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Radar Measurement of Tidal Winds at Stratospheric Heights over Arecibo

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ABSTRACT

Wind oscillations of tidal periods that showed a marked downward phase progression were detected at the lower stratosphere using the Arecibo radar. The amplitudes of 1–5 m s^{-1} were inferred for both diurnal and semidiurnal components, much larger than the values predicted by the classical tidal theory. The vertical wavelengths inferred were also less than the theoretical values; ~5 km for the diurnal component and 2–9 km for the semidiurnal component.

1. Introduction

The absorption of solar radiation at the ground and in the earth's atmosphere sets up a thermal forcing which can drive atmospheric oscillations with periods which are integral fractions of a day. The classical theory of atmospheric thermal tides on a spherical planet has been reviewed by Chapman and Lindzen (1970), while more recent developments have been reviewed by Forbes and Garrett (1979) and Kato (1980). The classical tidal oscillations are global in scale, the diurnal tide undergoing one, the semidiurnal two, complete oscillations around the earth.

Observations in the upper mesosphere and lower thermosphere [see reviews by Evans (1978) and Lindzen (1979)] have shown the presence of tidal oscillations which agree in their general characteristics with the theoretical predictions, although both theory and observation still need refinement.

There is strong evidence from rawinsonde data that the wind oscillation of period 24 h that occurs at tropospheric and stratospheric heights at low to midlatitudes is not the wavenumber 1 tide predicted by classical tidal theory and apparently observed in the upper mesosphere, but rather an oscillation linked to local, regional or continental topography (Wallace and Hartranft, 1969; Wallace and Tadd, 1974). They used 12 h differences of monthly mean winds taken over a 15-year period to determine a universal time "snapshot" of the diurnal wind over the globe at several altitudes. Complex wind variations observed at low altitudes tend to simplify with increasing altitude. Downward phase progression is observed at all heights above 1 km, indicating upward energy propagation, i.e., an energy source at the ground or in the boundary layer. Vertical wavelengths of the order of 10 km are inferred. Wind amplitudes of 1–5 m s^{-1} are inferred, much larger than the theoretical "classical" tidal amplitudes, which are considerably less than 1 m s^{-1} at these heights.

Fukao *et al.* (1978) analyzed a full day of stratospheric wind measurements from the Jicamarca radar (12°S) for diurnal and semidiurnal wind components. They inferred a diurnal wind oscillation with characteristics similar to those inferred from the averaged rawinsonde data: amplitude of several meters per second, downward phase progression, and a vertical wavelength of <10 km.

Wallace and Tadd (1974) were able to determine the semidiurnal wind component at a number of stations where balloon flights at 6 h intervals were available. They concluded that the mean semidiurnal winds had less local structure than the diurnal winds, had smaller amplitudes, and showed no vertical phase progression. They interpreted the data as being due chiefly to fundamental semidiurnal modes, and concluded that topographical forcing was not important for the semidiurnal component.

In this paper we present the results of a decomposition of a full day of zonal wind measurements at stratospheric heights taken with a 600 m height resolution at Arecibo into mean, diurnal and semidiurnal components.

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2. Observations

A 13-bit Barker-coded pulse with 4 μs bit length was used (e.g., Richaczek, 1969). The corresponding height resolution was 600 m. The transmitter was pulsed every millisecond, and at stratospheric heights the return from 20 successive pulses were summed (coherently integrated), giving an experimental bandwidth of 50 Hz. 128-point complex FFT's were done on the time series of these coherently integrated points, giving a frequency resolution of ~0.4 Hz. The resulting power spectra were summed for ~1 min before being written on tape. Data were taken at 30 heights spaced from 10–30 km at an interval of 600 m.

The Arecibo antenna was offset 15° from the zenith and pointed to the west for this experiment. For horizontal wind determinations this gives an experimental bandwidth of ± 35 m s^{-1} and a horizontal velocity resolution of ~0.5 m s^{-1}.

The need to Barker code the transmitted pulse complicates the data analysis. The Barker code has sidelobes, which means that the echo signal inferred at a given height can be influenced by the signal from heights as much as 52 μs, or ~7 km, away. The sidelobe pattern is known, and can be corrected for at all but the lowest heights, where the receiver was not turned on until a height of ~7 km in order to protect it from very strong ground-clutter echoes. Thus the velocities inferred up to heights of ~15 km should be treated with caution.

