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3.7.7 MESOSPHERIC WAVE NUMBER SPECTRA FROM POKER FLAT MST RADAR MEASUREMENTS COMPARED WITH GRAVITY-WAVE MODEL

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INTRODUCTION

This paper presents the results of a comparison of mesospheric wind fluctuation spectra computed from radial wind velocity estimates made by the Poker Flat MST radar with a gravity-wave model developed by VANZANDT (1982, 1985). The principal conclusion of this comparison is that gravity waves can account for 80% of the mesospheric power spectral density.

Two different hypotheses have been advanced in recent years to explain the origin and behavior of mesoscale (one to thousands of kilometers) wind fluctuations. One hypothesis is that these fluctuations are manifestations of two-dimensional (2D) turbulence (GAGE, 1979; LILLY, 1983). Two-dimensional turbulence consists of turbulent eddies with wind fluctuations constrained to occur primarily in a horizontal plane. The other hypothesis is that the observed wind fluctuations are due to a broad spectrum of gravity waves transporting momentum vertically (DEWAN, 1979; VANZANDT, 1982, 1985).

Both theories claim similar shapes for frequency and horizontal wave number spectra of horizontal wind fluctuations. Thus, observations that both frequency and horizontal wave number spectra follow power laws with an exponent of -5/3 cannot be used to determine what portion of the wind fluctuation energy spectrum is attributable to either type of motion.

There is a need to partition the spectrum between gravity waves and stratified, 2D turbulence in order to study the interaction of the fluctuations with the environment and to determine the horizontal and vertical transport of energy. For example, gravity waves are capable of vertical transport of energy and momentum and knowledge of gravity-wave flux at a given height would lead to estimates of mean flow drag, wave saturation and vertical transport of heat and chemical constituents at other altitudes.

Our approach to partitioning the spectrum has been to identify that part of the spectrum that is consistent with VANZANDT'S (1985) gravity-wave model as being due to gravity-wave motions. We have used vertical wave number (m) spectra to avoid the ambiguities mentioned earlier.

In the next section, a brief review of VanZandt's gravity-wave model formulation and the extension of that model to wind fluctuations along any oblique path will be presented. Then, spectra of vertical and oblique (15° off-zenith) radial winds measured near the summer mesopause by the Poker Flat

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MST radar will be described. Finally, the comparison between the observations and the model will be made and the implication that gravity waves dominate the fluctuating wind field near the mesopause will be discussed.

MODEL WAVE NUMBER SPECTRA

VANZANDT (1985) transferred the successful oceanic internal gravity-wave model spectrum of GARRETT and MUNK (1972, 1975) to the atmosphere. In this model, the gravity-wave energy spectrum is assumed to be the product of two independent factors: a frequency spectrum and a wave number spectrum. Since the total energy spectrum is assumed to be separable, integration over frequency yields the wave number spectrum and vice versa. These simple manipulations are all that is needed for spectra of horizontal or vertical wind fluctuations since the model was formulated from such observations. However, extending the model to wind fluctuations along any oblique path requires further use of gravity-wave theory.

Velocity perturbations measured along an oblique ray will be a combination of horizontal and vertical perturbations. The ratio of horizontal to vertical velocity perturbations for gravity waves is determined by the gravity-wave dispersion relation, which in this formulation is obtained from a WKB solution of the wave equation. With the dispersion relation and the zenith angle of the sampling vector, a formula is derived for the measured wave perturbation in terms of the absolute amplitude and frequency of a wave. For a detailed model derivation, see VANZANDT (1985).

Though, as pointed out by VANZANDT (1982), there is evidence for a uniform wave amplitude over a wide range of observing sites and seasons, we avoid any uncertainty in the amplitude by using the ratio of spectra taken at different zenith angles. The absolute amplitude cancels out of the ratio. If the fluctuations are due to gravity waves, then the ratio should agree with the ratio predicted by the gravity-wave dispersion relation. Thus, we are not investigating the universality of the gravity-wave model but rather the application of the gravity-wave dispersion relation to the observed fluctuations.

