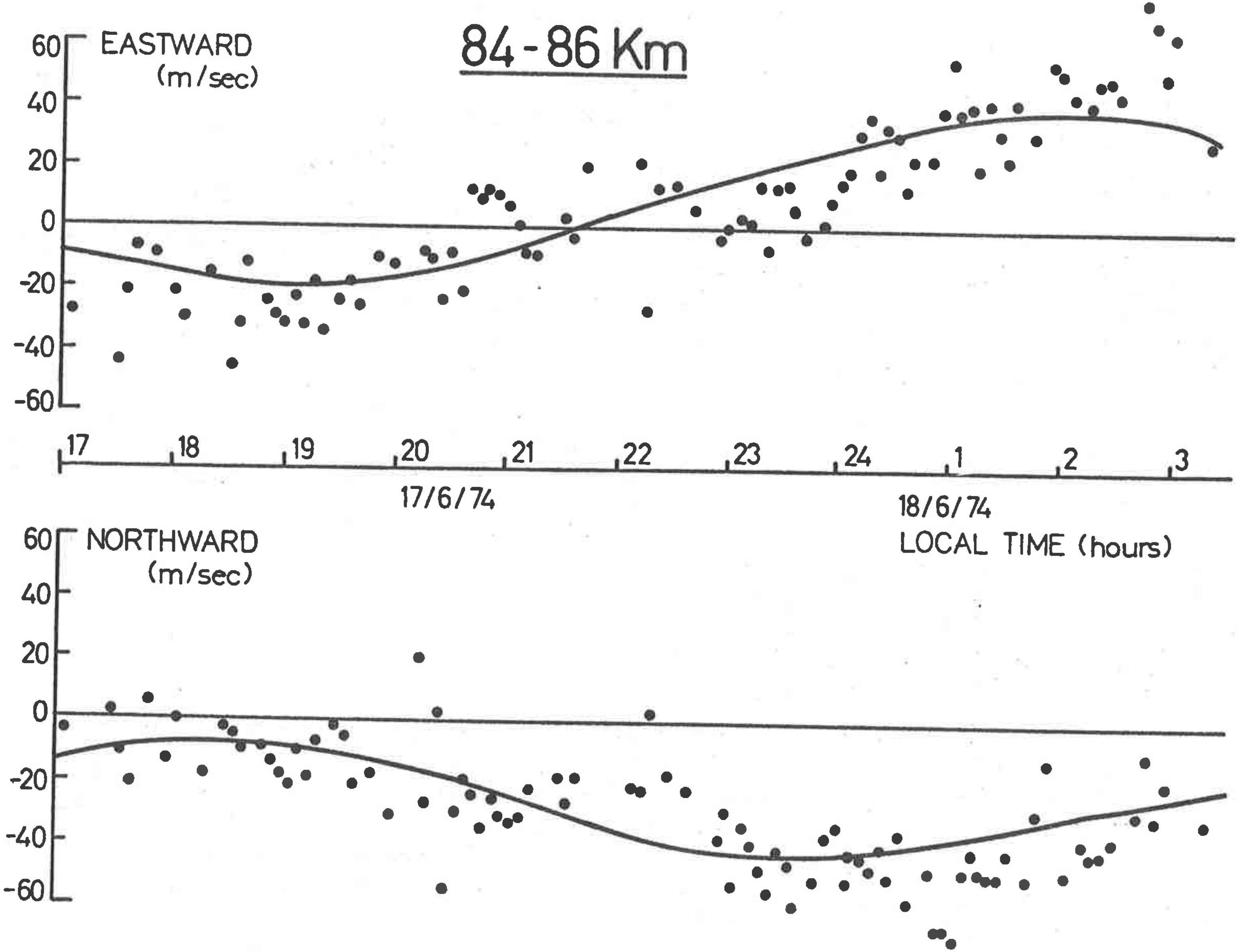

82-84 Km
60
EASTWARD
(m/sec)
40
20
0
-20
-40
-60
17
18
19
20
21
22
23
24
1
2
3
17/6/74
18/6/74
LOCAL TIME (hours)
60
NORTHWARD
(m/sec)
40
20
0
-20
-40
-60

FIGURE 5.4



84-86 Km
EASTWARD
(m/sec)
60
40
20
0
-20
-40
-60
17
18
19
20
21
22
23
24
1
2
3
17/6/74
18/6/74
LOCAL TIME (hours)
NORTHWARD
(m/sec)
60
40
20
0
-20
-40
-60

FIGURE 5.5

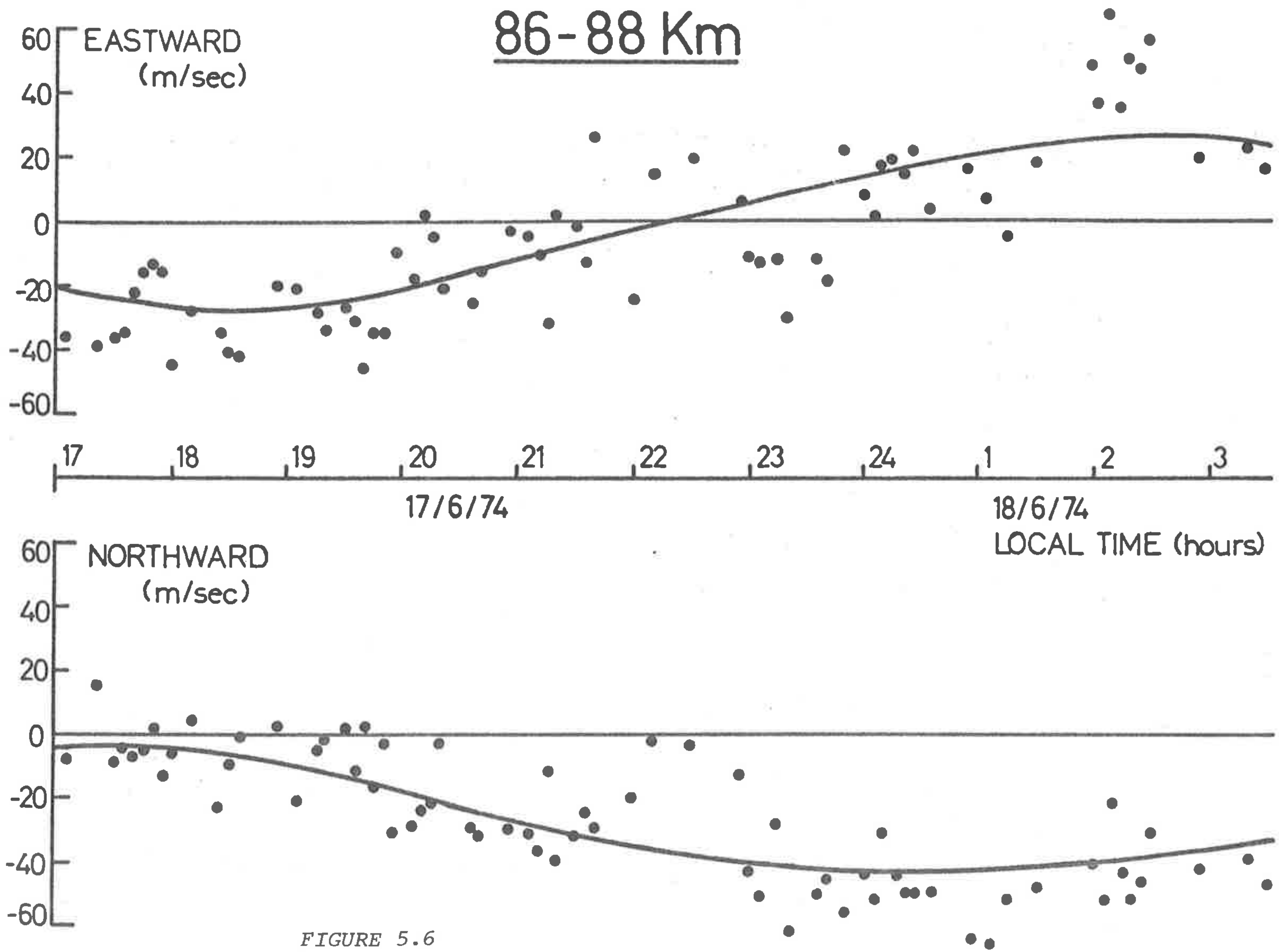

86-88 Km
EASTWARD
(m/sec)
60
40
20
0
-20
-40
-60
17
18
19
20
21
22
23
24
1
2
3
17/6/74
18/6/74
LOCAL TIME (hours)
NORTHWARD
(m/sec)
60
40
20
0
-20
-40
-60

FIGURE 5.6

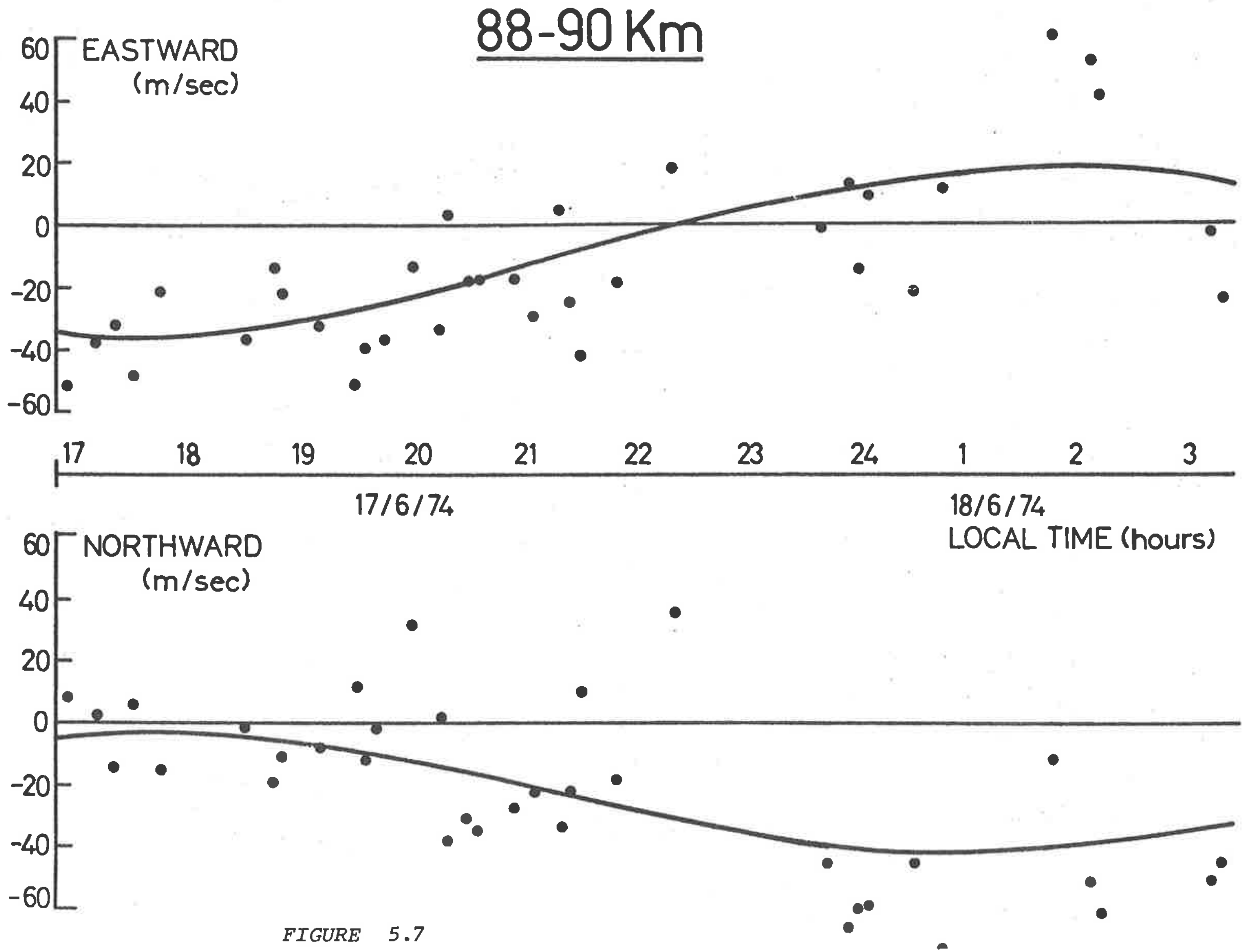

88-90 Km
60
EASTWARD
(m/sec)
40
20
0
-20
-40
-60
17
18
19
20
21
22
23
24
1
2
3
17/6/74
18/6/74
LOCAL TIME (hours)
60
NORTHWARD
(m/sec)
40
20
0
-20
-40
-60

FIGURE 5.7

the disadvantage of a linear receiving system where correct monitoring of the fading is only possible over a limited height range.

One feature of the harmonic fits to the zonal components is particularly interesting. There is a shear in the eastward drift which is evident for example by comparing the value at 1700 for each height. At the four ranges from 82 - 84 km upwards the values are respectively 5, -8, -21 and -34 m/sec Eastward. This is very strongly suggestive of an even shear with height and, more importantly, an even separation between the heights to which the drifts refer. It has been suggested that the D-region is stratified and that drifts can therefore only be obtained from the heights at which the strata lie. The present results would suggest that if stratification does hold, the strata are no more widely separated than 2 km.

5.3 A Rapid Reversal

A notable non-periodic short-term variation occurred in the E/W component of the drifts on June 18. The event, centred on 1500, does not appear particularly dramatic in FIGURE 5.1 due to the 3-hour smoothing of the data points. FIGURE 5.8 shows the event with no smoothing at all - successive 10 minute data points are plotted for the period

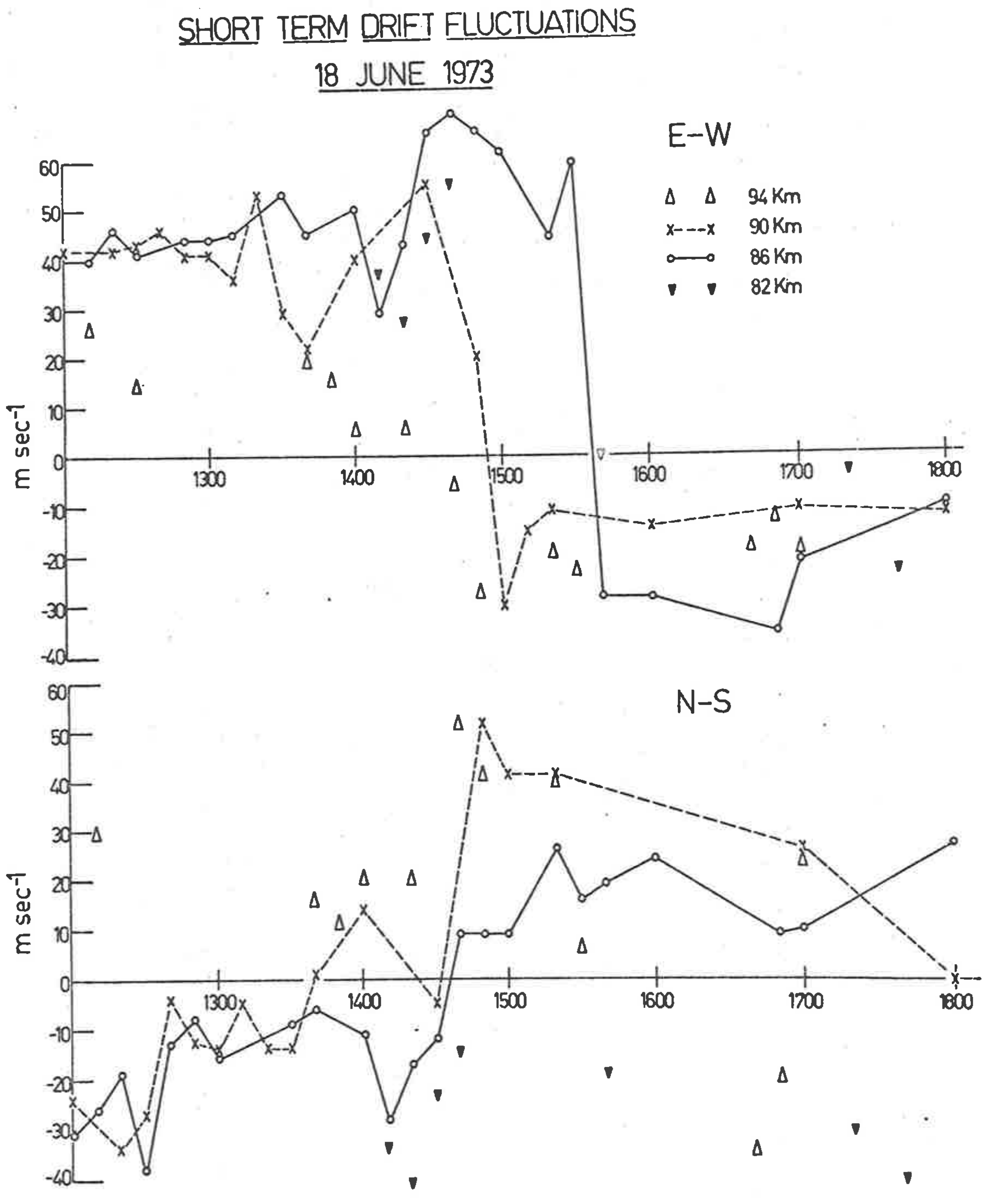


FIGURE 5.8 *A rapid change in the drifts illustrating the time resolution of the technique.*

1200 - 1800 on June 18. It can be seen that the E/W component at a height of 86 km changed from about 55 m/sec towards the East to about 25 m/sec towards the West in a time interval less than 10 minutes. Both before and after the reversal the drifts appear fairly constant. When the time of this reversal is examined at different heights, it is found that a downward progression occurs - the reversal takes place later at the lower heights. The apparent vertical velocity is about 1.6 m/sec. Over the same period no comparable change is observable in the N/S component.

5.4 Height Structure

The way in which the short-term features vary with height is of interest because, if they are largely caused by internal gravity waves, changes in the drift profiles over a short period should have a wave-like structure. The difficulty in studying these changes lies in obtaining a wide enough height coverage using a linear receiving system. The procedure adopted in the examples discussed in this section has been to average the drifts in half-hour intervals and then calculate difference profiles using heights common to successive intervals. The contributions to the differences from prevailing and tidal components should be small and the averaging, as well as maximising the height coverage, also serves to reduce the random errors in the final profile. The one disadvantage is that the velocity differences may represent different time changes at different heights if there is an uneven data rate.

In FIGURES 5.9 and 5.10 some of the vector velocity changes obtained in this way are shown. In all cases, differences could be obtained for at least six adjacent height intervals. The differences are between the half-hour averages either side of the stated time. The changes appear generally non-random and often indicative of waves. They are similar in nature to the results reported by MANSON, GREGORY and STEPHENSON (1973). There are times, as for example, the final profile in FIGURE 5.9, when the fluctuations are negligible. In many cases it is difficult to assign to them a vertical wavelength. The first two examples of FIGURE 5.9 seem to indicate λ_z>20 km whereas in the third example of FIGURE 5.10 a λ_z= 12 km is indicated. Similar profiles obtained for March are shown in FIGURE 5.11 and in all but the last example, where the changes are small, a λ_z in the vicinity of 8 - 12 km is appropriate.

Probably the most significant feature to arise from this study of the short-term variations is that in the

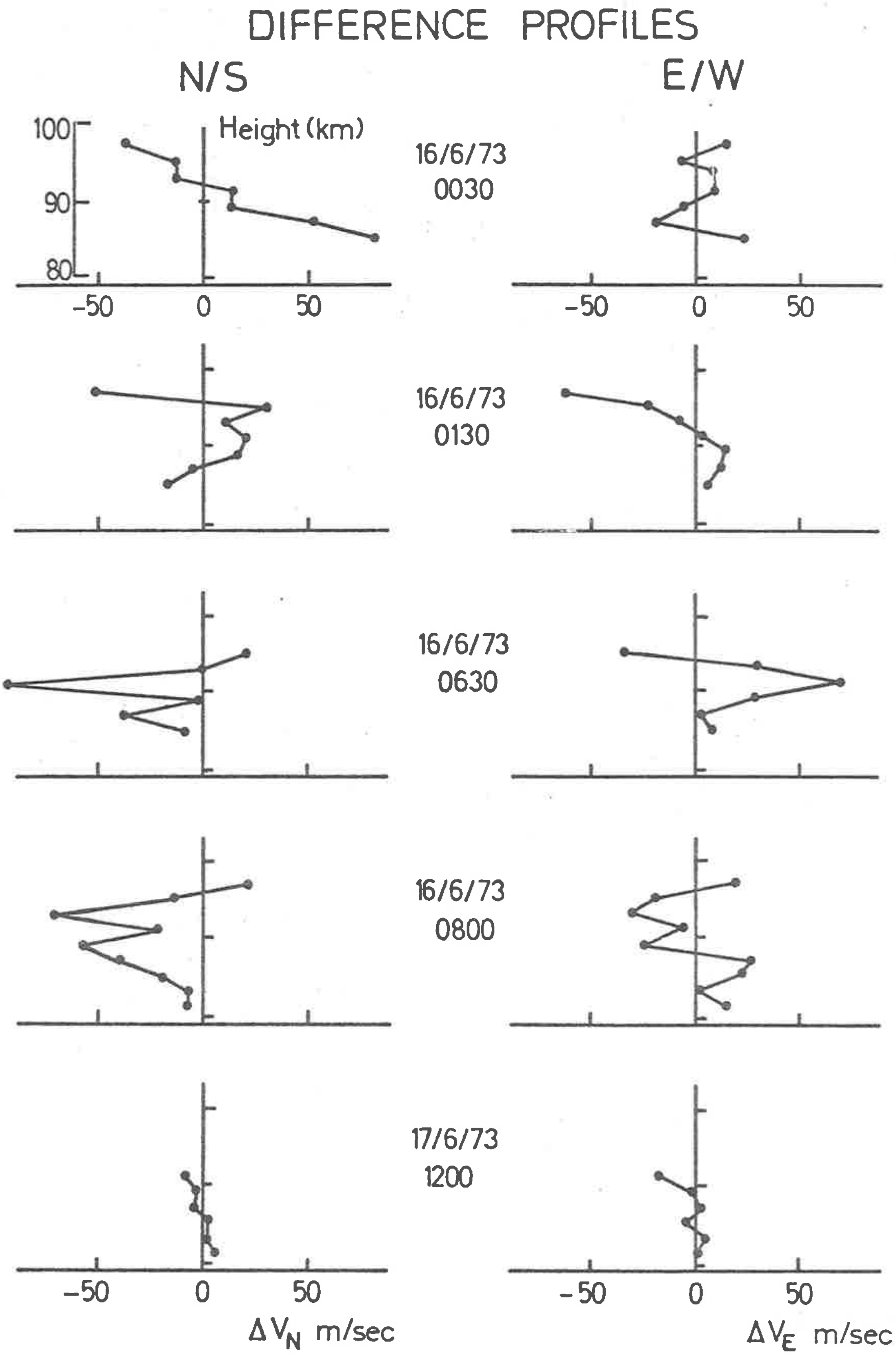


FIGURE 5.9 Differences between successive half-hour average profiles.

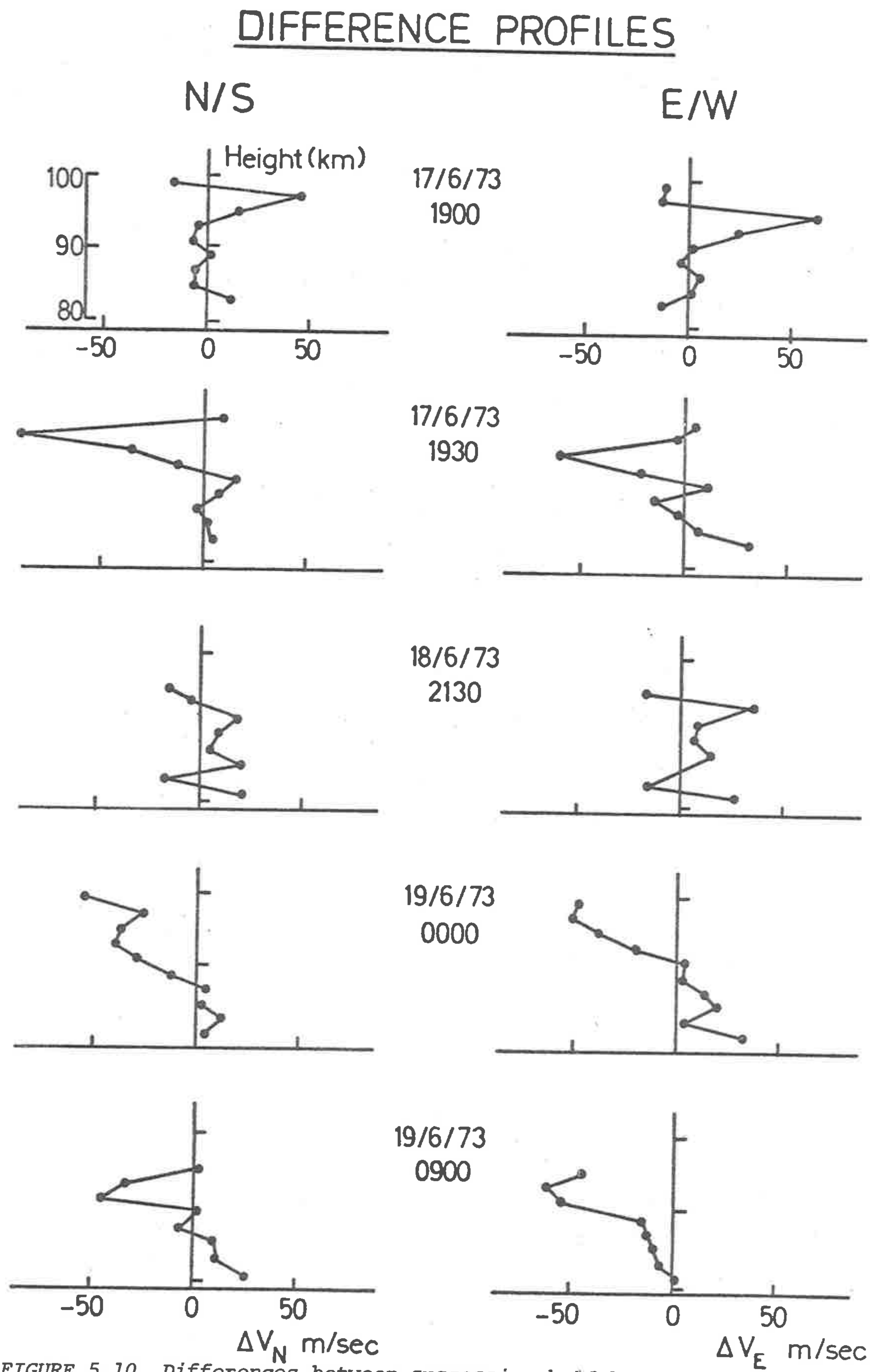


FIGURE 5.10 Differences between successive half-hour averages

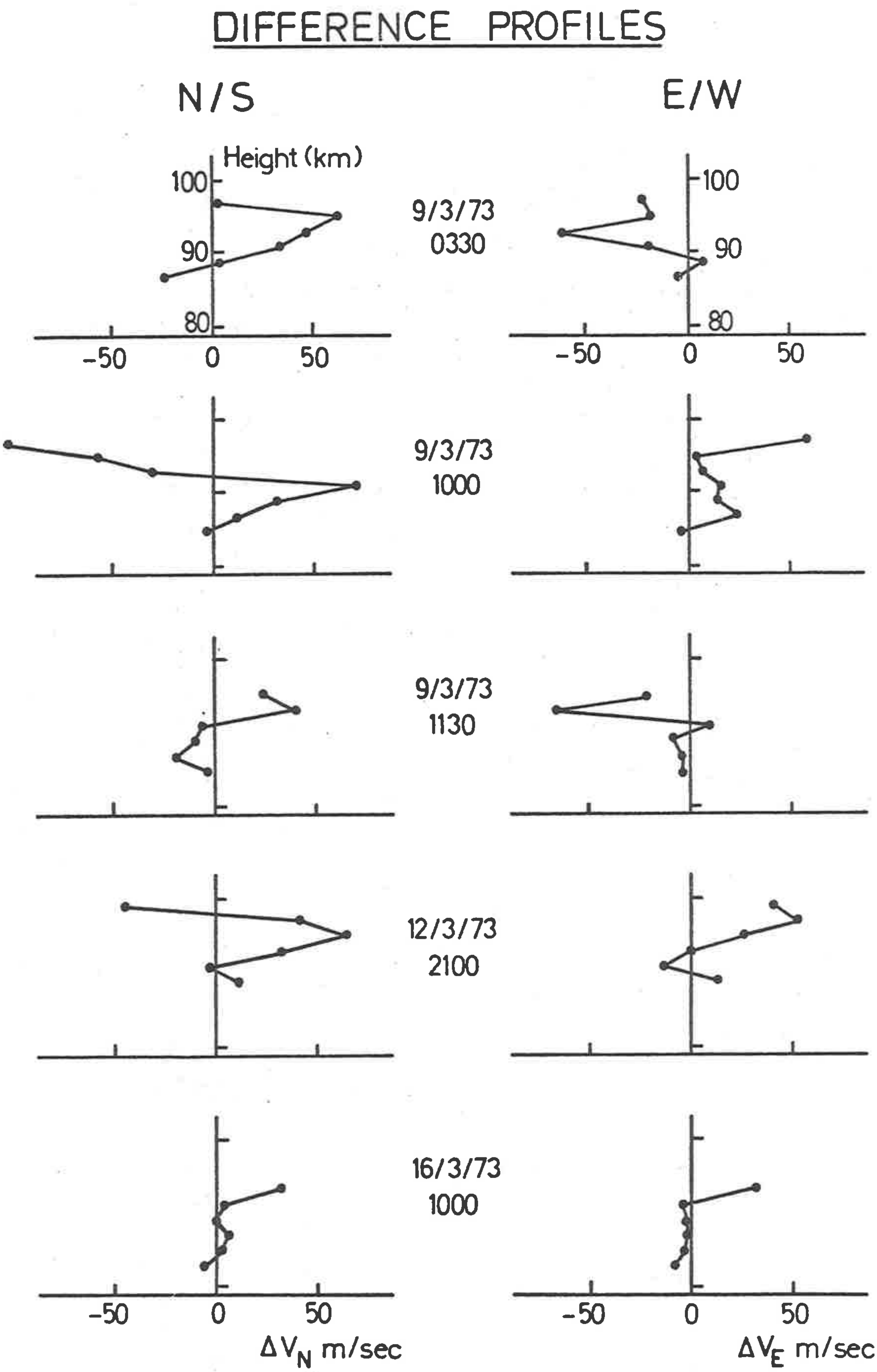


FIGURE 5.11 Differences between successive half-hour averages.