Fig. 1 shows a typical observed power spectrum (small open circles) where an atmospheric echo is discernible by its nonzero Doppler shift. The power spectra contain ground clutter at zero Doppler shift

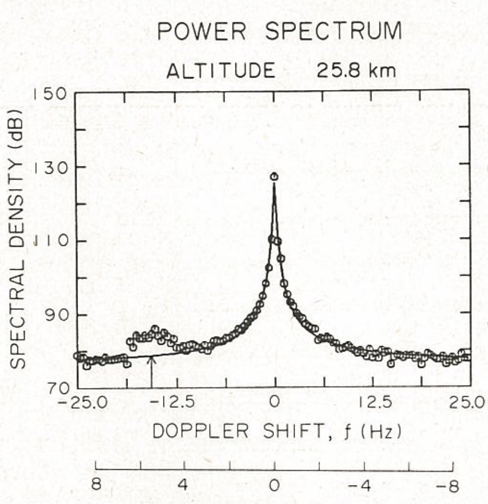

FIG. 1. Power spectrum (open circles) obtained by the Arecibo radar at 25.8 km at 0708 AST 22 August 1977. The ordinate is relative power (dB) with an arbitrary reference level. The abscissa is marked in terms of both the Doppler frequency and the equivalent radial velocity.

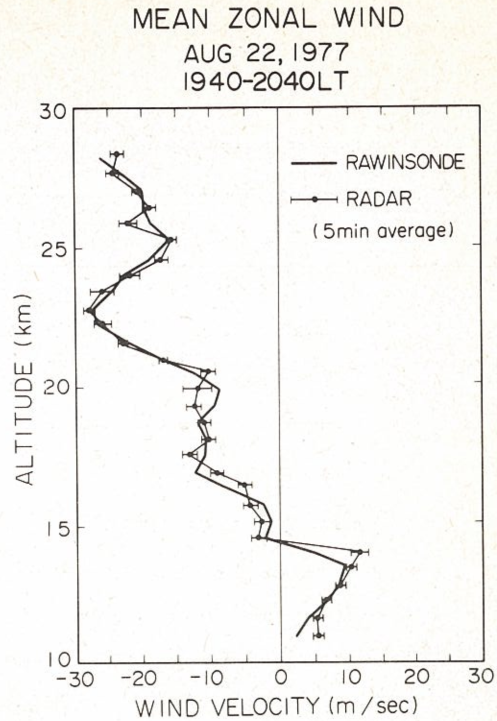

FIG. 2. Comparison of mean zonal winds observed by the Arecibo radar and by a rawinsonde launched at 1900 AST from San Juan, some 70 km east of Arecibo. Horizontal bars indicate standard deviations. Velocities are positive eastward.

that generally exceeds atmospheric echoes by 30–50 dB. It is apparent that the clutter component has a fairly long tail, the density of which decreases monotonously with increasing Doppler frequency f [Hz].

The clutter spectrum is generally well approximated by the simple formula

$$1/[C(f) - N] = af + b, \tag{1}$$

where $C(f)$ [dB] is the observed spectral density, whereas N [dB] is an asymptotic value for the infinite value of f, or the background noise power density.

Values of a, b and N are determined in the following way. First, N is analytically determined by Eq. (1) using three values of the spectrum; i.e., the values at and next to zero Doppler shift and the averaged value over the outermost 10 data points. Next, a and b are determined by fitting a line, in a least-mean-square sense, to the five points in the vicinity of zero Doppler shift. This procedure utilizes only a few values in the vicinity of zero Doppler shift and 10 values in the tail of the spectrum. Therefore, it is applicable to any spectrum with large echo power at nonzero Doppler shift, as is shown in Fig. 1. Any spectral component that exceeds the estimated clutter-plus-noise level is assumed to be a contribution due to atmospheric echoes.