It is important to note that Doppler shifting of the frequency spectrum has been neglected in the model formulation, since the gravity-wave dispersion relation used in the model assumes that the mean wind is zero. Doppler shifting should be considered in applying the model to frequency spectra. However, vertical wave number spectra are not affected by Doppler shifting because the velocity measurements along a beam are made nearly instantaneously and in a direction nearly orthogonal to the mean wind.

To apply the model to the data, the zenith angles of the radar beams, the slope of the horizontal velocity perturbations frequency spectra and the limits on the range of gravity-wave frequencies must be specified. The theoretical frequency limits are the inertial frequency, f, and the Brunt-Vaisala frequency, n. The approximate gravity-wave frequency spectra must be specified to permit integration over the frequency component of the total gravity-wave spectrum.

OBSERVED RADIAL WAVE NUMBER SPECTRA

Observations for this study are the radial wind velocities obtained by the Poker Flat MST radar from regions near the mesopause. Seasonal signal characteristics and radar system parameters are described in BALSLEY et al. (1983). We use high spatial resolution soundings (300 m) obtained during the STATE (STructure and Atmospheric Turbulence Environment) campaign in June 1983, and similar soundings obtained in July, 1984. In order to minimize cross talk, the vertical and oblique transmitters were turned off during alternate sampling periods. The sampling periods were approximately 1.5 minutes long. Thus, oblique velocities were obtained over 1.5-minute periods separated by 1.5 minutes during which the vertical beam was operating.

Mesospheric wind velocities were obtained from a vertical beam and two oblique (15° off-zenith) beams and from heights of 82 to 88 km. The two orthogonal, oblique beams were directed towards azimuths of 64° and 334° east of north and will be referred to as the east and north beams. The vertical beam directly measured the vertical wind while each oblique beam measured the projection of the total wind vector along the beam. This projection of the wind vector was a composite of the vertical wind times cos 15° (=.97) plus the horizontal wind along the beam azimuth, times sin 15° (=.26). Since vertical sample volumes were approximately 20 km from the oblique sample volumes at mesopause heights, the measured vertical wind did not necessarily equal the vertical contribution to the oblique measurements. Instantaneous horizontal winds could not be unambiguously determined with this system unless it was assumed that the wind field was homogeneous and uniform over the region sampled by the three beams.

Before transforming the velocity measurements to wave number spectra, spurious points due to aircraft and meteors in sidelobes and other types of interference were removed. An empirical editing procedure was used similar to those described by CARTER (1983). The velocity time series from each beam and at each height was examined for accelerations greater than 3 times a running average of acceleration. Any points exceeding that limit were removed from further processing.

In order to compute ratios of oblique to vertical spectral densities over the greatest range of wave numbers, individual profiles were selected that continuously covered the greatest range of heights. This height range was limited by the vertical system which is about 10 dB less sensitive than the oblique system. The final range of heights was also determined by the number of resultant spectra with equal numbers of points that could be averaged together to improve the confidence level. Thus, the height range (and corresponding wave number range) selected was a compromise between the greatest range available and the range continuously sampled most frequently.

The linear trend of each selected profile was removed leaving fluctuations about the mean shear. Then the profiles were "prewhitened" by differencing data points from successive heights. A cosine window was applied to the differenced data to minimize power leakage across the spectra. The differenced data was then transformed to the wave number domain using an FFT (Fast Fourier Transform) routine. The real and imaginary coefficients were squared and added to obtain a power estimate at each wave number. Finally, the wave number spectra were adjusted to reverse the effect of the differencing and the cosine windowing and then normalized such that the integrated spectra were equal to the variances of the original profiles yielding power spectral densities.