majority of cases the amplitude of the fluctuations increases with height. For each difference obtained a total velocity fluctuation ΔV was calculated such that $\Delta V^2 = \Delta V_E^2 + \Delta V_N^2$ with ΔV_E and ΔV_N the changes in the E/W and N/S component respectively. The average value of ΔV^2 was then calculated for each height using all the data from March and June. The resultant averages are plotted in FIGURE 5.12 along with a continuous curve, $ce^{h/H}$ where h is the height in kilometres and H = 8 km. The least number of values averaged was 91 for 98 - 100 km and at 88 - 90 km over 400 values were available. It is clear that $\overline{\Delta V^2}$ increases very nearly exponentially with a scale height of 8 km. The actual scale height between 80 and 100 km is closer to 6 km which implies that at these heights there is an energy input due to the short-term drift fluctuations since the energy density $\alpha \rho\ \Delta V^2\ \alpha\ e^{-h/6} e^{h/8} = e^{-h/24}$. This fall-off is much less rapid than that calculated by JUSTUS and WOODRUM (1973) using a daily difference technique although it is difficult to place much credence on their assumption of the constancy of tides on successive days. It is also interesting to compare this result with the energy input calculated for the diurnal tide on a particular day in July 1972 (STUBBS and VINCENT, 1973). The tidal energy input emerges as much

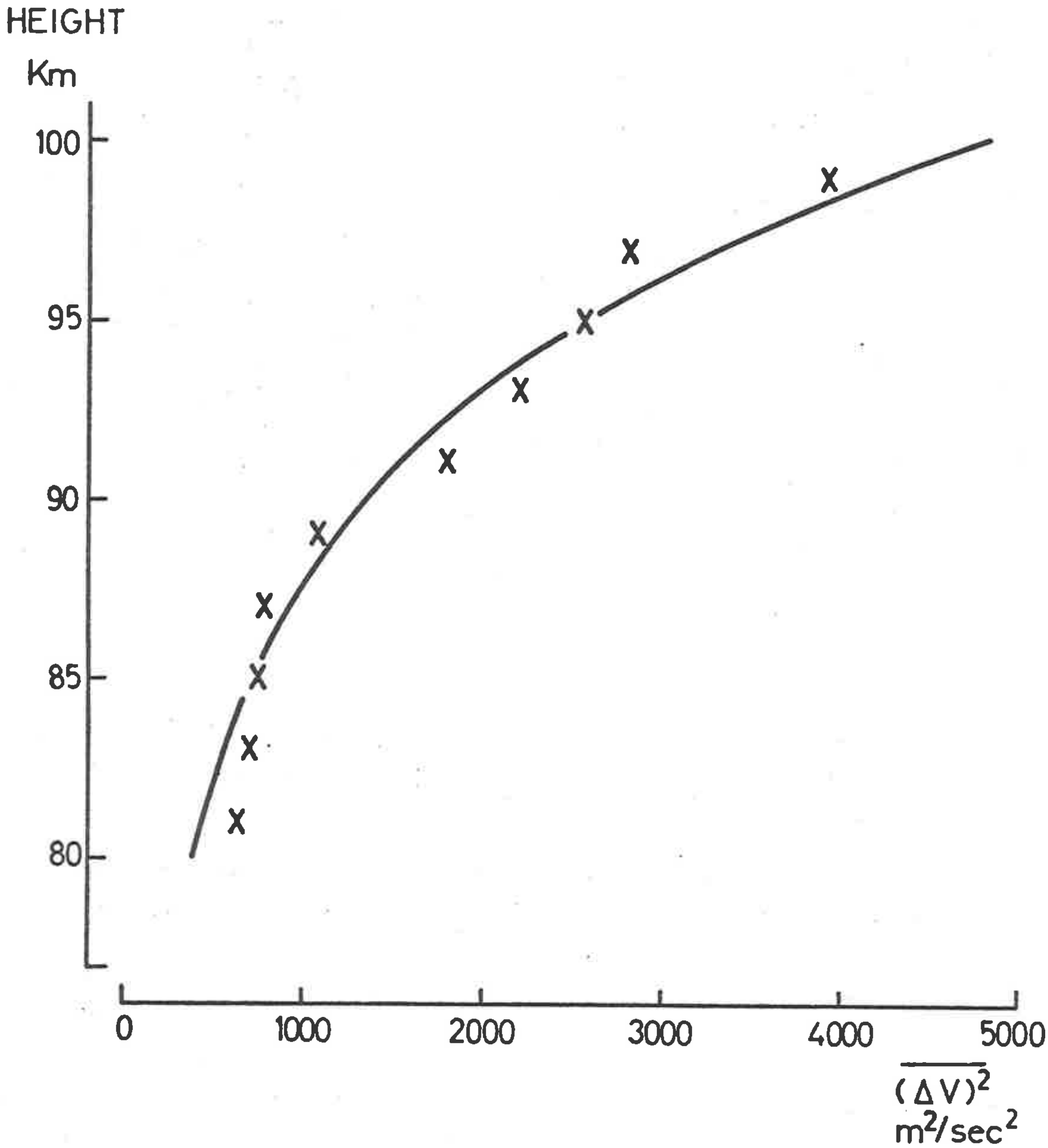


FIGURE 5.12 *The variation with height of velocity changes in a time of about 30 minutes. The crosses represent experimental averages for March/June 1973. The curve corresponds to an exponential growth with a scale height of 8 km.*

the more important. That this is more generally true is implied by the profiles of FIGURES 4.18 - 4.20 where there is no indication of the tidal amplitude increasing with height.

The actual energy density associated with the short term drift components can be approximated by the expression $\frac{1}{2}\rho\overline{\Delta V^2}$ and the value at 90 km is about $2 \times 10^{-3}\,J/m^3$. GOSSARD (1962) calculated energy density spectra associated with internal wave activity observed on micro barograph records of low-level pressure fluctuations. For the 24 June, 1957 the energy density of fluctuations with periods less than a few hours is very close to this figure. However this is the lowest example quoted and he points out that the flux associated with storms and frontal systems is often up to six times as great as this. That the energy density at the ground should be much greater than that at ionospheric levels is to be expected in view of the processes of dissipation and reflection encountered during the upward transport.

5.5. Summary

It has been seen that irregular fluctuations with periods less than six hours make an important contribution to the observed drifts. Near 86 km they can be of the same order as tidal and prevailing contributions and there is every indication that because of amplitude growth they are often a dominant contribution above 90 km. The periods and vertical structure of the fluctuations are in general consistent with the properties of internal gravity waves. A particular example was given of the potential of the drift technique for the observation of very rapid changes which would not be resolved by most other ionospheric probes.

CHAPTER SIX

A COMPARISON WITH WINDS FROM ROCKET TECHNIQUES

The setting up of the Woomera drift experiment as described in Chapter 2 was not completed until June, 1973. Hence for many of the comparisons in this chapter, only Buckland Park drift results were available. In these cases there is a horizontal separation of some 400 km between the two sets of measurements. The rocket trails or projectiles are within the sky area scanned by the Woomera drift system. The actual height overlap differs for the various types and times of rocket firings and will be discussed in the following section.

6.1 Nature of Firings

6.1.1 Dropsondes

The Woomera dropsonde is an expendable instrumental package which is ejected at rocket apogee and falls suspended beneath a metallized silk parachute. During its descent the sonde transmits information on temperature and density whilst its trajectory is deduced from radar tracking of the parachute. Hence profiles of temperature, density and horizontal wind are obtained. However the wind profiles only extend from 30 km to, at best, 60 km and so there is no overlap at any time of the day with the drift technique. Sonde firings are therefore of limited use for comparison purposes.

6.1.2 Falling spheres

In the falling sphere experiments conducted at Woomera, an inflatable passive sphere of 2m diameter is ejected just before rocket apogee. After inflation, it is tracked on the downleg of its trajectory by up to 4 radars. In general, sphere trials are scheduled for 30 minutes after local sunset (solar depression angle 6^0) enabling visual monitoring to be additionally employed. From the point of view of drift comparisons this is regrettable because wind information from the sphere experiment is only available up to 80 km and is least reliable near 80 km. In the evening all usable ionospheric echoes occur above 80 km. If the rockets were fired during the day there would be a possibility of a 20 km overlap of the heights covered by the two techniques. Another drawback of the twilight time slot is the occurrence of maximum radio interference at that time. It is hoped in the future to conduct a number of daytime falling sphere experiments, however all but one of the results in this chapter are for twilight firings.

6.1.3 Vapour trails

In the region from 80 - 200 km the best method of obtaining a wind profile is to release, from a rocket, chemicals in molecular form. The light from the released gaseous cloud is then due either to chemical reactions between the cloud and ambient atmospheric molecules or to the resonant scattering of incident sunlight. In the former case, observation is generally restricted to the night and in the latter, to twilight when there is a dark background. The trails used at Woomera are mainly lithium and occasionally trimethyl aluminium. They are not laid very frequently because of the cost factor. Recently, instrumentation has been developed to make possible the observation of a lithium trail during daylight hours. Use is made of a very narrow pass band optical filter because although the flux/wavelength interval is much greater for the release, it covers only a range of 1 pm compared to the foreground scatter continuum. A narrow band filter is mounted in front of a photomultiplier, and a chopper system coupled with a sensitive detector enables the lithium to be detected against the sky foreground. A scanning mirror is used to build up a frame of data which is stored on tape for processing. The trail is observed simultaneously from two sites for about 15 minutes.

6.2 Comparison of Results

6.2.1 Dropsonde

For the sonde of July, 1973 the Woomera drift equipment was operated with 2 km height steps, and there was good fading down to the 63 - 65 km interval. Sonde winds were obtained up to 60 km leaving just a 3 km 'gap' between the two sets of results, which are plotted in FIGURE 6.1. A simple extrapolation of the sonde profiles meets the drift profiles almost exactly so the agreement is as good as possible in view of the height discrepancy. There were several other sonde firings but in all cases either the sonde results or drift values were unsatisfactory from the point of view that the height gap between the two sets of data was greater than 10 km.

6.2.2 Spheres

A series of sphere comparisons carried out in 1972 is presented in FIGURES 6.2. - 6.8. In all cases, the ionospheric drifts were obtained at Buckland Park using a 5 km height gate. The crosses represent the average drift for the hour centred on the time of firing. Although there is no overlap, the resultant profiles are encouraging, with the drift values generally appearing a natural, though not unique, extension of the sphere profiles. A significant feature of the results is the size of the shears than can

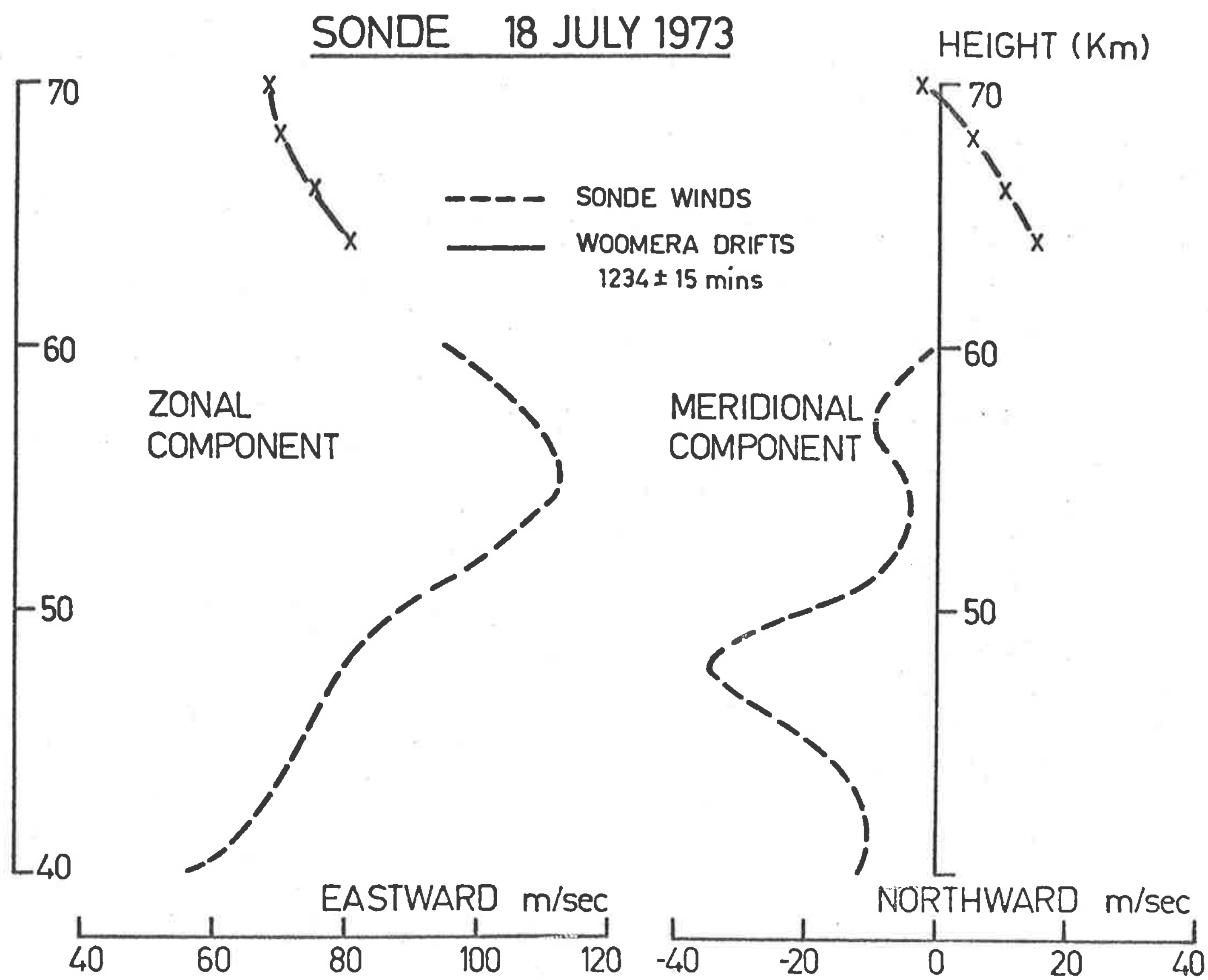

SONDE 18 JULY 1973
HEIGHT (Km)
70
60
50
40
SONDE WINDS
WOOMERA DRIFTS
1234 ± 15 mins
ZONAL
COMPONENT
MERIDIONAL
COMPONENT
EASTWARD m/sec
NORTHWARD m/sec
40
60
80
100
120
-40
-20
0
20
40

FIGURE 6.1

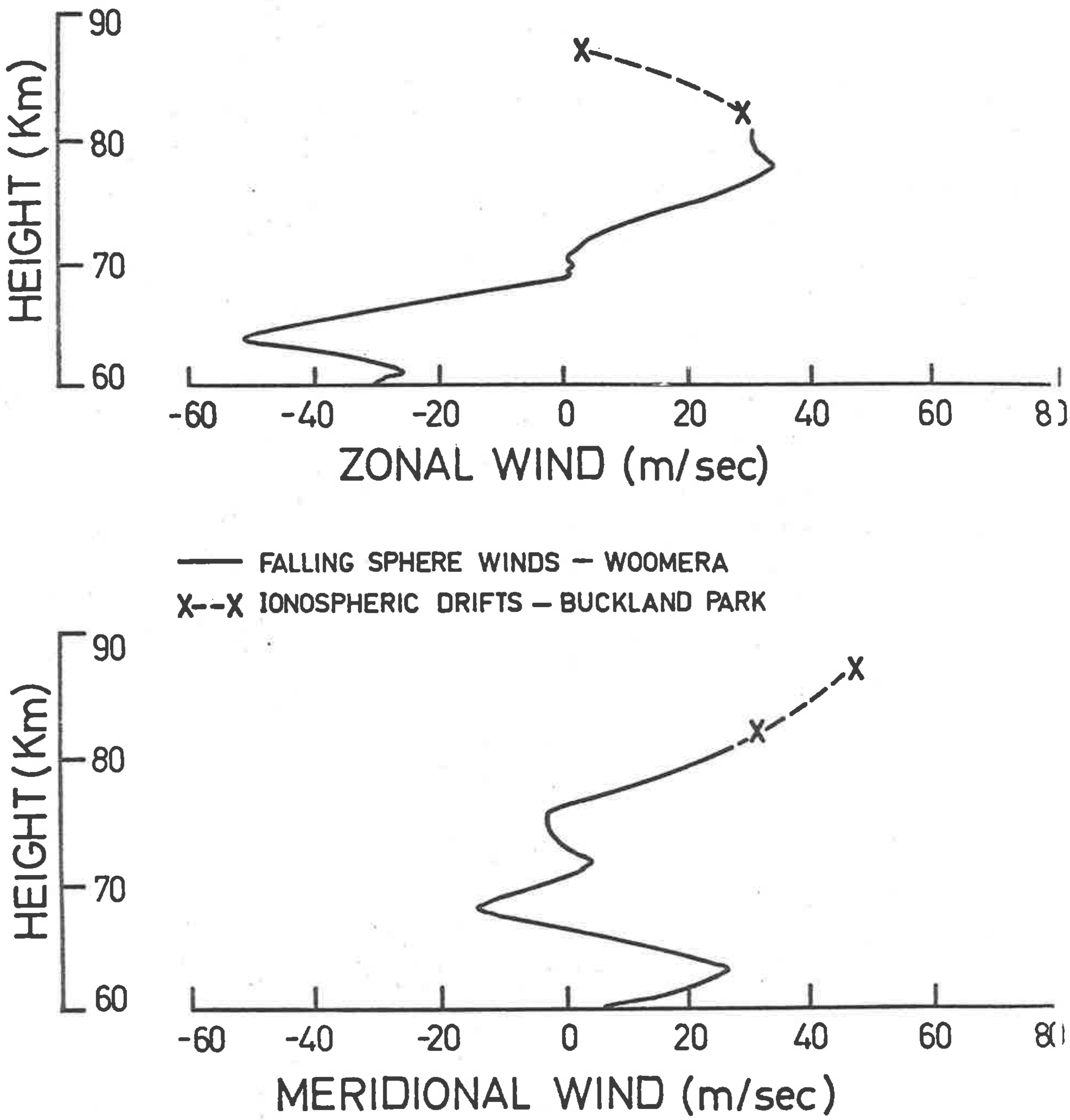


FIGURE 6.2 *Comparison for sphere of February 23, 1972. The time of firing was 1936 LT and the drift values are averages at 80 - 85 km and 85 - 90 km for 1936± 30 minutes*

