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INTERACTION OF ZONAL WINDS WITH THE EQUATORIAL MIDNIGHT PRESSURE BULGE IN THE EARTH'S THERMOSPHERE: EMPIRICAL CHECK OF MOMENTUM BALANCE

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Abstract. A minimum is observed at midnight in the time variation of averaged zonal winds measured in-situ near the equator at an average altitude of about 350 km, and we inquire whether observations of the equatorial midnight pressure bulge are consistent with this feature. The observed zonal wind maintains an eastward direction all night. It increases steadily in the early evening reaching a maximum of about 160 m/s at 2100 hrs local time (LT), dropping to a minimum of about 70 m/s around 2400 LT, then up to about 120 m/s around 0300 LT, decreasing thereafter and turning westward in the morning around 0500 LT. The nighttime eastward ion drift at Jicamarca, averaged over periods of similar solar activity (1970-71), shows the same signature, but generally with amplitude between 70 and 90% of the zonal wind amplitude. The 1978-81 Jicamarca ion drifts are similar except from 2400 to 0300 LT when they are about 30% larger than the in-situ zonal wind averages. The ion-drift is used together with the measured zonal winds to estimate the ion-drag effect in the momentum equation. The pressure term of this equation is estimated using in-situ measurements of the neutral temperature and density. The pressure term is found to be large before 2100 LT (~ 0.05 m/s^2), consistent with the steady increase in zonal wind before 2100LT, drops rapidly to near zero (roughly 0.005 m/s) before 2200 LT, and then increases slowly beginning at 2300 to about 0.015 m/s^2 just before 0300 LT when it is observed to reverse. The small pressure term observed from 2100 to 2400 LT is associated with the midnight pressure bulge and accounts for the decay of the wind from 160 m/s at 2100LT to 70 m/s at midnight. The larger value at 0300LT is consistent with the second maximum in the wind. It is found that the existing data sets are consistent in terms of zonal momentum balance. The striking similarity between the zonal wind and the eastward ion-drift is responsible for the small ion-drag term before midnight which is required to maintain the large zonal winds observed.

Introduction

This letter addresses the question of the effect of the equatorial midnight pressure bulge on the nighttime zonal winds in the altitude range between 300 and 400 km. It starts with a brief review of previous work, shows the evidence for a zonal wind effect, and analyzes the question in the light of previous data checking momentum balance locally. Only the nightime dynamics is discussed.

The equatorial midnight pressure bulge is known to affect the meridional winds (Friedman and Herrero, 1982), producing the reversals observed around midnight near 300 km altitude from the ground at Arecibo (Behnke and Harper, 1973) and from in-situ measurements on the Atmosphere Explorer-E (AE-E) satellite (Spencer et al., 1979). The early evening equatorward meridional wind at Arecibo increases until around 2200 LT when it abates, reversing to the poleward direction one to two hours after midnight. This led Harper (1973) to suggest the existence of a nighttime pressure maximum near midnight. In contrast, zonal neutral wind measurements at Arecibo (Burnside et al., 1981) seem to show little effect due to the local passage of the midnight pressure bulge. The zonal winds at Arecibo (18°N latitude, 67°W longitude) were found to be very different from those at Kwajalein (9°N latitude, 167°E longitude), and a longitude effect was suggested by Sipler et al., (1980). However, this effect may also be due to the latitude difference between the two stations. Recent, more extensive, zonal wind measurements in winter at Arequipa, Peru (12°S) show winds that peak at 2100 LT and blow eastward all night. Those data suggest a second though smaller maximum after 0300 hrs LT (J. W. Meriwether et al., 1985) consistent with the in-situ measurements discussed here.

