Atmospheric Tidal Motions over Australia below 20 Kilometers

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ABSTRACT

The diurnal and semidiurnal variations of the wind have been computed from 12 years of upper wind observations at Australian stations. The strong diurnal oscillation in the boundary layer over the continent is remarkably constant in phase and appears to interact with the sea breeze oscillations at the coast. Above a transition layer of varying thickness, a layer of roughly constant phase extends through much of the troposphere. Vertically propagating modes dominate the diurnal tide above 16 km, transferring energy upward.

In contrast to the diurnal tide, the semidiurnal tide (except in the boundary layer) is almost independent of height up to at least 13 km, also of season and geographical position.

1. Introduction

Atmospheric tides of solar diurnal and semidiurnal frequency (24 and 12 h period) are excited by daily variations in solar heating. Tidal theory (reviewed by Lindzen and Chapman, 1969, and Lindzen, 1971) predicts sets of modes for each frequency and zonal wavenumber, each mode with its characteristic distribution of excitation and resultant vertical and meridional variation. In the case of the semidiurnal tide, the excitation by radiative heating and cooling of ozone and water vapor is dominated by a single mode of global meridional scale which has a very large vertical wavelength (200 km), so that all the excitation in the mode below 100 km acts in phase. The solar semidiurnal tide should therefore be strong near the surface and very regular. The excitation of the diurnal tide by O_3 and H_2O, however, involves several different modes with comparable amplitudes. Some are "trapped" and others vertically propagating, with vertical wavelengths of 25, 12, and 7 km, so the thick O_3 source region leads to destructive interference, and excitation by surface heating assumes much greater relative importance. Furthermore, the modes are all of comparatively small meridional scale and are therefore sensitive to local variations, with the result that the diurnal tide may be expected to be comparatively irregular.

Indeed, when observations of the diurnal wind variation from many stations in the United States were compared (Hering and Borden, 1962), strong vertical variations of amplitude and phase were found, with large-scale patterns in the horizontal apparently related to topography. Studies of a larger area (Wallace and Hartranft, 1969; Wallace and Tadd, 1974) showed that topographic effects extended upward to at least 28 km.

Possible mechanisms for the topographic influence include forcing by thermal effects of land-sea contrast, terrain slope, and contrast in cloud cover or vegetation type, and also frictional effects. Since tidal theory does not yet incorporate such longitudinal variations of forcing, their relative importance is not yet established. However, because of the complexity of the diurnal tide, in studying such variations the need for the greatest possible data coverage is evident.

The present investigation aims to determine the diurnal wind variation over Australia as well as it can be deduced from available upper wind data, and to present an overall picture of its dependence on geographical position, height, and season. The diurnal variation within the boundary layer, as well as above it, is described since, besides being interesting for its own sake, the boundary layer is apparently a major source of excitation for the diurnal tide much higher up (Wallace and Tadd, 1974). Finally, the semidiurnal wind variation is also discussed.

2. Data and method

At most Australian upper wind stations, soundings are taken four times daily, at 0500, 1100, 1700, and 2300 GMT (sometimes 1 h earlier or later). The 2300 GMT observations, normally carried out using a radiosonde ascent, often reach 30 km, while the smaller balloons used at other times are rarely tracked as high as 18 km. Even at stations equipped with radar, it has often been the practice to carry out the 0500 GMT sounding by visual observation of a pilot balloon. It was found in a pilot study that above 14 km the 0500 GMT mean wind generally diverged sharply from the other three means, probably because the tendency to lose

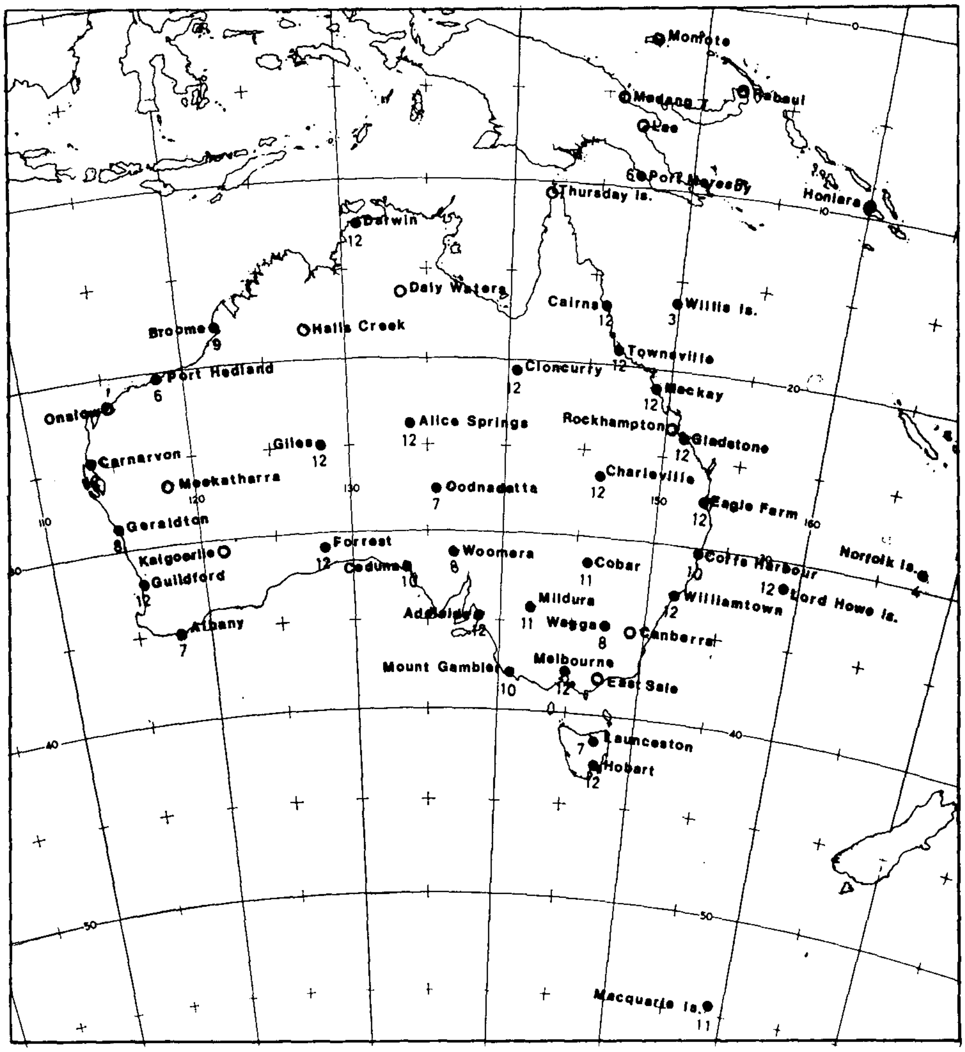

FIG. 1. Location of upper wind stations used in this study. The filled circles denote radar stations, with the length of radar record in years. There are two stations at Melbourne, Essendon (non-radar) and Laverton (radar).