Several power spectra were averaged together to narrow the confidence limits for the spectra. The averaging was done over the longest intervals for which vertical wind measurements over a sufficiently large and contiguous height interval were available. Since the radar echoes were not continuous in time, this averaging when signals were strong in the vertical beams means that our observed spectra are representative only of periods during which the intensity of 3-m scale-size, refractive index irregularities, which produce the backscattered radar echo, was greatest. The mesospheric irregularities are due to turbulent mixing of the electron-density gradient. Thus, the level of turbulence and/or the electron-density gradient was enhanced during the selected periods. Therefore, these spectra, obtained during periods of enhanced radar signal strength, probably represent an upper limit to average power spectra in the summer mesosphere over Poker Flat.

Spectra from three such periods are shown in Figures 1-3. Each spectral curve is labelled by the beam from which the velocities were obtained, with the number of spectra in each average given in parentheses. The spectra are plotted versus vertical not radial wave number. This means that for the oblique data, the vertical wave number is the radial wave number divided by cos 15°, which represents a difference of 1.04 between radial and vertical wave numbers. We have computed 95% confidence intervals for the curve in each figure with the smallest number of averaged spectra using the degrees of freedom for a Bartlett spectral window and the procedure of JENKINS and WATTS (1968). Successive velocity profiles are not completely independent so the actual degrees of freedom are less than we have used and the confidence limits somewhat broader.

The spectra in the figures as well as additional spectra from July 1984, exhibit several points of similarity. The magnitudes of the oblique spectra are generally within a factor of 3 of each other and the spectral slopes from separate beams are approximately parallel. Indeed, an average of eight such spectra result in equal power spectral densities for the two oblique beams as shown in Figure 4. However, the slopes of the vertical power spectra are not as steep as the oblique spectra and we have sought an instrumental bias as an explanation for this discrepancy.

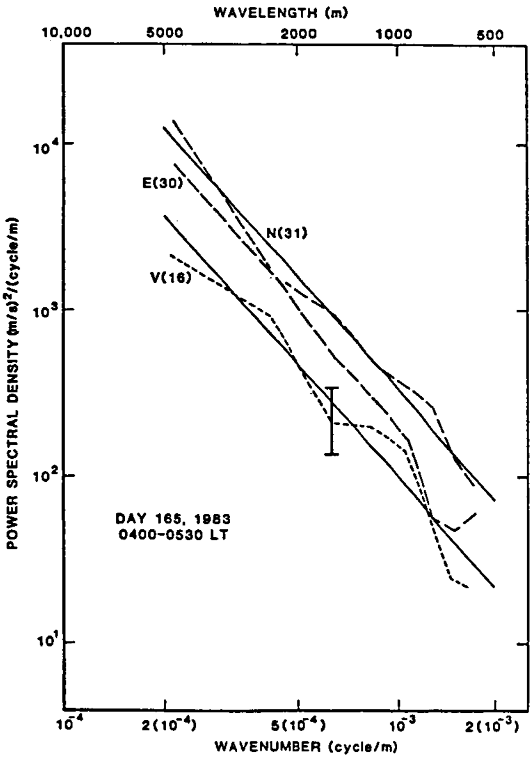

Figure 1. Vertical wave number spectra for Day 165 (June 14), 1983. Dashed lines are spectra computed from observations along East, North and Vertical beams. Each plotted spectrum is the average of the number of spectra in parentheses obtained between 0400 and 0530 local time. Solid lines are model curves.