Equatorial Zonal Wind Data

The zonal wind data used here were obtained on the Dynamics Explorer-2 (DE-2) satellite, and the measurements taken at different local times have been averaged at low and midlatitudes and published by Wharton et al., (1984). Figure 1 shows the average zonal wind at the equator as a function of local time. This represents an average over the latitude range from $-10°$ to $10°$ at an average altitude of about 350 km. A Fourier series fit to 4th order on the data is shown by the solid curve passing through the data points, which have been obtained from Figure 1 of Wharton et al. (1984) averaging every 14 minutes. The reduced chi-square value (Bevington, 1969) for the fit shown is about 0.95. As discussed by Wharton et al. (1984) sampling effects due to the aliasing of the 24 hr. period with a six month period variation are small for the winds at the equator. In addition, the minimum around

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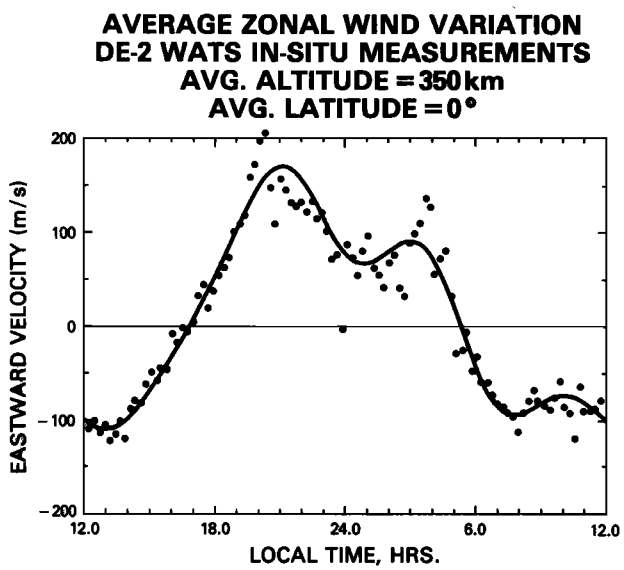

Fig. 1. Seasonally averaged zonal wind variation with local time at the geographic equator.

2400 LT followed by the maximum at 0300 LT is consistent with the ion-drift data (see below) and with the ground-based zonal wind measurements at Arequipa (Meriwether et al., 1985).

Momentum Balance

We inquire whether previously observed nighttime pressure variations are consistent with the midnight minimum and other features of the zonal velocity field. The analysis is based on zonal momentum balance of measured quantities at the specified altitude. Measured values are used for all the terms in the momentum equation except for the viscosity term which can be estimated from the other terms and compared with previous viscosity estimates.

The zonal wind U_x at the equator is driven locally by the zonal component of the pressure gradient $\partial p/\partial x$, and moderated by ion-drag and viscosity. That is,

$$\frac{\partial U_x}{\partial t} = -\frac{1}{\varrho}\frac{\partial p}{\partial x} + \frac{\mu}{\varrho}\frac{\partial^2 U_x}{\partial z^2} - \nu_{ni}(U_x - V_{ix}), \quad (1)$$

where ν_{ni} is the neutral-ion collision frequency, V_{ix} the zonal ion-drift, μ the viscosity coefficient, and ϱ the mass density (e.g., Rishbeth, 1972). This equation must be satisfied locally. Hence, the consistency of independent measurements of the parameters in this equation may be checked by direct substitution from the data. In eq. (1), the Coriolis force term has been omitted because it is negligible near the equator. The advective term is also omitted since it is always less than about 30% of $\partial U_x/\partial t$ considering the equivalence of zonal displacement x and local time t at the equator. That is, $x = \Omega Rt$, where Ω and R are the Earth's angular velocity and radius, respectively.

