STUDIES OF *D*-REGION DRIFTS DURING THE WINTERS OF 1970–72

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[Manuscript received 11 April 1973]

Abstract

Partially reflected radio waves have been used to study *D*-region drifts in the midwinter month of July for the years 1970–72. The drifts are shown to agree well with wind estimates at similar heights obtained using meteor radar techniques. The behaviour of the drifts is consistent with a strong tidal influence, and the nature of these tides is discussed with emphasis on the phase and amplitude variations in time and height. Comparison of the monthly mean drift profiles with wind models leads to a particularly good representation of the expected zonal drift profile. No evidence is found for any anomalous winter behaviour.

I. Introduction

When the ionosphere is illuminated from below by radio waves, the resulting weak partial reflections from the *D*-region can be used to deduce the drift velocities of ionospheric irregularities at these heights (see e.g. Fraser 1968). A comparison of drifts measured in this way with Doppler-radar determinations of the drift of meteor trails at similar heights was made by Rossiter (1971) to resolve uncertainty as to whether the measured velocities are indicators of neutral air motions. As the question was not completely resolved by his work, further studies of a similar type have been carried out and the results are described in the present paper.

Rossiter's (1971) observations, which were made over a 6 hr period centred on noon, showed that the quality of agreement between the drifts and meteor measurements varied seasonally, the greatest discrepancies occurring in winter. He found that drifts on several days, with the maximum number in July, showed little apparent tidal rotation although the meteor data, which had been analysed by a particular type of model-fitting procedure (Groves 1959), showed significant tidal variations. Furthermore, the drifts were predominantly towards the east and the mean zonal drift velocities were substantially greater than the wind models proposed by other workers, particularly at heights near 90 km. Rossiter suggested that the drift results in winter may be influenced by eastward-propagating internal gravity waves, perhaps generated by the eastward motion of cold fronts. His results raised several questions regarding the suitability of the partial reflection technique for studies of the neutral air motions, and the relative roles played in *D*-region dynamics by atmospheric tides and internal gravity waves. Stubbs (1973) has recently published results obtained in summer and autumn which show, with an improved experimental arrangement and observations extended over the full 24 hr, that the partial reflection method does give reliable measurements of the neutral wind in the *D*-region, the drifts showing strong tidal features and agreeing well with the meteor results.

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The work described here is a study of partial reflection drifts during the winter month of July for three consecutive years. This month was chosen for detailed study because it was the period when Rossiter (1971) found the poorest agreement between drift and meteor data. Drift observations were made over as many hours of the day as possible and particular attention was paid to the height region around 90 km. The present drift recording and analysis techniques, together with the somewhat different method of analysing the meteor data from that used by Rossiter, are described in Section II. Following a comparison of the drift and meteor results in Section III, the overall results are discussed in Section IV.

II. Experimental Techniques

(a) *Partial Reflection Drifts*

The present drift results were obtained using the Buckland Park transmitting and receiving arrays (Briggs *et al.* 1969). The observational techniques were progressively improved during the three years. For the original work described by Rossiter (1971), the transmitter peak power was 5 kW and the receiving arrangement consisted of three single dipoles at the corners of a right-angled triangle of short side 91 m. Reflections from a 5 km range interval were recorded on chart and digitized manually to 10 levels at time spacings of 1 s, the drifts being then calculated from the full correlation analysis of Briggs *et al.* (1950) and Phillips and Spencer (1955).

For all the observations reported here, the transmitter power was increased to 25 kW with a 30 μs pulse length. In July 1970 two different forms of data recording and analysis were used:

(1) The first system of recording was the three-receiver arrangement used by Rossiter, although the sides of the triangle were increased to 182 m, as Golley and Rossiter (1970) had shown that drift velocities are systematically reduced in magnitude when the size of the sampling triangle is too small. Receiver separations of at least 160 m seem desirable for partial reflection work, and of the order of 300 m for *E*-region total reflections. The amplitude fading was again recorded on chart and manually digitized.

(2) In addition to the three-receiver measurements a number of observations were made using the spatial correlation technique due to Briggs (1968). This method, which made use of the full array of 89 dipoles available at Buckland Park, gave an independent check of the measured velocities and indicated whether any observational bias was introduced by the first method. The technique allowed velocities to be calculated each second and only those records with constant values over a 3 min interval were accepted. In general there were no significant differences between velocities measured by the two methods. Because of the relatively low sensitivity of the 89 receivers used for the spatial correlation method it was only possible to observe reflections from above 90 km. Since the rejection rate for this method was similar to that for the three-receiver technique (50%), and the measured velocities were similar, it was decided not to use the full array in subsequent years.

The quantity of data obtained in 1970 was limited because of the time required for the analysis of the chart records and the extensive computer analysis involved in the spatial correlation technique. For these reasons the drift data were averaged into two 10 km height intervals centred on 85 and 95 km respectively.

In July 1971 the receiving arrangement remained effectively a right-angled triangle of side 182 m but each of the original dipoles was replaced by four dipoles in parallel, thereby increasing the signal-to-noise ratio. The major improvement, however, was the recording of many records on magnetic tape, thus allowing the amplitude fading to be sampled more frequently than for the chart records; the tape records weie in fact sampled every 0·2 s and digitized to 64 levels. Only one height range could be sampled at any one time and, since a right-angled triangle was used, some directional errors may still have been present. These errors were substantially reduced, however, because for the tape data it was possible to record the echo amplitudes at four dipole groups situated at the corners of a square thereby allowing the drift velocities to be calculated by analysing all four possible receiving triangles. The final drift velocity was then taken as the average of the four resultant drifts.