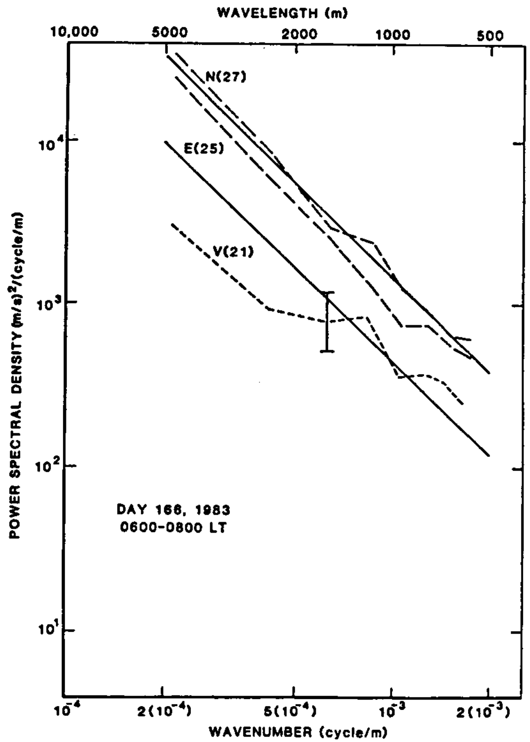

Figure 2. Same as Figure 1 but for Day 166 (June 15), 1983.

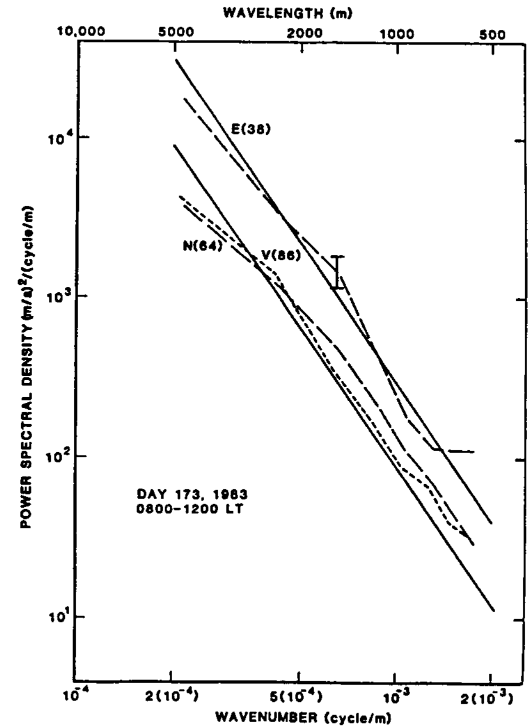

Figure 3. Same as Figure 1 but for Day 173 (June 22), 1983.

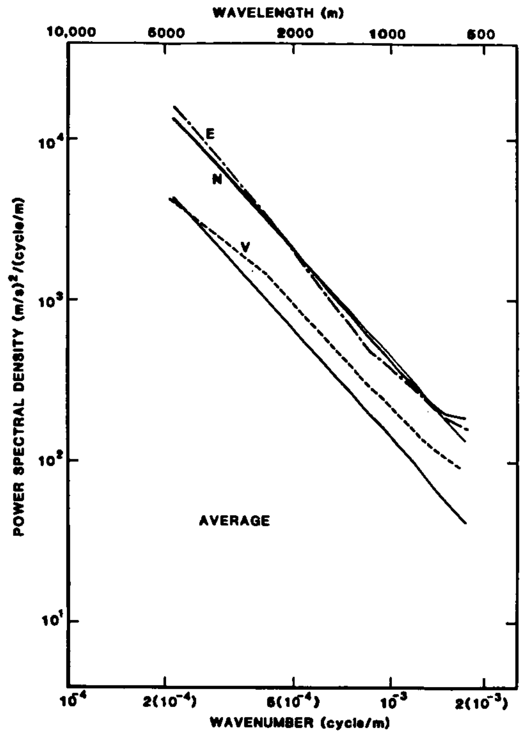

Figure 4. Vertical wave number spectra obtained by averaging all the spectra obtained over short periods in June 1983 and July 1984. The thin line through the East and North spectra is the model curve.