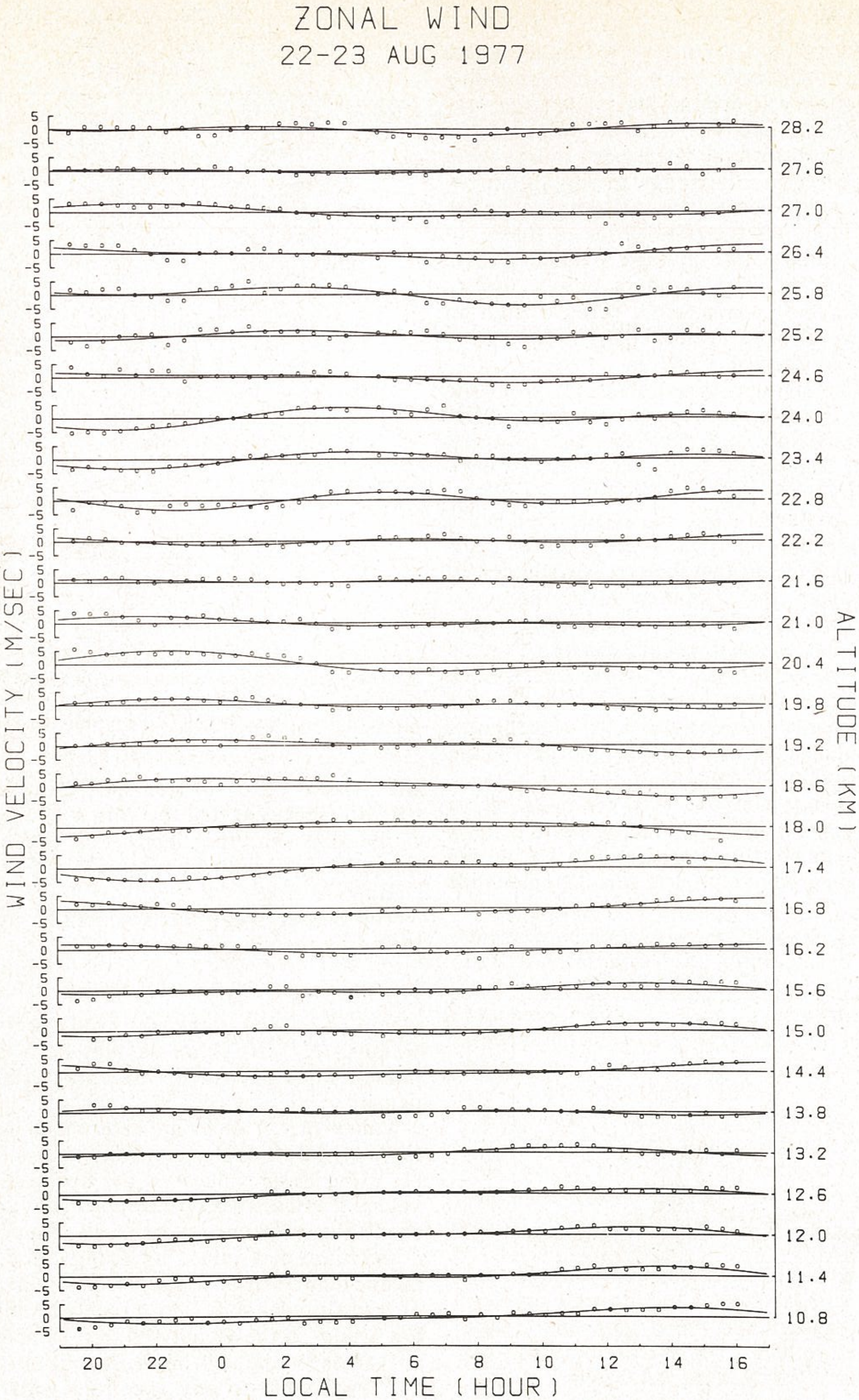

FIG. 3. Temporal variation of the zonal wind (mean winds are not included). Velocities are positive eastward. The ordinates are zonal velocities, whereas the abscissa is the observation period. Dots denote 1 h averages plotted every 30 min, and the solid lines are the best-fitted combination of the 24 and 12 h sinusoids.

Statistical fluctuations of the observed spectra seem to be <3 dB in the present observation, so that only echoes 3 dB higher than the estimated clutter-plus-noise level are used, weighted with the inverse of the clutter-plus-noise level, to calculate mean Doppler shifts. The vertical arrow in Fig. 1 indicates the mean Doppler shift thus estimated.

3. Results

Fig. 2 shows the mean zonal wind velocity versus height. The velocities are positive eastward (westerly). The horizontal bars are not error bars, but rather show standard deviations that indicate the magnitudes of the zonal wind fluctuations around the mean value. Rawinsonde wind determinations for this day are in general agreement with the radar values. In most cases the rawinsonde points fall within the range of fluctuation observed by the radar.

Fig. 3 shows the temporal variation of the zonal wind versus height after the mean winds have been subtracted. The ordinates are zonal velocities, ranging ± 5 m s^{-1}, and the abscissa is local time during the observation period, from 1900 AST 22 August to 1700 AST 23 August 1977. The dots are 1 h velocity averages plotted every 30 min. The solid lines are the combination of 24 and 12 h sinusoids which give the best fit to the data at each height. It can be seen that a combination of 24 and 12 h components generally gives a good fit to the data, though there are clearly shorter period oscillations present at some altitudes. The best-fit amplitudes range from 1 to 5 m s^{-1} and 1 to 2.5 m s^{-1} for the diurnal and the semidiurnal oscillations, respectively. These amplitudes are generally one order larger than the values predicted by the classical tidal theory (Chapman and Lindzen, 1970; Lindzen and Hong, 1974). However the present results are consistent with those inferred from the similar 1-day observation by the Jicamarca radar (Fukao *et al.*, 1978), and are comparable to the tidal amplitudes obtained from averaged rawinsonde data (Wallace and Tadd, 1974).

Fig. 4 shows contour plots of (a) the diurnal component and (b) the semidiurnal component versus height and time. Contours are drawn at 2.5 m s^{-1} intervals. Shaded areas indicate eastward (westerly) winds. No smoothing was done in the vertical direction.