the balloon in strong jet-stream winds biased the mean in favor of light winds. It was also found that, except near the surface, the amplitude of the diurnal variation rarely much exceeded 1 m s^{-1}, a small fraction of the total wind at many levels and of its day-to-day variability, and sometimes smaller than the difference between one 6 yr mean and another. Precautions must therefore be taken to ensure sufficient homogeneity of the observations; otherwise the estimates of the tidal variations may be seriously biased. At the same time, one needs a large set of observations to obtain significant results.

The data available for the final study consisted of all Australian upper wind observations from land stations from 1962 to late 1973, approximately 8×10^5 soundings, on magnetic tapes. The stations are shown in Fig. 1. Three different analyses were made, for use at different heights.

a. Surface to 3 km

In the boundary layer, diurnal variations are large and the observations virtually complete at each station, so no problem of bias arises. Monthly means were computed from all the observations at each of the four hours, regardless of type. A check was imposed on the vector shear between levels, to eliminate mispunched data. The monthly means were then averaged in threes to obtain seasonal means, which were plotted as hodographs, initially of the total velocities (Fig. 2), and later of the diurnal component alone with the overall mean and the semidiurnal component removed (Fig. 3). This analysis was performed for the entire 12 yr period and in some cases also for 6 yr and 3 yr periods separately.

The seasonal mean wind at one of the six-hourly observing times t, $\mathbf{V}(t)=u(t)+iv(t)$, where u, v are the

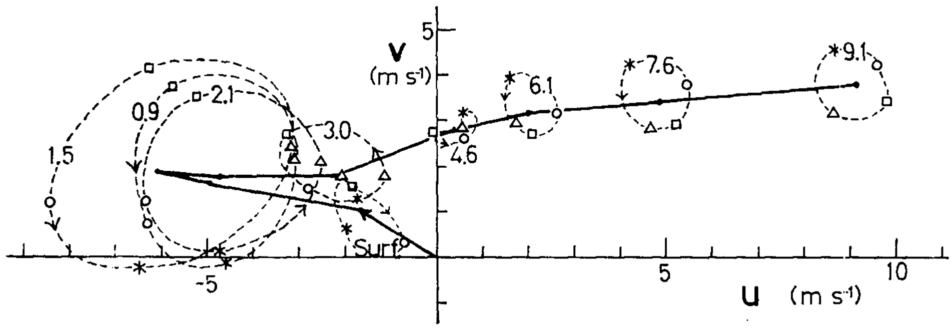

FIG. 2. Hodographs of the mean wind at Alice Springs (elevation 545 m) in December–February, 1962–73. Each dashed circle shows the daily variation at a height which is given in kilometers above mean sea level (△ 0500 GMT, □ 1100, ○ 1700, ∗ 2300). The thick line joining the centers of the circles indicates the diurnal mean.

eastward and northward components, may be written

$$\mathbf{V}(t)=\mathbf{V}_m+\mathbf{V}_{da}\exp(i\omega t)+\mathbf{V}_{dc}\times\exp(-i\omega t)+\mathbf{V}_{sd}\exp(2i\omega t), \quad (1)$$

where ω is the diurnal angular frequency and the terms are respectively the mean, diurnal anticlockwise and clockwise circular rotations, and the semidiurnal variation. Since in the Southern Hemisphere the diurnal variation is usually close to a uniform anticlockwise rotation, convenient parameters to describe the diurnal variation are the amplitude and phase of $\mathbf{V}_{da}$, which is given by

$$\mathbf{V}_{da}=\tfrac{1}{4}[\mathbf{V}(2300)-\mathbf{V}(1100)+i(\mathbf{V}(1700)-\mathbf{V}(0500))]. \quad (2)$$

The phase was expressed as the local time (LT), t_{da} hours, at which the component was northward, i.e.,