In July 1972 all observations were recorded on magnetic tape and a range-scanning gate enabled information to be obtained simultaneously from several 5 km height intervals. Instead of the previous right-angled triangle an approximately equilateral arrangement was used, with each receiving antenna consisting of four dipoles connected in parallel. This recording arrangement has been described in more detail by Stubbs (1973).

(*b*) *Meteor Trail Drifts*

For each ionized trail tracked by the Adelaide meteor radar system (Weiss and Elford 1963), the recorded information includes the direction cosines of the specular reflection point, the height of the reflection point, and the "line of sight" component of the trail drift. One observation by itself is not sufficient to determine the horizontal wind velocity and at least two echoes at different locations but at nearly the same height and time are required. Since these conditions rarely occur, some form of averaging process is necessary in order to determine the wind components. Two types of analysis are in general use at Adelaide at present:

(1) One technique is due to Groves (1959) and consists of a harmonic analysis of all the meteor data from a given observing period, by a least squares fitting procedure. The data are usually analysed in 24 hr periods and a prevailing wind plus 24, 12, and 8 hr harmonic components are fitted. The mean value and the amplitude and phase of each component are allowed to have polynomial variations with height. Rossiter (1971) compared his drift results with meteor data as analysed by this approach since it leads directly to the required tidal information. However, the wind fields for a given period reconstructed from this analysis may show apparent rotation introduced by data taken outside the relevant time and height interval.

(2) An alternative technique is to take the data from a particular height and time interval and perform a simple component averaging, making the assumption that the velocity is constant within the data interval. Information from other intervals is ignored. In order to get as large a number of meteor trails as possible, and therefore a significant average, the height and time intervals are usually chosen to be 5 km and between 1 and 3 hr respectively. This technique of determining the meteor winds has been chosen in the present work, firstly because the Groves (1959) analysis introduces considerable smoothing, and secondly because the simple averaging procedure treats the meteor and drift data in much the same fashion, thus allowing a more direct comparison to be made.

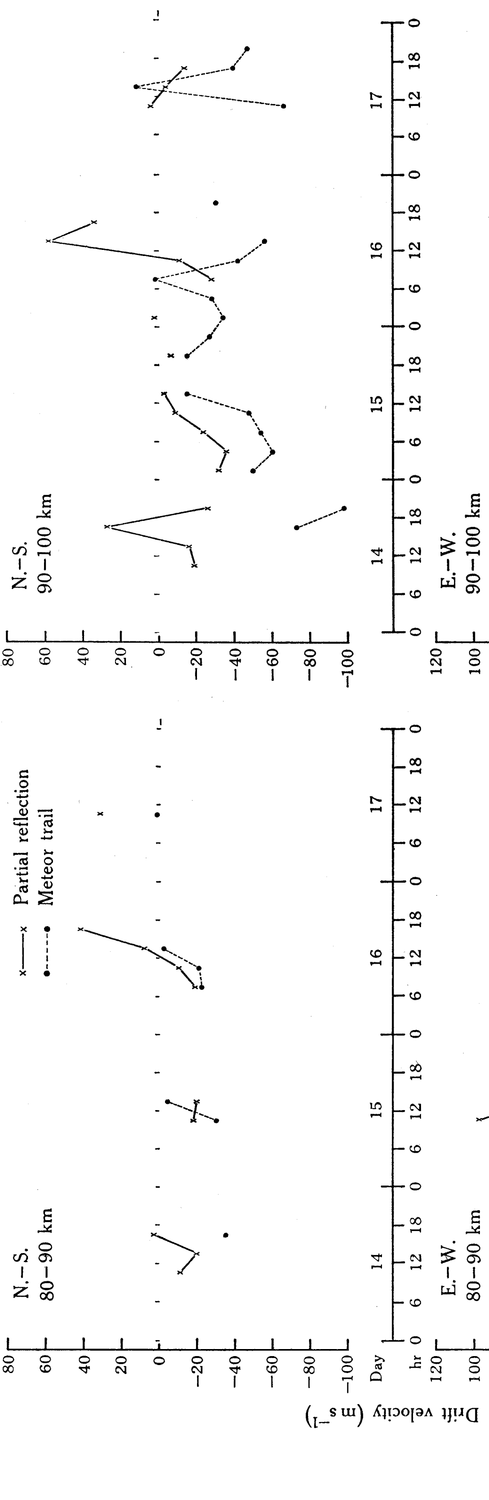

Fig. 1. July 1970, 80–100 km in 10 km intervals.

Figs. 1–3.—Comparisons of 3 hr averages of north–south and east–west components of partial reflection and meteor trail drifts during July 1970, 1971, and 1972.

