Planetary wave signatures in the equatorial atmosphere–ionosphere system, and mesosphere-E- and F-region coupling

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Abstract

Upward transport of wave energy and momentum due to gravity, tidal and planetary waves from below and extratropics controls the phenomenology of the equatorial atmosphere–ionosphere system. An important aspect of this phenomenology is the development of small- and large-scale structures including thin layers in the mesosphere and E-region, and the formation of wide spectrum plasma structures of the equatorial F-region, widely known as equatorial spread F/plasma bubble irregularities (that are known to have significant impact on space application systems based on trans-ionospheric radio waves propagation). It seems that the effects of tidal and gravity waves at mesospheric and thermospheric heights and their control of ionospheric densities, electric fields and currents are relatively better known than are the effects originating from vertical coupling due to planetary waves. Results from airglow, radar and ionospheric sounding observations demonstrate the existence of significant planetary wave influence on plasma parameters at E- and F-region heights over equatorial latitudes. We present and discuss here some results showing planetary wave oscillations in concurrent mesospheric winds and equatorial electrojet intensity variations in the Indian sector as well as in the mesospheric airglow and F-layer height variation in Brazil. Also presented are evidences of planetary wave-scale oscillations in equatorial evening pre-reversal electric field (F-region vertical drift) and their effects on equatorial spread F / plasma bubble irregularity development and day-to-day variability.

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1. Introduction

Vertical coupling in the atmosphere–ionosphere system through atmospheric waves (tidal gravity and planetary waves) that propagate to ionospheric heights from the lower regions (tropo–stratosphere) of their generation, and the associated interactive
processes, controls the dynamics and phenomenology of the ‘quiet time’ ionosphere. Over the equatorial latitudes the unique feature of the horizontal orientation of the Earth’s magnetic field lines leads to a condition in which the dynamo electric fields generated by these waves play the leading role in associated ionospheric response sequences extending to the upper layers of the ionosphere. While the vertical coupling processes driven by tidal forcing and the consequent dynamo electric fields are basically responsible for shaping the major phenomenology of the equatorial region, the interactive processes involving planetary waves (PW) and gravity waves are believed to play a significant role in the day-to-day and short-term variabilities widely observed in the important aspects of this phenomenology: equatorial electrojet (EEJ) current, F-layer plasma drifts and evening pre-reversal electric field (PRE) that controls the equatorial spread F (ESF) and related processes. However, evidences on direct association between the lower atmosphere wave forcing and the ionospheric response signatures continue to be sparse in this respect. Major sources of perturbations that can obscure a clear identification of the cause–effect relationship in the vertical coupling processes, in this context, are known to be solar ionizing radiation flux variations and enhanced solar wind–magnetosphere–ionosphere interaction during events of space weather disturbances when the penetrating/disturbance dynamo electric fields and the disturbance wind system dominate the equatorial region. Thus, only quiet time observational data sets will be useful in the investigations of the variability arising from the vertical coupling processes. The entire vertical coupling process can be seen as a two-stage process: (a) dynamic processes by which the wave energy from the lower heights/regions (troposphere and stratosphere) propagates to higher levels in middle atmosphere and lower thermosphere (MLT) regions (including the lower E-region heights); (b) the subsequent sequences in which the electrodynamic processes take over at height levels near and above the E-layer extending well into the F-region. In this paper, we intend to present some new results, and briefly evaluate some relevant results available recently on these two aspects and discuss their possible consequences to our investigations towards a better understanding of the quiet time variability of the equatorial F-region plasma dynamics and ESF irregularity processes.