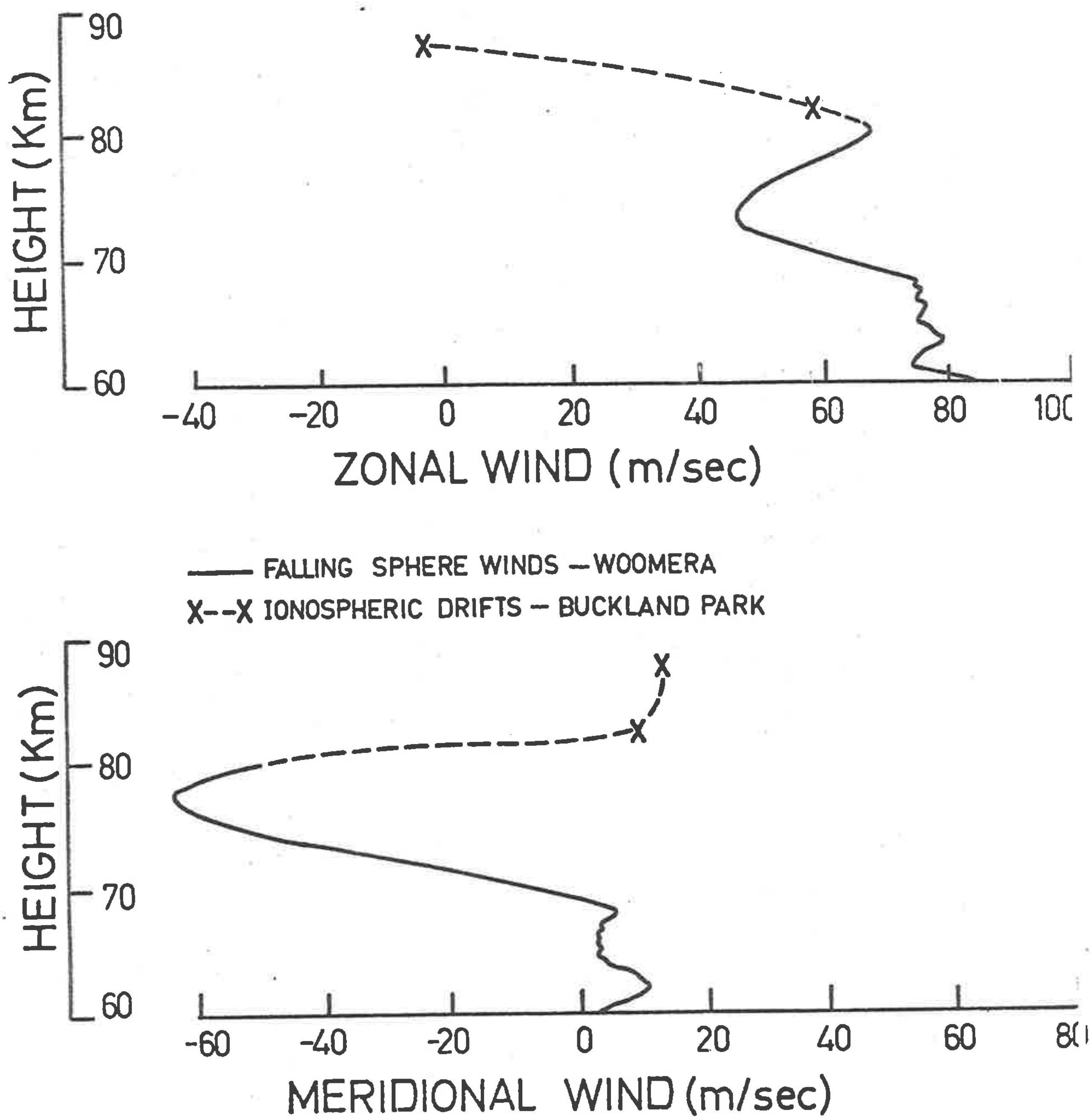


FIGURE 6.3 *Comparison for sphere of April 19, 1972*

Time of firing = 1831 LT

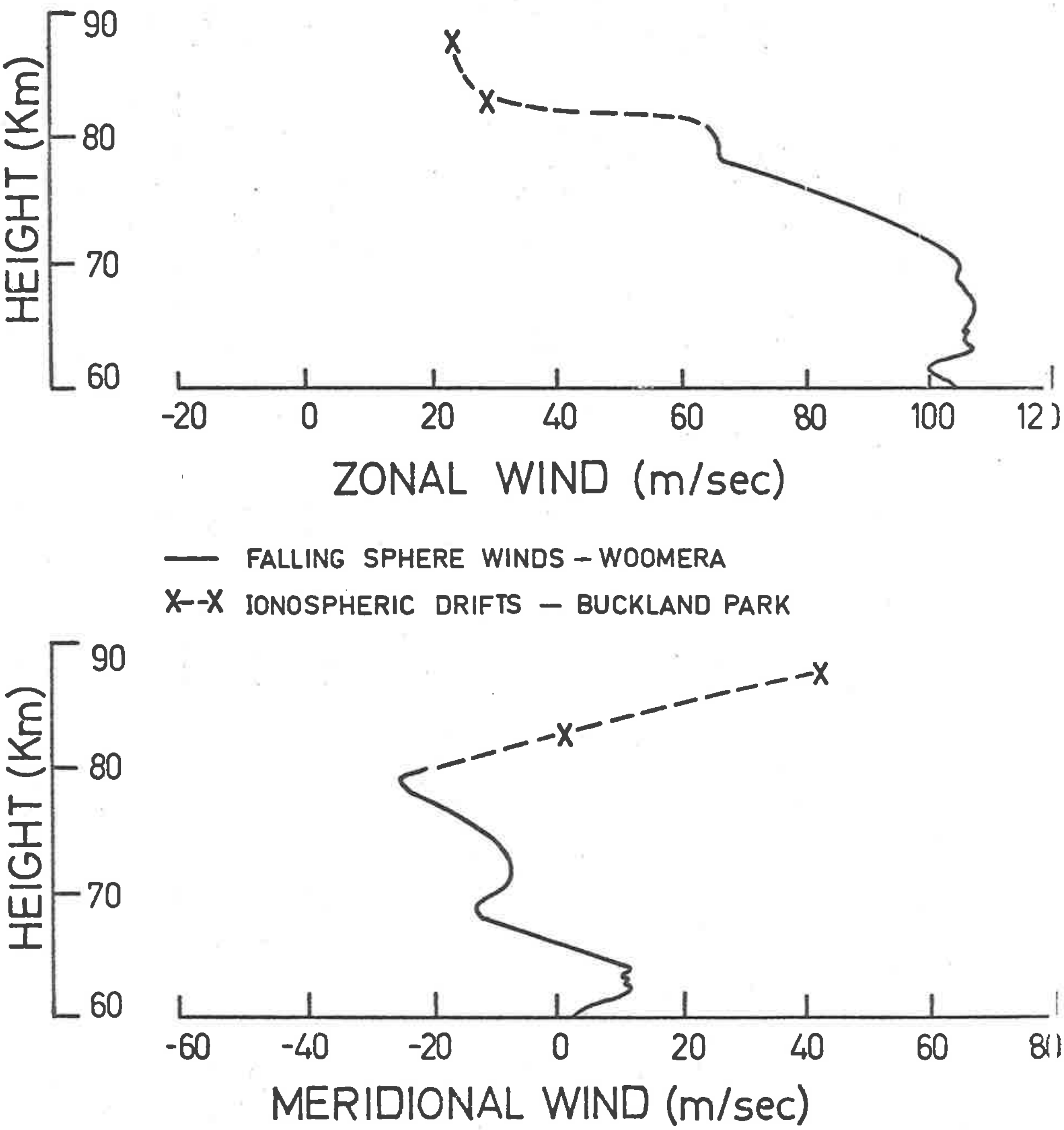


FIGURE 6.4 *Comparison for sphere of May 18, 1972.*
Time of firing : 1807 LT

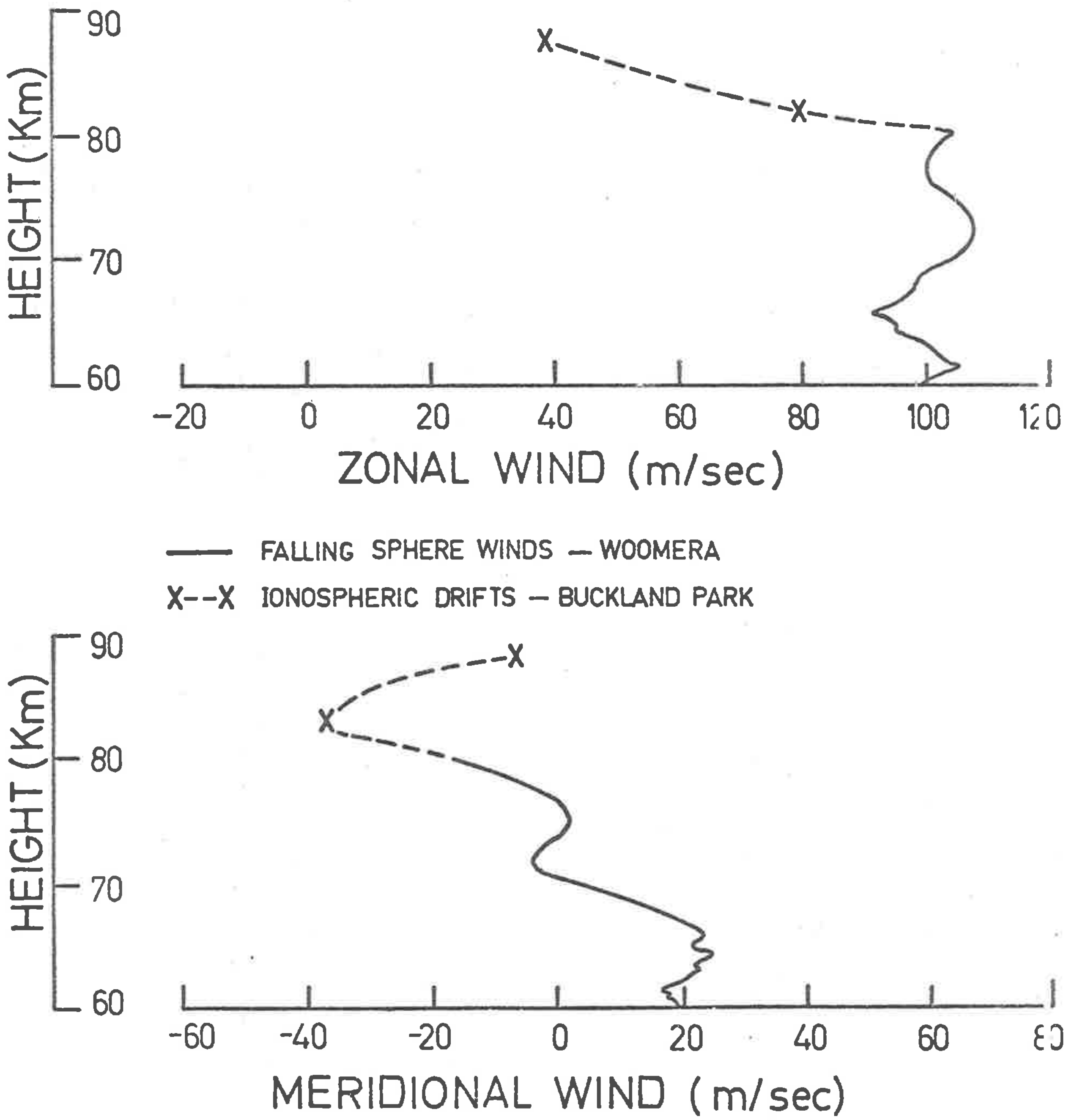


FIGURE 6.5 *Comparison for sphere of June 15, 1972.*
Time of firing: 1801 LT

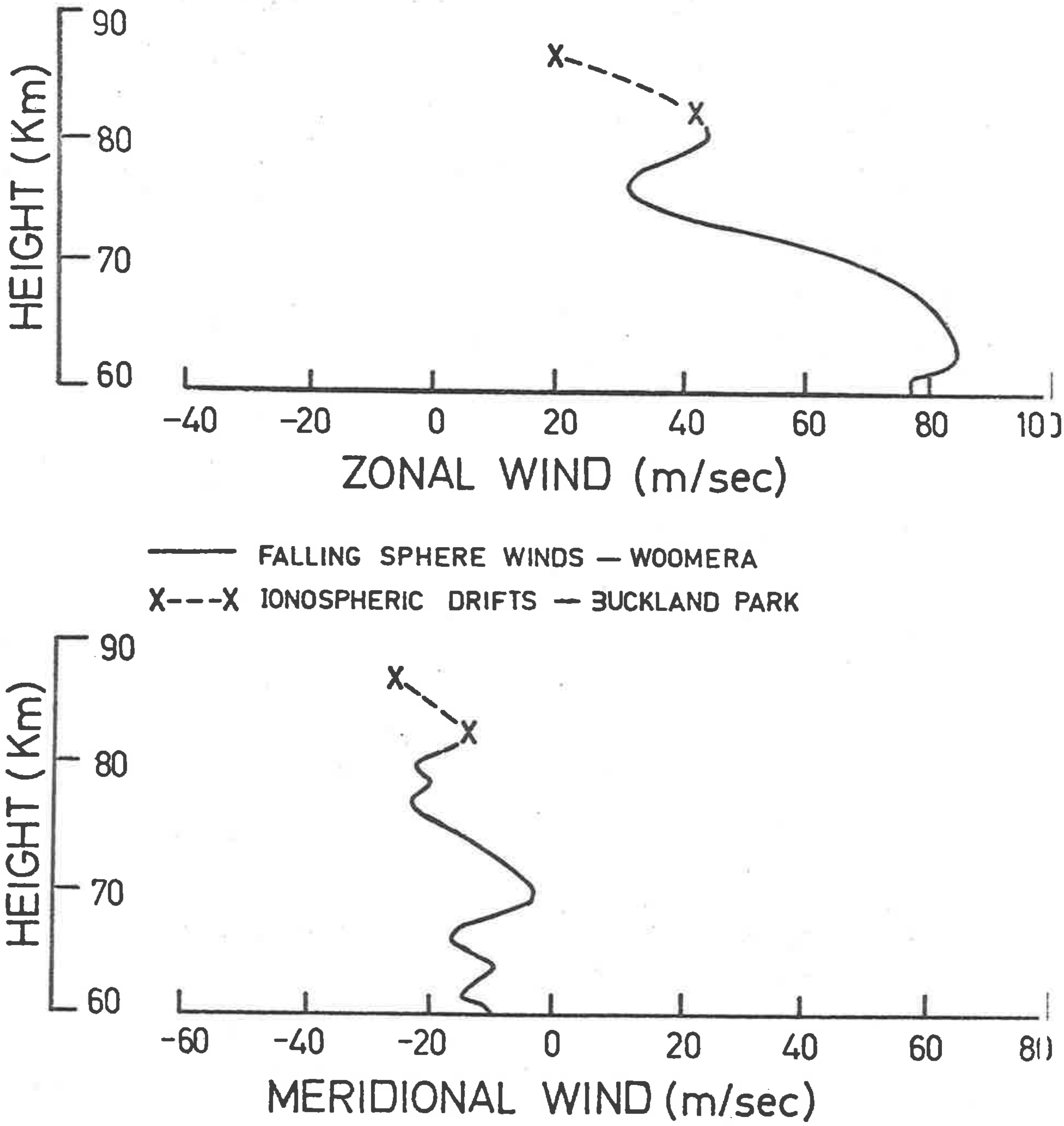


FIGURE 6.6 *Comparison for sphere of September 13, 1972.*
Time of firing 1845 LT

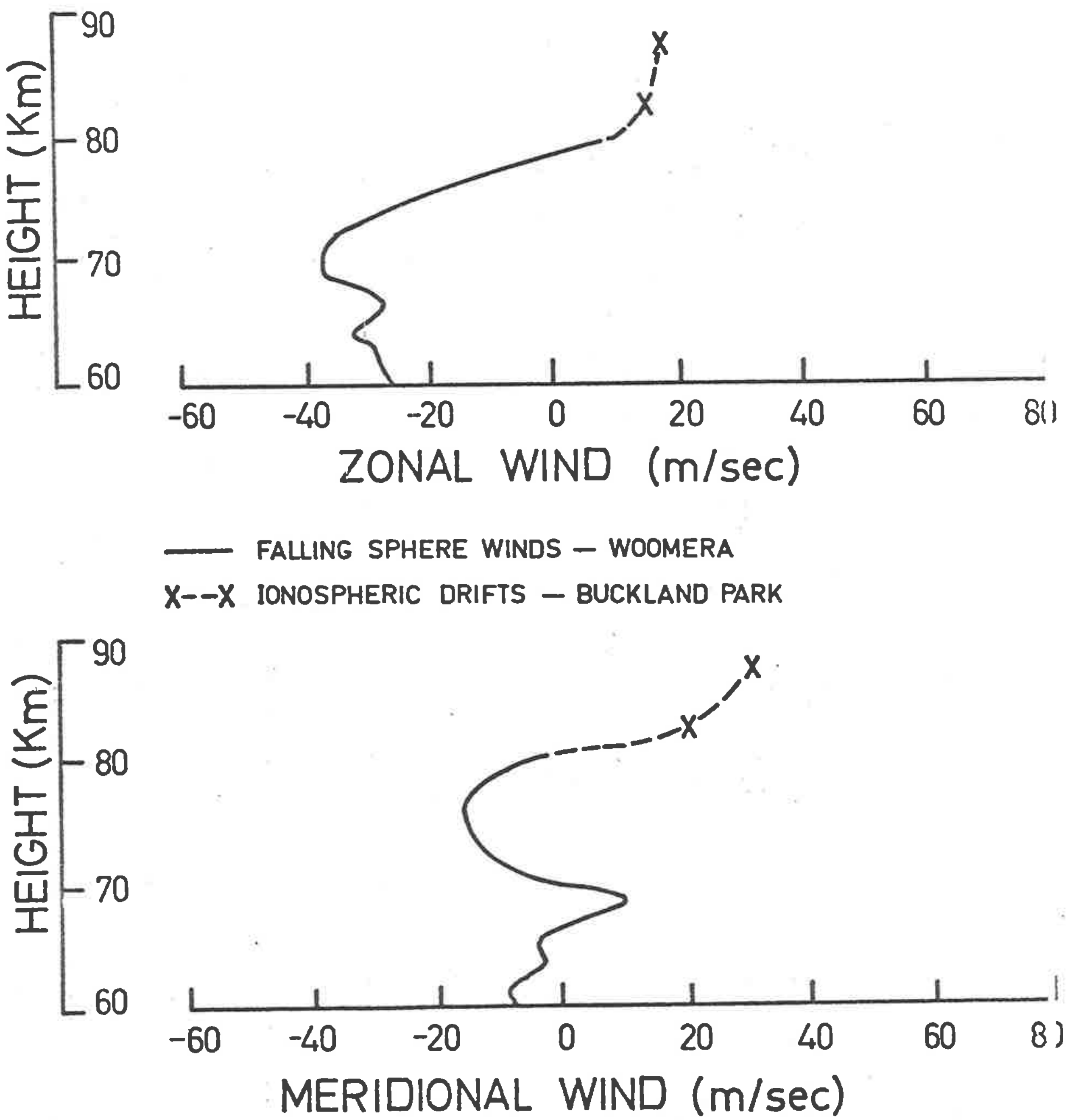


FIGURE 6.7 *Comparison for sphere of November 16, 1972.*
Time of firing: 1934 LT

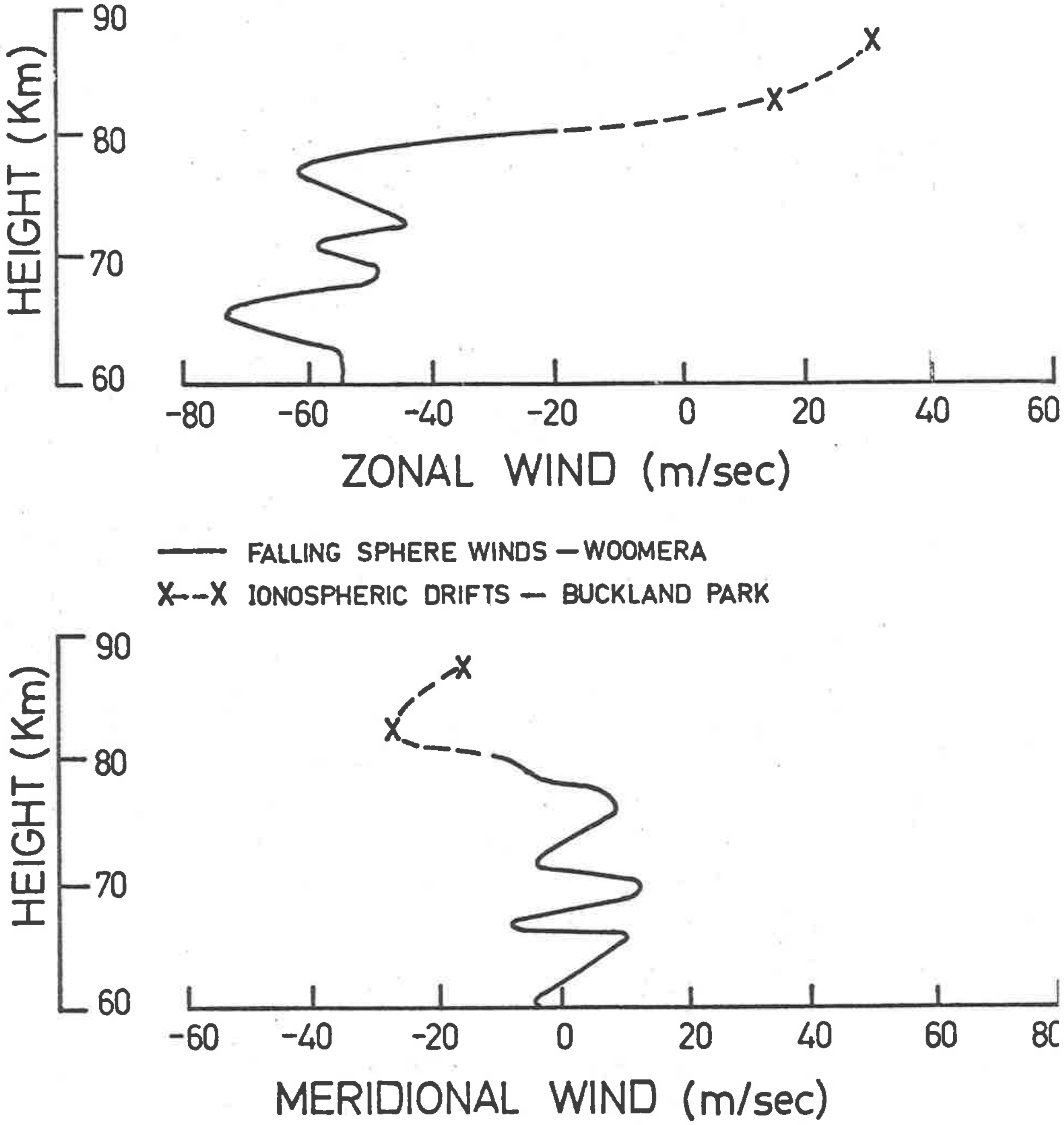


FIGURE 6.8 *Comparison for sphere of December 13, 1972.*
Time of firing: 2000 LT

sometimes occur in the region. Most of the composite profiles have a distinctly oscillatory character but it is rather meaningless from a single sounding to assign the oscillations to tidal or irregular sources. In FIGURE 6.2 though, the two components have a vertical structure that could be associated with the diurnal tide: the main features of the zonal profile are some 8 - 10 km displaced downward with respect to the meridional profile and both curves have an apparent half-wavelength in the vicinity of 16 km. This interpretation seems the more likely in view of the fact that the sounding was in late February.

For the sphere experiment in August, 1973, Woomera drifts were available but only down to 85 km because of noise. The results are shown in FIGURE 6.9. Again, the overall trends are quite consistent although the inflection in the upper 2 km of the zonal component of the sphere winds implies a fairly sharp structure in the intervening 5 km. By the same token, it would not require a shear as great as occurs in meridional component near 90 km.

The results for the one daylight sphere comparison to date are shown in FIGURE 6.10. The rocket was fired at 1230 CST on March 21, 1974 so the sphere was in the volume of interest at about 1232. Woomera drift records for the 6 minute

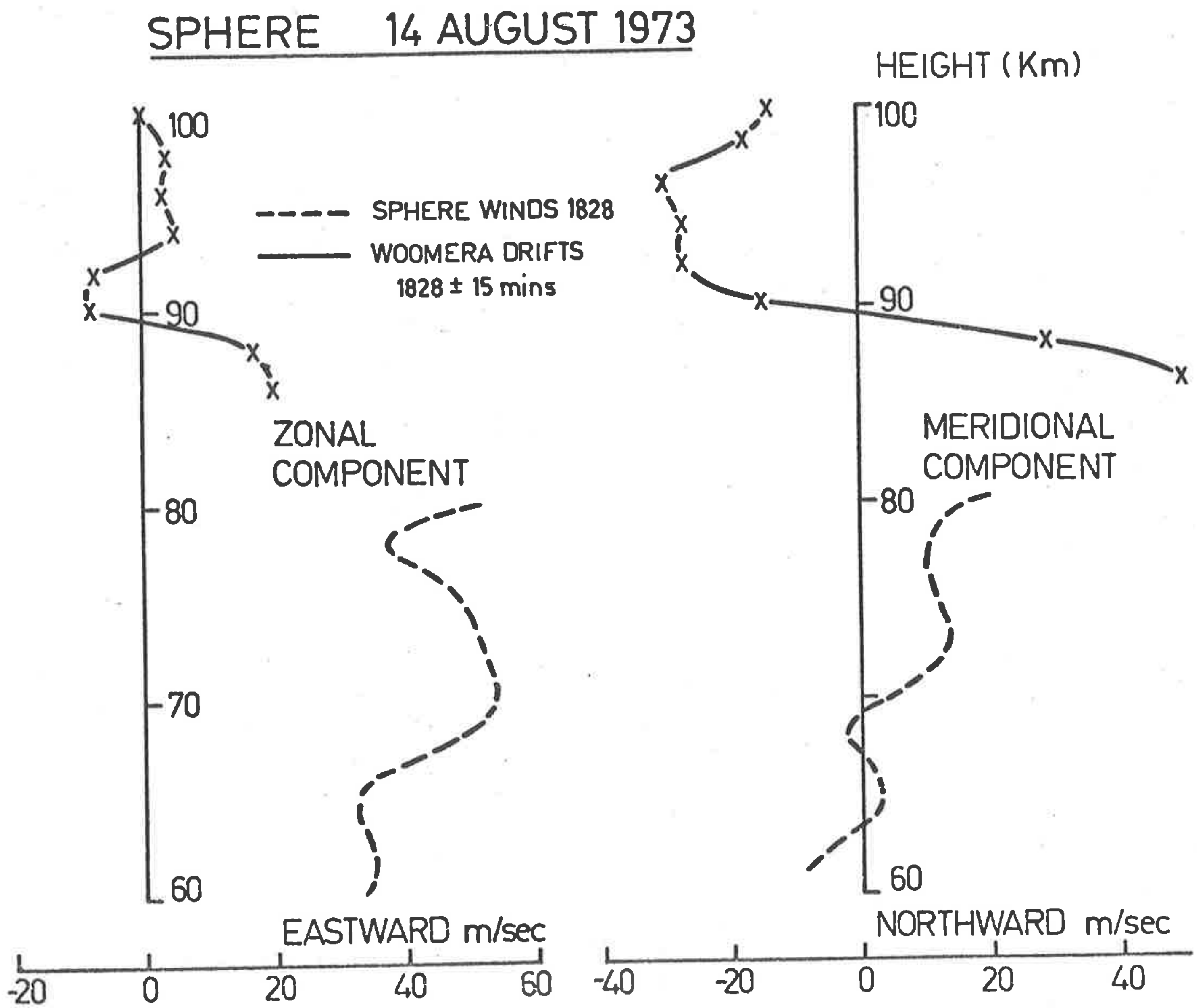

SPHERE 14 AUGUST 1973
HEIGHT (Km)
100
90
80
70
60
SPHERE WINDS 1828
WOOMERA DRIFTS
1828 ± 15 mins
ZONAL
COMPONENT
MERIDIONAL
COMPONENT
EASTWARD m/sec
NORTHWARD m/sec
-20
0
20
40
60
-40
-20
0
20
40

FIGURE 6.9

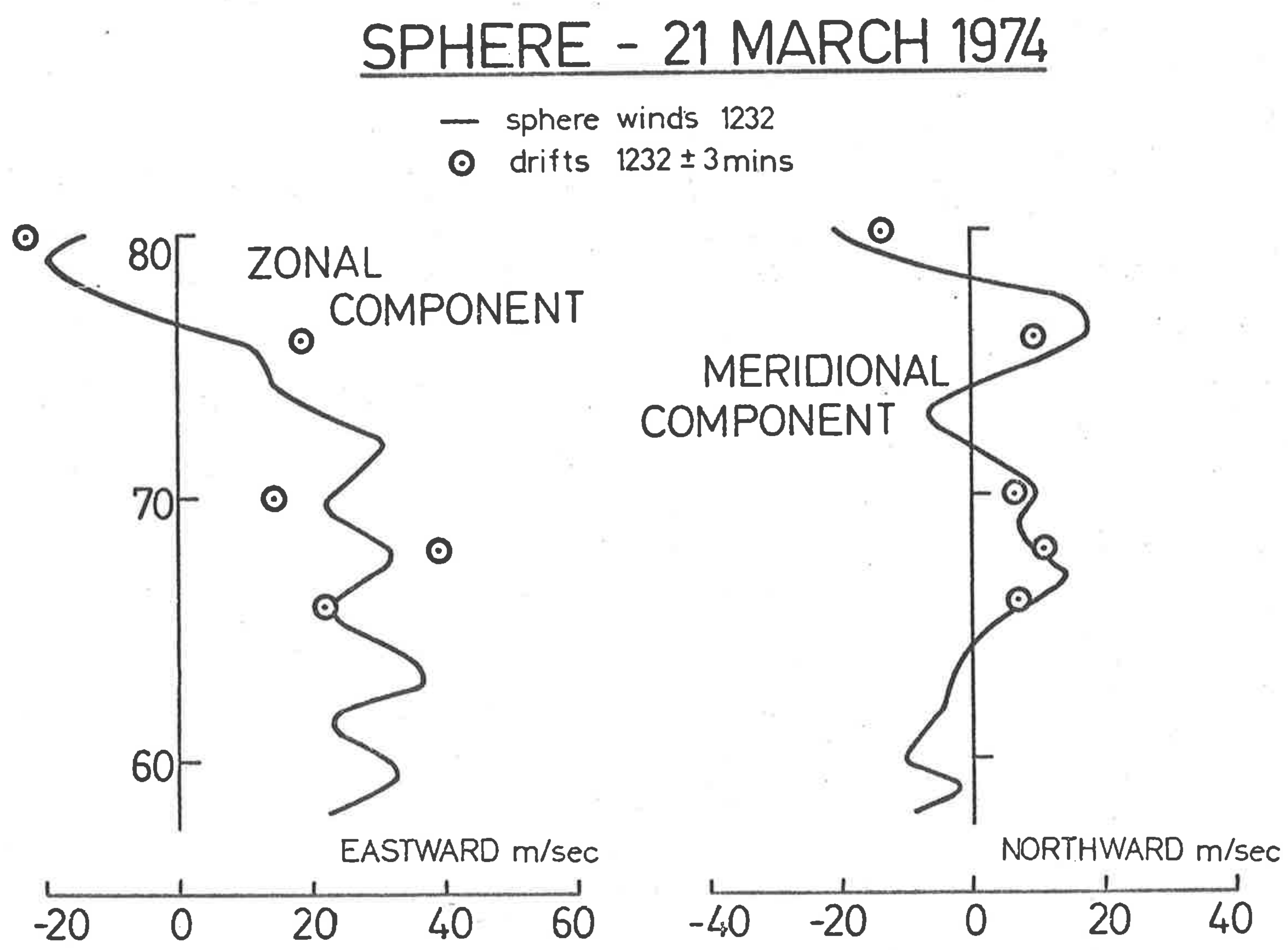


FIGURE 6.10 Common volume daylight sphere comparison.

period 1229 - 1235 were analysed in 3 minute blocks starting at 1229, 1230, 1231 and 1232. Drift values were obtained for 5 of the 10 possible heights during this period and the averages are represented by the circled dots in FIGURE 6.10. At all heights and for both components the two techniques are in good agreement.

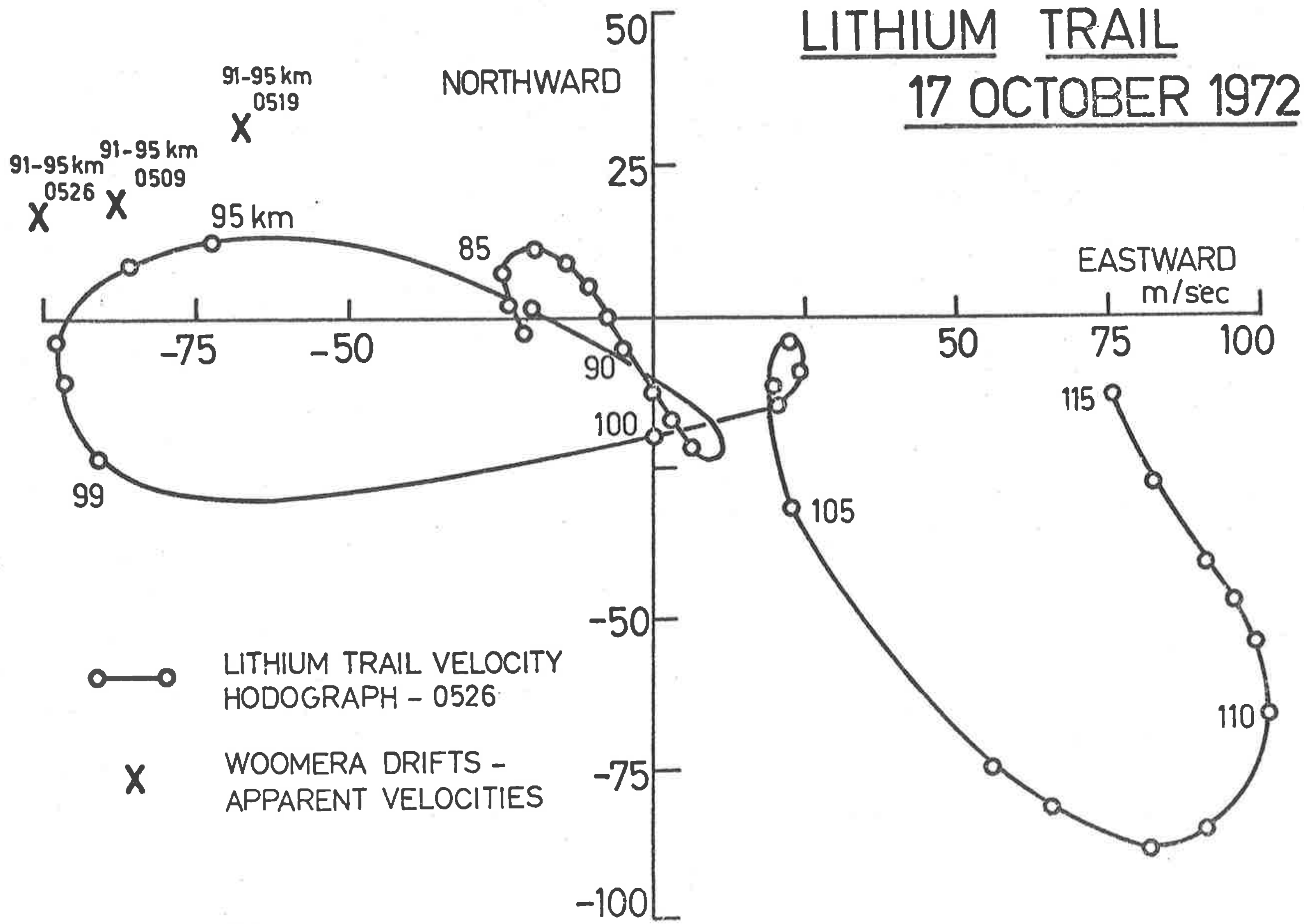
6.2.3 Vapour trails

In October, 1972 an early comparison was made by taking to Woomera a very simple version of the drift equipment incorporating a low power transmitter, fixed height reception and a small chart recorder. The rocket was fired early in the morning and a lithium trail laid at 0526 local time (FIGURE 6.11). Fading of the ionospheric echo at 93 km was recorded from 0500 to 0540. Although the records were noisy because of the low power used, it was possible to calculate apparent velocities for several five-minute segments where the fading was deep. The resultant velocities are consistent in magnitude and direction and are in fair agreement (FIGURE 6.12) with the wind deduced from the lithium trail, remembering that apparent velocities invariably overestimate the actual drift.

A more intensive comparison was made for the lithium trail of April, 1973. FIGURE 6.13 is a sequence of photos

FIGURE 6.11

Photograph of the lithium trail - October 17, 1972.

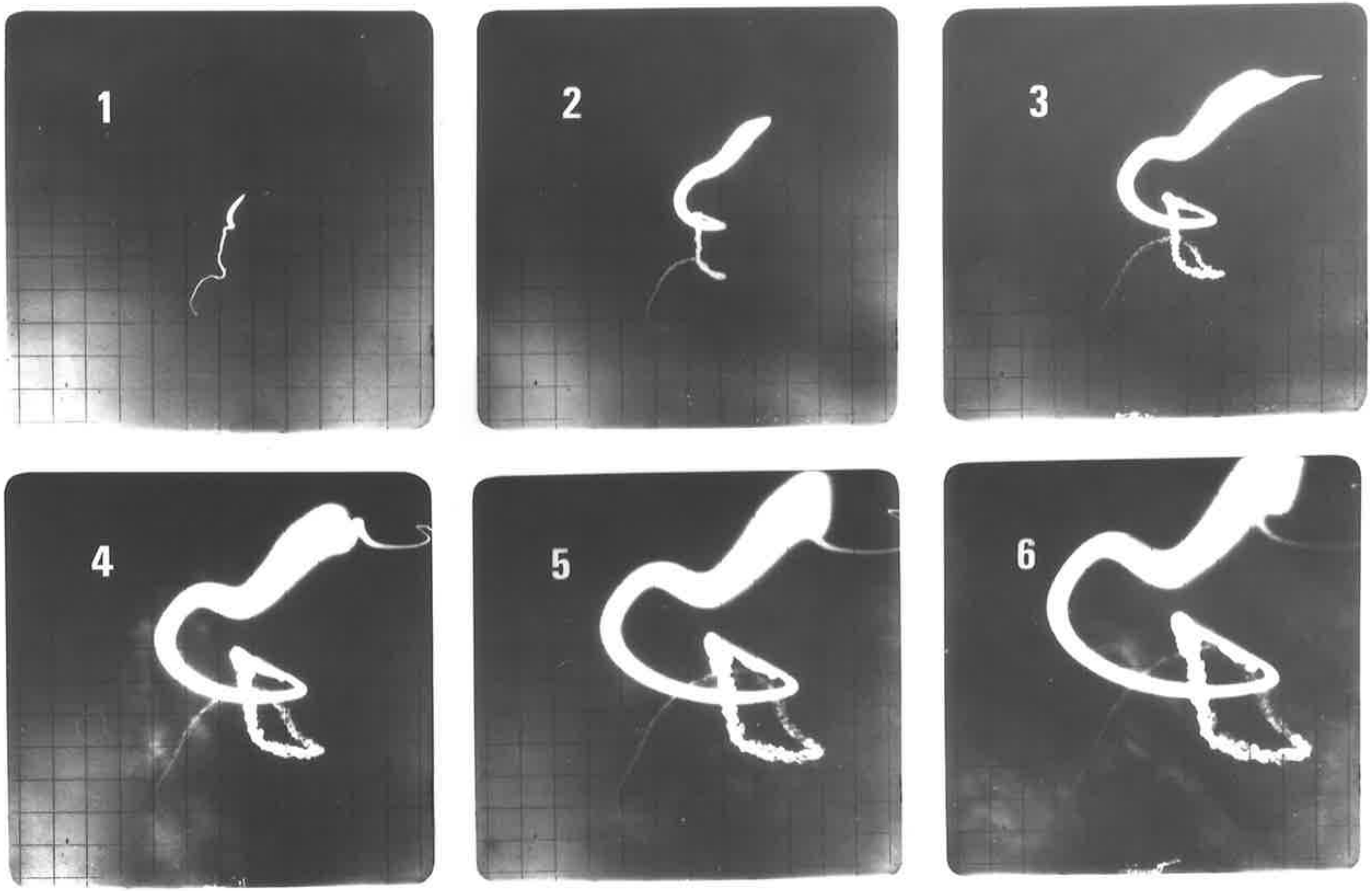


LITHIUM TRAIL
17 OCTOBER 1972
NORTHWARD
EASTWARD
m/sec
50
25
-50
-75
-100
-75
-50
50
75
100
91-95 km
0519
91-95 km
0509
91-95 km
0526
95 km
85
90
100
99
105
110
115
LITHIUM TRAIL VELOCITY
HODOGRAPH - 0526
X
WOOMERA DRIFTS -
APPARENT VELOCITIES

FIGURE 6.12.