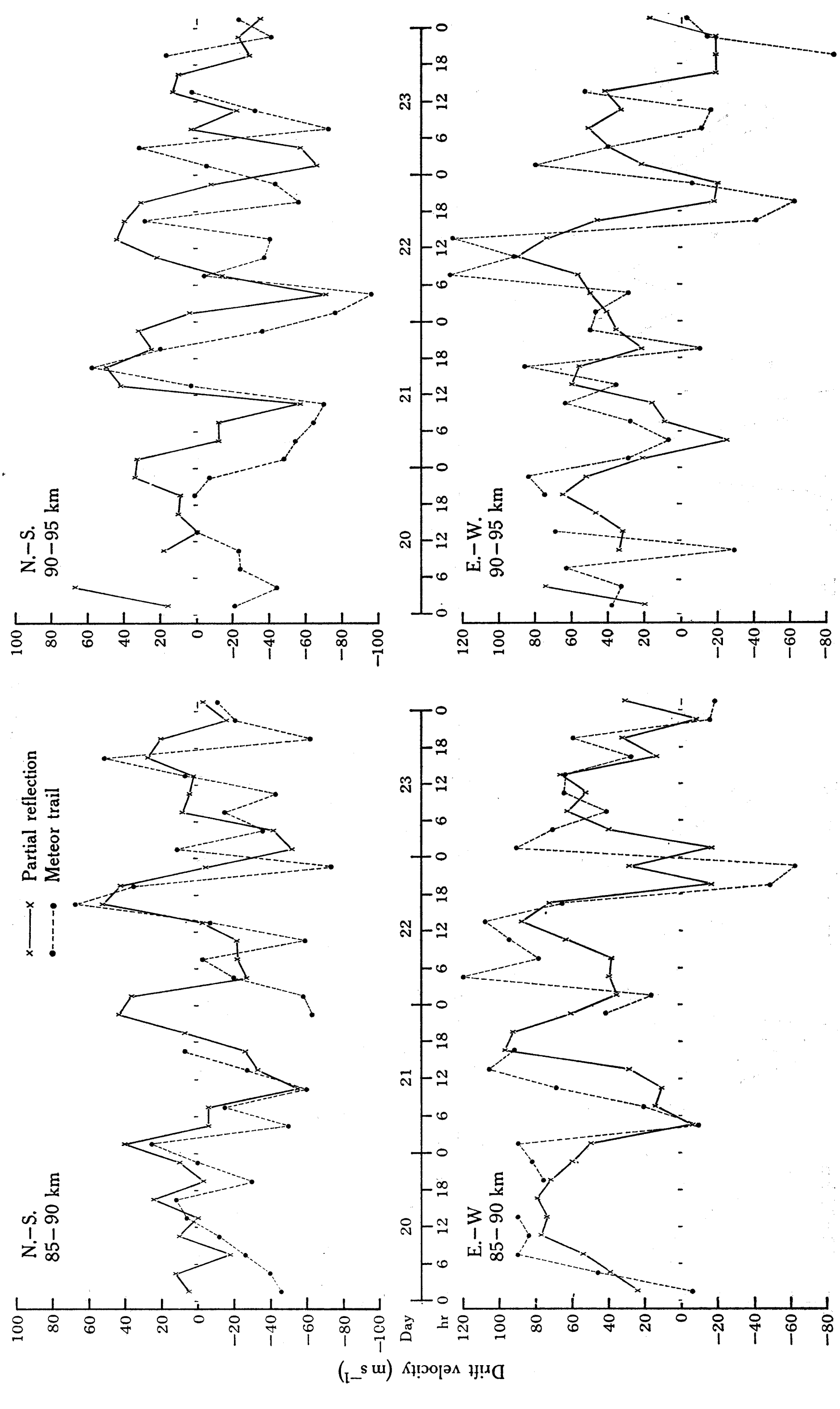

Fig. 2. July 1971, 85–95 km in 5 km intervals.