The term $\partial U_x/\partial t$ and the magnitude of U_x in eq. (1) may be obtained directly from Figure 1. The pressure gradient is determined by the variations in temperature and density. The major nighttime temperature variations have been found to be those associated with the midnight temperature maximum. Figure 2 (upper frame) shows the average nighttime neutral temperature variation T at the equator near 350 km altitude averaged around the 1978 autumnal equinox using the AE-E satellite data (Herrero and Spencer, 1982). The neutral density n of the principal constituent, atomic oxygen, exhibits similar variations, except for a delay of two to three hours in the occurrence of the corresponding maximum. Average AE-E density measurements at this altitude are reproduced well by the MSIS model (Hedin, 1983) and are shown also in Figure 2. We note that the AE-E data used correspond to late 1978, and the solar activity then was comparable to that for the ion-drift (of Figure 3) and zonal wind measurements of Figure 1. The bottom frame of Figure 2 shows the pressure variation obtained from the product of T and n and the gradient term $-1/\varrho(\partial p/\partial x)$.

Figure 3 shows the eastward ion-drift measurements at Jicamarca. The solar maximum data taken in 1970-1971 and 1978-1981 have been averaged separately and plotted therein. Also shown are the 1974-77 drifts which correspond to solar minimum. These are discussed in more detail by Fejer et al., (submitted to J. Geophys. Res., 1985). We note the similarity between the night-time zonal wind of Figure 1 and the ion-drifts, particularly from 1800 to 2400 LT for all three data sets, and throughout the entire night for the 1970-71 drifts. Both quantities reach their maximum values simultaneously at 2100 hrs LT, then go through a minimum near or after midnight, both having secondary maxima almost at the same time between 0300 and 0400 hrs LT. We use this similarity to set V_{ix} approximately equal to a given fraction of U_x during the night, using the measured values of U_x and V_{ix}. Comparison of

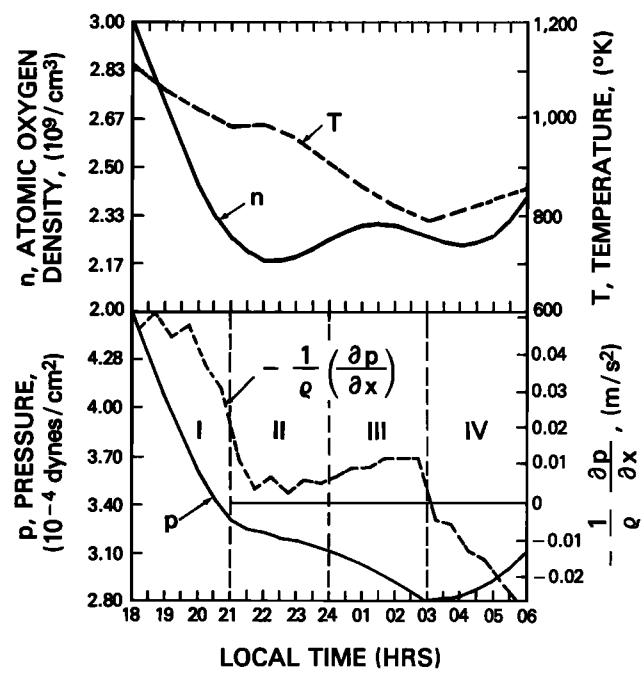

Fig. 2. Nighttime equatorial temperature, density, pressure and forcing term $(-1/\varrho\ (\partial p/\partial x))$.

U_x in Figure 1 with V_{ix} in Figure 3 gives on the average $V_{ix} \cong 0.8\ U_x$ roughly for the 1970-71 ion drifts and $V_{ix} \simeq 0.9\ U_x$ or more for the 1978-81 data. With this simplification the ion-drag term may be represented by U_x/τ_i, where $\tau_i = 1/0.2\ \nu_{ni}$ for the 1970-71 data is the decay time corresponding to ion-drag. The viscosity term may be cast into the same form assuming a vertical profile of the velocity of the form $U_x = U_{ox} + (U_{x\infty} - U_{xo})(1 - \exp(-h/H_v))$, where h is the height above a reference altitude somewhere above 150 km and the viscous scale height (H_v) is of the order of 100 km (Rishbeth, 1972). If h is several times as large as H_v, then the viscosity term may be approximated by U_x/τ_v, where $\tau_v \cong (\varrho\ H_v^2/\mu) \exp(h/H_v)$. Thus, equation (1) may be rewritten as

$$\frac{\partial U_x}{\partial t} = -\frac{1}{\varrho}\frac{\partial p}{\partial x} - \left(\frac{1}{\tau_v} + \frac{1}{\tau_i}\right) U_x. \qquad (2)$$