FIGURE 6.13

Photographic sequence of lithium trail - April 19, 1973


1
2
3
4
5
6

of the trail at about 100-second intervals showing how it is distended by the wind. Drifts were obtained at Buckland Park over the height range 80 - 120 km. The individual drift results for 30 minutes either side of the mean trail time are plotted in FIGURE 6.14. A curve of best fit has been drawn for each component in order to construct a hodograph for comparison with the trail hodograph. Both curves suggest a wavelength of around 30 km and a growth in amplitude with height. These properties, and the sense of phase shift between the two components, are consistent with the diurnal tide, a dominant characteristic of the autumn months. The comparison with the rocket results appears in FIGURE 6.15. The salient feature of the motions at both sites is the tidal spiral, anticlockwise with increasing height. The lithium trail hodograph is like the smooth drift profile with a wavelike perturbation in the lower 20 km. This could indicate that the drift results failed to pick up this short-term effect or that it was in fact associated only with the Woomera site.

The first lithium trail tracked by the daylight scanning system was on the 21st June, 1973. The trail was observed between 0850 and 0901. Drift records were taken

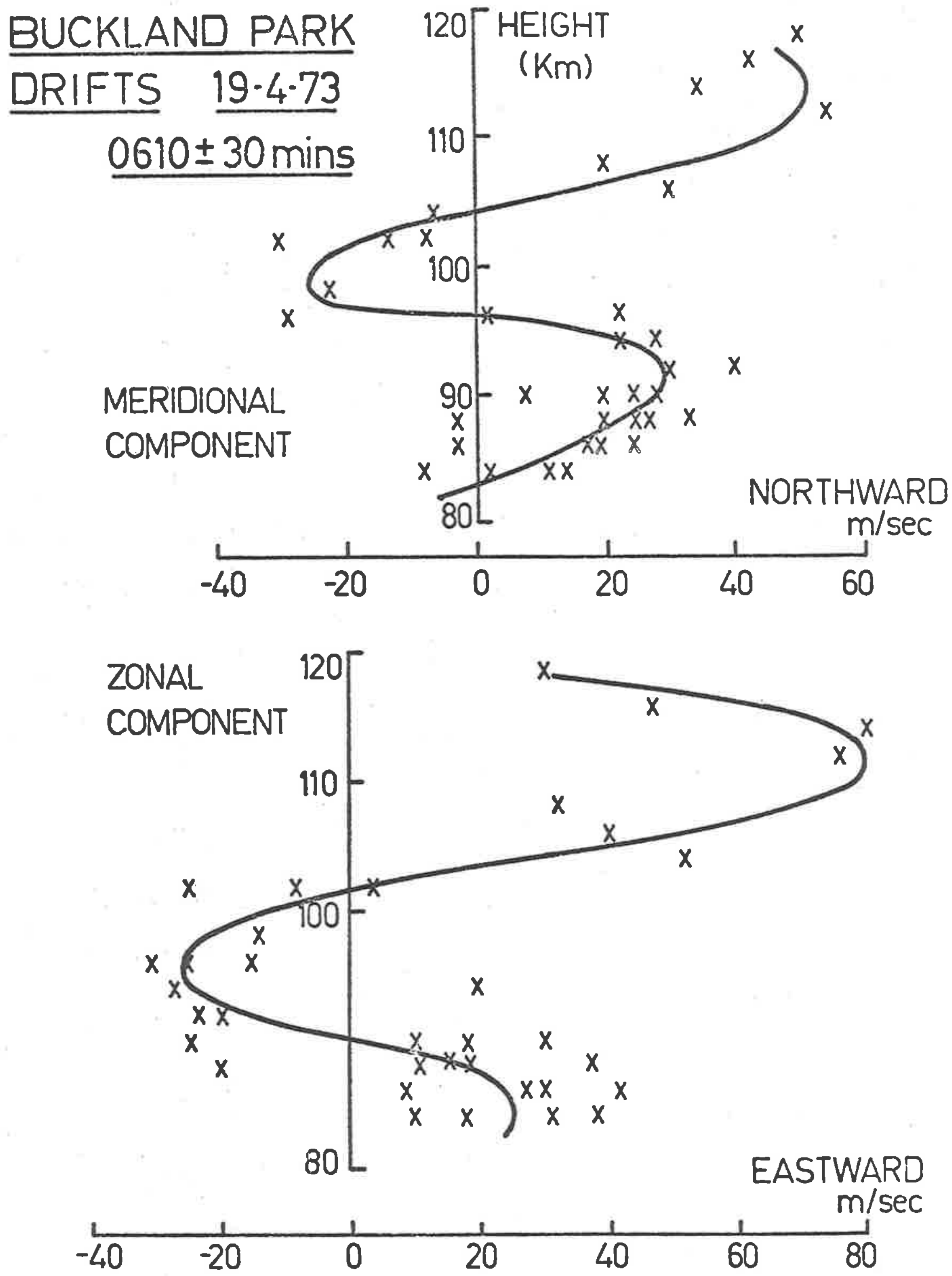


FIGURE 6.14 Buckland Park drifts for comparison with lithium trail of April 19, 1973.

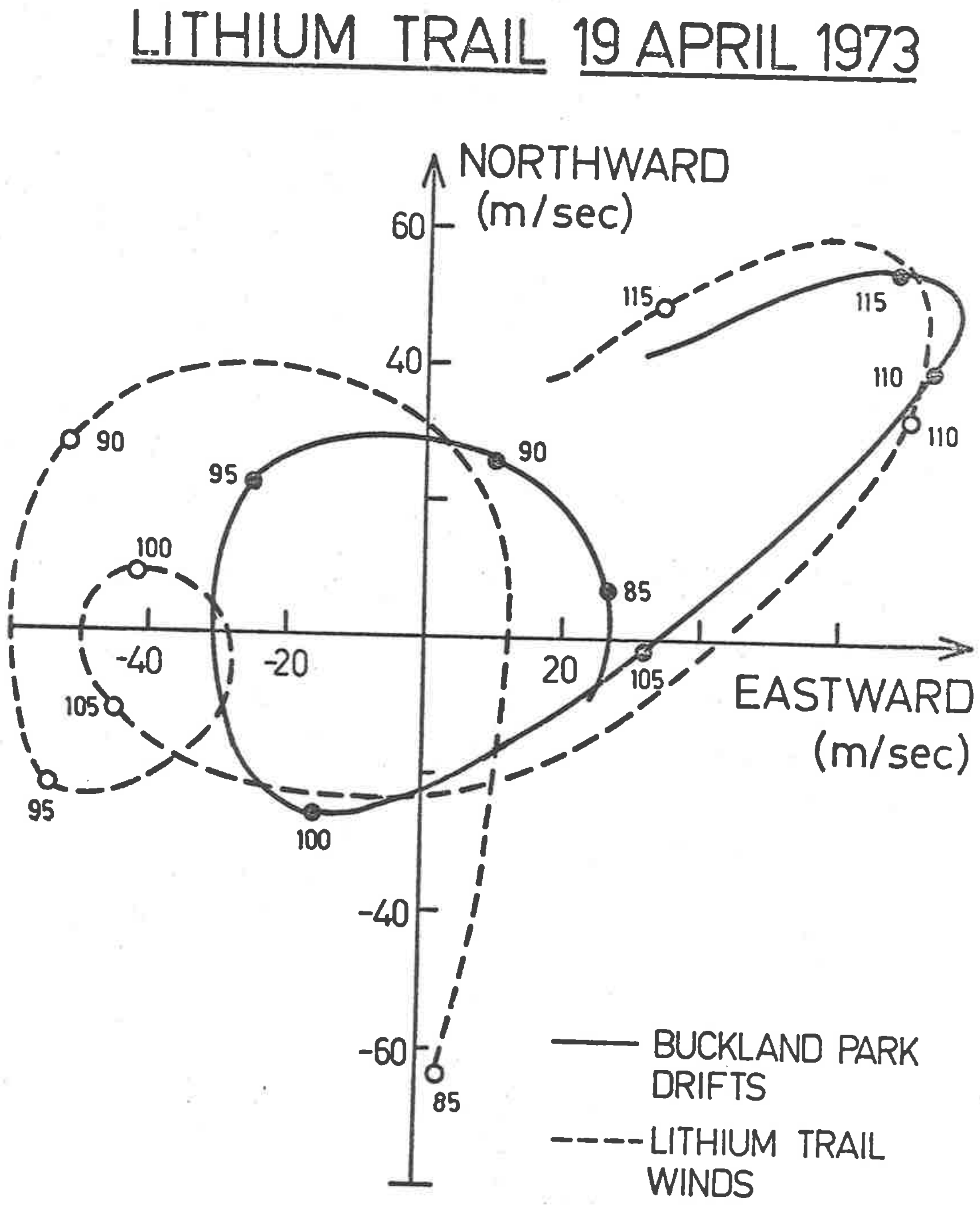


FIGURE 6.15 *A comparison of Buckland Park drifts and Woomera winds deduced from the lithium trail of April 19, 1973.*

continuously at Woomera during the experiment and then treated in the following way. From 0849 to 0903 the fading at intervals of a minute in overlapping 3 minute segments was analysed using each of the four possible right-angled triangles. The average of the various triangles was then plotted for each height to determine a mean value for the period of interest. The results at two of the heights are shown in FIGURE 6.16. It can be seen that the values generally vary in a systematic fashion, often quasi-sinusoidally, but particularly at the greater height the extent of this variation can be quite large. This emphasises the desirability for comparisons, of having drift data as close as possible to firing times. The actual comparison of the resultant drift hodograph and that deduced from the trail is the subject of FIGURE 6.17. For the 10 km of height overlap there is a singular lack of agreement. Notwithstanding the generally assumed infallibility of rocket results, they are in this case particularly surprising. For example, the zonal value at 90 km of 80 m/sec Westward compares with the mean prevailing component for June of 20 m/sec Eastward. The smooth anticlockwise rotation suggests that the difference is most likely to be due to a large tide although these are not common in winter. In fact in the strongest case observed in the drifts (FIGURE 3.11)

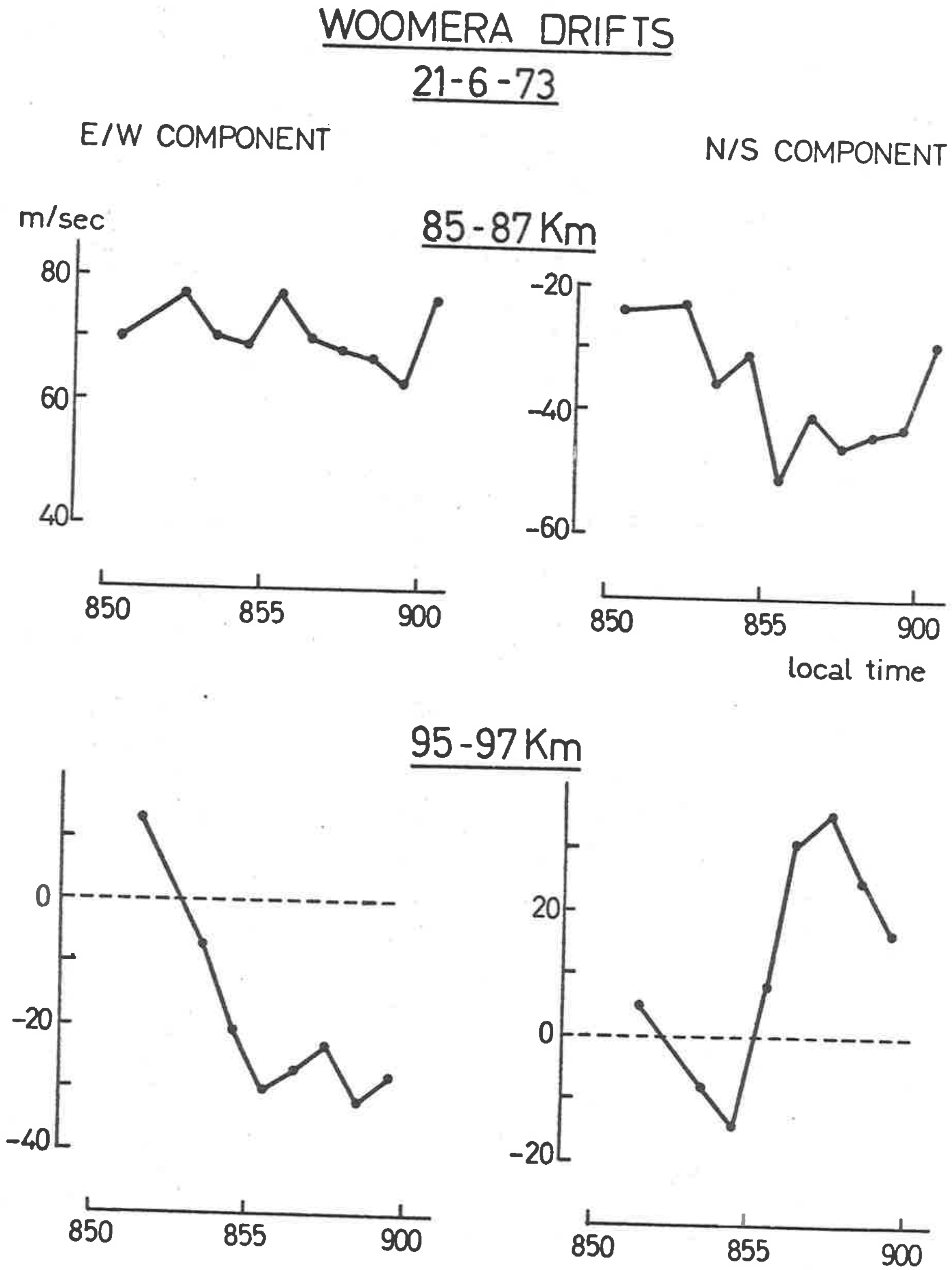


FIGURE 6.16 *Drift fluctuations near the firing time for the lithium trail of June 21, 1973.*

LITHIUM TRAIL 21 JUNE 1973

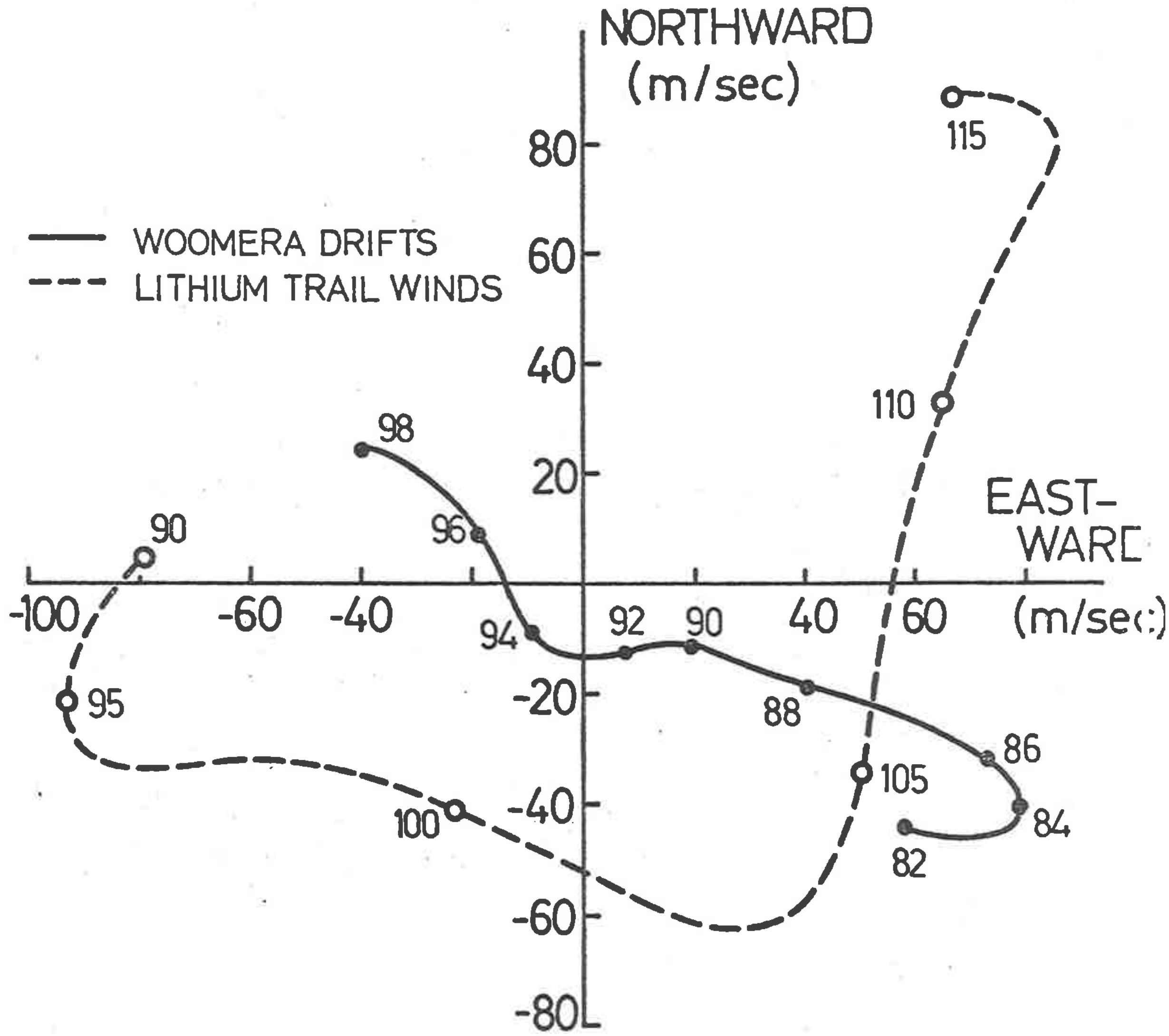


FIGURE 6.17 A comparison of Woomera D-region drifts and Woomera winds deduced from the lithium trail of June 21, 1973.

the phase is such that the contribution is strongly Eastward in the late morning. It is worth noting that on other days in the month when the meteor system was running no similar Westward value was obtained. The disagreement remains puzzling and although a check of the daylight scanning system using a visually tracked twilight trail is planned, it has not yet occurred.

The experiment of September 19, 1973 was beset with problems. Three-minute drift records were taken at Buckland Park at 1420, 1425, 1445, 1450 and 1455 and from 1428 to 1445 fading was continuously recorded at Woomera to coincide with the trail. However the Woomera drift records were unusable due to a recorder malfunction and additionally only one of the two optical observation sites was fully operative. This means that the trail heights must be viewed with caution. The two hodographs, using the averaged Buckland Park drift values, are shown in FIGURE 6.18. In detail they are not at all coincident although they lie basically in the same quadrant. Only near 94 km are the values close. The differences could be due to a real difference in the techniques or, just as easily, the separation of the sites and the lack of values just after the actual firing time.

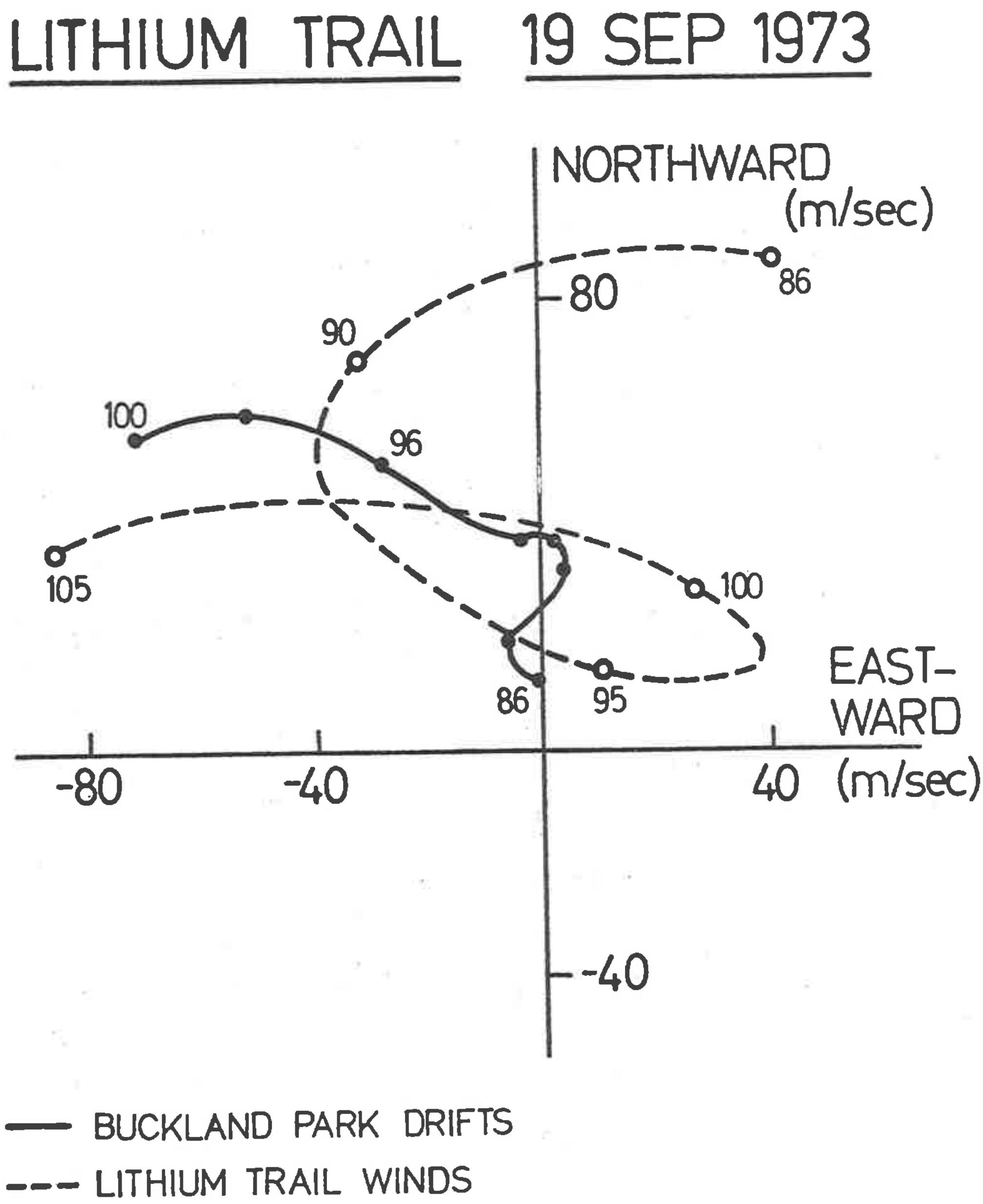


FIGURE 6.18 *A comparison of Buckland Park drifts and Woomera winds deduced from the lithium trail of September 19, 1973.*

6.3 Summary

The comparisons described in this chapter have featured varying degrees of compatibility between the techniques as far as time, height and horizontal separation are concerned. Of the common-volume comparisons, only the June daytime lithium trail produced significant discrepancies. In contrast to this, the one daylight sphere experiment produced a good height overlap and particularly encouraging agreement. Of those comparisons with small breaks in height continuity there were no glaring inconsistencies at the closest approach. For the two lithium trails where only Buckland Park drifts were used, the results could be reasonably interpreted as suggesting similar mean and tidal features at the two sites but with definite differences due to localised short-term effects. This will be considered more fully in the ensuing chapter.

CHAPTER SEVEN

AN ADELAIDE/WOOMERA COMPARISON

In view of the correspondence between the drifts and neutral winds, comparisons between drift measurements from distant sites should yield valuable information on the horizontal scale of the wind systems at ionospheric heights. In this chapter some early results of such a comparison between Adelaide (Buckland Park) and Woomera will be presented. The respective co-ordinates are [34° 38'S,138° 37'E] and [30° 45'S, 136° 18'E].

7.1 Previous Work

As was discussed in Chapter 3, many of the early comparisons between drifts and other techniques involved large horizontal separations, assuming stability of the neutral wind over distances up to 1000 km. Recently, two groups have conducted experiments involving widely spaced sites with the aim of discovering the features common to each. MULLER and KINGSLEY (1974) have reported a series of meteor wind observations utilising three sites in England. With combinations of beams they were able to obtain a minimum ionospheric separation of 236 km and a maximum of 843 km. They found that the major constituents of the wind, the tidal oscillations and planetary waves, had scales in the horizontal of at least several hundred kilometres. They ascribed the differences to turbulence and gravity waves. At 800 km separation a slightly increased discrepancy was obvious but no more than would be expected from a tropospheric weather system origin for the planetary waves. The recent Russian work (KOKOUROV et al, 1971; KAZIMIROVSKIY and KOKOUROV, 1972) has been with total reflection drifts. However although they have used two sites, only one has been used for reception so that the reflections are oblique and vertical. Thus there is, as well as a 50 km horizontal separation, a vertical separation of from 3 to 10 km depending on ionospheric conditions. They found similar statistical properties for the two sets of results but definite differences in synchronous measurements. The height difference however makes it hard to draw conclusions from their work.

7.2 Specific Examples

The Adelaide/Woomera comparison work is still at an early stage. Some continuous common sequences were obtained in June, August and October 1973 and February, 1974. A problem arises from the difficulty of communication between the sites. The best and most continuous results may be obtained for different heights at the two sites because of either differing receiver gain settings or different ionospheric conditions. It should also be borne in mind that the arrangement of receiving aerials is not the same at Woomera as it is at Buckland Park. However in both cases the basic aerial separation is sufficient to minimise the 'triangle size' effect.

Four examples characterised by the best available data density and continuity at both sites, are shown in FIGURES 7.1 and 7.2. In all cases there are times when the agreement is good. For example, in the meridional component at the bottom of FIGURE 7.1 the consistency from 2100 to 2300 is quite remarkable. On the other hand, after 2300 the results are too sparse to determine whether there exist real differences as seems indicated. Certainly all the examples indicate a generally close correlation between the results.

FIGURE 7.3 is of interest because it shows a prolonged systematic difference between the two sets of results. Only the zonal component is drawn at each of two heights, no similar difference being evident in the meridional component. This would imply that the difference was due to a latitudinal shear in the mean wind since, if the effect were related to a tidal amplitude difference, a similar

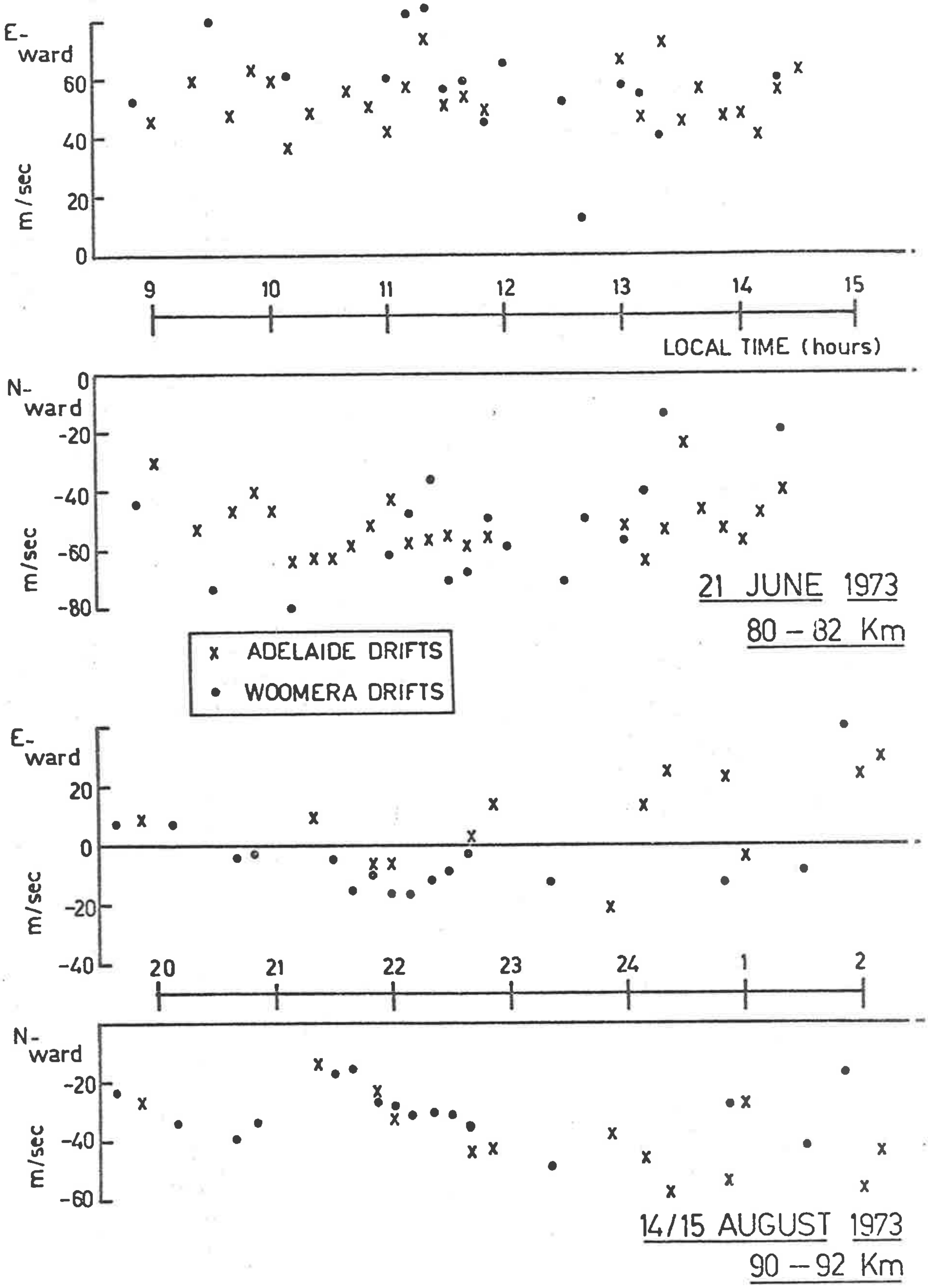


FIGURE 7.1 Comparisons of individual drift determinations at Buckland Park (Adelaide) and Woomera.