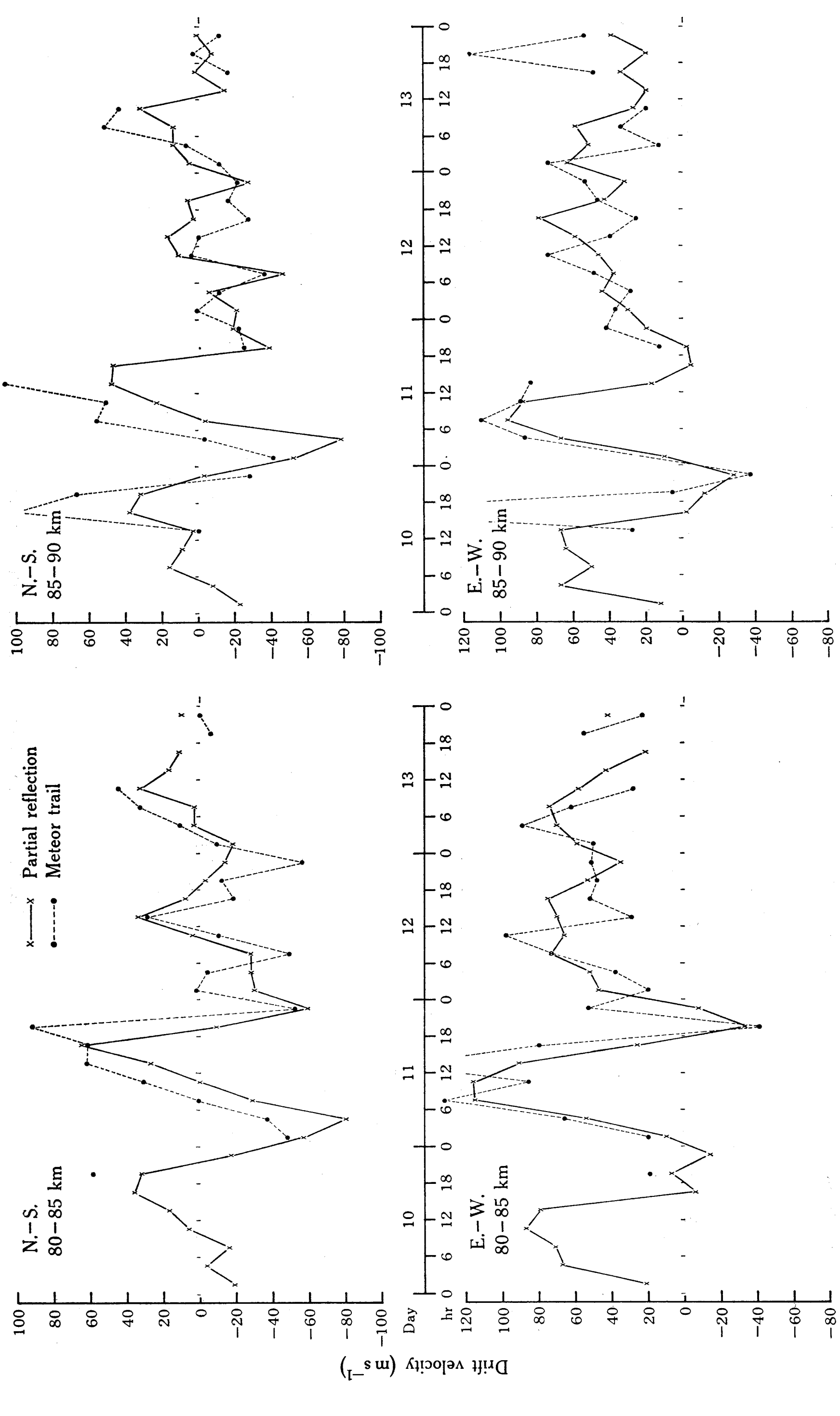

Fig. 3(*a*). July 1972, 80–90 km in 5 km intervals.

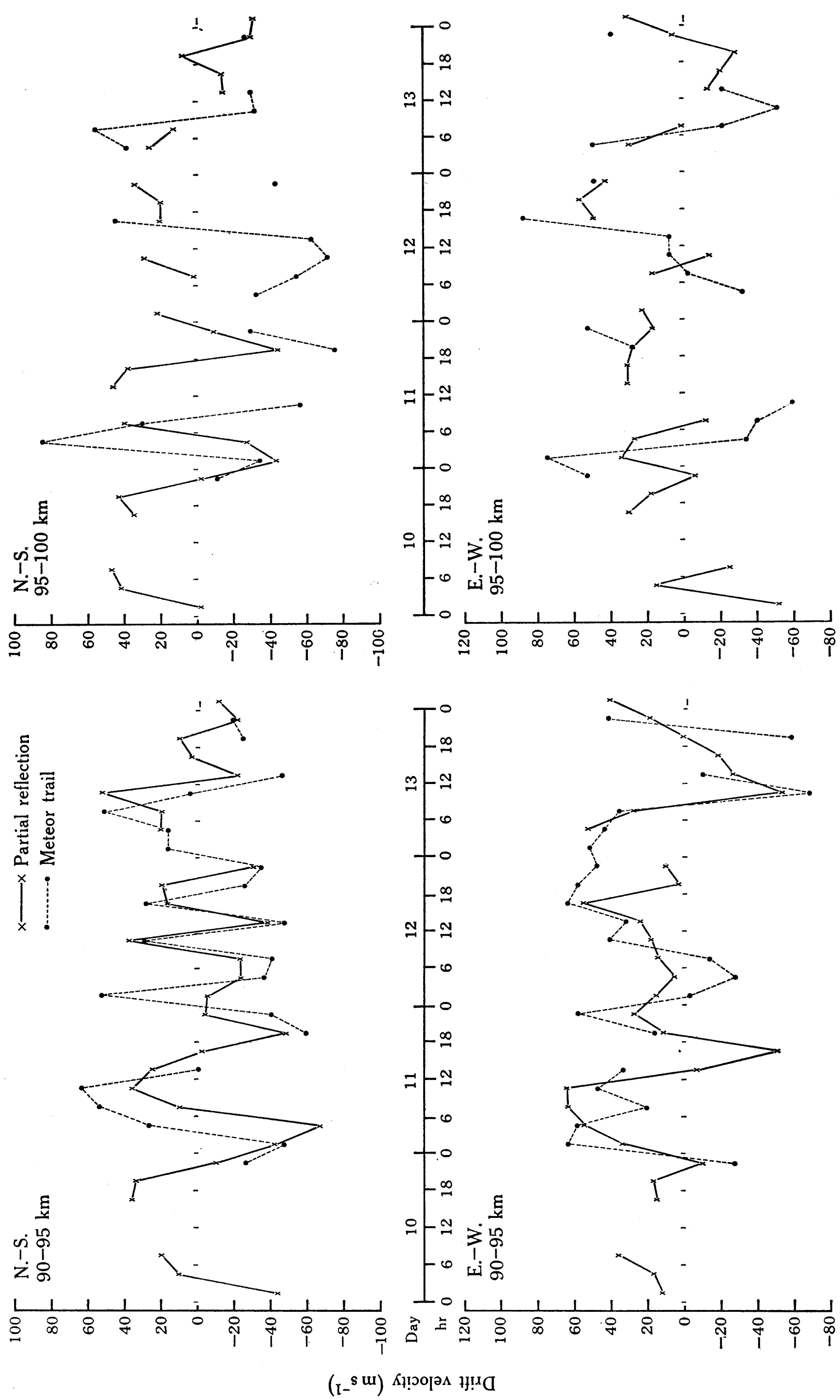

Fig. 3(*b*). July 1972, 90–100 km in 5 km intervals.

III. Results

(a) *Comparisons of Drift Data*

The purpose of this section is to consider the relative behaviour of the meteor and drift motions over the three winter periods. Features of the data which are of a geophysical interest are discussed in Section IV. In order to display the results for all three years in the same way it was found most convenient to group the data into 3 hr periods and compute the average wind for each period by the two methods. The strong diurnal variation in the meteor rates resulted in many hours in the late afternoon when less than two meteor echoes occurred in a particular 5 km height interval. The likelihood that at least two meteor echoes occurred in a given height interval was therefore increased by using the 3 hr data periods.