The contours of eastward and westward winds indicate a marked downward phase progression with height in both diurnal and semidiurnal components. The vertical phase progression of the semidiurnal oscillation was not observed in the rawinsonde data (Wallace and Tadd, 1974). This indicates an upward flux of energy from below. It seems probable that the oscillations we observed are the non-migrating tidal winds excited at the ground or in the boundary layer as suggested by Wallace and Hartranft (1969). The consistent phase progression in both components over a large number of independent heights indicates that the fitting procedure detects true atmospheric oscillations.

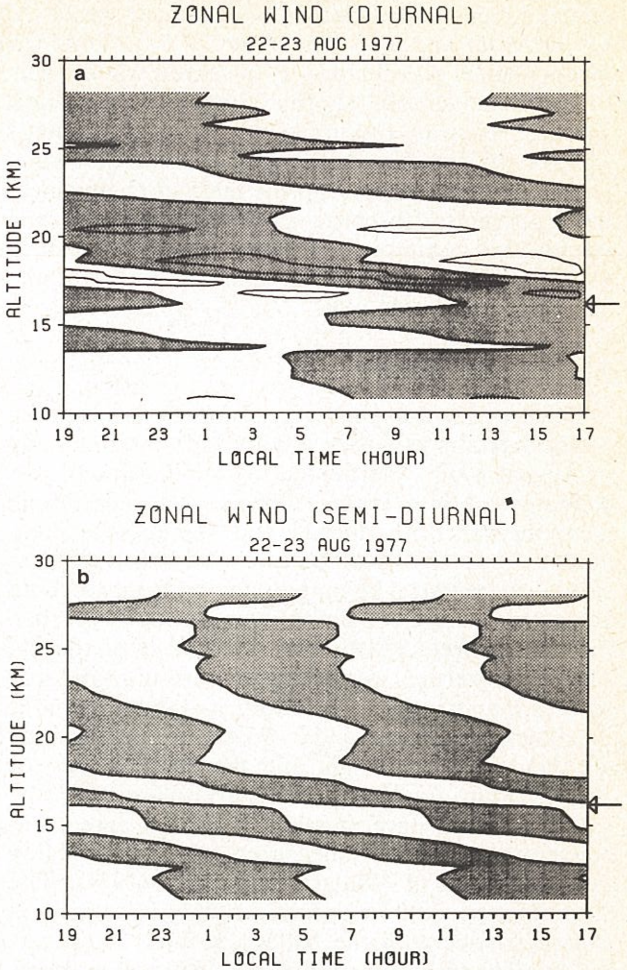

FIG. 4. Time-height sections of (a) the diurnal component and (b) the semi-diurnal component. Contours are drawn at 2.5 m s^{-1} intervals. Shaded areas show eastward winds. The tropopause observed by the rawinsonde is indicated by the arrow.

It should also be noted that there are considerable fluctuations superimposed on the phase progressions that are generally downward with height. This may indicate that many competing modes coexist. It will help us understand the nature of the observed wind oscillation to estimate a wavelength, though it should be defined for a single mode. We infer a mean vertical wavelength by assessing the vertical distance over which the phase progresses by 24 or 12 h. As for the diurnal component, the mean wavelength is inferred to be 5–6 km over the 15–25 km altitude. The vertical structure seems to be more complicated for the semidiurnal component. The inferred mean wavelength is 8–9 km over the 20–30 km altitude, and 2–3 km below 20 km. Thus the mean vertical wavelengths of observed waves seem to be considerably shorter than the theoretical values. As was discussed in the observation's section, the data from below 15 km should be treated with some caution, though the marked change near the tropopause observed in both the diurnal and semidiurnal components cannot be easily explained as a sidelobe contamination from the Barker code, and this is probably real.

4. Concluding remarks

The lower stratospheric zonal wind was measured by the Arecibo radar for a period of about one day in August 1977. The fluctuating wind velocity was decomposed into mean, diurnal and semidiurnal components. Both diurnal and semidiurnal components showed a marked downward phase progression with height. The amplitudes inferred for both components were one order of magnitude larger than the values predicted by the classical tidal theory. The mean vertical wavelengths were considerably shorter than the theoretical values—about 5 km for the diurnal oscillation and 2–9 km (the shorter value for the 15–20 km height and the longer value for above 20 km) for the semidiurnal oscillation.

It is perhaps unreasonable to ask for agreement between one day radar measurements and the highly averaged data of Wallace and Tadd (1974). The diurnal component we inferred agrees in its general characteristics with the balloon results. However, the semidiurnal component we inferred suggests local structure which is variable in time in order to average out in the mean balloon results. Clearly, more radar observations are desirable. These will offer the possibility of correlating the observed tidal wind oscillation with regional meteorological variables on a day-to-day basis in order to investigate the tidal forcing.

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