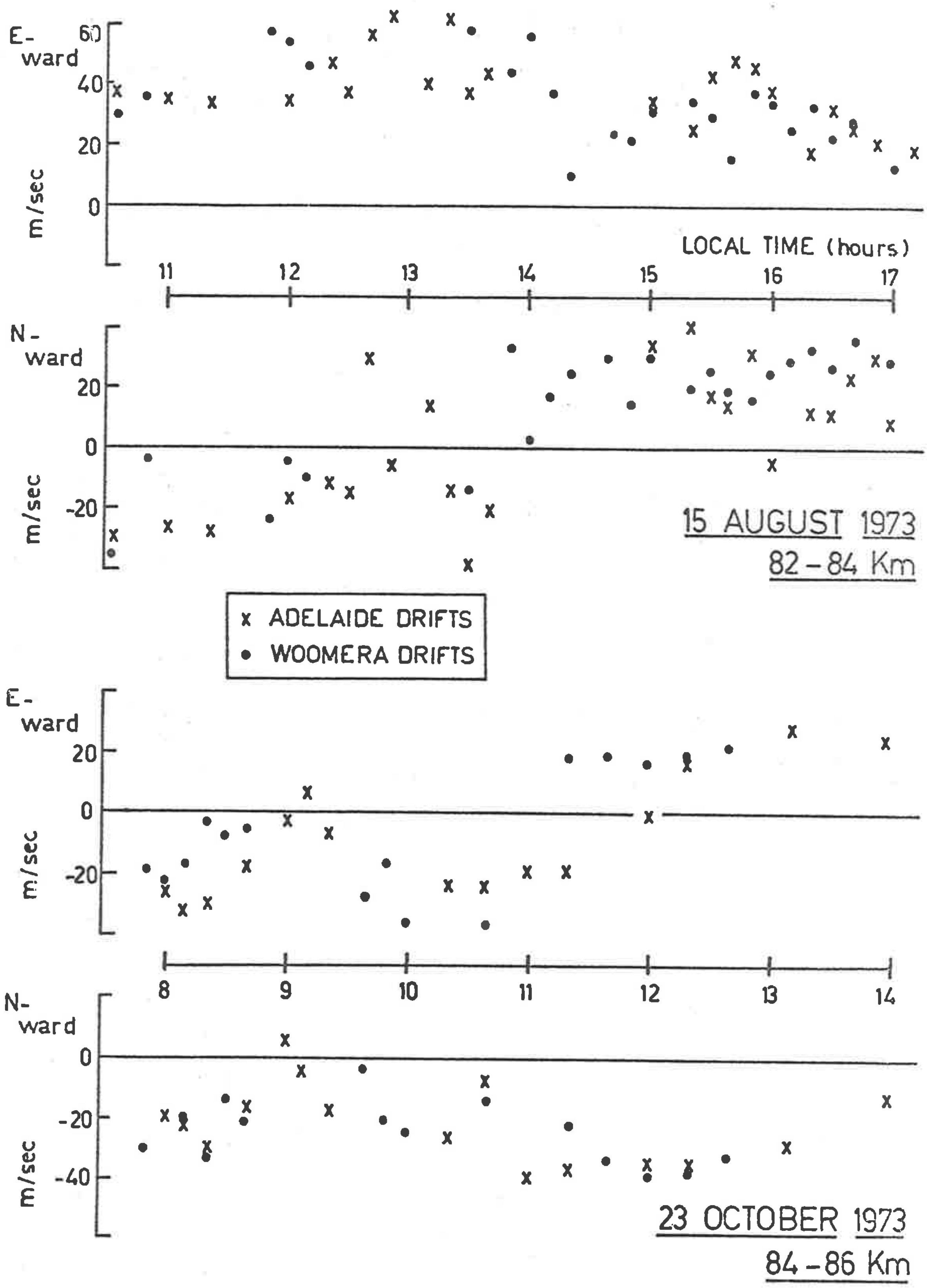


FIGURE 7.2 Comparisons of individual drift determinations at Buckland Park (Adelaide) and Woomera.

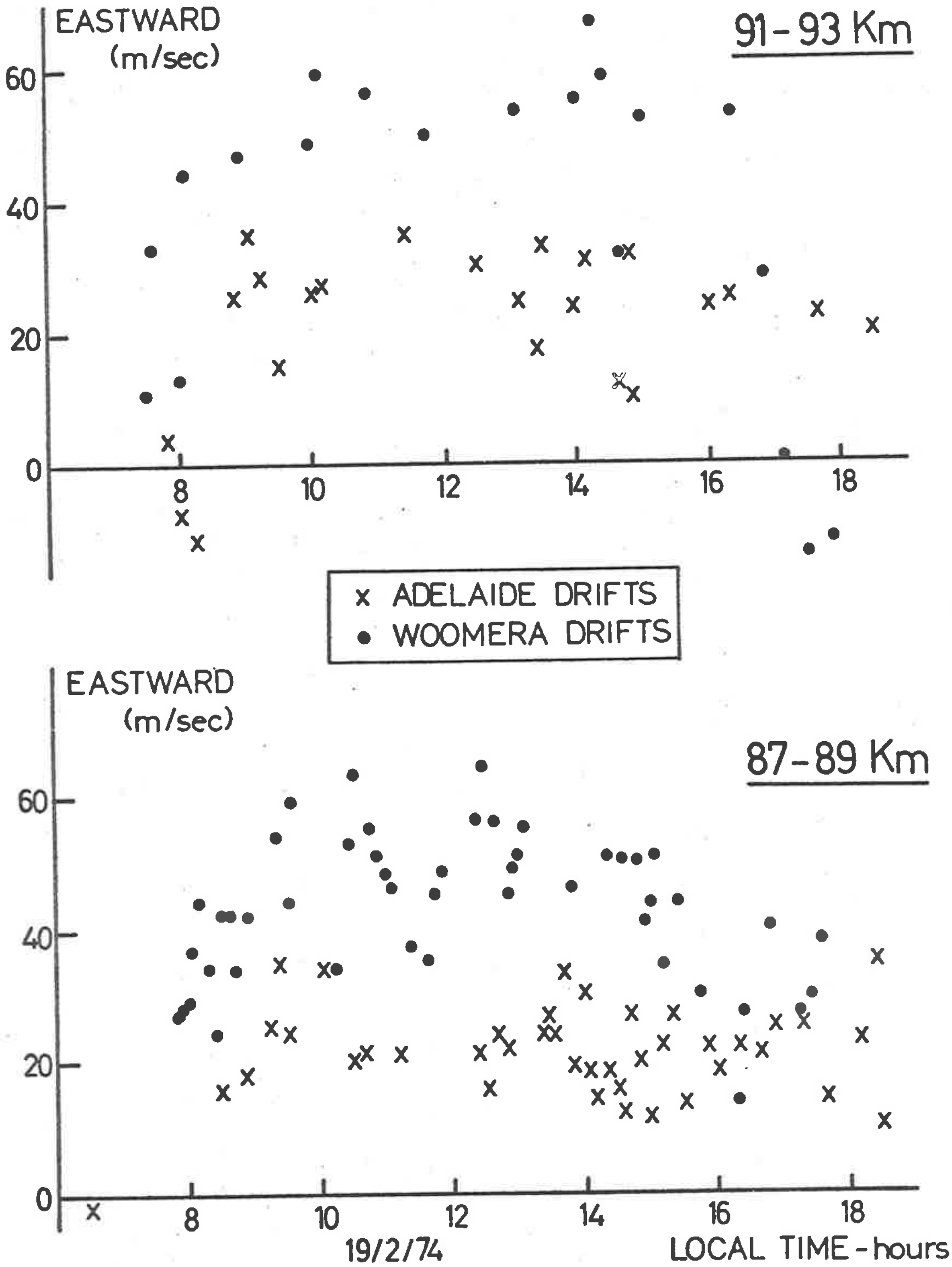


FIGURE 7.3 A comparison of individual drift determinations at Adelaide and Woomera showing systematically different behaviour.

discrepancy would be expected in the meridional component. However in other ways the difference in FIGURE 7.3 is more suggestive of a larger tide at the Woomera site, because at either end there is an indication that the Woomera drifts are crossing the Buckland Park values and would probably become more negative. The phase would be typical for the diurnal tide in February. Unfortunately no night-time values were available to decide between the possibilities.

7.3 General Features

In order to use as much of the data as possible and to look for any general trends, half-hour averages at the two sites were compared for each 2 km height interval. These comparisons have been used to produce the scatter plots of FIGURES 7.4 and 7.5. Although two height intervals have been grouped in the diagrams the actual comparisons are for the one height range. Woomera values at 81 - 83 km for example have not been compared with Adelaide values at 83 - 85 km. The main features of the plots are as follows:

(i) The degree of correlation decreases with increasing height. Although there is not a great deal of observable difference between the plots of FIGURE 7.4 the distribution between 89 and 93 km at the top of FIGURE

ADELAIDE-WOOMERA COMPARISON: JUN, AUG, OCT, 1973.
TIME RESOLUTION: 30 min.

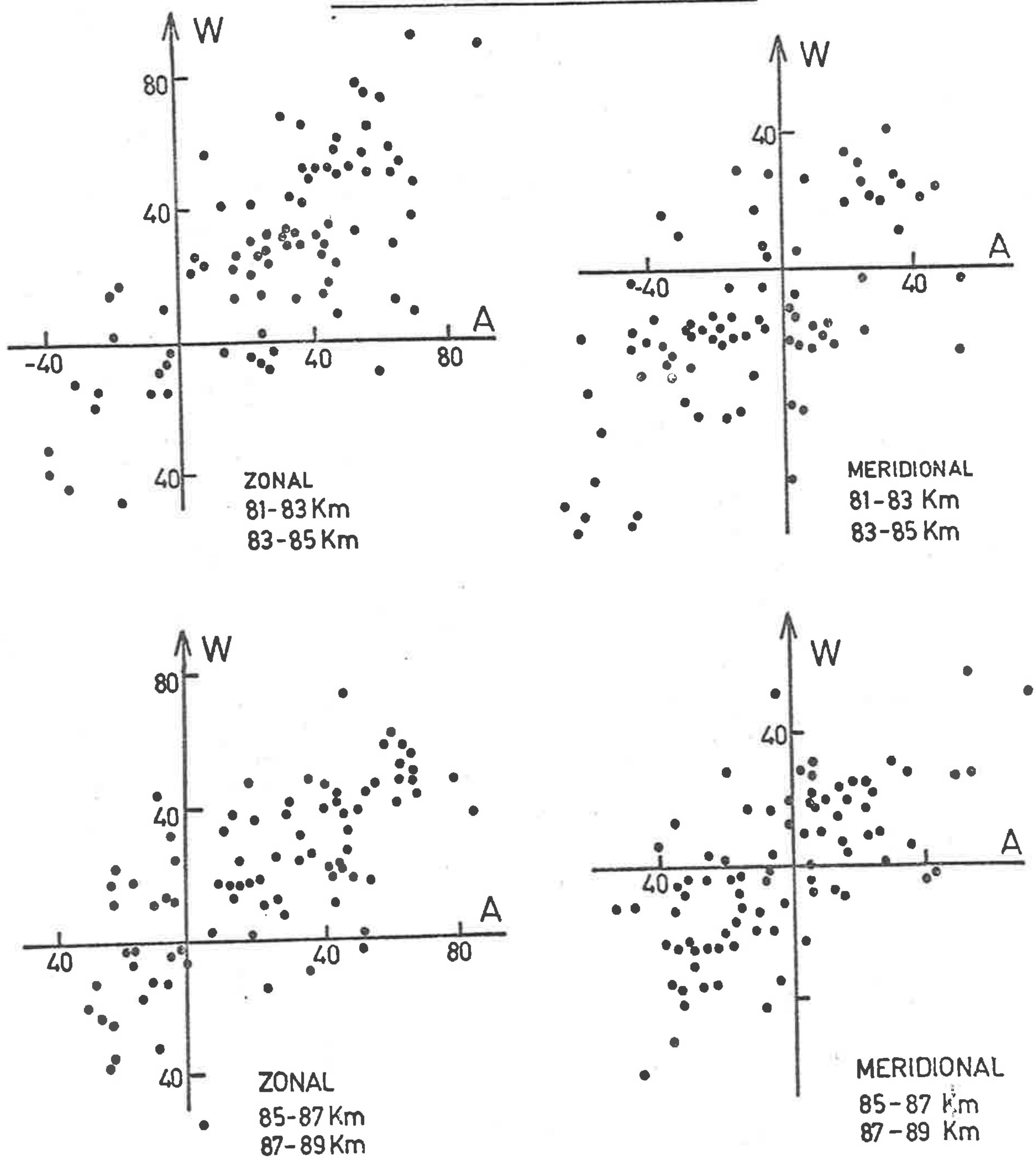


FIGURE 7.4 *A general comparison of half-hour average drifts at Adelaide and Woomera.*

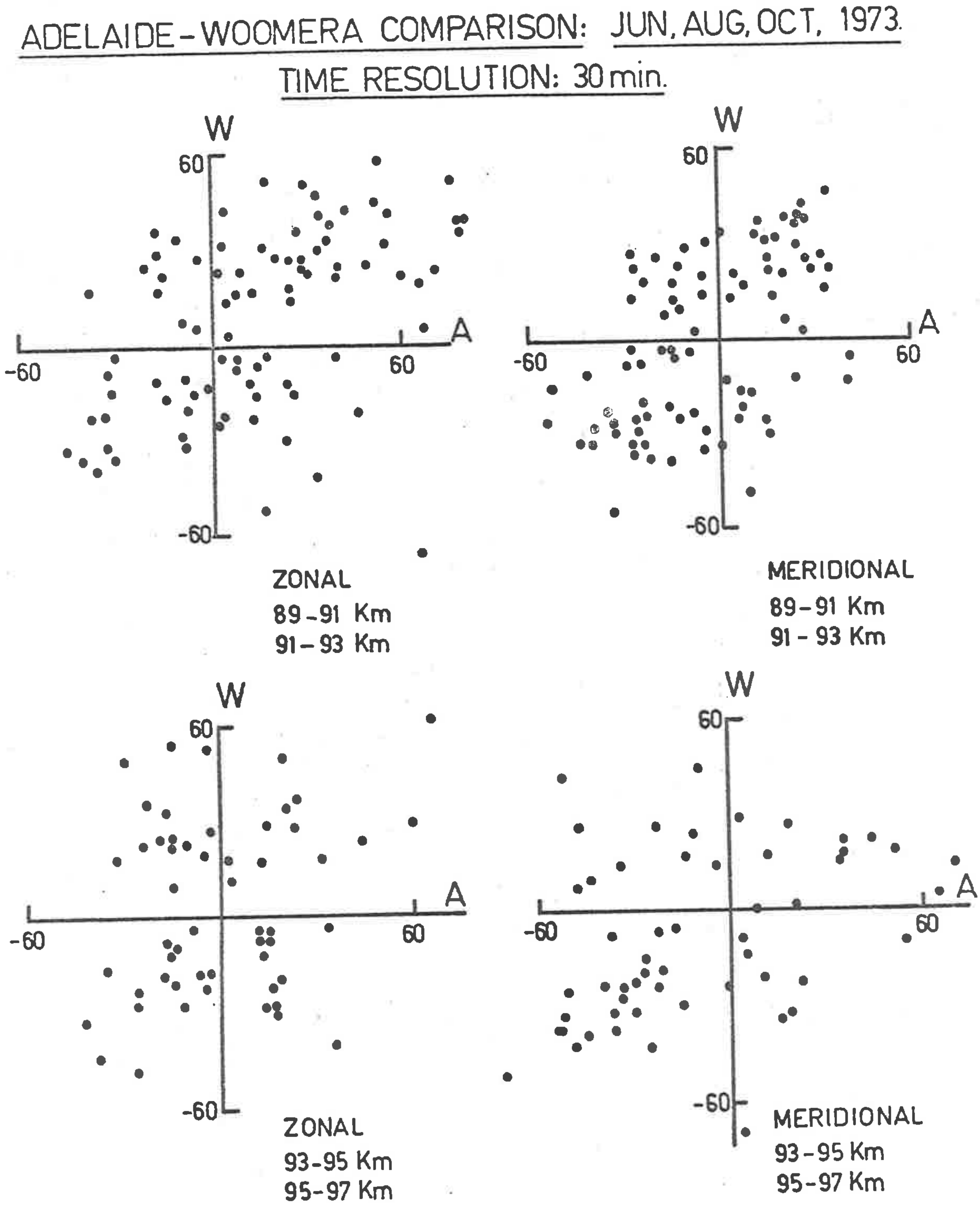


FIGURE 7.5 A general comparison of half-hour average drifts at Adelaide and Woomera.

7.5 is noticeably more scattered and above 93 km there is little correlation. The actual correlation coefficients have been plotted in FIGURE 7.6.

(ii) There is no evidence of a consistent difference in magnitude between the two sites. In the cases where the correlation is high, the line W = A appears a good estimate of the preferred distribution.

(iii) There is no significant difference between the zonal and meridional components as far as the extent of correlation is concerned.

7.4 Summary

From the limited amount of data so far available for Adelaide/Woomera comparison purposes, it is clear that there is often a very good correspondence, even for individual soundings, between drift values at the two sites, particularly for heights between 80 and 90 km. Between 90 and 100 km the values may often be completely different and it is tempting to equate this difference to the increased influence of irregular components at these heights, whilst preserving the horizontal stability of the mean and periodic motions.

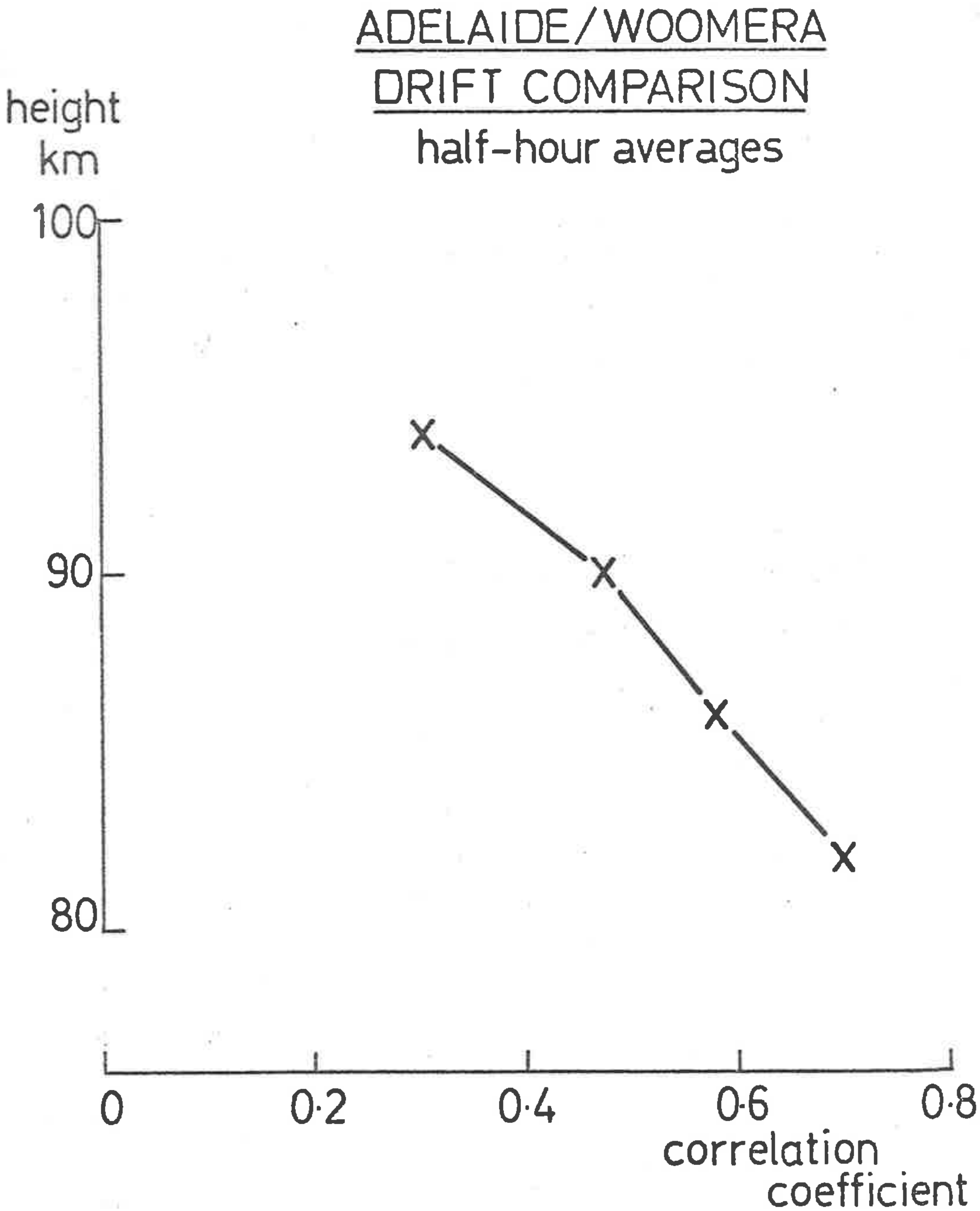


FIGURE 7.6 *The variation with height in the degree of correlation between the Adelaide and Woomera D-region drifts.*

A comparison of FIGURES 5.12 and 7.6 encourages such a view. However the need for detailed continuous observations enabling the day-to-day comparison of means and tidal amplitudes and phases is apparent.

CHAPTER EIGHT

GROUND PATTERN PARAMETERS

In previous chapters the emphasis has been on the drift values and on how they compare with neutral wind measurements. In the process of correlation analysis several other parameters are defined and calculated. In some cases they fluctuate strongly from one record to the next but on the average show interesting consistent variations vertically, diurnally and seasonally. These variations and the physical interpretation of the parameters themselves will be considered in this chapter.

8.1 Pattern Size

From the ground diffraction pattern, an ellipse is defined at the 0.5 level of the two dimensional spatial correlation function. The parameter used in this section as an indicator of the size of these ground pattern irregularities is the scale S where $S = \sqrt{ab}$ a and b being the lengths of the major and minor axes respectively.

8.1.1 Height Variation

In FIGURE 8.1 the way in which the pattern scale was found to vary with height is demonstrated for two different sets

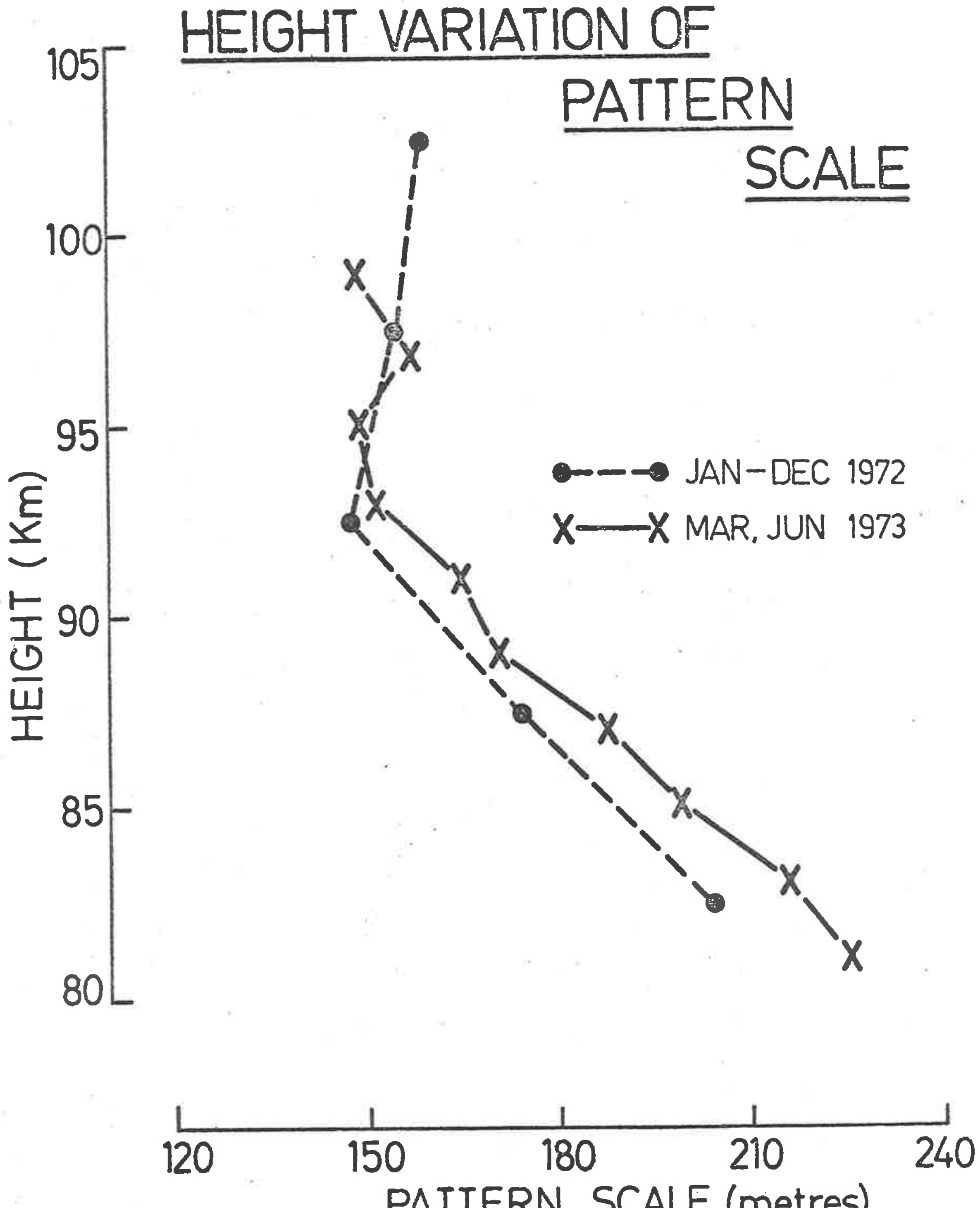


FIGURE 8.1 *The scale of the ground diffraction pattern as a function of the height from which echoes are accepted.*

of data. For the 1972 data, 5 km height gates were used and so five height groupings were appropriate for the averaging of the full year's data. In the 1973 case, ten 2 km intervals were averaged using the data from March and June. In all cases except the upper two heights of the 1973 data, sufficient values were averaged to reduce the error in the mean to less than 2 metres. The two striking features of the plots are the strong decrease in scale to a minimum between 90 and 95 km and the consistency in magnitude and trend of the two sets of data.

8.1.2 Diurnal variation

The pattern scale was also found to vary throughout the day. The average for a particular hour of the day was calculated at each height range. The 1972 results for three height intervals are displayed in FIGURE 8.2. The mean errors are typically 3 or 4 metres. At 80 - 85 km there is an enormous change of scale during the day with a pronounced minimum during the daylight hours. At 85 - 90 km this change is much less marked and there is an emergence of peaks near 0600 and 1700. This feature then dominates the diurnal variation for the 90 - 95 km data.

8.1.3 Seasonal variation

A third form of averaging applied to the pattern size data was a grouping into months in order to isolate any

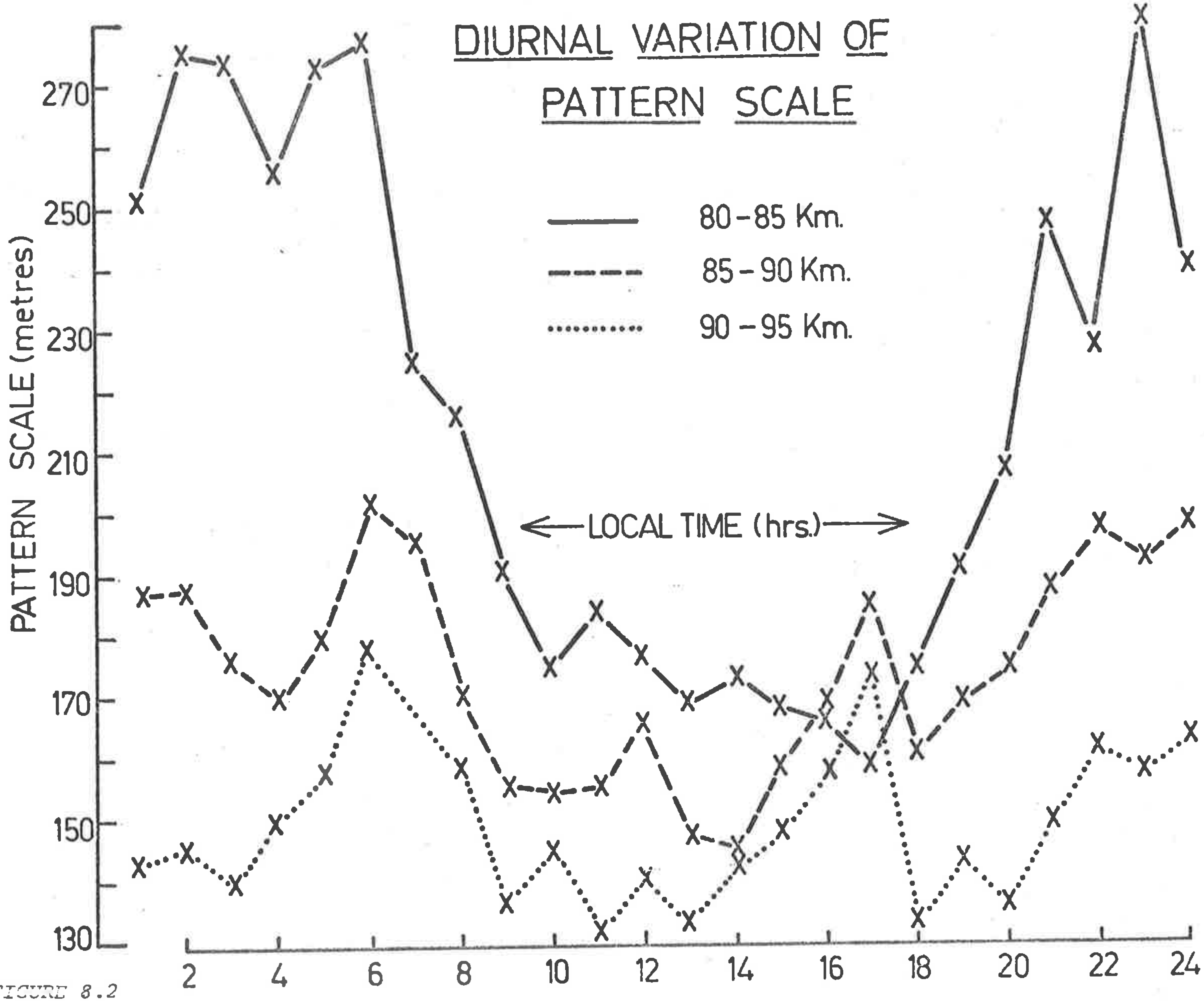

DIURNAL VARIATION OF PATTERN SCALE
80-85 Km.
85-90 Km.
90-95 Km.
PATTERN SCALE (metres)
270
250
230
210
190
170
150
130
LOCAL TIME (hrs.)
2
4
6
8
10
12
14
16
18
20
22
24

FIGURE 8.2

systematic seasonal changes. The results are shown in FIGURE 8.3. For the 85 - 90 km and 90 - 95 km ranges there is an almost purely semiannual effect with maxima at the equinoxes, the spring equinox being the larger. The variation for 80 - 85 km is similar but with more of an annual character due to a somewhat confused 'minimum' in winter. FRASER and VINCENT (1970) also discovered a strong seasonal variation in pattern size with equinoctial maxima, larger in spring.