The results for 1970 are displayed in Figure 1. The data for the 80–90 km height interval in particular can be seen to be sparse and although the two sets of points tend to overlap it is obviously impossible to infer the presence or absence of tidal behaviour from such a restricted set of data. The results for 90–100 km are more continuous but not much more conclusive. The variations in drift velocity from one time interval to the next appear to suggest appreciable tidal motions and the drifts do not seem to be directed significantly more to the east than the winds deduced from the meteor data. However, the latter do appear to suggest a stronger southward component than the drifts indicate. The reason for this is not clear but may be associated with a strong shear in the meridional wind over the 10 km height interval, and a tendency for the majority of meteor observations to come from the lower heights.

It is easier to draw conclusions about the relative behaviour of the two sets of observations from the results for 1971 shown in Figure 2. At 85–90 km, the agreement is often good in both the zonal and meridional components but there is a tendency for the meteor wind determinations to fluctuate more than the drift values from one interval to the next. On none of the days of observation do the drifts show any evidence of the anomalous behaviour observed by Rossiter (1971), i.e. predominant eastward motion with little evidence of a tidal rotation. On the contrary, the drifts clearly show the presence of a significant diurnal tide on all days and even by eye it is apparent that the zonal drift motion leads the meridional component by about 6 hr, which gives the expected anticlockwise rotation in the southern hemisphere if the diurnal tide is indeed predominant. Very similar observations can be made from the results for the 90–95 km interval, although perhaps the detailed agreement is not as good. One point which is worth making is that for the hours around noon both the drifts and meteor wind motions tend to have large eastward components.

In Figure 3 the data for 1972 are presented for four height intervals. The least number of data points was obtained in the 95–100 km height interval and at best there is only fair agreement between the results from the two techniques. At all other heights the results show good agreement. The generally eastward motion near midday, as noted above, is again obvious in the 1972 results but there is little evidence of any tendency for the drifts to remain constant when the meteor winds exhibit tidal behaviour. In particular, attention is drawn to the very strong tides evident in both curves on 11 July and the comparative lack of a strong tide on later days. The overall sinusoidal nature of all curves is readily apparent.

In order to arrive at a quantitative measure of the quality of agreement between the two sets of data, the correlation coefficient of the two curves was calculated for the 85–90 and 90–95 km intervals in 1971 and 1972. Points where only two or three meteors had been used in a 3 hr wind determination were ignored. In six of the eight cases, with an average of 25 degrees of freedom, the resultant coefficients were significant at better than the 0·5% level. The meridional components at the two

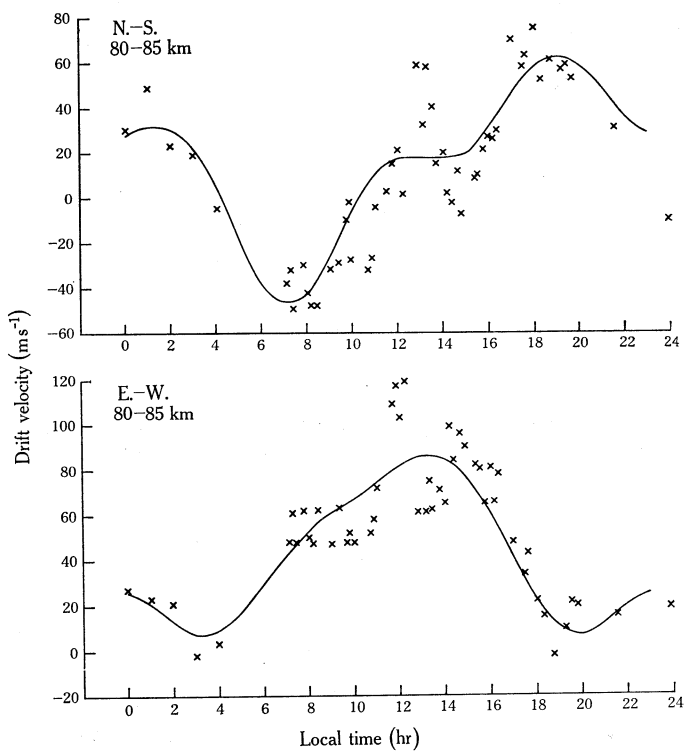

Fig. 4.—Partial reflection drifts at 80–85 km on 9 July 1972. The curves are the best fits to the individual drift determinations for a mean wind plus 24, 12, and 8 hr harmonic components.

heights for 1971 yielded correlations that were significant at the 10% level. These lower correlations reflect not only the greater variability of the meteor results but also short periods of complete disagreement, as for example between 2100 hr on 21 July and 0300 hr on 22 July at 85–90 km.