$$t_{da}=\frac{12}{\pi}\arctan(u_{da}/v_{da})-1+\lambda+r, \quad (3)$$

where λ is the longitude in hours and r the rise-time of

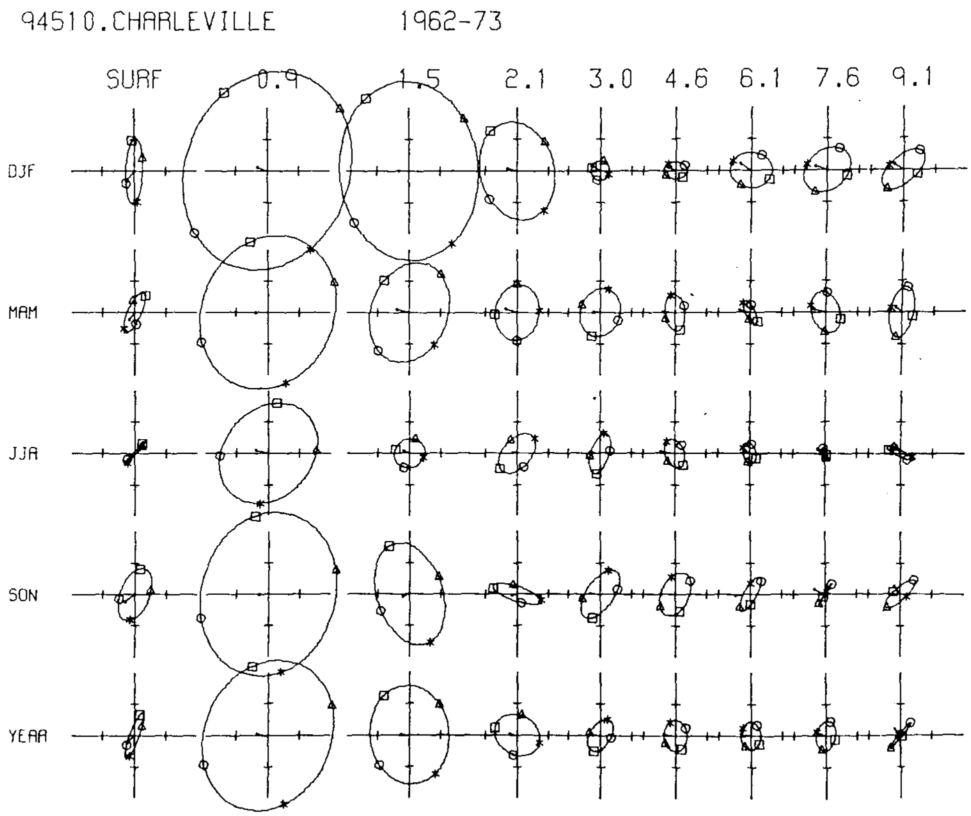

FIG. 3. Hodographs of the diurnal wind variation at Charleville (elevation 304 m). Symbols and orientation of axes as in Fig. 2. The vectors are the semidiurnal component $\mathbf{V}_{sd}$ as defined in (4). The pips on each set of axes are at ± 1 m s^{-1}. The number at the top of each column is the height in kilometers.

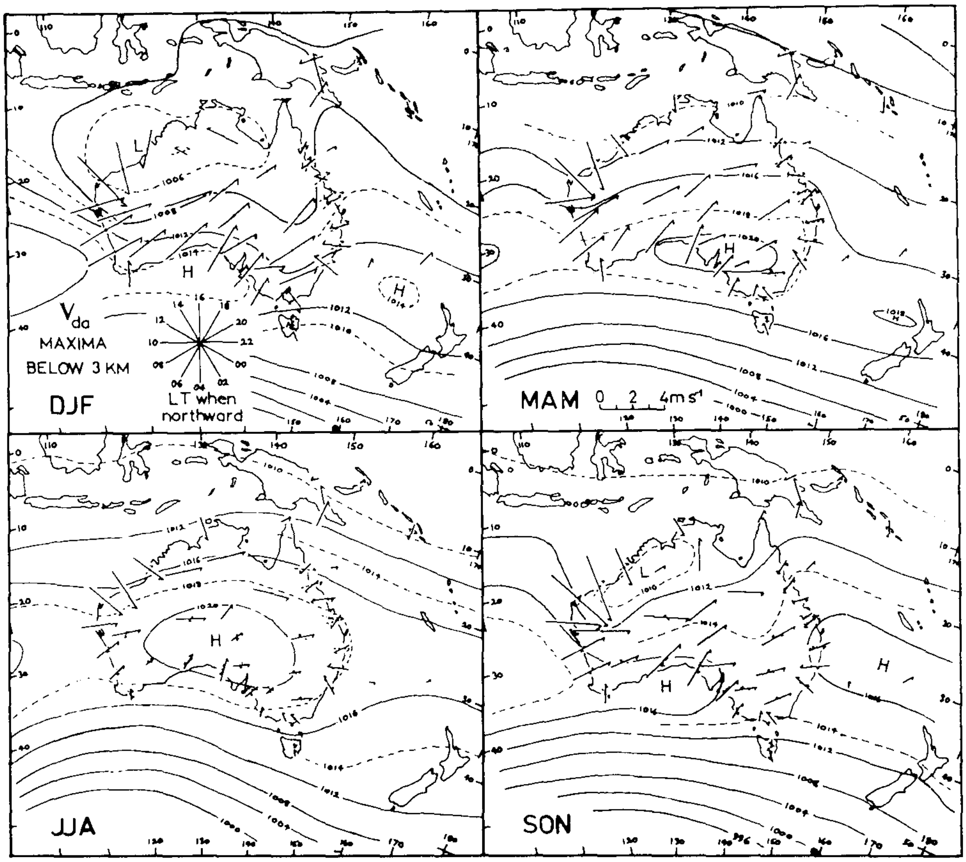

FIG. 4. The anticlockwise diurnal wind component at 1600 LT at its lowest maximum of amplitude (solid vectors) in 1962–73. The dashed vectors represent second maxima below 3 km. Climatological charts of mean sea level pressure for the mid-season months are superimposed, after Taljaard *et al.* (1969).

the balloon, which was assumed to be released 15 min before the hour and to rise at 5.6 m s^{-1}.

The semidiurnal component,

$$\mathbf{V}_{sd}=\tfrac{1}{4}[-\mathbf{V}(0500)+\mathbf{V}(1100)-\mathbf{V}(1700)+\mathbf{V}(2300)], \quad (4)$$

was similarly expressed in amplitude-phase form, but was also plotted as a vector on the hodographs illustrated in Fig. 3.

b. 3 to 15 km

The first analysis begins to break down at 3 km because many pilot balloon soundings terminate there, resulting in biased mean winds higher up. A second analysis was therefore made by the same method but with only radar-observed winds included in the means. Because radar observations at 0500 GMT were comparatively rare during the early half of the data period, and several stations only commenced radar observations after 1962, for the sake of homogeneity the data sample was restricted to the six years 1968 to 1973. Hodographs of the diurnal component were plotted and the amplitudes and phases of $\mathbf{V}_{da}$ and $\mathbf{V}_{sd}$ computed as above.

c. Up to 20 km

With increasing height it becomes progressively rarer to find an adequate number of observations at all four observing times, but at a few stations the 12 h difference vector method employed by Wallace and Hartranft (1969) was usable up to 20 km. The vector differences between the mean winds at 1100 and 2300 GMT were computed for each month and averaged to obtain seasonal means. All radar observations from 1962 to 1973 were included, except those at stations which began radar observations after 1962; 2300 GMT observations made before 1100 GMT observations began were discarded. Stations with less than 3 years of radar record at both times were not analyzed. To try to eliminate a strong bias effect associated with the subtropical jet, monthly means were discarded if less than about 10% of soundings reached the height concerned. This simple criterion was not completely successful in separating out the biased from the satisfactory means, and a more sophisticated way of selecting the observations could be better.

The 12 h difference vector provides an estimate of the amplitude and phase of the diurnal tide if the latter is

TABLE 1. The boundary layer maximum of the diurnal wind variation over the Australian continent.

	DJF	MAM	JJA	SON
Mean phase* (LT when northward)	1947±15 min	1947±17 min	2028±28 min	2038±22 min
Mean height of maximum above surface* (m)	590	490	320	460
Mean amplitude of maximum** (m s^{-1})	2.84±0.16	2.01±0.20	1.17±0.14	2.29±0.19

* 12 stations, excluding Halls Creek and Daly Waters.
** 11 stations south of 23°S.

assumed to be a uniform anticlockwise rotation, i.e., if $\mathbf{V}_{dc}$ is assumed to be negligible compared with $\mathbf{V}_{da}$.