8.1.4 Discussion

The relationship between the ground patterns and the actual ionospheric irregularities is a matter of some contention. It depends to a large extent on the model which one assigns to the reflecting layer. If it is assumed to be a continuous undulating surface from which several discreet rays are reflected then it is possible to show (BRIGGS, 1972) that for certain surface 'depths' the scale of the ground diffraction pattern can be much less than the horizontal scale of the ionospheric 'ripples'.

Whilst this may be the most appropriate model for total reflections from the E- and F-regions, it seems likely that for D-region partial reflections a scattering model with reflections from several bounded irregularities may be more appropriate. For example, phase path records from D-region

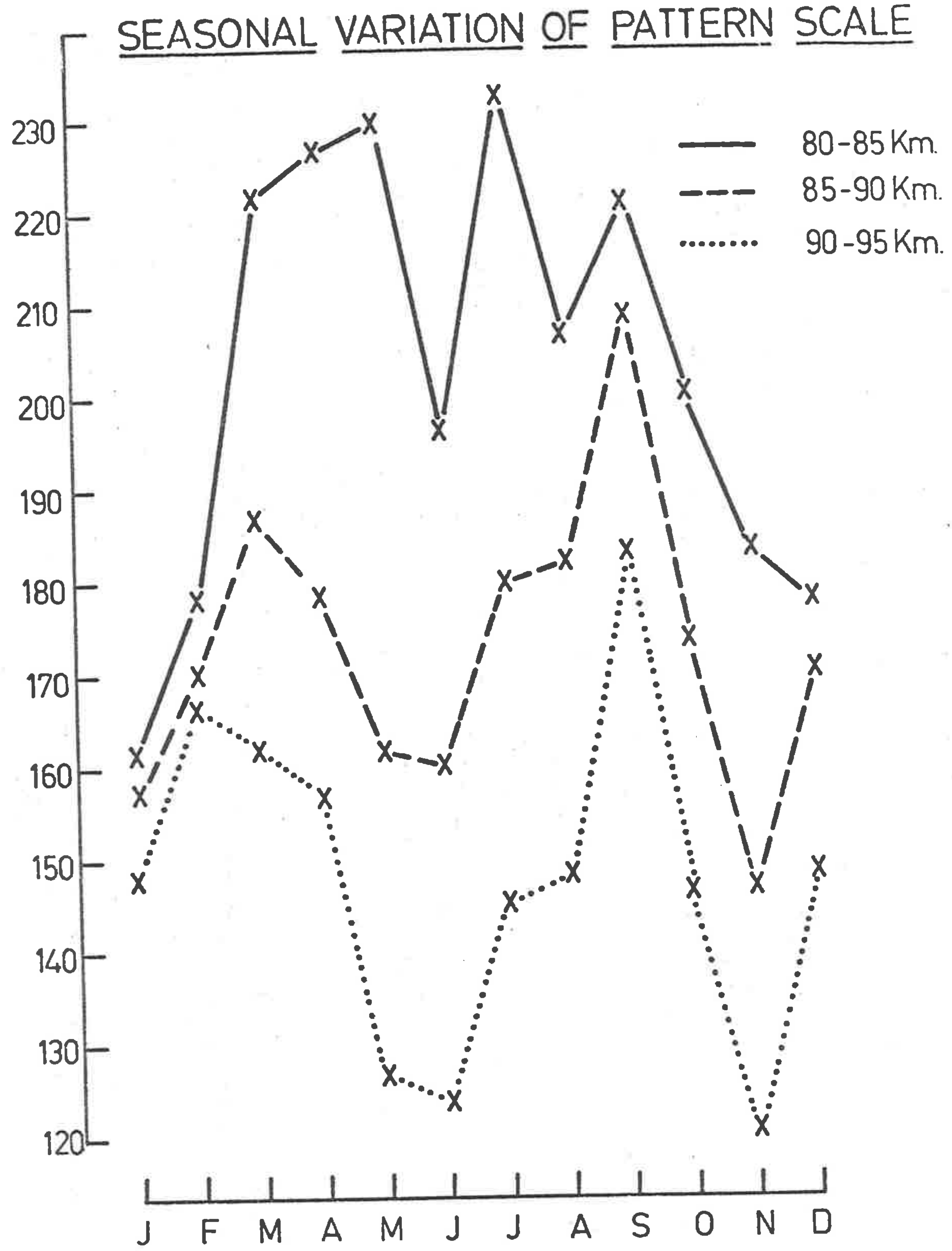


FIGURE 8.3 *The vertical scale is in metres. The initials are for the months of 1972.*

partial reflections usually have quite a different character from those for E-region total reflections. Smooth, slowly varying phase records are the exception rather than the rule for the D-region. In this case, it seems not unreasonable to assume that smaller ground patterns indicate a more disturbed ionosphere with finer structure and generally smaller irregularities. For two irregularities with the same vertical scale, the one with the smaller horizontal scale would scatter over a wider range of angles and this would then be transformed at the ground to the smaller scale diffraction pattern. Similarly any anisotropy in the irregularities should be reproduced in the ground pattern. VINCENT (1973) has shown that, at 2 MHz, irregularities with vertical scales near 30m may be preferentially selected.

A recent experiment conducted by LINDNER (1973) is of interest because it is related to the present work but involved different techniques. A comparison of his conclusions with the present results strengthens the case for a correspondence between ground and ionospheric scale sizes. The work, also carried out at Buckland Park, used direction-of-arrival information to deduce the angular spread of D-region echoes. His conclusions with regard to height, diurnal and seasonal trends were briefly:

(i) 'a significant feature of all of the graphs is the fairly steady increase of angular spread with increasing height up to about 90 km, with a sharp increase being evident in several profiles at about 85 km. The few results taken above 90 km show that a decrease of angular spread is evident.'

(ii) 'although there is a larger scatter of values, at night the angular spread is often significantly smaller than the average daytime values.' Echoes were only monitored for one night.

(iii) 'no systematic seasonal trends are found below 85 km but above this there is evidence of a small systematic increase of angular spread until winter, when a sudden increase occurs in June.' No results were taken in summer.

In all three cases, the trends are entirely compatible with FIGURES 8.1 to 8.3 if it is accepted that the smaller irregularity should produce a larger angular spread. Given the uniform vertical scale, this should be true for most irregularity models.

A possible interpretation of the variations with height and season is in terms of wave activity. FRASER and VINCENT (1970) cited the work of MULLER (1968b) to explain the seasonal variability, the inference being that larger irregularities would be less stable in the presence of waves. MULLER found that the amplitude of gravity waves, as measured from meteor trail evidence of their shear strength, was greatest at the solstices and least at the equinoxes. However ELFORD and ROPER (1967) have reported the apparently conflicting finding that the seasonal variation of the turbulent dissipation rate had maxima at the equinoxes. It is also difficult to see how waves can be invoked to explain the diurnal variation of the pattern size in the 80 - 85 km range.

8.2 Pattern Elongation

The elongation of the ground pattern is defined by the ratio a/b, with a and b again the major and minor axes of the correlation ellipse. The elongation is found to increase with height from 75 km to 100 km in a fairly linear fashion, as demonstrated in FIGURE 8.4. The crosses are the 1972 averages for six height intervals. A similar trend was observed by CLEMESHA (1963) also with a spaced receiver experiment but in Nigeria, close to the Equator.

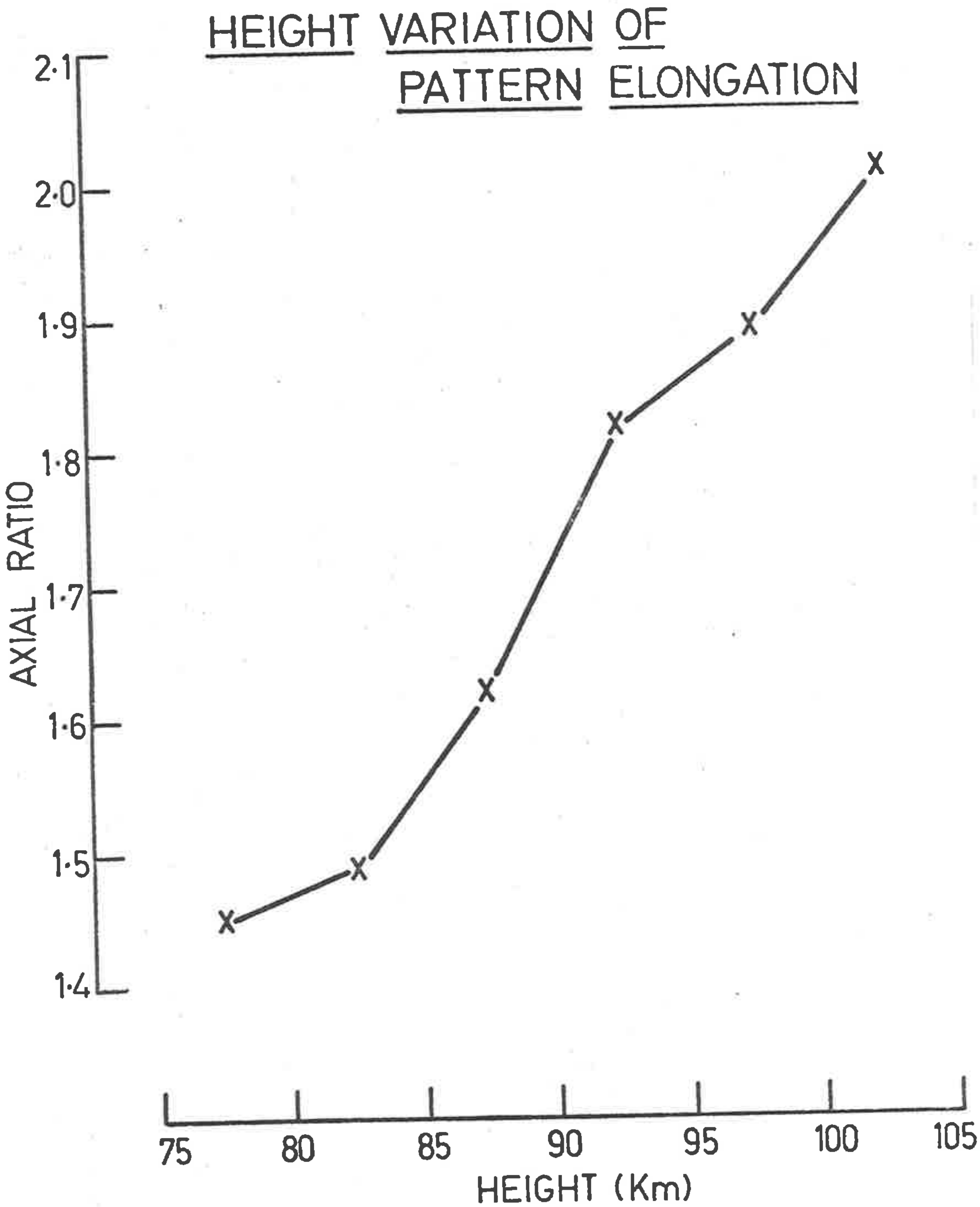

HEIGHT VARIATION OF
PATTERN ELONGATION
AXIAL RATIO
2·1
2·0
1·9
1·8
1·7
1·6
1·5
1·4
75
80
85
90
95
100
105
HEIGHT (Km)

FIGURE 8.4

He found that the axial ratio increased from less than 2 at 70 km to 8 at 110 km. He explained this in terms of anisotropic diffusion along the magnetic field lines. The variation in FIGURE 8.4 is only from 1.4 to 2.0 in 25 km but the influence of the earth's magnetic field would be a much smaller effect at 35^0 S. The presence of magnetic control is also suggested by the histograms of FIGURE 8.5. Here the angle between the major axis of the correlation ellipse and the direction of magnetic North has been used as a measure of pattern alignment. The values at each height have been grouped into twelve intervals of width 15^0. Thus the two groups least aligned with the magnetic field are from 75^0 to 90^0 and 90^0 to 105^0. The numbers involved in the averaging were least in the top two and the bottom histogram and hence the less smooth histograms. Nevertheless there is a clear change in character over the height range. At 100 - 105 km the distribution is virtually symmetrical about 0^0 with a ratio of about 2 : 1 from 0^0 to 90^0. This tendency for field alignment changes to a much flatter distribution by 85 - 90 km and at 75 - 80 km a new symmetry about 55^0 is evident. The reason for this is not apparent although it may be related to the mean winds.

The axial ratio fluctuates during the day but shows a

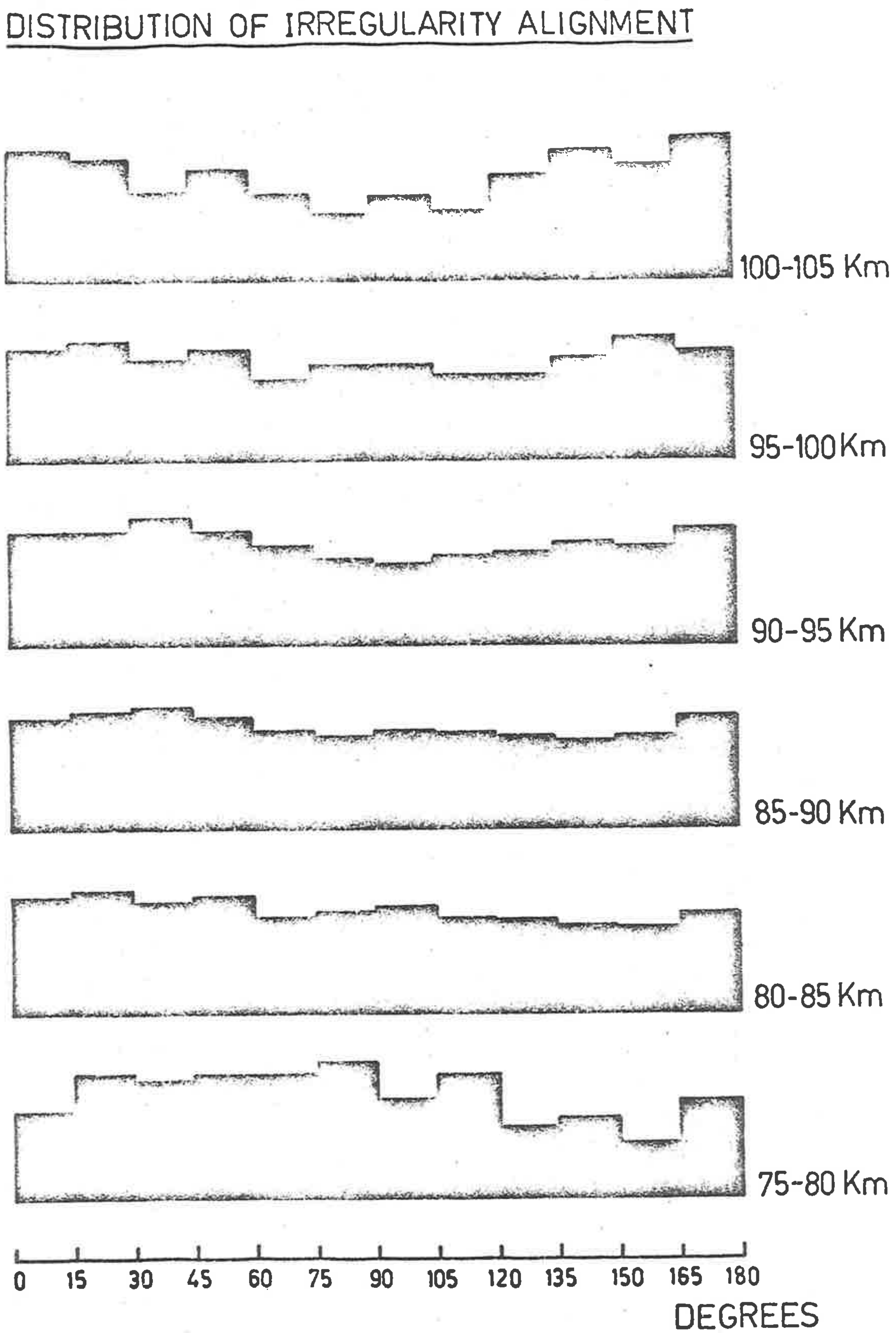


FIGURE 8.5 *The orientation of the diffraction pattern with respect to Magnetic North.*

dominant 24-hour variation over most of the height range. This diurnal variation is plotted in FIGURE 8.6. In all cases there is a minimum value for the axial ratio near noon and a broad maximum around midnight. No evidence was found for any seasonal variation in the elongation.

8.3 Characteristic Velocity

The characteristic velocity V_c, which is used in the calculation of the FCA 'true' velocity V_t, has been subjected to several interpretations. It has sometimes been regarded as a measure of the magnitude of the random changes occurring within the moving diffraction pattern, in keeping with the original reason for its introduction. FOOKS (1965) identified V_c with the speed of movement of wavelike components making up the fading pattern. RASTOGI et al, (1970) regarded the ratio V_c/V_t as a measure of the irregularity lifetime. The ratio might also be considered as a measure of the importance of internal changes as compared with the overall motion. Computer modelling by PITTEWAY, WRIGHT and FEDOR (1971) has tended to justify the original view of V_c by showing that it is directly related to random changes in the model, at least in the case of scattering from small irregularities which have mean and random components of motion.

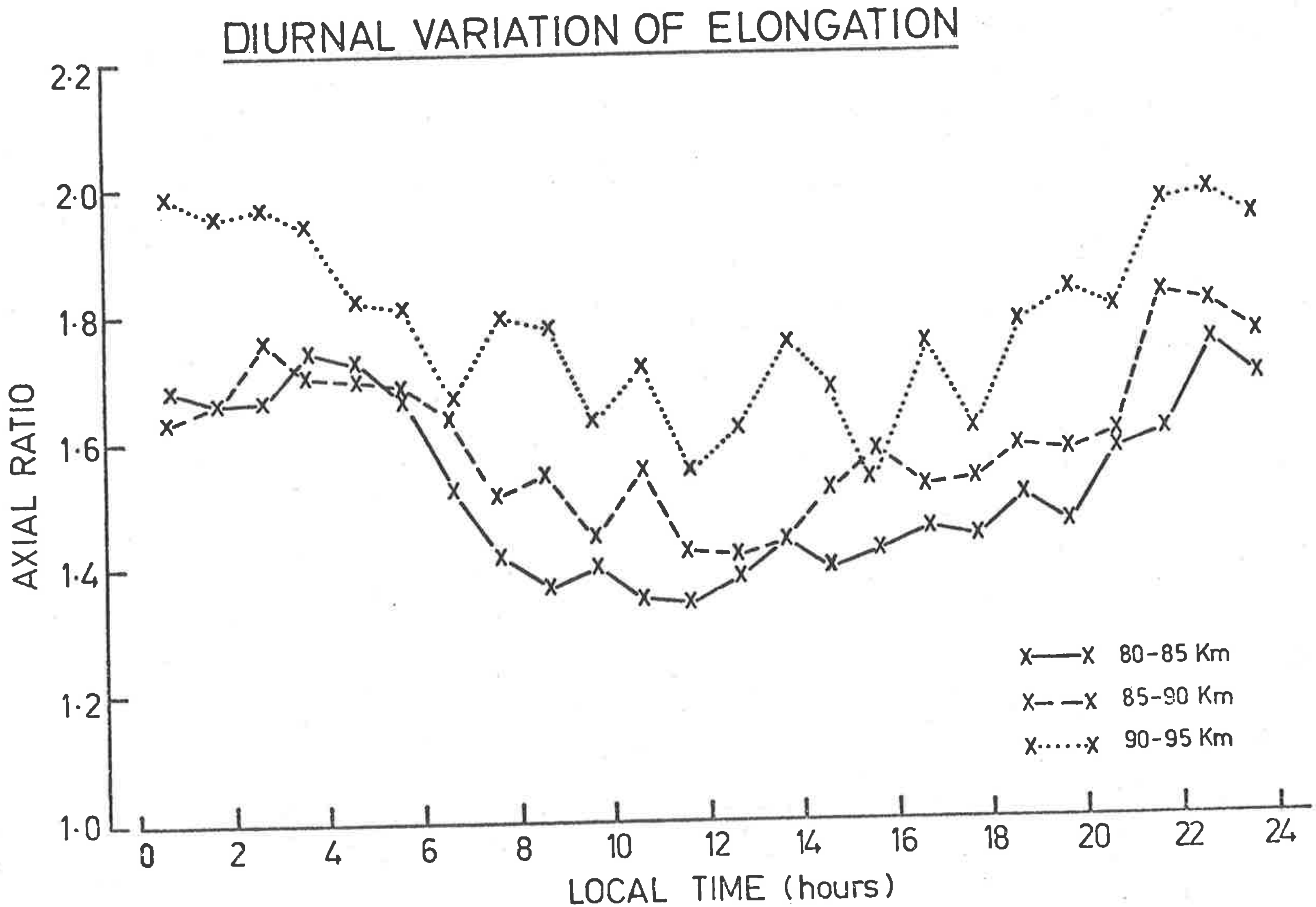

DIURNAL VARIATION OF ELONGATION
AXIAL RATIO
2·2
2·0
1·8
1·6
1·4
1·2
1·0
0
2
4
6
8
10
12
14
16
18
20
22
24
LOCAL TIME (hours)
80-85 Km
85-90 Km
90-95 Km

FIGURE 8.6

As with the other parameters discussed in this chapter, an attempt was made to trace systematic variations in the value of V_c. The results are shown in FIGURES 8.7 and 8.8. The errors associated with the individual points range from 1 to 4 m/sec so much of the fine structure, particularly at 85 - 90 km, is quite real. There are no very clear features of the diurnal variations however, except perhaps for a morning peak at 90 - 95 km and a daytime minimum at 80 - 85 km. The monthly averages of FIGURE 8.8 exhibit such differing behaviour at the three height ranges that it would be desirable to see a sequence of several years' results in order to determine which features were consistently repeated. At 80 - 85 km the maxima are centred on May and November which correspond to the minima for 90 - 95 km. In fact the behaviour at 80 - 85 km suggests a connection with the pattern size variation. The two are virtually in antiphase. Thus, a large V_c is indicative of considerable gravity wave and turbulent activity which in turn results in a reduction of the average pattern size. However this simple interpretation obviously breaks down for the other heights. It is difficult to compare the results with those of LEE and BOWHILL (1963) and SPRENGER and SCHMINDER (1967). They both found seasonal variations in V_c but were using total reflections. The large

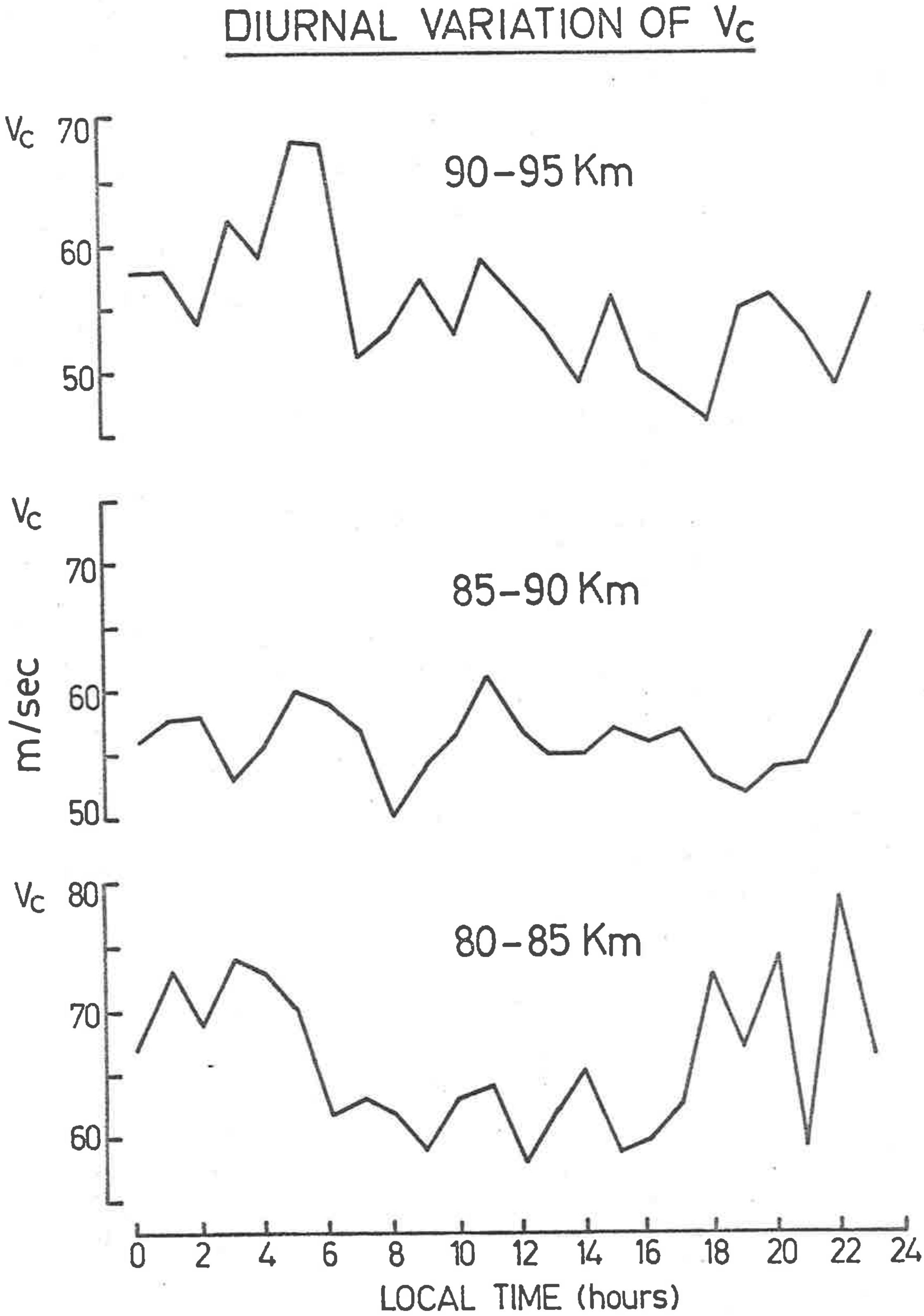
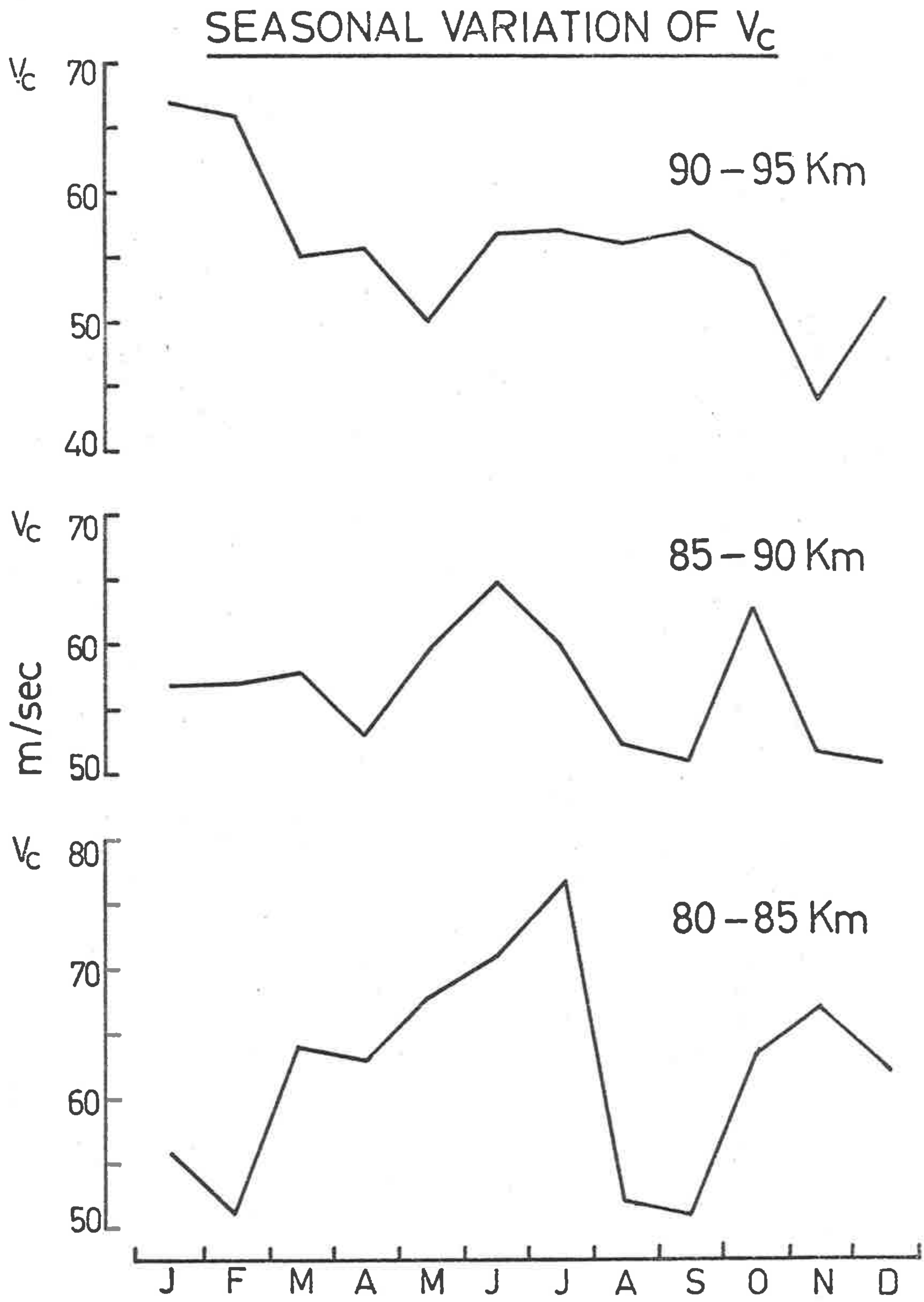

DIURNAL VARIATION OF Vc
Vc
70
60
50
90-95 Km
Vc
70
60
50
m/sec
85-90 Km
Vc
80
70
60
80-85 Km
0 2 4 6 8 10 12 14 16 18 20 22 24
LOCAL TIME (hours)

FIGURE 8.7



SEASONAL VARIATION OF V_c
V_c
70
60
50
40
90 – 95 Km
V_c
70
60
50
m/sec
85 – 90 Km
V_c
80
70
60
50
80 – 85 Km
J F M A M J J A S O N D

FIGURE 8.8

values which have appeared in late summer at 90 - 95 km in FIGURE 8.8 are a feature of their summer results.