The relative smoothness of the drift results when compared with the meteor wind fluctuations has already been indicated. Possible reasons for this are discussed in Section IV, but it may be noted here that this smoothness is partly due to a more uniform data rate for the drift observations, and also indicates the consistency of individual 3 min drift determinations. All data points from the records analysed for the 80–85 km altitude interval on 9 July 1972 are plotted in Figure 4, together with the analytic curves resulting from a least squares fit to the points assuming a

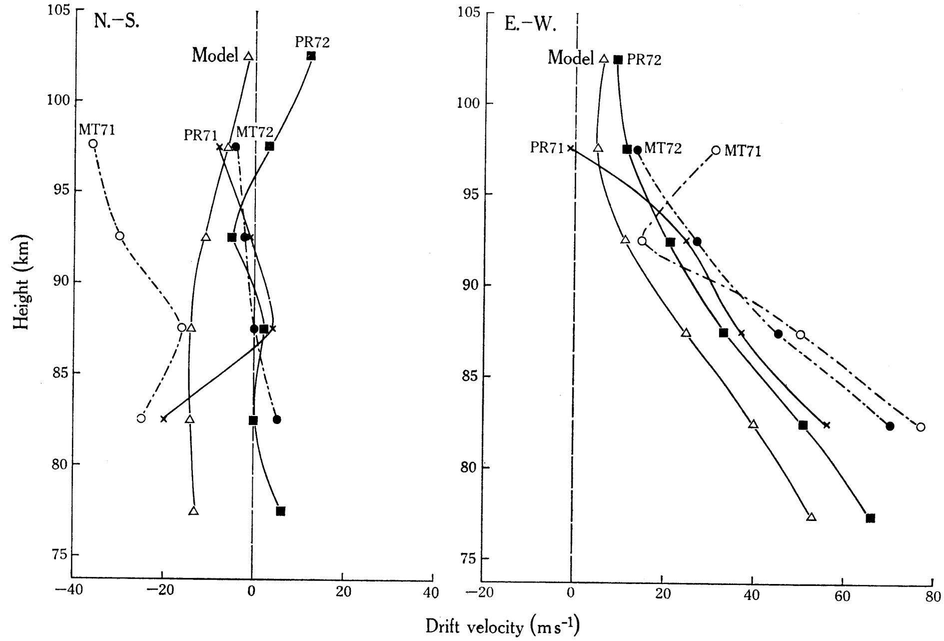

Fig. 5.—Comparison of the Groves (1969) wind model with the mean winds for July 1971 and 1972 as determined by the partial reflection (PR) and meteor trail (MT) drift techniques.

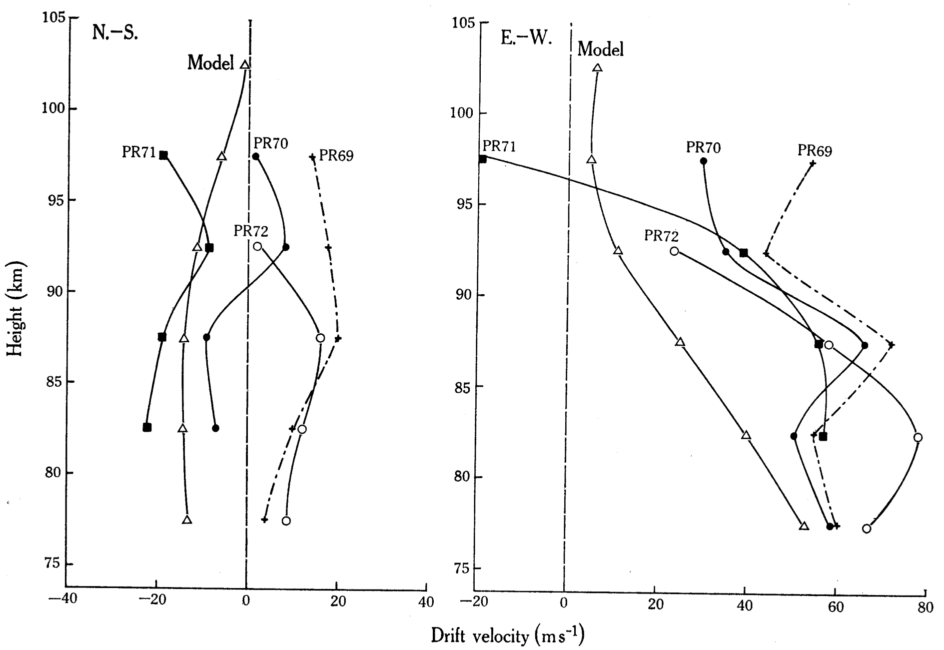

Fig. 6.—Comparison of the Groves (1969) wind model with the mean partial reflection (PR) drift profiles for July 1969–72 as determined using only data obtained between 1000 and 1400 hr LT.

prevailing wind plus 24, 12, and 8 hr harmonic components. It is noteworthy that drift measurements taken within a relatively short time of one another do not show a wide scatter, although there are significant overall departures from the mean trend. This latter point is considered in Section IV.

(*b*) *Mean Height Profiles*

The results for the two years with the most data are compared in Figure 5 with a recent model from (Groves 1969) for the mean July winds. It will be seen that the expected shear in the zonal wind in particular is well reproduced in all curves, the drift magnitudes being slightly in excess of the model predictions. The meridional model is not as closely followed, the partial reflection drift curve for 1971 and both the drift and meteor profiles for 1972 indicating that the model may overestimate the meridional component in much of the 80–100 km height range. On the other hand, the discrepancy may simply indicate year-to-year variations in the winds (Elford and McAvaney 1971), since the model has been compiled from data taken over a number of years.

In view of the fact that Rossiter (1971) found the drifts for the hours 1000–1400 to be consistently more eastward than those predicted by the model for heights between 80 and 100 km, profiles for the three years were recalculated including only information obtained between these hours (Fig. 6). The resultant zonal profiles lie considerably more to the east than the profile taken from the model, especially near 90 km, which is very similar to the findings of Rossiter, his curves for 1969 being included in Figure 6 for comparison. It seems quite likely that the apparent anomaly in his drift profiles was due to results only being available for the middle of the day.