The hodographs such as Fig. 3 suggest that V_{da} does generally dominate in the region of observation, and is therefore likely to do so also at the heights where we use analysis (c). (An exception is Hobart, where the clockwise component appears to dominate from 3 to 11 km.) Outside the boundary layer, the observed deviations from uniform anticlockwise rotation have practically no consistency from station to station. Formal error estimates, computed from the observed wind variances or, in the case of annual means, from the scatter of the monthly means also, confirm that in general the observed $\mathbf{V}_{dc}$ is not significantly different from zero except in the boundary layer, where the diurnal variation is larger and the day-to-day variation smaller. Since many of our measurements of $\mathbf{V}_{da}$ with the same errors as $\mathbf{V}_{dc}$, certainly are significant, ignoring $\mathbf{V}_{dc}$ is justified.

3. The diurnal wind variation below 3 km

At almost all stations strong diurnal variation of the wind is observed near the surface. The anticlockwise

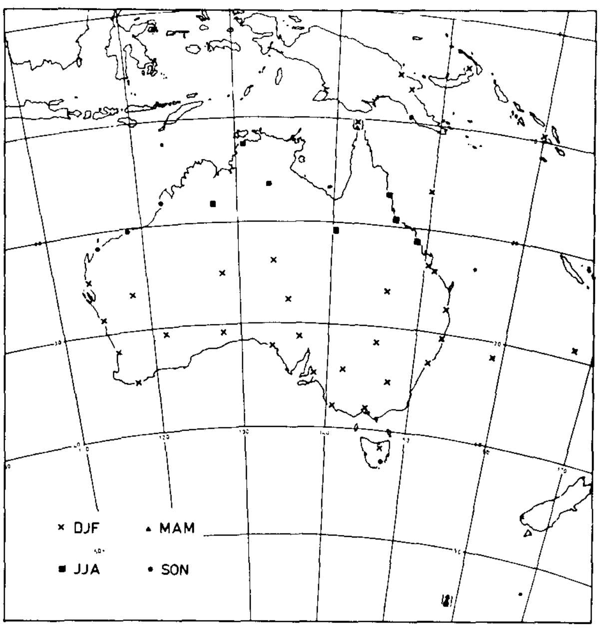

FIG. 5. Season of maximum amplitude of the boundary layer diurnal oscillation. The low-level sea-breeze oscillation on the east coast is ignored.

component at its maximum amplitude is plotted in Fig. 4, for the four seasons December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON). When a second maximum also occurs below 3 km, this is shown dashed in Fig. 4. The vectors are $\mathbf{V}_{da}\exp(i\omega t)$ at 1600 LT, and can be read as harmonic dial vectors giving the time at which the component is northward. The oscillation is strongly influenced by topography and land-sea contrast and will be discussed in appropriate groups of stations.

a. The inland stations

There are 14 stations on the continent more than 100 km from the coast. While the diurnal variation at the surface is affected by even slight terrain slope, almost all show a very similar, large diurnal variation in the lowest 1 to 3 km above the ground, throughout the year. This variation reaches a maximum amplitude of 0.6–3.6 m s^{-1} at a few hundred meters, where it is fairly close to a uniform anticlockwise rotation, northward at 2000 LT. The only exceptions are the two northernmost inland stations where in DJF the phase below 1.5 km is 1200 LT. Fig. 2, a typical set of hodographs for a central Australian station, shows that the phenomenon gives rise to a low-level maximum of wind speed at night, sometimes called the nocturnal jet.

The boundary layer oscillation (BLO) varies greatly with season. South of 23°S it is strongest in summer (DJF) (Fig. 5), but north of 23°S it is strongest in JJA, probably because the "monsoonal" cloudiness in summer reduces the solar radiation reaching the surface and the greater soil moisture content increases the heat capacity of the ground, both these effects tending to reduce the amplitude of diurnal variation. Fig. 3 clearly shows the summer maximum of amplitude, and also a seasonal change in vertical scale; in summer the maximum is higher up and the BLO is present with almost constant phase up to 3 km compared with only 1.5 km or so in the other seasons. This variation appears to correspond to the variations in height of the mixed layer. These changes are summarized in Table 1, which also shows that there is negligible seasonal variation of the average phase of the BLO. The errors in this table are rms errors of the means, estimated from the observed scatter. The amplitude and height seasonal changes become stronger with increasing latitude until for some stations the winter BLO is so shallow as to be virtually unresolved in height by the rather coarsely-spaced observations. Because of the resolution difficulty, the heights and amplitudes in Table 1, estimated by graphical interpolation in the vertical, can be only qualitatively correct.