In Figure 2b we distinguish four regions where the values of $\partial p/\partial x$ are markedly different, and in each region we add up the observed acceleration $\partial U_x/\partial t$, the observed pressure gradient term and ion-drag term in equation (2), using the empirical ion density model of Chiu (1975) for the latter. In region I, $\partial p/\partial x$ is very large and negative, presumably driving the wind eastward against a diminishing ion-drag and producing the large $\partial U_x/\partial t$ observed in Figure 1 from 1800 to 2100 LT. In region II, $\partial p/\partial x$ is small, and in III, the gradient increases again though never becoming as large as in I. Finally, the pressure gradient becomes positive (negative $-1/\varrho(\partial p/\partial x)$) in region IV where the wind reverses to westward, consistent with the reversal of U_x to westward at 0500 LT in Figure 1. In the following paragraphs we look at these regions in more detail.

Table I shows the values of the terms in equation (2) for each time region. The local time shown in parentheses below each region number corresponds roughly to the center of that region where $\partial U_x/\partial t$ is nearly constant (Figure 1). The ion-drag decay time τ_i is given in the first row of Table I. It has been computed as indicated above, using $V_{ix} \simeq 0.8\ U_x$ and the ionospheric density model of Chiu (1975) and the expression for ν_{ni} in Rishbeth (1972). Figure 3 shows that V_{ix} can be larger than $0.9\ U_x$, giving a τ_i two times as large as that in the table. This provides an indication of the uncertainty in our discussion below.

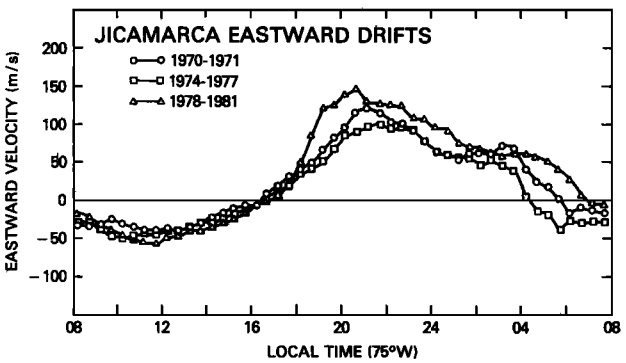

Fig. 3. Incoherent scatter radar measurements of the equatorial F-region eastward ion-drift.

TABLE I

	I (1930 LT)	II (2230 LT)	III (0130 LT)	IV (0430 LT)
τ_i(hrs)	0.81	2.1	6.2	7.2
U_x/τ_i (m/s²)	0.035	0.014	0.004	0.002
$\partial U_x/\partial t$ (m/s²)	0.020	−0.015	0.010	−0.025
$-1/\varrho(\partial p/\partial x)$ (m/s²)	0.050	0.004	0.015	−0.012

The second row in Table I gives the corresponding ion-drag acceleration U_x/τ_i, with U_x coming from Figure 1 at the corresponding time. The observed local acceleration $\partial U_x/\partial t$ has been obtained from the solid curve of Figure 1 and is given in the third row. The last row gives the forcing term $-1/\varrho(\partial p/\partial x)$ obtained from Figure 2b. The viscosity factor U_x/τ_v is not estimated a priori here since we have no observations of the vertical distribution of the velocity. However, any significant contribution from it would appear as a deficit in the momentum balance whose magnitude would then give an estimate of the viscosity scale height H_v.