IV. Discussion

In many ways the present results are not inconsistent with the findings of Rossiter (1971). As noted above, the mean height profiles for the hours 1000–1400 are in good agreement with his winter results, even when the profiles for the full 24 hr data conform well with the model values. Although the preceding comparisons have shown that on the average there is good agreement between the drift and meteor observations, it is nevertheless true that there are often short-term differences, with the meteor values being generally the more variable. There is less variability in the summer and autumn periods and consequent better agreement between the winds deduced by the drift and meteor techniques (Stubbs 1973). However, it is well known that the wind fields in the upper atmosphere are considerably more subject to variation in winter than in other seasons (Murgatroyd 1971). The question remains therefore as to whether the drift technique is inherently incapable of detecting the short-term variations in the wind fields or whether the meteor wind results are affected to a much greater extent by the fluctuations.

Part of the variability in the upper atmosphere arises from the enhanced amplitude of internal gravity waves in winter. Rocket measurements of mesospheric temperatures at Wallops Island (38° N.) show that temperature fluctuations during this season are usually much larger than in summer (Nordberg *et al.* 1965), the fluctuations having peak-to-peak amplitudes of the order of 20% and vertical wavelengths of ~ 10 km. These short-term variations appear to be caused by gravity waves with periods of $\gtrsim 1$ hr (Theon 1968; Theon *et al.* 1969). The vertical scales agree well with those of similar fluctuations in the wind as measured by rocket-laid vapour trails (Kochanski 1964); the wave amplitudes as deduced by Kochanski were tens of metres per second.

It is suggested here that the winds determined by the meteor technique are the ones that are most influenced by the short-term quasi-periodic components described above. This is because successive trails observed with the Adelaide meteor equipment may be separated by horizontal distances of up to 400 km. Internal gravity waves with vertical wavelengths of the order of 10 km and periods of an hour or so will have very large horizontal wavelengths, typically of the order of hundreds of kilometres (Hines 1960) so that the gravity wave–wind systems will be slightly tilted. Since 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 will be inconsistent, any meteor wind determinations from only a few widely spaced meteors within a given time interval should be suspect.

By contrast, for the partial reflection experiment, the bulk of the returned power comes from within an area of radius 25 km and the wind fields are probably more homogeneous over this region. It is certainly true, however, that some smoothing of the wind profiles will result from the use of pulses of finite length (typically 4 km). Because the partial reflection process often involves scattering from a considerable volume, at least at the higher altitudes, it is probable that the returned power will not come uniformly from within the 5 km gate interval and this will result in a certain degree of averaging of the drift motions. Additionally, the 3 min record length may result in some averaging. However, this does not mean that short-term variations, such as might be caused by internal gravity waves, will not be observed. For example, in Figure 4 there is a pronounced quasi-periodic fluctuation in both the north–south and east–west drift components occurring between about 1100 and 1800 hr, the fluctuation amplitude being of the order of 20–30 m s^{-1} and the apparent period approximately 3 hr. It is planned to study such variations further by using somewhat shorter radar pulses and taking more frequent drift observations.

In view of Rossiter's (1971) findings, consideration was given to the presence of cold fronts and the extent to which their presence could be correlated with discrepancies between the two sets of results. In 1970 cold fronts were present on all days from 13 to 17 July, while in 1971 fronts were present only on days 21 and 22. The 1972 weather maps are harder to interpret since there appears to be a fairly rapid movement of fronts but, with the exception of day 10, cold fronts were present for some time on each of the days from 9 to 13. There appears to be no clear connection between the two sets of data. For example, 22 July 1971 is a day when there are a number of discrepancies between the two sets of results, while for 11 July 1972, when a cold front was also passing over Adelaide, there is excellent agreement. It is concluded that periods of disagreement between the two methods cannot be firmly associated with a particular cold front, although the presence of a large number of such fronts during the winter months probably contributes to increased wave activity at ionospheric heights.

The tidal behaviour of the drifts shows some interesting features that are worth considering. The most prominent tide occurred on 11 July 1972 and the results of a harmonic analysis for this day are shown in Table 1. The zonal and meridional components for the diurnal tide are not quoted separately as the tide was highly circular at all three heights, i.e. the amplitudes derived from the east–west and north–south components were similar with a phase difference of close to 6 hr. The tidal phase in Table 1 is the local time of maximum northward velocity. The salient feature of the results is the growth of the semidiurnal tide with height and the corresponding decrease in the diurnal tidal amplitude. The diurnal mode is nevertheless the dominant factor at all heights. Whilst the amplitudes are not as large on other days over the periods of observation, this general behaviour is observed in the majority of cases, the diurnal tide being usually the stronger but with the semidiurnal component increasing in importance with altitude. Similar winter behaviour has been reported by Roper (1966). He found from a spectral analysis of 1961 meteor observations taken at Adelaide that, whereas the diurnal tide was strongly dominant at 83 km and still the more important at 91 km, the semidiurnal tide had become the dominant feature of the spectrum at 97 km.

TABLE 1

TIDAL FEATURES OF DRIFT RESULTS FOR 11 JULY 1972

Height (km)	Diurnal tide amplitude ($m\,s^{-1}$)	Diurnal tide phase (hr)	Semidiurnal tide amplitude ($m\,s^{-1}$)	Prevailing ($m\,s^{-1}$) E.–W.	Prevailing ($m\,s^{-1}$) N.–S.
90–95	40	11·5	27	29	−17
85–90	51	13·4	16	40	−6
80–85	66	14·7	—	48	3

The phase of the diurnal tide shown in Table 1 is also of interest since it changes by only slightly more than 3 hr over a height range of about 10 km. If the vertical wavelength were constant with height, this would correspond to a vertical wavelength of the order of 70 km, which is much larger than the theoretically predicted value of about 25–30 km for the propagating (1, 1) mode at a latitude of 35° (Lindzen 1967). Similar calculations for the other days in July 1971 and 1972 show that on only one day (out of eight) is a wavelength as short as 30 km indicated. The overall mean vertical wavelength is ~ 50 km.