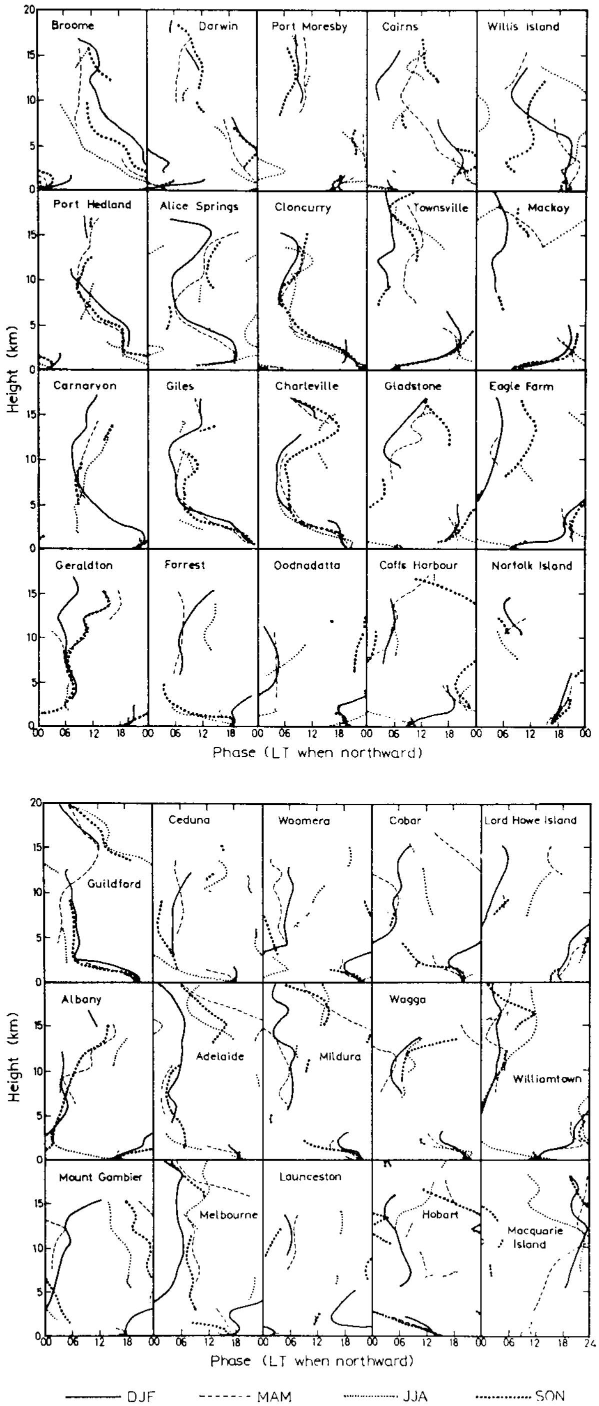

FIG. 6. Phase of the anticlockwise diurnal wind component as a function of height for the four seasons (not plotted when amplitude <0.25 m s^{-1}).

A similar nocturnal low-level jet is very well developed in the Great Plains region of the USA. Buajitti and Blackadar (1957) attempted to explain it as an inertial oscillation forced by the diurnal variation of friction—the low eddy viscosity at night giving rise to high winds at a few hundred meters. However, they could not account for the observed amplitude and phase. Later attempts at modeling (Estoque, 1963; Holton, 1967; Krishna, 1968; Bonner and Paegle, 1970; Sheih, 1972) showed that another forcing mechanism, rejected by Buajitti and Blackadar, was also important, namely diurnal variation of the geostrophic wind due to the thermal effects of terrain slope. Bonner and Paegle used observations of the geostrophic wind to show that the combination of the two mechanisms adequately explained the observed wind.

The frictional forcing mechanism gives rise to a diurnal wind vector opposing the mean geostrophic wind in the afternoon, while the thermal forcing leads to upslope motion in the afternoon. Comparing the mean sea level pressure charts with the afternoon diurnal wind vectors in Fig. 4, we see that the constancy of phase of the BLO over most of Australia is sometimes hard to account for. North of 23°S, and also at Meekatharra, Giles, Alice Springs, and Oodnadatta, frictional forcing can probably explain the observed phase. Elsewhere, drastic seasonal changes in the direction of the geostrophic wind produce little response in the diurnal wind vectors. At Woomera, west of the Flinders Ranges, and especially at Charleville, Cobar, and Wagga, where the ground slopes gently up to the "Great Dividing Range" to the east, it may well be that thermal forcing is important, as in the Great Plains. This mechanism, however, can hardly dominate the winter diurnal wind variation at Mildura, which is in the center of a flat plain with no topography in excess of 100 m for 200 km around. At Forrest, 100 km from the sea, sea breezes contribute to the diurnal variation in summer (Clarke, 1955) but are not likely to control it in winter. It is plausible that the discrepancy between the mean geostrophic wind direction and the phase for all the southern inland stations (Kalgoorlie, Forrest, Woomera, and Mildura) in winter is due to correlation between the geostrophic wind direction and the strength of the frictional forcing. That is, on days when the geostrophic wind is in the climatological direction (westerly) it is likely to be cloudy so the diurnal variation is weak.

b. The coastal stations

Fig. 4 shows that at 1600 LT the diurnal wind vector at its height of maximum amplitude is always onshore at coastal stations. We shall call a diurnal wind variation with this phase a "sea-breeze oscillation," without seeking to imply that a frontal process is necessarily involved. Frontal sea breezes, of course, are one of the most conspicuous features of the climate in the well-populated coastal regions and have received

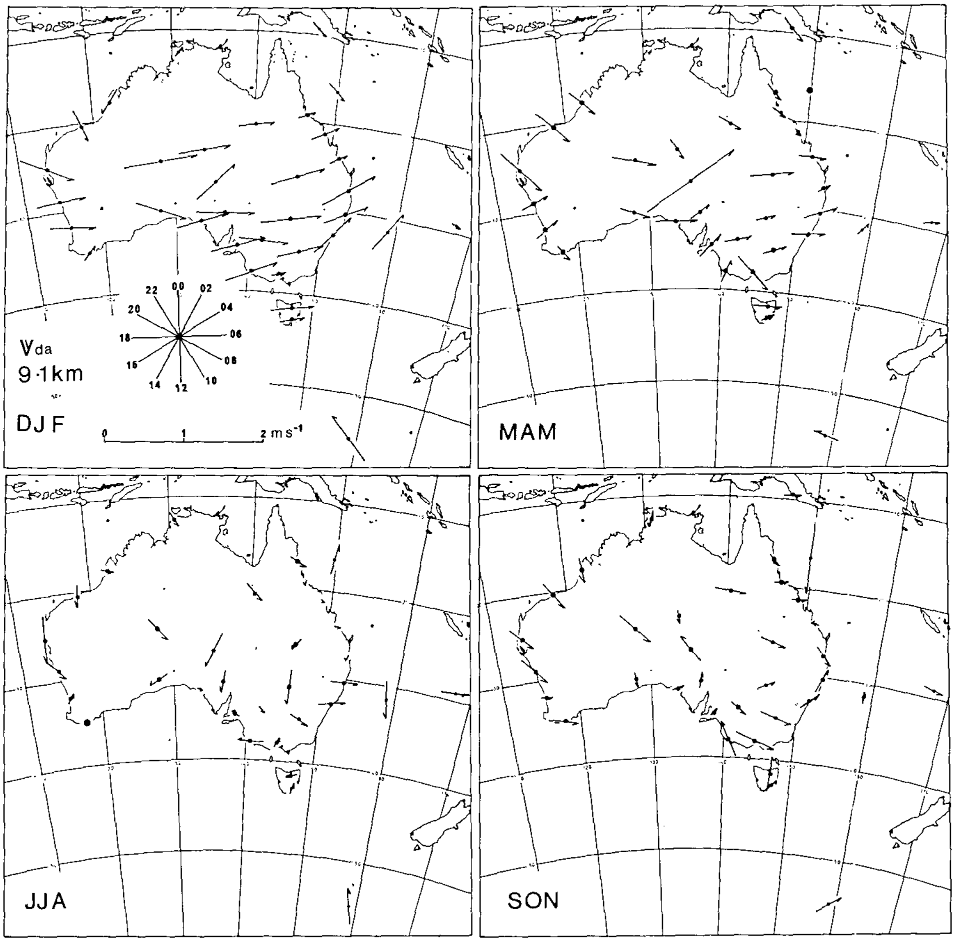

FIG. 7. Harmonic dial vectors representing the anticlockwise diurnal wind component at 0000 LT, averaged over 7.6, 9.1, and 10.7 km, 1968–73.

considerable observational study, which has been reviewed by Gentilli (1971). Some effort has also been devoted to modeling them (Neumann and Mahrer, 1971; Clarke, 1973; Walsh, 1974).