It might be noted in concluding this chapter that BRIGGS and VINCENT (1973) have shown that the backscatter distribution from isotropic irregularities in fact depends only on transmission and reception parameters and contains no information on the nature of the irregularities themselves. However it does appear from the present work that the irregularities are anisotropic in both plan and elevation, and the existence of large variations is sufficient indication that the results cannot be purely instrumental.

CHAPTER NINE

CONCLUSIONS

Partially reflected radio waves have been used to study the drift of ionospheric irregularities in the height range 60 - 105 km. The returned radio amplitude has been sampled at three spaced receiving points and analysed using the technique of 'full correlation analysis'. Two experimental sites have been used for the observations which primarily cover the years 1972 and 1973. In this chapter the main aspects of the work are summarised and several suggestions made for the fruitful continuation of the study.

9.1 Meteor Comparison

A year's results from the Adelaide drifts experiment was compared with corresponding measurements made using reflections from meteor trails. The two observing systems involved a similar height coverage and temporal averaging but the meteor system was sensitive over a much greater horizontal extent. The basic agreement between the two sets of results was found to be good. There were however times of significant difference. Some of these could have been introduced, particularly at times of low data rate, by the various forms of averaging used to facilitate comparison. It was felt that in many cases though, the discrepancies could be explained in terms of small-scale wave activity within the wide horizontal coverage of the meteor system. The greater variability of the meteor results and an apparently seasonal variation in the quality of agreement were suggestive of this. There was also evidence that the drifts were systematically too small but this minor effect has an instrumental explanation.

Overall, it was difficult not to conclude that both systems were in fact measuring the neutral air wind. There had been two particularly strong objections to the drift technique - that it was unable to detect the neutral air tides, and that the derived wind vector was systematically displaced due to the influence of directional internal gravity waves, Eastward in the case of Adelaide. In fact, with continuous data over the full 24 hours, the tides emerged as a major component of the drifts, and at no stage were the drifts consistently more Eastward than the meteor results.

9.2 Analysis of the Drifts

The drifts were analysed in terms of mean, periodic and irregular contributions.

9.2.1 Prevailing Component

The prevailing component of the drifts was found to exhibit many of the characteristics associated with neutral air winds. The zonal component was generally much larger than the meridional component with the predominant flow being Eastward. In surveying the annual variation of the mean zonal drifts, a change in character with height was apparent. Below 95 km the mesopheric circulation pattern, strongly Eastward in winter and weakly Westward in summer, was followed. Above 95 km there was evidence of the winter Westward flow associated with the lower thermosphere.

One notable feature of the mean drifts was the way in which they fluctuated from day to day. In some instances this variation was distinctly wave-like and the periods suggested the presence of planetary waves.

9.2.2 Tidal Components

Fluctuations with periods of 24 and 12 hours were found to be a major component of the drifts and the sense of rotation of the wind vector almost always anticlockwise. The zonal and meridional components were analysed independently but the tidal amplitudes were usually similar indicating a basically circular motion of the wind vector. Two outstanding features of the results were the relative dominance of the diurnal component and the variability of the amplitudes. On the average however there were indications of a semiannual variation in the tidal scalar amplitudes with maxima at the equinoxes, larger for the 24-hour component in the autumn, and for the 12-hour component in the spring.

The 24- and 12-hour components exhibited somewhat differing phase properties. In the case of the 24-hour tide, there was generally a clear 6-hour phase shift between E/W and N/S components. Except for the winter months, the phase showed remarkable stability throughout the year with the time of Eastward maximum being close to local noon. On the other hand, the phase of the 12-hour component varied considerably from month to month and even within a month. Except for the winter months, the phase lag of the Northward component was close to 3 hours. The 8-hour component was usually too small in amplitude to determine the phase reliably. In the few cases where detailed height information was available, the diurnal tide appeared to have a wavelength of some 35 km. The lack of amplitude growth with height implied a considerable energy input at these heights from tidal sources.

The results were compared to current theory of atmospheric tides and found to agree particularly well except in the case of the phase of the semidiurnal tide. The lack of detailed information in height and latitude prevented the isolation of particular tidal modes. A significant feature of the tidal momentum transport was the poleward transport by the diurnal tide during the late spring and early summer. No connection was found between the tidal amplitudes at 88 km and the amplitude of the Sq current fluctuations.

9.2.3 Irregular Components

When the mean and tidal components had been removed from a sequence of drift results, significant residual values remained. Although unpredictable in magnitude and period, these residual fluctuations were often of a form implying the presence of internal gravity waves. The way in which the drifts changed in a fixed short period over a range of heights was examined, and here also the changes were often wave-like with apparent vertical wavelengths in the vicinity of 8-20 km. At 86 km the magnitude of the residual fluctuations was about 15 m/sec in both the E/W and N/S components which made the contributions of the irregular component to the drifts comparable with those of the mean and tidal components. It was however shown that a feature of the short-term changes was an exponential amplitude growth with height. This implied that their contribution to the energy input near 90 km was less than that of the tides and that in the region between 90 and 100 km the irregular components may dominate the dynamics. The presence of these large short-term features does not preclude the isolation of the mean and tidal components, providing drift values are determined frequently enough. The potential of the drift technique for the observation of very rapid changes was illustrated.

9.3 The Woomera Experiment

A new experimental site at Woomera for the determination of D-region drifts was described. The site was chosen to be under the volume usually probed by rocket wind-measuring techniques. It was thus suitable for essentially common-volume comparisons with the drifts. Its horizontal displacement from Adelaide made possible a study of the horizontal scale of the drifts.

9.3.1 Rocket Comparisons

Three types of rocket experiment were discussed - dropsondes, falling spheres and vapour trails. In the case of the most common firings - dropsondes and twilight spheres - there was unfortunately no height overlap with the drift technique. In view of this the agreement of the results presented was as convincing as could be expected, even when drift values were only available from Adelaide before the Woomera equipment was operational. One daytime falling sphere experiment was conducted making possible a 20 km height interval in common with the drifts. The concurrence of the two sets of values was excellent. The quality of agreement for the varpour trail experiments was more varied although with one exception the differences could be accounted for in terms of site separation and increased small-scale wave amplitudes upwards of 80 km.

9.3.2 Wind Systems Scale

Preliminary results were presented of the degree of correlation between the D-region drifts at Adelaide and Woomera. Some examples were given of very close agreement even in successive individual values. At other times there was evidence of marked divergence in the results. This was both random and systematic. In the latter case, the difference could have been due to latitudinal changes in the mean wind or tidal amplitude. The main general feature of the comparisons was a decrease in correlation with increasing height. This would be explained if the regular components were quite local and independent and increasing in amplitude with height. No 24-hour segments of drift values were available for the Woomera site to test whether in fact the prevailing and periodic amplitudes were closely related to the corresponding Adelaide values.

9.4 Ground Diffraction Pattern

In the process of calculating the drifts, other parameters related to the ground diffraction pattern were evaluated. The average behaviour of these parameters was particularly interesting in view of a possible direct relationship between the ground pattern and the ionospheric irregularities. The pattern size was found to decrease from 80 km to a minimum at about 92 km and then increase slowly up to 100 km. The size had a particularly strong diurnal variation at 80 - 85 km. A seasonal variation was evident at all heights, almost purely semiannual between 85 and 95 km with maxima at the equinoxes. The pattern elongation was found to increase linearly with height and this, together with the alignment distribution, suggested a degree of magnetic control over the irregularities that increased with increasing height.

9.5 Future Work

Throughout the work described in this thesis two problems related to the receiver gains were encountered.

Firstly, because no form of automatic gain control was employed, the gains had to be personally monitored and adjusted, particularly for several hours around sunrise and sunset. For routine observations, some automatic gain control is desirable. In order not to affect the amplitude fading, the time constant should probably be of the order of five minutes. The disadvantage then is the necessity to leave the transmitter on much longer than the actual recording time, whereas this has been avoided in the past to reduce interference to some important local operations near 2 MHz. Some compromise will probably be necessary. The second problem is more fundamental - the limited dynamic range of the receivers due to the use of linear gain. This is why it has been necessary to concentrate primarily on one height interval. There is a great range of echo strengths between 60 and 100 km and often the gains can only be correctly set for optimum fading from a small part of this interval. Where more detailed height information has been obtained by soundings with several different gain settings, it has been at the cost of the continuity at any one height. The best answer theoretically, seems to lie in the use of logarithmic receivers. Computer calculations on typical records have shown that FCA is relatively insensitive to non-linear transformations of the data. Practically, however, it may prove simpler to switch the present receivers through a preset series of stepped gains or to add extra banks of linear receivers, each operating over part of the height range.

There is little doubt that the drifts can provide a wealth of meteorological information at D-region heights. On the basis of the present results, some financial support has already been obtained for the routine continuation of the observations. If the above instrumental improvements can be effected, this routine information will be obtainable for all of the D-region. It should then be ideal for the study of the propagation of tides and internal gravity waves in the region. Ths cost of computing would be the limiting factor in deciding the frequency of drift determinations.

A further improvement suggested by the work of the Canadian group under Gregory is the simultaneous recording of echo structure information since it is possible sometimes to arrive at a distorted picture by assuming the equivalence of echo range and height. DAE equipment has been set up at Buckland Park and it could be incorporated into the drift work, especially for studies of vertical structure. Direction-of-arrival information at least could be recorded.

The study of coupling with the lower atmosphere would be of interest. Occasions when planetary wave activity is evident in the drifts provide an opportunity to compare the fluctuations with those of parameters in the stratosphere. It has been suggested (QUIROZ, private communication) that stratospheric warmings sometimes pass close to Adelaide in winter and that a close examination of the drift results at these times could prove fruitful.

It was pointed out that 2 MHz transmissions may preferentially select irregularities of a particular vertical size. It would be interesting to see whether pattern sizes deduced from 6 MHz partial reflections exhibit the same variations as those described in Chapter 8. The use of 6 MHz would also extend the heights that could be studied with partial reflections.

Rocket comparisons of two sorts should be continued. Daylight sphere and vapour trail experiments are still needed to clarify the degree of instantaneous correspondence between drifts and neutral winds, particularly above 90 km (vapour trails). There does not seem to be any further need to use other than Woomera drifts in future comparisons or to attempt further comparisons with dropsondes or twilight falling sphere firings.

The continuation of the Adelaide/Woomera drift comparison is important. The power of the Woomera transmitter is being increased, and once continuous records are being taken at the two sites, it will be possible to correlate individual features of the drifts such as the phase of a particular tidal component. Latitudinal studies of tidal modes are of considerable importance because of their influence on the momentum and energy budgets of the upper atmosphere. With this in mind a third observing station has been proposed - at Darwin (12° S) where tidal theory predicts large amplitudes.

In the author's opinion, the work described in this thesis should considerably enhance the status of the partial reflection drift technique as a method of studying the complicated dynamics of the ionospheric D-region. Its wider and fuller use is strongly recommended.

BIBLIOGRAPHY

ALLEYNE. H. ; KEENLISIDE. W. ; KENT. G.S. ; Mac DOUGALL. J.W. and SCHOLFIELD. A.J.	1974	JATP 36 171
AWE. O.	1961	JATP 21 142
AWE. O.	1964a	JATP 26 1239
AWE. O.	1964b	JATP 26 1257
BARBER. N.	1957	JATP 11 299
BARTMAN. F.L. ; CHANEY. L.W.; JONES. L.M. and LIU. V.C.	1956	J. App. Phys. 27 706.
BATTEN. E.S.	1961	J. Meteorol. 18 283
BAUER. S.J.	1958	Pure & App. Geophys. 40 235.
BELROSE. J.S. and BURKE. M.J.	1964	JGR 69 2799
BELROSE. J.S. and CETINER. E.	1962	Nature 195, 688
BEYNON. W.J.G. and GOODWIN G.L.	1967	JATP 29 1181
BEYNON. W.J.G. and JONES E.S.O.	1965	Proc. Roy. Soc. A288 558
BEYNON. W.J.G. and WRIGHT. J.C.	1969	JATP 31 119

BLAMONT. J.	1963	Plan. Space Sci. <u>10</u> 89
BLAMONT. J. and de JAGER. C.	1961	Ann. Geophys. <u>17</u> 134
BLAMONT. J.E. and TEITELBAUM. H.	1968	Ann. Geophys. <u>24</u> 387
BRETHERTON. F.P.	1966	Quart. J. Roy. Met. Soc. <u>92</u> 466
BRIGGS. B.H.	1967	Instruction Manual for ionospheric drift measurements' - ADP 42 Dept of Phys, U of A.
BRIGGS BRIGGS. B.H.	1972	Geofysiske Publikasjoner <u>29</u> 121.
BRIGGS. B.H.; ELFORD. W.G. ; FELGATE. D.G.; GOLLEY. M.G. ROSSITER. D.E. and SMITH. J.W.	1969	Nature <u>223</u> 1321.
BRIGGS. B.H. ; PHILLIPS. G.J. and SHINN. D.H.	1950	Proc. Phys. Soc. <u>B63</u> 106
BRIGGS. B.H. and SPENCER. M.	1954	Rep. Prog. Phys. <u>17</u> 245
BRIGGS. B.H. and VINCENT. R.A.	1973	Aust. J. Phys. <u>26</u> 805.

BROWN. G.M. and CHAPMAN. J.W.	1972	JATP 34 1445
BROWN. N.	1973	PhD thesis, University of Adelaide.
BULL. G.V.	1964	Can. Aeronautics and Space J. 10 238
BUTLER. S.T. and SMALL. K.A.	1963	Proc. Roy. Soc. A274 91
CHAPMAN. S.C.	1924	Quart. J. Roy. Met. Soc. 50 165
CHAPMAN. S.C. and LINDZEN. R.S.	1970	'Atmospheric Tides' D. Reidel Publ. Comp. Holland.
CHAPPELL. D.W.G. and HENDERSON C.L.	1956	JATP 8 163
CHARNEY. J.G. and DRAZIN. P.G.	1961	JGR 66 83
CHIMONAS. G. and HINES. C.O.	1970	JGR 75 875
CLEMESHA. B.R.	1963	JGR 68 2363
COLGROVE. F.D. ; JOHNSON. F.S. and HANSON. W.B.	1966	JGR 71 2227
CONOVER. W.C. and WENTZIEN. C.J.	1955	J. Met. 12 160
DAVIS. T.N.	1960	JGR 65 3497

DIEMINGER. W.	1952	JATP <u>2</u> 340
ELFORD. W.G.	1959	Plan. Space Sci. <u>1</u> 94
ELFORD. W.G.	1968	AFCRL Special Report 75, P 11
ELFORD W.G. and McAVANEY. B.J.	1971	'Five year survey of mean prevailing winds near 90 km'. Presented at 15th IUGG General Assembly, Moscow.
ELFORD. W.G. and ROPER. R.G.	1967	Space Research <u>7</u> 42
ELLYETT. C. and WATTS. J.M.	1959	J. Res. Nat Bureau Standards <u>63D</u> 117
EVANS J.V.	1972	JATP <u>34</u> 175
FEDOR. L.S.	1967	JGR <u>72</u> 5401
FEDOR. L.S. and PLYWASKI. W.	1972	JATP <u>34</u> 1285
FEJER. J.A. and VICE. R.W.	1959	JATP <u>16</u> 291
FELGATE. D.G.	1970	JATP <u>32</u> 241
FELGATE. D.G. and GOLLEY. M.G.	1971a	AJP <u>24</u> 1397
FELGATE. D.G. and GOLLEY. M.G.	1971b	JATP <u>33</u> 1353

FELGATE. D.G.; HUNTER. A.N.; KINGSLEY. S.P. and MULLER H.G.	1974	'Comparative Studies of E-region ionospheric drifts and meteor winds' Plan. Sp. Sci.
FOGLE. B. and HAURWITZ. B.	1966	Space Sci. Reviews <u>6</u> 279
FOOKS. G.F.	1965	JATP <u>27</u> 979
FRASER. G.J.	1968	JATP <u>30</u> 707
FRASER. G.J. and KOCHANSKI. A.	1970	Annals de Geophys <u>26</u> 675
FRASER. G.J. and VINCENT. R.A.	1970	JATP <u>32</u> 1591
GARDNER. F.F. and PAWSEY. J.L.	1953	JATP <u>3</u> 321
GOLLEY. M.G. and ROSSITER. D.E.	1970	JATP <u>32</u> 1215
GOLLEY. M.G. and ROSSITER. D.E.	1971	JATP <u>33</u> 701
GOSSARD. E.	1962	JGR <u>67</u> 745
GOSSARD. E. and PAULSEN. M.R.	1968	JATP <u>30</u> 885
GREEN. J.S.A.	1965	Proc. Roy. Soc. <u>A288</u> 564
GREENHOW. J.S.	1950	Phil. Mag. <u>41</u> 682
GREENHOW. J.S.	1952	JATP <u>2</u> 282

GREENHOW. J.S.	1959	JGR 64 2129
GREENHOW. J.S. and NEUFELD. E.L.	1955	Phil. Mag. 46 549
GREENHOW. J.S. and NEUFELD. E.L.	1961	Quart. J. Roy. Met. Soc. 87 472
GREGORY. J.B.	1956	Aust. J. Phys. 9 324
GREGORY. J.B.	1961	JGR 66 429
GREGORY. J.B.	1965	J. Atmos. Sci. 22 18
GREGORY. J.B. and MANSON. A.H.	1967	JGR 72 1073
GREGORY. J.B. and MANSON. A.H.	1969a	JATP 31 683
GREGORY. J.B. and MANSON. A.H.	1969b	JATP 31 703
GREGORY. J.B. and REES. D.T.	1971	J. Atmos. Sci. 28 1079
GREGORY. J.B. and VINCENT. R.A.	1970	JGR 75 6387
GROVES. G.V.	1959	JATP 16 344
GROVES. G.V.	1969	J. Br. Interplan Soc. 22 285
GROVES. G.V.	1972	Plan. Space Sci. 20 2099

GUHA. D. and GELLER. M.A.	1973	JATP 35 1853
GUSEV. V.D. and MIRKOTAN. S.F.	1956	Raditekhnika i elektronita 1 743
HARNISCHMACHER. E.	1968	'Winds and Turbulence in the Stratosphere, Mesophere, Ionosphere' p 227; N. Holland Pub. Co.
HAUBERT. A.	1972	JATP 32 1337
HAUBERT. A. and DOYEN. G.	1966	Ann. Geophys. 22 338
HAUG. A. and HOLT. O.	1968	'Drift measurements by use of D-region partial reflections The Auroral Observatory, Tromso, Norway.
HINES. C.O.	1960	Can. J. Phys. 38 1441
HINES. C.O.	1963	Quart. J. Roy. Met. Soc. 89 1
HINES. C.O.	1968	JATP 30 837
HINES. C.O.	1972	Phil. Trans. A271 457
HINES. C.O. and RAO. R.R.	1968	JATP 30 979

HINES. C.O. and REDDY. C.A.	1967	JGR 72 3
HODGES. J.R.	1967	JGR 72 3455
HOOK J.L.	1970	Plan. Space Sci. 18 1623
IYENGAR. R.S.	1970	JATP 32 955
JONES. D. and MAUDE. A.D.	1965	Nature 206 177
JONES. I.L.	1958	JATP 12 68
JUSTUS. C.G. and WOODRUM. A.	1973	J. Atmos. Sci. 30 1267
KAZIMIROVSKY. E.S. and KOKOUROV. V.D.	1972	Phil. Trans. Roy. Soc. Lond. A271 499
KELLEHER. R.F.	1966	JATP 28 213
KELLOGG. W.W. and SHILLING. G.F.	1957	J. Meteorology, 8 222
KENT. G.S. and WRIGHT. R.W.H.	1968	JATP 30 657
KIM. T.S. and CURRIE. B.W.	1958	Can. J. Phys. 36 160
KOCHANSKI. A.	1963	JGR 68 213

KOCHANSKI. A.	1964	JGR 69 3651
KOKOUROV. V.D.; KAZIMIROVSKY. E.S. ZAKHAROV. V.N. and ZHOVTYY. YE. I	1971	Geomag. and Aeronomy 11 612
KRAUTKRAMER. J.	1943	Deutsche Luftfahrt forschung F.B. Nr 1761
KRAUTKRAMER. J.	1950	Archiv Elekt Ubertragung 4 133
LEE. H.S. and BOWHILL. S.A.	1963	JATP 25 659
LILLER. W. and WHIPPLE. F.L.	1954	JATP Supp 1, 112
LINDNER. B.	1973	Ph.D thesis, University of Adelaide.
LINDZEN. R.S.	1967	Mon Weather Rev. 94 295
LINDZEN. R.S.	1967	Quart. J. Roy. Met Soc. 93 18
LINDZEN. R.S.	1968	Proc. Roy. Soc. A303 299
LINDZEN. R.S.	1969	JATP 31 449
LYSENKO. I.A. ; PORTNYAGIN. YU.I ; SPRENGER. K. ; GREISIGER. K.M. and SCHMINDER. R.	1972	JATP 34 1435

MACDOUGALL. J.W.	1966	JATP 28 1093
McGEE. C.R.	1966	JATP 28 861
MANNING. L.A. ; VILLARD. O.G. and PETERSON. A.M.	1950	Proc. Inst. Rad. Engrs. 38 877
MANRING. E. ; BEDINGER. J.F. ; PETTIT. H.B. and MOORE. C.B.	1959	JGR 64 587
MANRING. E. ; BEDINGER. J. and KNAFLICH. H.	1962	JGR 67 3923
MANSON. A.H.	1968	JATP 30 677
MANSON. A.H. ; GREGORY. J.B. and STEPHENSON. D.G.	1973	JATP 35 2055
MATSUSHITA. S.	1967	Handbuck der. Physik 49 1
MIRKOTAN. S.F.	1958	Raditekhnika i elecktronika 3 570
MITRA. S.N.	1949	Proc. Inst. Elect. Eng. Part III 96 441
MULLER. H.G.	1966	Plan. Space Sci. 14 1253
MULLER. H.G.	1968a	JATP 30 701

MULLER. H.G.	1968b	Plan. Space Sci. <u>16</u> 61
MULLER H.G.	1972	Phil. Trans. Roy Soc. Lond. <u>A271</u> 585
MULLER H.G. and KINGSLEY S.P.	1974	'On the scale sizes of wind systems in the meteor zone' (PSS preprint)
MURGATROYD. R.J.	1957	Quart. J. Roy. Met. Soc. <u>83</u> 417
MURPHY. C.H. ; BULL. C.V. and EDWARDS. H.D.	1966	JGR <u>71</u> 4535
NEWELL. R.E. and DICKINSON. R.E.	1967	Pure Appl. Geophys <u>68</u> 162
NORDBERG. W. ; KATCHEN. L. ; THEON. J. and SMITH. W.S.	1965	J. Atmos. Sci. <u>22</u> 611
NORDBERG. W. and SMITH. W.	1964	NASA T.N. D-2107 March.
OLIVIER. C.P.	1942	Proc. Amer. Phys. Soc. <u>85</u> 93
OSBORNE . D.G.	1966	JATP <u>28</u> 45
PANT. P.S.	1956	JGR <u>61</u> 459

PANT. P.S.	1956	JGR <u>61</u> 459
PAWSEY. T.L.	1935	Proc. Camb. Phil. Soc. <u>31</u> Part I, 125
PHILLIPS. G.J.	1951	Proc. Inst. Elect Engrs. Part III <u>98</u> 237
PHILLIPS. G.J.	1952	JATP <u>2</u> 141
PHILLIPS. G.J. and SPENCER. M.	1955	Proc. Phys. Soc. <u>B68</u> 481
PIERCE. A.D. and CORONITI. S.C.	1966	Nature <u>210</u> 1209
PIGGOTT. W.R. and THRANE. E.V.	1966	JATP <u>28</u> 311
PITTEWAY. M.L.V. ; WRIGHT. J.W. and FEDOR. L.S.	1971	JATP <u>33</u> 635
POKROVSKIY. G.B. and TEPTIN. G.M.	1972	Geomag. and Aeronomy <u>12</u> 124
RAO. P.B. and RAO. B.R.	1963	'Proc. Intern. Conf. on the Ionosphere' (ed Strickland) pp 363-369 Inst. of Phys and Phys Soc. London.
RASTOGI. R.G. ; CHANDRA. H. and DESHPANDE. M.R.	1970	JATP <u>32</u> 999
RATCLIFFE. J.A.	1959	JGR <u>64</u> 2102

RATCLIFFE. J.A. and PAWSEY. J.L.	1933	Proc. Camb. Phil. Soc. 29 Part II, 301
REES. D.	1969	J. Br. Interplan Soc. 22 275
REVAH. I.	1970	JATP 32 1313
ROBERTSON. D.S. ; LIDDY. D.T. and ELFORD. W.G.	1953	JATP 4 255
ROPER. R.G.	1965	NASA Report X-651-65-102
ROSENBERG. N.W. ; GOLOMB. D. and ALLEN. E.F.	1963a	JGR 68 3328
ROSENBERG. N.W. ; GOLOMB. D. and ALLEN. E.F.	1963b	JGR 68 5895
ROSSITER. D.E.	1970a	Aust. J. Phys. 23 103
ROSSITER. D.E.	1970b	Nature, 228 1299
SALES. G.S. and BOWHILL. S.A.	1962	JATP 24 451
SASTRI. J.H. and RAO. B.R.	1971	JATP 33 507
SIEBERT. M.	1961	Adv. Geophys. 7 105
SPIZZICHINO. A. ; DELCOURT. J. ; GIRAUD. A. and REVAH. I.	1965	Proc. Inst. Elect. Engr. 53 1084

SPRENGER. K. and SCHMINDER. R. 1967 JATP 29 183

SPRENGER. K. and SCHMINDER. R. 1968 JATP 30 693

SPRENGER. K. and SCHMINDER. R. 1969 JATP 31 1085

STENING R.J. 1970 Plan. Space Sci. 18 121

STROUD. W.G. ; NORDBERG. W. and WALSH. J.R. 1956 JGR 61 45

STUBBS. T.J. 1973 JATP 35 909

STUBBS. T.J. and VINCENT. R.A. 1973 Aust. J. Phys. 26 645

TEPTIN. G.M. 1971 Izv. Atmos. Ocean Phys. 7 555

TESTUD. J. 1973 Space Res. 13 393

TITHERIDGE. J.E. 1962a JATP 24 269

TITHERIDGE. J.E. 1962b JATP 24 283

VINCENT. R.A. 1969 JATP 31 607

VINCENT. R.A. 1972 JATP 34 1881

VINCENT. R.A. 1973 Aust. J. Phys. 26 815

WITT. G. 1962 Tellus 4 1

WRIGHT. J.W.	1968	JATP 30 919
WRIGHT. J.W.	1972	JATP 34 1365
WRIGHT. J.W. and FEDOR. L.S.	1967	Space Res. 7 67
WRIGHT. J.W. and FEDOR. L.S.	1969	JATP 31 925
YERG. D.G.	1955	JGR 60 173
YUEN. P.C. ; WEAVER. P.F. ; SUZUKI. R.K. and FURUMOTO. A.S.	1969	JGR 74 2256

APPENDIX I

Reprint of a paper:

"The measurement of winds in the D-region of the ionosphere by the use of partially reflected radio waves"

T.J. Stubbs.

Journal of Atmospheric and Terrestrial Physics, 1973, Vol. 35, pp. 909–919. Pergamon Press. Printed in Northern Ireland

The measurement of winds in the *D*-region of the ionosphere by the use of partially reflected radio waves

T. J. Stubbs
Department of Physics, University of Adelaide, Australia

(*Received* 7 *November* 1972)

Abstract—The fading of radio waves reflected from the upper *D*-region of the ionosphere has been studied using the spaced receiver technique and full correlation analysis. The resulting ionospheric drift velocities are shown to provide reliable estimates of the motion of the neutral air in the reflection region. Strong tidal components are clearly evident, the diurnal tide being predominant. The drifts agree well in both magnitude and direction with radio measurements of drifts of meteor trails made at the same location and times.

1. Introduction

Most of the observations of ionospheric drifts have in the past been made using totally reflected radio waves. The height of reflection is then determined by the frequency used and the prevailing ionospheric conditions and is likely to vary unpredictably. Recently several workers have shown that use can also be made of weak partial reflections from the *D*-region (Awe, 1961; Fraser, 1965; Fraser and Kochanski, 1970; Rossiter, 1970). These result from the weak back-scatter of the incident radiation by small-scale irregularities in the lower ionosphere. The reflections come from a range of heights between 60 and 100 km. The use of these partial reflections has the advantage that a given height can be selected by a time gate and so the height of observation is directly under the experimenter's control. This means that the position of the gate can be varied to select the most suitable height for interesting comparisons with other techniques such as rocket vapour trails and meteor trail drifts.

The work described in the present paper is a further contribution to the study of drifts in the *D*-region using partially reflected radio waves.