A substantial contribution from the evanescent $(-1, 1)$ mode may account for the larger than expected tidal wavelengths. It is possible, however, that the increase may be due to the propagating mode being severely dissipated at the altitudes of observation. Certainly, the marked decrease in amplitude with height on 11 July is in contrast with the expected behaviour of amplitude growth with height required to compensate for the density decrease with altitude, so that the wave energy density remains constant. When amplitude values u from Table 1 for the diurnal tide are combined with data for the density $\bar{\rho}$ taken from the U.S. Standard Atmosphere Supplement (1966) to form the energy density $E \sim \bar{\rho}u^2$, it is found that E decreases exponentially with height, having a "scale height" of 3·6 km (Fig. 7). Hines (1965) has shown that for an exponential decay of energy the vertical flux divergence is given by

$$dF/dz = -\rho u^2 V_g/2h_0\,,$$

where V_g is the vertical group velocity and h_0 the scale height. The rate ε of energy loss per unit mass is then

$$\varepsilon \sim u^2 V_g/2h_0\,.$$

Assuming the magnitude of the group velocity to be similar to the observed vertical phase velocity ($\sim 0\cdot 5$–$1\ \mathrm{m\,s^{-1}}$), the rate of energy loss at 90 km on 11 July is found to be $\sim 0\cdot 18$–$0\cdot 36\ \mathrm{W\,kg^{-1}}$. On only one day (22 July 1971) did the diurnal tidal energy density remain constant with height. On the other days the exponential scale height was $\sim 4\cdot 6 \pm 1$ km and the estimated average energy dissipation was $\sim 0\cdot 1\ \mathrm{W\,kg^{-1}}$.

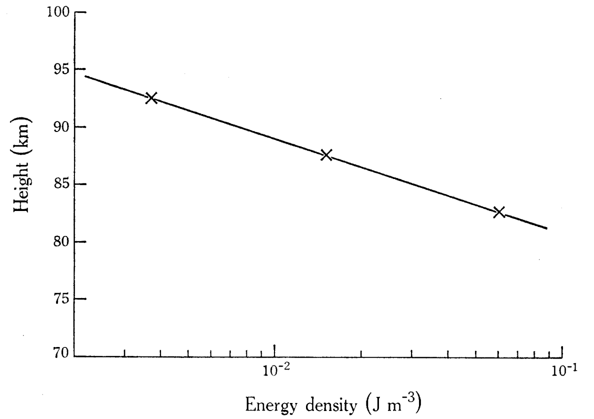

Fig. 7.—Energy density, plotted on a logarithmic scale, for three height intervals in the diurnal tide on 11 July 1972.

Although the above results must be treated with caution since they are taken on a limited number of days and over a relatively small height range (10–15 km), they indicate that the diurnal tide is a significant source of energy for the 90 km altitude region. Elford and Roper (1966) have shown that there is a marked correspondence between the strength of the diurnal tide and the strength of turbulence in the lower thermosphere as measured at Adelaide. The energy loss rates inferred above are quite comparable with other estimates made for the dissipation of internal gravity waves (Hines 1965; Zimmerman and Rosenberg 1972).

If the energy loss rate on 11 July is as large as is estimated then the energy is probably being dissipated by some form of turbulent mechanism. An increase in the eddy viscosity in the mesosphere and lower thermosphere might account for the very much smaller amplitude of the diurnal tide on days succeeding 11 July (see Fig. 3) since the amplitude of the tide is markedly dependent on the level of dissipation (Hines 1972). Alternatively, the variations in the amplitude could be due to corresponding changes in the source region (e.g. in the water vapour content of the troposphere) or in the temperature structure of the lower atmosphere (Lindzen 1968).

More quantitative examples of the variability of the diurnal tide are included in Table 2, which shows the amplitude and phase for the altitude range 85–90 km in July 1971. The tidal wind field is somewhat elliptical on all days but both components indicate day-to-day phase changes of up to 6 hr and amplitude differences of the order of 30 m s^{-1}.

TABLE 2

PARAMETERS OF DIURNAL TIDE AT 85–90 KM FOR 20–23 JULY 1971

The phase is the time of maximum eastward and northward velocity respectively

Day	E.–W. component amplitude (m s^{-1})	E.–W. component phase (hr)	N.–S. component amplitude (m s^{-1})	N.–S. component phase (hr)
July 20	24	14·7	11	20·3
21	51	18·5	41	23·9
22	21	12·3	34	17·1
23	34	10·4	29	14·7

V. Conclusions

This work has shown that even in winter the partial reflection drifts can adequately represent both the diurnal and monthly mean behaviour of the wind fields in the mesosphere and lower thermosphere. The comparison with the meteor wind estimates has indicated the influence of small-scale wave activity but in general the drifts vary in a much smoother fashion. It is clear that, if drift readings are obtained with sufficient density, both tidal and gravity wave effects can be isolated.

VI. Acknowledgments

The helpful comments of Dr. B. H. Briggs and Dr. W. G. Elford are gratefully acknowledged. This work forms part of a program of observations supported by the Australian Research Grants Committee and the Radio Research Board. One of us (T.J.S.) received support from a CSIRO Postgraduate Scholarship.

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