The present results indicate that the structure of the mean diurnal wind variation over coastal stations is determined sometimes as high as 5 km by the interaction of the sea-breeze oscillation and the frictionally-forced BLO.

If the geostrophic wind is offshore, then both oscillations act in phase, blending together to produce a diurnal variation almost indistinguishable in our climatological data from that observed inland. The only noticeable difference is that the variation at the coastal station is elongated from a circle into an ellipse perpendicular to the shore. Even this effect is not always conspicuous. Such blending occurs at the west coast of Western Australia, long famous for the strength and regularity of its sea breezes, the northwest coast, where the sea-breeze oscillation is even stronger, Port Moresby in Papua, New Guinea, and part of the South Australian coast.

On the east coast of Queensland, where the geostrophic wind (the trade) is onshore, the two oscillations oppose each other. At the surface and 0.3 km, sea-breeze phase is observed. The diurnal variation ellipse is highly elongated, in contrast to the open ellipses observed on the west coast. At 0.9 km and above, the diurnal variation has the phase of the continental BLO and is comparatively circular. This BLO has a maximum amplitude of about 1 m s^{-1} near 2 km, and sometimes extends upward to 5 km. The rest of the east coast is similar, the sea-breeze oscillation being dominant near the surface and a BLO of roughly opposite phase above 0.5–1.5 km. The amplitude of the east coast sea-breeze oscillation varies little with season, but the BLO varies the same way as inland (Fig. 5).

As Clarke's investigations showed (Gentilli, 1971), the propagation of the sea breeze inland depends strongly on local topography. In our data, we observe a sea-breeze oscillation at Rockhampton, 30 km inland on the coastal plain, with about half the coastal amplitude at the surface but falling off much more rapidly with height. At Launceston, Tasmania, on the other

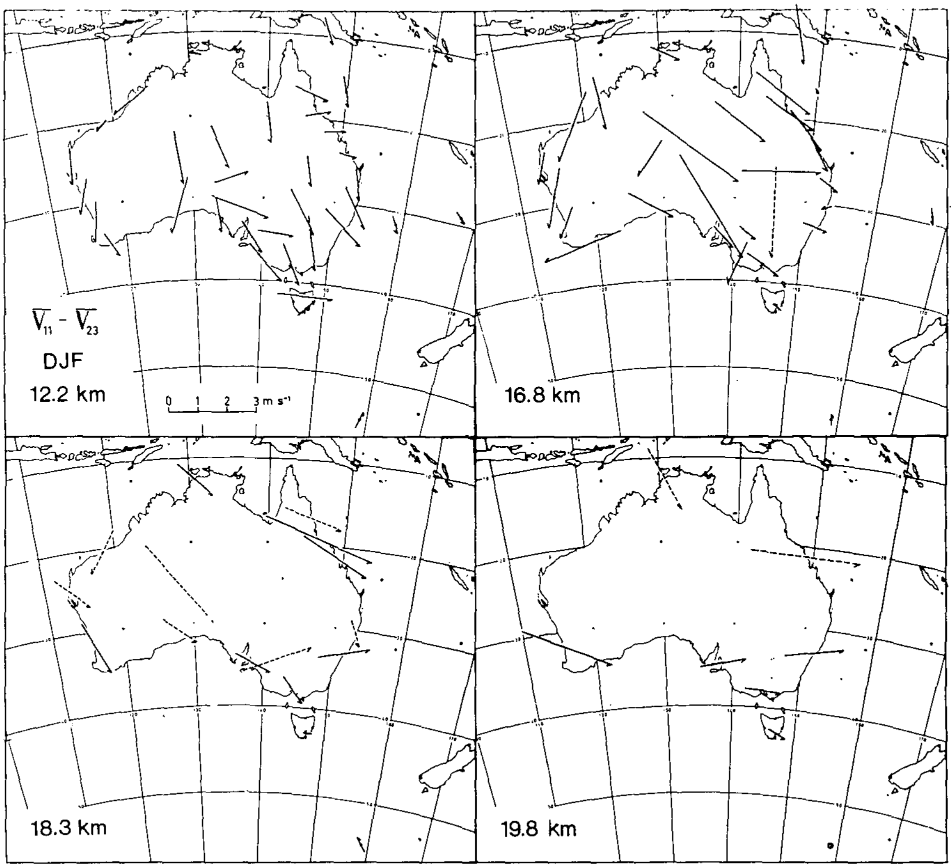

Fig. 8. Twelve-hour difference vectors (1100–2300 GMT) for summer (DJF) at 12.2, 16.8, 18.3, and 19.8 km, 1962–73. Dashed vectors are marginally below the 10% completeness limit at 1100 GMT.

hand, 50 km up a valley at an elevation of 166 m, the sea-breeze oscillation extends upward to 1 km, except in winter.

The extreme topographic influence at Lae in New Guinea should also be mentioned. This station lies between two high mountain ranges which force the diurnal wind variation to be practically a straight line up-and-down the valley up to almost 3 km. The oscillation has sea-breeze phase at the surface, gradually becoming later with increasing height.

c. *Small islands*

The diurnal oscillation at low levels over small islands is weak and often lacks a clear maximum of amplitude in the boundary layer. The phase in the lowest 3 km is about 1800 LT at Lord Howe and Norfolk Islands, and 2000 LT at Willis and Thursday Islands.

d. *Secondary maxima below 3 km*

At many stations, both coastal and inland, second maxima of amplitude with different phase occur below 3 km. These are shown as dashed vectors in Fig. 4. Such a secondary oscillation occurs over the southwestern part of the continent in JJA and SON at average heights of 1.8 and 2.3 km respectively. It is evident from the plots of phase as a function of height (Fig. 6) that it is the lower extremity of an oscillation which extends with fairly constant phase through much of the troposphere as described below.