A maximum acceleration of about 0.02 m/s² is observed around 1930 LT, the pressure gradient contributing about 0.05 m/s² and the ion-drag term about 0.035. The deficit of roughly 0.005m/s² could be as large as 0.02 m/s² with τ_i of about 1.6 hrs consistent with the uncertainty expected above. In region II the pressure term is negligibly small, about 0.004 m/s², not sufficiently strong to counteract the ion-drag effect of about 0.014 m/s². Thus, the observed wind decays to about 70 m/s in three hrs (Figure 1). This corresponds to a decay time longer than 3 hrs. If $\partial p/\partial x$ were zero, then U_x would decay more rapidly than observed due to the combined effect of τ_i and τ_v since $\tau_i \sim 2.1$ hrs. Again, τ_i could be twice as large, 4 hrs, allowing a deficit of roughly 0.01 m/s² for viscosity. In region III we see that the forcing term of 0.015 m/s² overrides the decay term giving a net acceleration of about 0.01 m/s² roughly the same as observed. In region IV the forcing term goes through zero quickly becoming negative after 0300 LT. Here the ion drag term is negligible (.002 m/s²) and τ_v aids the forcing term in reducing the wind to zero to produce the large acceleration of about −0.025 m/s² after 0300 LT. Thereafter the increasing magnitude of the gradient develops the reversal to westward at 0500 LT. In this region, the deficit of about 0.010 m/s² which is much larger than the ion-drag contribution of 0.002, is assigned to U_x/τ_v. On this basis, we estimate $\tau_v \simeq 1.5$ hrs and obtain $H_v \simeq 80$ km with $h \cong 160$ km using the relation given preceding equation (2) above. This number is consistent with a-priori nighttime estimates of H_v (Rishbeth, 1972).

Summary and Conclusions

An empirical check of momentum balance has been made using the average zonal winds and ion-drifts

measured under similar solar activity conditions combined with the nighttime average-pressure variations given by the AE-E data and the ion density given by the model of Chiu (1975). The analysis shows that the nighttime pressure variation obtained from temperatures and densities measured on AE-E is consistent with the observed variations in the zonal wind, and that the zonal wind decay time due to ion drag and viscosity reasonably accounts for the observed decay in velocity leading to the midnight minimum. We find that the F-region eastward ion-drift plays an important role in reducing ion-drag (or equivalently, accelerating the neutral wind) between evening and midnight as predicted by Rishbeth (1971) and Heelis et al. (1974). It is significant to note that the midnight pressure bulge does not exhibit a maximum in the zonal direction. Instead it is characterized by a very small gradient before midnight, which is consistent with the simultaneous zonal wind decay. The small pressure gradient observed between 2100 and 2400LT appears as a necessary feature to obtain the corresponding decay in U_x. If there were a significant pressure maximum at midnight, the zonal pressure gradient term would become negative before midnight resulting in a zonal wind reversal from eastward to westward at or before midnight, contrary to observation. On the other hand, the absence of the midnight pressure bulge (enhancement) would lead to a significantly larger pressure term (positive) between 2100LT and midnight, driving higher zonal winds, contrary to observation. Furthermore, if the ion-drift were not as significant as observed with $0.8\ U_x \lesssim V_{ix} \lesssim 0.9U_x$ before midnight for both 1970-71 and 1978-81 data, the ion drag decay time would be shorter, inconsistent with the observed 3 hr decay from 2100 to 2400LT. Thus, we find that zonal momentum balance is satisfied using the data above. We note that the plasma drifts in the F-region are the result of the combined effects of the ionospheric E and F region winds weighted by the corresponding Pedersen conductivities (e.g., Anderson and Mendillo, 1983), and may account in part for the difference in behavior between the 1970-71 and 1978-81 drifts near and after midnight.

These results demonstrate the utility of averaged data in thermosphere dynamics to check the consistency of independent measurements. They also point to the need for measurements of the vertical variation of U_x in order to estimate the viscosity more accurately. But more importantly, they portray the effect of the midnight pressure variation on the zonal winds and the importance of the eastward ion drift in thermosphere dynamics.

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