2. Experimental Technique

The results were obtained using the facilities of the Buckland Park experimental station (Briggs *et al.*, 1969). The 2 MHz transmitting aerial consists of a square arrangement of four half-wave dipoles capable of transmitting linear, ordinary or extraordinary polarisation. In order to obtain maximum amplitude of partial reflections from the 80–100 km height interval, the sense of polarisation of the radio pulses was usually arranged to be ordinary during the day and extraordinary at night. Although the extraordinary wave is always preferentially reflected from a discontinuity, it is also highly absorbed by the lower *D*-region in the daytime. The transmitted pulses were of Gaussian shape with an approximate width between the half-power points of 25 μsec. The repetition rate was 50 Hz and the peak power, 25 kW.

The receiving aerials were selected from the Buckland Park array as indicated in Fig. 1. Two features worthy of note are the grouping of four aerials connected in parallel to form each receiving location, and the choice of two approximately equilateral receiving triangles. The use of four dipoles in parallel, and consequent narrowing

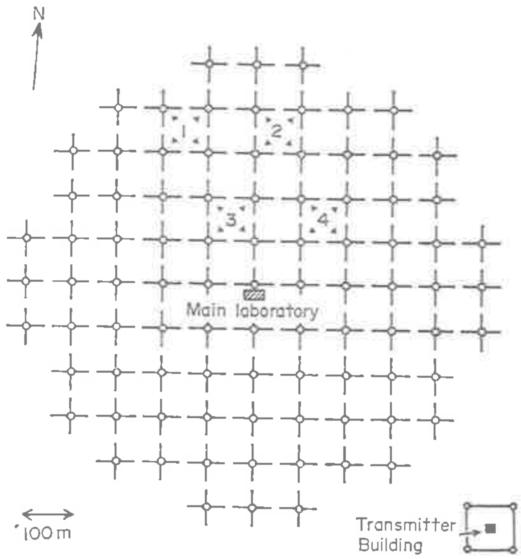


Fig. 1. The Buckland Park aerial array showing the aerial groupings used to obtain two approximately equilateral receiving triangles.

of beamwidth, has the two-fold advantage of increasing the signal-to-noise ratio and decreasing the response to off-vertical echoes. The advantages of an equilateral receiving configuration have been indicated both theoretically (Barber, 1957) and experimentally (Golley and Rossiter, 1970). If the spacing is too large, the fading at adjacent aerials is uncorrelated and the analysis breaks down. Golley and Rossiter (1971) found that the optimum correlation was 0·5 and this determines the best triangle size. Naturally ionospheric conditions vary widely, but for average *D*-region conditions this corresponds to a spacing of about 160 m. The final arrangement was chosen with these criteria in mind. The effective spacings in the two nearly equilateral receiving triangles (marked 123 and 234 in Fig. 1) were 204, 182 and 204 m.

The height of observation was selected by a gating pulse which stepped automatically to a new height interval with each transmitter pulse, restarting at the lowest height when the chosen cycle was complete. The width of the transmitted pulse set a limit of approx. 4 km on the height resolution, and a gate width corresponding to a 5-km height interval was used in all records. Generally, a cycle of five heights fully exploited the dynamic range of the receivers, which meant that the fading at each height was sampled at a frequency of 10 Hz. The data were digitized, and recorded on 7-channel magnetic tape. This made possible the use of 64 digital levels. A 3-min record length was chosen as the best compromise between the need for an adequate number of fading cycles to reduce statistical sampling errors, and the desire for effectively instantaneous velocity determinations.

3. Methods of Analysis

When a radio wave is totally or partially reflected from a certain level of the ionosphere, the reflected wave has imposed upon it an irregular diffraction pattern. At ground level the pattern moves with a velocity equal to twice the horizontal velocity of the ionosphere (FELGATE, 1970). The sampling of this pattern by a triangular arrangement of closely spaced receiving stations has been widely used for the observation of ionospheric drifts. Similar amplitude fluctuations are recorded by all three receivers with time displacements dependent on the magnitude and direction of the ionospheric velocity.

Various methods of analysis have been used to derive the velocity from the fading records. The fluctuations of the received signal are, in general, due to a combination of two factors—an overall steady motion of the pattern past the receivers, and random changes within the pattern as it moves. The simplest approach is to ignore the changing form of the diffraction pattern and attribute the fading solely to the average horizontal motion. The velocity can then be calculated from the time shifts between records and the respective spacings of the receiving locations. This has become known as the 'method of similar fades', and velocities derived in this way are termed 'apparent'. The time displacements for maximum similarity of the fading records can be calculated either manually using only corresponding peaks in signal strength or by the use of cross-correlation functions.

The technique of 'full correlation analysis' developed by BRIGGS *et al.* (1950) and PHILLIPS and SPENCER (1955) attempts to allow for both the contributions to the fading in deriving ionospheric parameters from those of the moving diffraction pattern. A parameter, designated V_c, is derived from the spatial and temporal correlation functions and this, together with the geometric form of the diffraction pattern, is used to produce a modified velocity, known as the 'true' velocity. The interpretation of V_c in terms of the physical properties of the ionosphere is uncertain but a high value of V_c may correspond to the case where the structural changes of the ionospheric irregularities are significant and the apparent velocity is then likely to be in error. However 'full correlation analysis' is only practical on a large scale if adapted for a fast computer (FOOKS, 1965), and the apparent velocities are still widely used.

The spaced receiver technique has the disadvantage that the spatial sampling is sparse and this necessitates sampling over a considerable interval of time. Thus the problem arises of possible variations in the bulk motion during the interval sampled. The various methods of analysis can therefore best be tested by comparison with velocities derived from a detailed spatial sampling of the diffraction pattern at two closely spaced times. An experiment conducted by GOLLEY and ROSSITER (1970) using a large number of aerials to sample the pattern, indicated that, despite the various assumptions of 'full correlation analysis', the 'true' velocity is a good estimate of the motion of the diffraction pattern provided the sampling triangle is not too small. In general, the 'apparent' velocity over-estimates the speed.

4. NATURE OF RESULTS

Records were taken on some 40 days from January to May 1972. Where possible they were taken throughout the full 24 hr, the aim being to look for evidence of tidal influence, both diurnal and semi-diurnal, in the drift results. The receiving equipment was continuously monitored and the receiver gains adjusted to give maximum continuity of data from the 85–95 km height range. This choice was largely governed by the prevailing echo structure. In the daytime the echo from the E-region was usually centred at a height of approx. 100 km, with D-region partial reflections returned from between 60 and 95 km. At night, 2-MHz radio waves are totally reflected from the F-region near a height of 230 km, but quite strong partial reflections nearly always occur between 80 and 100 km. On many occasions, the nighttime echo centred on about 87 km (extraordinary polarisation) was comparable in magnitude with the daytime signal from the same height (ordinary).

All records were subsequently analysed using full correlation analysis to derive the 'true' velocity, the velocity of the ground diffraction pattern being halved to obtain the ionospheric velocity. Records were rejected when the mean signal amplitude was either too low or close to the receiver saturation level. Noise was often a problem, particularly at night, but was easily detected after analysis by false peaks in the cross-correlation functions at zero time lag or by unduly sharp peaks in the autocorrelation functions, again at zero time lag. Records were rejected where the maximum of any one of the three cross-correlation functions failed to exceed a value of 0·2.

Ionospheric conditions were not always suitable for wind observations. On occasions, the cross-correlation functions were oscillatory, ruling out the possibility of determining unambiguously the true time shifts between receivers. On other occasions, the occurrence of strong but short-lived echoes from a sporadic E-layer made gain setting impossible because the signal alternately fell to zero or saturated the receivers. Sometimes two separate echoes contributed to the fading from a particular height interval and generally the analysis broke down in these cases. An additional problem, occurring less often, but sometimes for several hours at a time, was interference due to multiple-hop E- and F-region echoes from the previous transmitter pulse.

In general, records were obtained at least every 20 min and often every 10 min. For routine work the results for only one receiving triangle were analysed, but the second provided a useful consistency check. On 23 January 1972, 74 wind determinations were obtained between 80 and 95 km. For each of these, the value of 'true' velocity magnitude obtained using one triangle was subtracted from the corresponding value for the other triangle and this difference was squared. The 74 squared differences were then averaged to obtain the root mean square velocity difference. The directions were treated in the same way. The values thus obtained can be taken as an indication of the reliability of individual wind values. This analysis yielded an uncertainty in magnitude of 10 m/sec and in direction of 14°.

5. Comparison with Meteor Winds

In this section, the results of the partial reflection drifts are compared with the winds determined by the Adelaide Meteor System (Weiss and Elford, 1963). The meteor radar is generally run for 3 days each month but was used for a 6-day period in March 1972. The technique is most reliable at the heights where the number of detectable ionisation trails is largest. At 27 MHz this height is approx. 80–95 km. For the comparisons extending over several days, winds were determined on a 3-hourly basis. The two determinations might be expected to differ for several reasons. In a given interval it is possible that meteor echoes may occur only at a particular height, for example 84 km, whereas the bulk of the radio power returned may be from a lower height. Similarly, the influx of meteors is by no means constant and the only

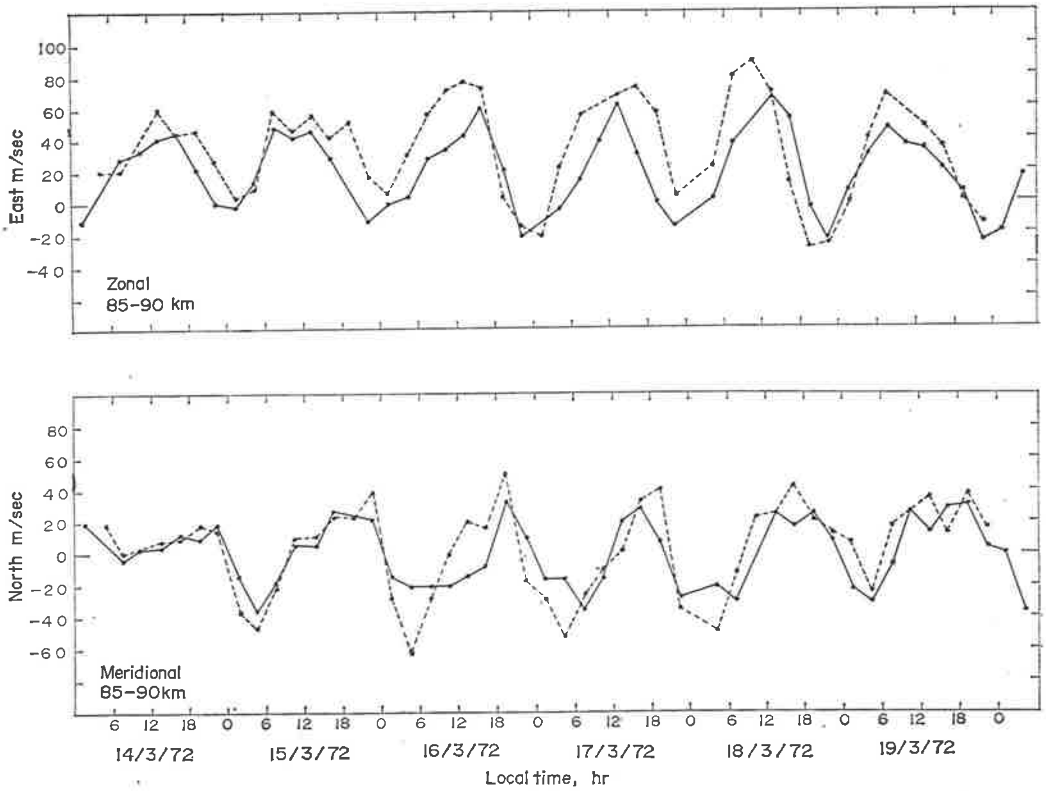


Fig. 2. A 6-day comparison of ionospheric drifts (full lines) and meteor radar wind estimates (broken lines) with each discrete point representing a 3-hourly average.

echoes may occur early in a particular 3-hr interval whereas ionospheric conditions may dictate that drifts can be determined only in the last hour. It is theoretically possible for the two reflection points to differ horizontally by some 150 km although the two techniques basically probe the same volume.

The longest simultaneous determination by both methods is shown in Fig. 2. The full lines represent the ionospheric drifts whilst the dashed lines follow the velocities given by the meteor radar, each discrete point being the appropriate 3-hourly average. The occasions where no such determination was possible by either method are relatively few. A total of 339 individual drift readings from the 85–90 km interval were obtained over the 6 days so each point on the continuous curve represents, on the average, seven drift values.

An obvious feature of the curves is the dominance of the 24-hr oscillation. In both the meteor and radio cases, the wind is towards the east shortly after noon, local time, on most days. The meridional component has a northward peak roughly 6 hr later. This behaviour is consistent with the expected anti-clockwise rotation of the diurnal tide in the southern hemisphere. Further tidal features of the drifts for individual days are discussed in the following section.

Some idea of the prevailing winds, as indicated by the ionospheric drifts, can be obtained from Fig. 2. In the case of the zonal winds, there is a constant bias towards the east. The magnitude of this eastward prevailing wind in fact averages 22 m/sec over the 6-day period. The meridional oscillation, on the other hand, shows no significant overall trend towards north or south, the mean drift being 2 m/sec to the south.

It is interesting to note two of the results of a harmonic analysis of the ionospheric drift data displayed in Fig. 2. Firstly, the average magnitude of the diurnal tide is close to 30 m/sec whereas the semidiurnal oscillation has an amplitude just in excess of 10 m/sec. Similar results have been established by analysis of the Adelaide meteor results (Elford, 1959) and are in contrast to the situation at higher latitudes (e.g. Jodrell Bank) where the semi-diurnal tide is predominant. Secondly, a comparison with corresponding ionospheric drifts from the 90–95 km height indicates that the phase of the diurnal tide at the upper height level leads the 85–90 km phase by between 1 and 2 hr (as can be seen, for example, by a careful comparison of Figs. 5(d) and (e)).

Similar agreement between the two techniques was found for the other months and also for other heights where results of both techniques were sufficiently continuous. Figures 3 and 4 show the results for February and May 1972, respectively. The quality of the agreement differs little for these 2 months. There appears to be a less purely diurnal tide in the early winter month of May. It is noticeable that the prevailing part of the February meridional component becomes consistently more northward over the 3-day period. It is possible that such variations offer evidence of planetary waves and it is hoped in the future to study these long-term variations in greater detail.

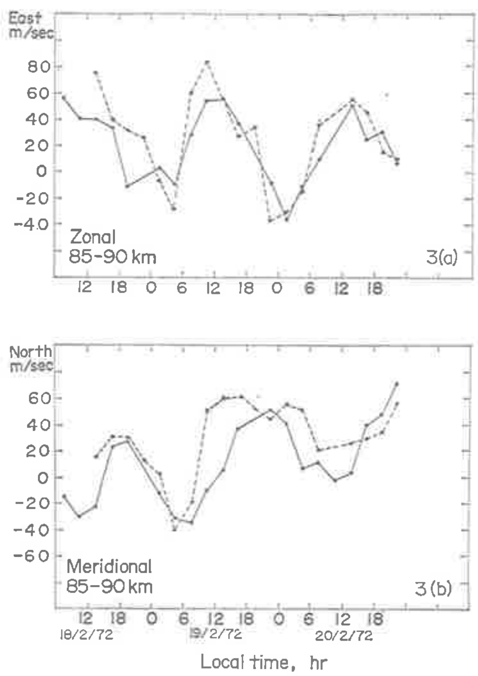


Fig. 3. A 3-day comparison of ionospheric drifts (full lines) and meteor radar estimates (broken lines) for February 1972.

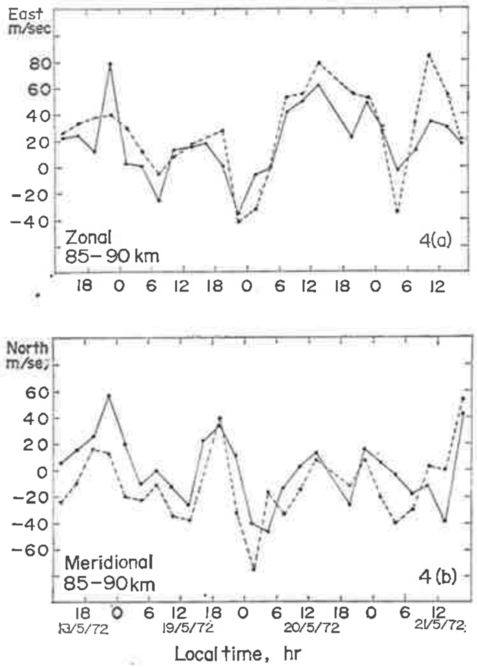


Fig. 4. A 3-day comparison of ionospheric drifts and meteor radar wind estimates for May 1972.

The general agreement between the ionospheric drifts and the meteor results is striking. There is, however, a tendency for the ionospheric drifts to be smaller in magnitude by an average of about 10 per cent. This is consistent with the triangle size effect. When the pattern size exceeds the aerial spacing, the drifts will be less than the real ionospheric velocity but there will be no compensating tendency for them to be large when the pattern size is small. Preliminary analysis in fact indicates that the pattern size is often largest in the early morning and that it undergoes a semiannual variation with a maximum in March. This cannot be isolated conclusively from 5 months' data but would explain the more even agreement in magnitude for May than for March.

6. Tides on Individual Days

A convenient way of displaying the tidal features of the drifts is to plot the tip of the wind vector for each hour of the day and this has been done for six examples in Fig. 5. Figure 5(a) is a case where the tide is predominantly diurnal but does not complete the rotation within the 24 hr. This might be explained by a change in the prevailing wind or tidal amplitude but may just be the result of random fluctuations of neither prevailing nor tidal origin. In Fig. 5(b), the 24-hr oscillation is more nearly circular with a prevailing wind of about 20 m/sec towards the east. It is of interest to compare this with Fig. 5(c) which represents the same day but for the higher height interval of 90–95 km. It would appear that on this particular day the tidal amplitude was greater at the greater height. However a comparison of Figs.

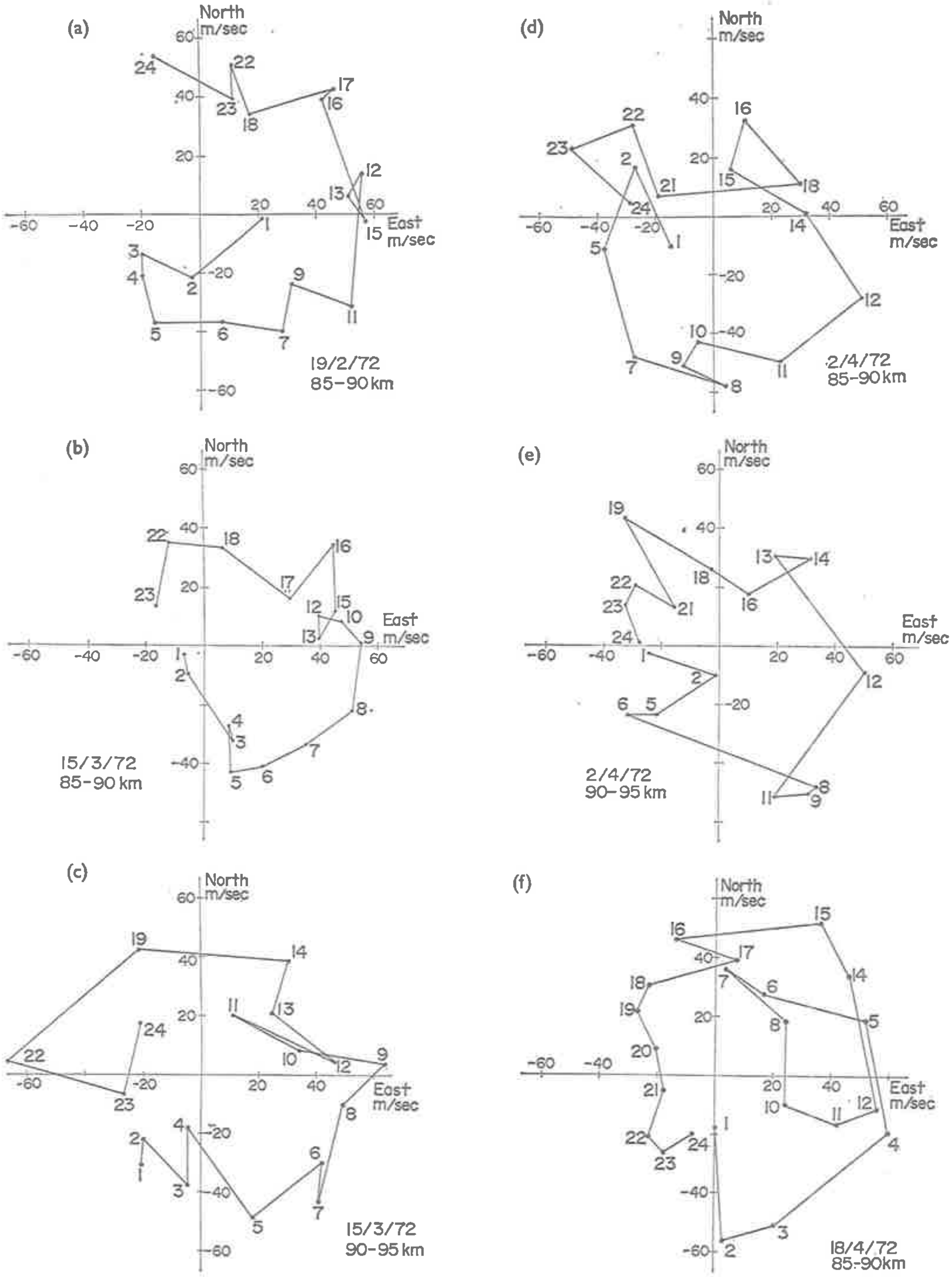


Fig. 5. Polar plots of ionospheric drifts with the position of each hourly average drift indicated by a point in the direction of the observed movement. The magnitude of the drift is indicated by the distance from the origin.

5(d) and (e) shows that this is not always the case; for this example the diurnal tide again dominates the wind changes but the magnitudes are similar at both heights.

The occurrence of 12- as well as 24-hr tidal components is clearly shown in Fig. 5(f). It is interesting to note that in most cases the phase of the diurnal tide is such that the daytime winds are generally eastward and furthermore the phase of the semi-diurnal tide with respect to the diurnal often results in the wind changes being rather small in the middle of the day. This is very much in keeping with the results obtained for example by ROSSITER (1970) whose observations refer mainly to the period between 1000 and 1500 hr LT.

These six vectors plots are representative of the tidal influence that was evidenced on virtually all the days where results were obtained for most of the 24 hr.

It is intended, when data have been collected over 12 continuous months, to publish details of tidal strengths and trends together with information on the annual variation of the prevailing wind. In particular, the midwinter month of July has been studied carefully over the years 1970–1972.

It is hoped in future experiments to compare the drifts with common-volume measurements by rocket techniques and also to extend results to the lower D-region. The results discussed in this paper have been obtained using a 5-km gate width dictated by the nature of the transmitted pulse. This is rather too large to search for gravity waves which might be expected to have vertical wavelengths of about 8 km at the heights in question. Studies are planned using 2-km gate widths and a correspondingly narrower pulse.

7. DISCUSSION

The results presented in this paper should be seen in the light of previous comparisons between ionospheric drifts measured by spaced receivers and by other techniques. In many such comparisons, the two sets of data have been separated in space or time. GREENHOW and NEUFELD (1955) compared the daily mean prevailing wind measured by the Jodrell Bank meteor system with the mean prevailing wind based on a survey of Cambridge E-region drift results made by BRIGGS and SPENCER (1954). The same data were used by JONES (1958) to compare the phases of the semi-diurnal tide. However the two sets of results were obtained in different years. No conclusive results were obtained by MIRKOTAN (1961) in a comparison with meteor-radar winds, but KOCHANSKI (1964) found good agreement at about 110 km between radio measurements of the plasma cloud associated with a luminous trail, and the neutral gas motions as determined optically. SPRENGER and SCHMINDER (1968) have investigated in some detail both the average and periodic behaviour of winds at comparable heights obtained from meteors at Jodrell Bank and from low frequency totally reflected radio waves in North Germany. The station separation was about 900 km. MÜLLER (1968) compared Sheffield meteor data with ionospheric drifts from Aberystwyth, but whereas the average height of the drift data was 103 km, the meteors occurred primarily in the 80–100 km region, averaging 95 km.

WRIGHT (1968) compared drifts from sporadic E-layers with the distortion of luminous trails from rockets and projectiles launched in the same area. The agreement was poor in magnitude and direction but was better if the 'factor of 2' was ignored. More recently, WRIGHT *et al.* (1971) used their multifrequency drift measuring equipment in conjunction with the Garchy radar meteor system. The results were not encouraging but heights covered by the two methods barely overlapped. VINCENT (1972) used a spaced phase path technique to obtain total reflection drifts which agreed on the average with meteor 'winds'. A recent experiment by LYSENKO *et al.* (1972) confirms the overall impression created by the total reflection work: that whilst there are some discrepancies in detail, the technique reliably indicates the principal features of the neutral air circulation.

Fewer comparisons have been made using the technique of partial reflections. FRASER and KOCHANSKI (1970) compared ionospheric drifts obtained at Birdlings Flat, New Zealand with other spatially distant wind measurements, and with circulation models. With several reservations, they concluded that a wind interpretation of the drifts was indicated.

As in the present experiment, ROSSITER (1970) made a common-volume comparison between ionospheric drifts (total and partial reflections) and the Adelaide meteor radar measurements. The results were taken over a small part of the day and the scatter of the individual comparisons indicated that there was a basic unreliability in individual drift determinations. There was also a tendency for the drifts to be systematically too small, a fact which may well have resulted from the use of a triangle of side 91 m. The drift equipment has since greatly improved, as was outlined in Section 2. Groups of aerials in an almost equilateral configuration replace the individual dipoles used in the earlier work. The information rate has been increased by a factor of 20 and the number of digitizing levels by a factor of 10. All records are now rigorously rejected which do not comply with the assumptions of full correlation analysis. These improvements in technique have resulted in a much better agreement between the ionospheric drifts and the meteor drifts than was found by Rossiter.

The experimental evidence, in the case of both total and partial reflections, now seems to justify the two basic assumptions of the spaced receiver technique: namely, the interpretation of the diffraction pattern parameters in terms of physical properties of the reflection region (RATCLIFFE, 1956) and the assumption that the electrons in the lower ionosphere do not in general move independently of the neutral air under the influence of magnetic and electric fields.

8. CONCLUSION

The partial reflection experiment has been shown to be capable of measuring the neutral air motion in the D-region. As many as possible of the uncertainties in the experimental arrangement and analysis have been eliminated and results taken intensively during both day and night. A comparison has been made with winds determined by the meteor radar technique and good agreement has been found. It should be remembered that the meteor technique is itself indirect and with its own uncertainties, such as the susceptibility of individual ionized trails to small-scale turbulent eddies, but the method is generally accepted as well suited to the measurement of neutral winds.

It may be that there are occasions when the partial reflection results are influenced by other than the air motion but the present results indicate that these cases are not as common as has been suggested. It may be concluded that the technique provides a reliable and relatively inexpensive method for synoptic studies of neutral air motions in the height range 65–100 km.

Acknowledgements—The author would like to thank Drs. B. H. BRIGGS, W. G. ELFORD and R. A. VINCENT for helpful suggestions and discussion. He is grateful to Dr. ELFORD for allowing him to analyse the meteor data. The work was supported by the Australian Research Grants Committee, the Radio Research Board, and the University of Adelaide, and the author was in receipt of a C.S.I.R.O. Studentship.

REFERENCES

AWE O.	1961	*J. atmos. terr. Phys.* **21,** 142.
BARBER N.	1957	*J. atmos. terr. Phys.* **11,** 299.
BRIGGS B. H., ELFORD W. G., FELGATE D. G., GOLLEY M. G., ROSSITER D. E. and SMITH J. W.	1969	*Nature, Lond.* **223,** 1321.
BRIGGS B. H., PHILLIPS G. J. and SHINN D. H.	1950	*Proc. Phys. Soc.* **B63,** 106.
BRIGGS B. H. and SPENCER M.	1954	*Rep. Prog. Phys.* **17,** 245.
ELFORD W. G.	1959	*Planet. Space Sci.* **1,** 94.
FELGATE D. G.	1970	*J. atmos. terr. Phys.* **32,** 241.
FOOKS G. F.	1965	*J. atmos. terr. Phys.* **27,** 979.
FRASER G. J.	1965	*J. atmos. Sci.* **22,** 217.
FRASER G. J. and KOCHANSKI A.	1970	*Annls Géophys.* **26,** 675.
GOLLEY M. G. and ROSSITER D. E.	1970	*J. atmos. terr. Phys.* **32,** 1215.
GREENHOW J. S. and NEUFELD E. L.	1955	*Phil. Mag.* **45,** 471.
JONES I. L.	1958	*J. atmos. terr. Phys.* **12,** 68.
KOCHANSKI A.	1964	*J. geophys. Res.* **69,** 3651.
LYSENKO I. A., PORTNYAGIN YU. I., SPRENGER K., GREISIGER K. M. and SCHMINDER R.	1972	*J. atmos. terr. Phys.* **34,** 1435.
MÜLLER H. G.	1968	*J. atmos. terr. Phys.* **30,** 701.
PHILLIPS G. J. and SPENCER M.	1955	*Proc. Phys. Soc., Lond.* **B68,** 481.
RATCLIFFE J. A.	1956	*Rep. Prog. Phys.* **19,** 188.
ROSSITER D. E.	1970	*Aust. J. Phys.* **23,** 103.
SPRENGER K. and SCHMINDER R.	1968	*J. atmos. terr. Phys.* **30,** 693.
VINCENT R. A.	1972	*J. atmos. terr. Phys.* **34,** 1881.
WEISS A. A. and ELFORD W. G.	1963	*Proc. IRE* **24,** 197.
WRIGHT J. W.	1968	*J. atmos. terr. Phys.* **30,** 919.

Reference is also made to the following unpublished material:

MIRKOTAN S. F.	1961	Nice Symposium URSI-CIG.
WRIGHT J. W., PLYWASKI W. and FEDOR L. S.	1971	NOAA Tech. Mem. ERLTM-OD6, Boulder, Colorado.

APPENDIX II

Reprint of a paper:

"Studies of D-region drifts during the winters of 1970 - 1972"

T.J. Stubbs and R.A.Vincent.

Stubbs, T., & Vincent, R. (1973). Studies of D-Region Drifts During the Winters of 1970?72. *Australian Journal of Physics, 26*(5), 645-660.