Research indicates that low resolution Doppler-shifted frequency spectra coupled with the clutter rejection scheme used at Poker Flat resulted in an increase in the magnitude of vertical velocity spectra by approximately a factor of 2. The computer at the Poker Flat MST radar samples the received signal, then performs coherent integration, accumulates 64 coherently integrated time series points at all heights and performs FFTs on those time series. This results in Doppler-shifted frequency or, equivalently, velocity spectra. In routine operation, the number of coherent integrations is set to cover the velocity range -35 to +35 m/s for oblique sampling and -4 to +4 m/s for the vertical velocities. Thus, spectral resolution is normally 1/8 m/s per spectral point on vertical spectra. During the STATE campaign, however, vertical velocities were obtained from spectra covering the range -36 to +36 m/s with 64 points for a resolution of about 1 m/s. The clutter rejection scheme used at Poker Flat consists of removing the mean of the 64-point time series gravity-wave dispersion relation, followed by a Hanning windowing of the time series before the FFT is computed. This procedure leaves a 3-point wide notch in the spectra at 0 m/s which is subsequently interpolated to restore some of the signal power at dc. This procedure does remove clutter signals but is also removes part of the signal power for velocities near zero. This effectively pushes near-zero velocities away from zero, increasing the variance of fluctuating vertical velocities and, therefore, artificially enhancing vertical power spectral density. When the spectral resolution is 1/8 m/s, as it usually is, this effect is not noticeable, but when the resolution is the same as the limits of the measured vertical velocities, the power spectral density can be enhanced by a factor of two. This enhancement is apparent in Figure 4.

DISCUSSION

The thin solid lines in the Figures are the model spectra. The lines were placed as follows: first a straight line was fitted to the largest amplitude oblique spectra. Then, the model ratio was computed using N = 0.02 rad/s, f = 1.32 (10-4) rad/s (inertial period = 13.2 hrs for the latitude of Poker Flat) and a frequency spectral slope of -5/3. Finally the computed ratio was used to construct the expected vertical velocity spectra. The agreement between the observed and model vertical spectra is excellent for high wave numbers in Figures 1-3. The log (base 10) of the average observed ratio was 0.3 at the 10^{-3} cycle/m point (1 km wavelength), while the model predicted a log ratio of 0.53. There is also an indication of considerable day-to-day variability and anisotropy of the spectra in the east and north beams.

The agreement between the observed spectra and the gravity-wave model, while not perfect, does provide justification for a gravity-wave interpretation. With this comparison, we can begin to assess the relative contributions of gravity waves and 2D turbulence to the fluctuation spectra. Since 2D turbulence does not generate vertical motions, on average, the vertical velocity spectra are primarily due to gravity waves. Then that portion of the oblique spectra that agrees with the vertical spectra plus the log of the model ratio is probably contributed by gravity-wave motions. The excellent agreement shown in the figures suggests that most of the velocity fluctuations are gravity-wave motions and that there is little contribution from 2D turbulence. The observed vertical velocity spectra may be twice as great as the actual spectra due to data-acquisition processing and thus, the actual log of the ratio between the averages of the 8 sets of oblique and vertical spectra may be as great as 0.6. The difference between this ratio and the model-predicted log ratio of 0.53 suggests that a surplus of up to 17% of the probable gravity-wave spectral amplitude could be attributed to other processes such as 2D turbulence.

The fluctuation spectra exhibit azimuthal anisotropy that varies from day to day. For instance, in Figure 3, the north and vertical spectra are nearly equal, implying that the north beam is recording only vertical fluctuations while horizontal fluctuations are principally aligned east-west. With measurements in only two directions, determination of the exact shape of the azimuthal dependence is impossible. This anisotropy has an impact on the agreement between the model and the observations, but the difference cannot be quantified until similar measurements are made over a more complete set of azimuths.

The observed anisotropy can result from anisotropy in wave sources or from the selective transmission of gravity waves through an anisotropic wind field at lower levels. However, anisotropy argues against domination of the spectra by 2D turbulence since such turbulence should produce azimuthally isotropic, horizontal velocity fluctuations.

Isotropy is recovered when spectra are averaged for a sufficiently long period. The average east and north spectra in Figure 4 are approximately equal in contrast to the short-term averages presented in the other figures. These observations seem to suggest that the fluctuation spectrum is isotropic over long time scales but that large anisotropies can exist for brief periods. This long term isotropy is most likely not a consequence of 2D turbulence but due to a saturation amplitude limit to be discussed elsewhere.

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