4. The diurnal wind variation in the free atmosphere

Fig. 6 shows the variation with height of the phase of the anticlockwise diurnal component of the wind, except where the amplitude is negligible (<0.25 m s^{-1}). In general, above the boundary layer the phase first changes rapidly, then settles down to a fairly constant value. The altitude where the phase stops changing rapidly is greatest at low latitudes. South of about 26°S, the phase transition often occupies only 1 km; north of 26°S, however, most stations show a steady progression to earlier phase with increasing height, for some 2 to 5 km. Wallace and Hartranft (1969), while they did not observe a constant phase layer, observed the phase progression at low and middle latitudes and pointed out that it implies an upward

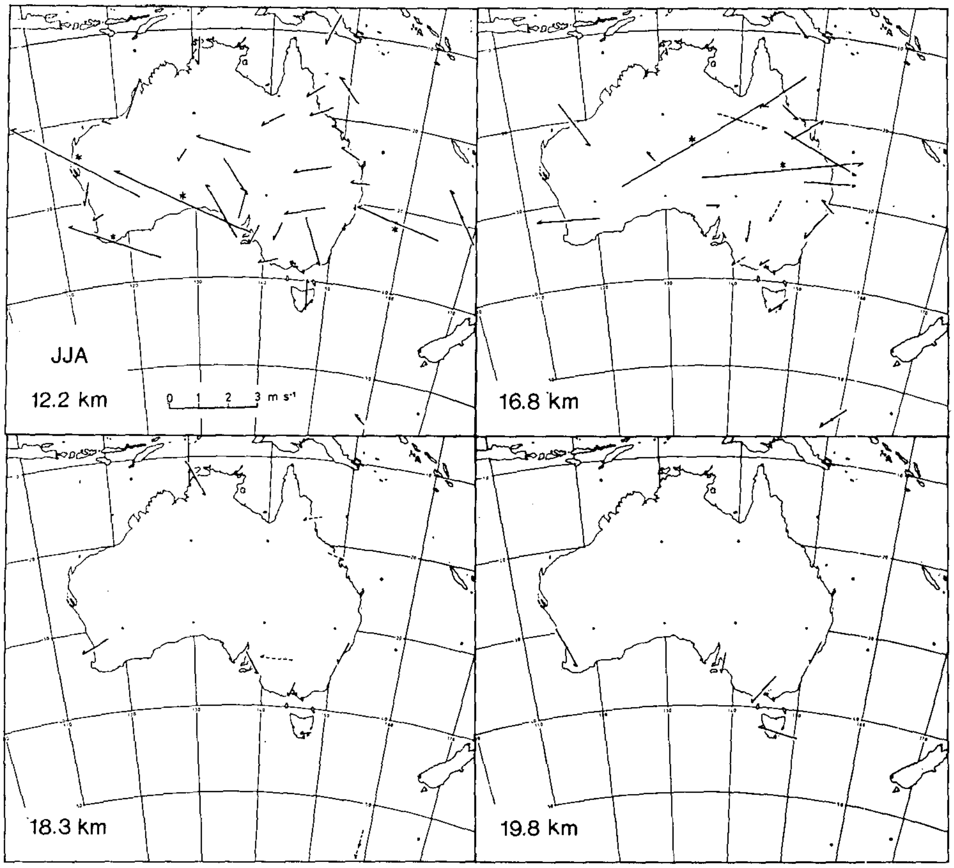

FIG. 9. As in Fig. 8 but for winter (JJA). See text.

flux of energy from the boundary layer due to the diurnal cycle.

Exceptions to this pattern are the east coast stations from Gladstone south, Lord Howe and Norfold Islands, and Albany, which all show a steady *increase* of phase with height for 5 or 10 km in the troposphere, apparently implying a downward energy flux.

The diurnal oscillation in the layer of comparatively constant phase is illustrated in Fig. 7 by harmonic dial vectors computed from the 1968–73 radar data. To reduce the noise, the vectors have been averaged over three adjacent levels. South of the tropics, there is considerable seasonal variation, the diurnal variation being strongest and most uniform in summer (DJF). In DJF, the mean phase for the 21 continental stations south of the tropics is 0500±20 min LT and there is a slight trend to earlier phases on the east coast and later on the west coast. Tasmania is in the same tidal régime as the mainland, but Macquarie Island (phase 20.9h) is in the high-latitude régime.

The variations of phase between 10 and 15 km fall into four main zones: i) Macquarie Island: phase about 2200 LT; ii) southeastern Australia, Tasmania, Lord Howe and Norfolk Islands: phase about 0600 LT; iii) the remainder of the continent south of 19°S: phase increases by about 6 h between 10 and 15 km, averaging about 0930 LT; and iv) the far north: phase about 0930 LT. In this height range, therefore, what systematic changes of phase with height are observed imply a downward flux of energy.

Twelve-hour difference vectors for the levels up to 20 km are given in Figs. 8 and 9. The large amplitudes shown for certain stations at 12.2 and 16.8 km in JJA appear to result from the observational bias mentioned above. The stations for which 19.8 km difference vectors are shown may be assumed to be free from this effect. Anticlockwise rotation of the difference vectors with increasing height (implying an upward flux of energy) is generally apparent above 16 km, but not always above 200 mb (about 12 km) as Wallace and Hartranft reported. There are large phase variations between the two seasons investigated. Wallace and Tadd (1974) found similar large seasonal changes at the 100 mb level and above.

5. The semidiurnal wind variation

The semidiurnal component $\mathbf{V}_{sd}$, although only about 0.3 m s^{-1} in amplitude, proved to be easily detectable at most stations. In accordance with the theoretical prediction, it is very much more uniform than the diurnal component $\mathbf{V}_{da}$. Some deviations from the regular amplitude and phase occur in the lowest 2 km. Apart from these, no significant variations of phase with altitude, season or geographical location were found. Because of the lack of seasonal dependence seen in charts such as Fig. 3 and in the results of Wallace and Tadd (1974), only annual mean data were analyzed in detail.

The V_{sd} vectors at 3 km (Fig. 10) are clearly consistent with a uniform phase of 0630 LT and an amplitude of 0.3 m s^{-1}. Some of the scatter of phase seen in this figure is due to occasional shifts of observing schedules 1 h earlier or later; it was not considered worthwhile to attempt to correct for this. Estimating a vertical mean phase at each radar station from 2 km to the maximum height at which the measured phase was still within 1.5 h of the mean, i.e., 8 to 17 km, the overall mean phase was 0632±5 min LT, the standard devation less than 0.5 h. (Two stations where the phase varied too much to define a mean were omitted.) This is consistent with the observations of Harris *et al.* (1962, 1966), Finger *et al.* (1965) and Wallace and Tadd (1974), and is definitely later than the theoretical predictions of 0540 (Lindzen, 1971) or 0600 (Lindzen and Hong, 1974). Lindzen points out that a similar unexplained phase discrepancy occurs in the solar semidiurnal variation of surface pressure, the best established of all atmospheric tidal elements. Lindzen and Blake (1971) suggest that it probably arises because the theory ignores daily variations of the O_3 distribution.

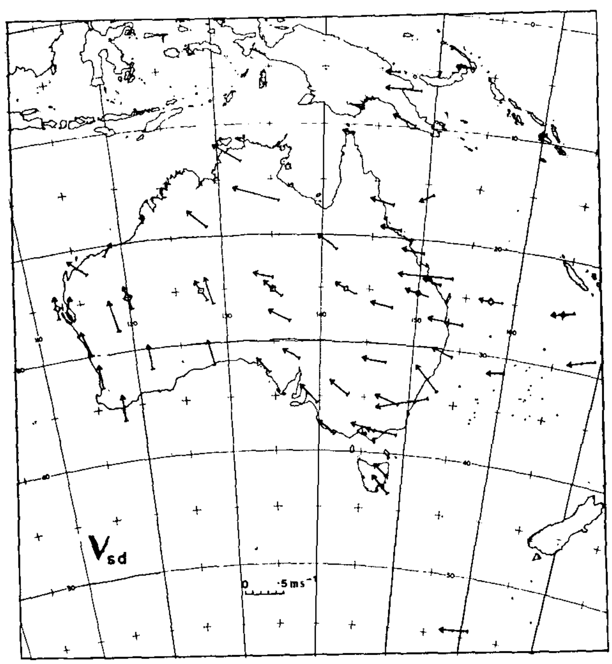

FIG. 10. Semidiurnal wind component vectors $\mathbf{V}_{sd}$, as defined in (4), at 3.0 km. Annual mean, 1962–73. Vectors with a phase of 0630 LT and amplitude of 0.3 m s^{-1} are shown along 25°S for comparison, distinguished by diamonds.

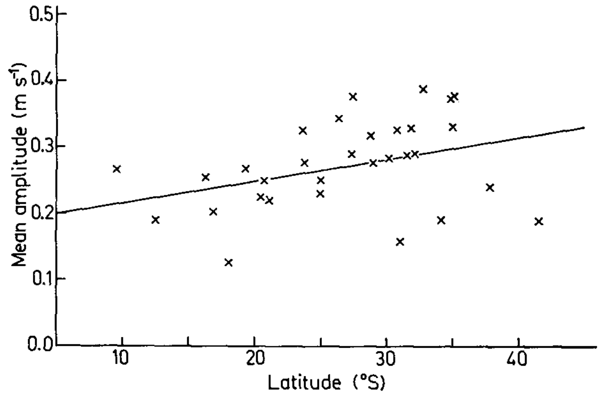

FIG. 11. Vertical mean amplitudes of $\mathbf{V}_{sd}$ from 3.0 to 9.1 km, plotted against latitude.

The vertical variation of amplitude is not a monotonic increase from the ground to 10 mb, as Wallace and Tadd imply. The present data indicate that the amplitude has a maximum, typically about 0.4 m s^{-1}, at or below 2 km, then *decreases* gently up to about 13 km. Here the mean amplitude appears to start increasing with height but is difficult to estimate without bias and with useful precision. The results of Harris *et al.* (1962, 1966) for stations at 30–40°N suggest a similar pattern, which is reasonably consistent with theoretical predictions (Lindzen and Hong, 1974) of constant amplitude or of a very shallow minimum in the region below 20 km.

If we estimate vertical mean amplitudes from 3 to 9 km, where the measurements are reliable and free from boundary-layer effects, a significant positive correlation with latitude is found (Fig. 11) (r=0.368 for 31 stations, 4 stations with clearly unsatisfactory data being omitted). This is to be expected theoretically (e.g., Lindzen, 1971), but for a quantitative comparison with theory it would be necessary to consider the zonal and meridional components separately.

The seasonal mean amplitudes at 3 km from all 39 stations with usable data at that height (Table 2) suggest slightly greater amplitude in winter, consistent with the predictions of Lindzen and Hong. The effect does not appear to be concentrated in any particular latitude zone.

6. Conclusions

In the boundary layer, strong diurnal wind variations are found, forced largely by the diurnal variation of

TABLE 2. Mean amplitude of the semidiurnal tide at 3 km.

Season	Amplitude (m s^{-1})
DJF	0.279±0.019 (rms)
MAM	0.296±0.020
JJA	0.319±0.015
SON	0.314±0.016

friction over land and by land-sea contrast at the coast. The continental oscillation, while its amplitude varies greatly with season at middle latitudes, has surprisingly constant phase, which at some stations cannot be explained by considering frictional forcing in the climatological mean situation alone. The diurnal variation to the west of the "Great Dividing Range" should be analyzed as a function of general synoptic situation to determine whether or not forcing by the thermal effects of terrain slope is also important there. On the east coast, where the continental boundary layer oscillation and the sea-breeze cycle are opposed in phase, the sea-breeze cycle is observed near the surface (almost constant with season) and the continental-type cycle above, extending to higher altitude than it does inland.

Above the boundary layer, the most widespread feature of the diurnal variation is a deep layer of rather constant phase, in the south typically extending from 1 km above the boundary layer to 10 km or more, in the north somewhat higher. The systematic phase changes below this layer (in the north) or above it (in the south) are usually such as to transport energy into it. Above 16 km, however, the dominance of vertically propagating modes, transferring energy upward, is generally apparent. Strong seasonal variation is evident.

Thus the main conclusions of recent studies, which qualitatively confirm the theory, are borne out. Quantitative comparison with theory still awaits extension of the latter to include realistic surface forcing.

The semidiurnal wind variation above the boundary layer up to about 15 km depends very little on height, season, or geographical position, except for slight changes in amplitude. The only significant disagreement with classical tidal theory is a phase discrepancy of about half an hour.

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