2. PW oscillations in the MLT E- and F-region

Evidence of PW in the mesosphere has come from extended wind observations by medium frequency (MF) radars operated at different geographic locations (Harris, 1994; Haris and Vincent, 1993; Gurubaran et al., 2001). These waves are very large horizontal scale atmospheric oscillations in neutral wind, density and pressure that propagate zonally and vertically from the troposphere–stratosphere regions of their generation to the MLT region and perhaps beyond. Their oscillation periods have been found to lie in the 2- to 20-day range and beyond (Forbes, 1996, Forbes et al., 1995), the main periods being near the normal Rossby modes, i.e., 2-, 5-, 10- and 16-days, as well as at other (secondary) periods resulting from the superposition of these periods. 2-day waves in the MLT region is now well established (see, e.g., Pancheva et al., 2004), as are the 5-, 10- and 16-day periods (Forbes et al., 1995). An important feature of the PW is that their occurrence is of episodic nature so that attention needs to be focused on specific intervals. For example, a study of the 2-day waves in mesospheric zonal and meridional winds over Adelaide by Harris (1994) showed their occurrences for 2–3 week duration generally following mid-summer, and Clark et al. (2002) presented mesospheric wind measurements showing occurrence of 7-day wave of zonal wave number $S = 1$ that lasted more than 20–25 days during August–September 1993. PW of quasi-2-day and 3- to 5-day periodicities in equatorial mesospheric winds have been reported by Gurubaran et al. (2001) and Vincent (1993), respectively, and in mesospheric airglow intensity by Takahashi et al. (2002). Influence of PW on ionospheric parameters had been indicated earlier when Brown and Williams (1971) showed correlation between stratospheric pressure variations and electron density. More recently, direct evidence of PW role on mid-latitude sporadic E-layer generation, based on $f_0$Es data from an extended longitudinal chain of stations was provided by Haldoupis and Pancheva (2002).

Manifestation of the PW effects in the equatorial ionosphere came to be widely recognized after Chen (1992) demonstrated 2-day oscillations in the intensity of the equatorial anomaly and Forbes and Leveroni (1992) demonstrated 16-day oscillations in the EEJ intensity and the F2-layer peak density. In view of the theoretical difficulty to predict their upward propagation to ionospheric
heights (Hagan et al., 1993; Forbes et al., 1995) these results were attributed to the electrodynamic signature of the PW interaction with the dynamo region (lower E-region) of the ionosphere. The wave oscillations in the EEJ over Huancayo, Peru, analyzed by Parish et al. (1994) for an interval in 1985 confirmed the similar periodicities, namely, 2-, 5-, 10- and 16-day, to those found earlier by Forbes and Leveroni (1992) for an interval in 1979. Thus the oscillation at some prominent PW periods in the EEJ appears to be reasonably well established. We have attempted in this paper, as a first step, to verify the possible association of such oscillations in the EEJ strength with the oscillations in the MF radar winds using their concurrent long duration measurements realized in the EEJ region over India. Fig. 1 shows, as an example, the power spectrum obtained for the EEJ strength (lower panel), represented by $\Delta H$, the difference of magnetic field $H$-component variation over an EEJ station (Trivandrum: 8.48°N, 76.95°E; dip lat: 0.28°) and that over an off-EEJ station (Alibag: 18.68°N, 72.86°N; dip lat: 12.96°). In the upper panel is shown the power spectrum of the zonal wind at 88 km, concurrently measured by an MF radar at Tirunveli (located close to Trivandrum). We may note in these spectra (which correspond to an observational interval from 15 March to 26 April 1994) prominent spectral peaks near the periodicities at ~6.5-day and ~14-day in both the EEJ strength and zonal wind. The prominent periodicities vary with the observational epoch as other such plots have demonstrated. These results might seem to indicate that PW found at mesospheric heights could propagate up to the dynamo region where they modulate the ionospheric current system associated with EEJ at the PW time scales (more discussion on this aspect will follow later).

Additional results based on concurrent data on EEJ and MF radar winds for extended intervals are presented in Fig. 2. Here, the contour plot shows the time evolution of normalized power spectrum determined by “Morlet wavelet” analysis for EEJ strength (upper panel) and zonal and meridional winds at 88 km (lower panel) for the first half of 1999 (left panels) and second half of 1993 (right panels). The data availability dictated the selection of these intervals. We may note in this figure that during the late northern summer month of August 1993 the EEJ strength has significant spectral power for periodicities near 2-, 3-, 5- and 6.5-days. Correspondingly, the zonal wind also presents significant power at these periodicities, suggesting, thereby, that the mesospheric PW observed at 88 km might have propagated to dynamo region and/or influenced the wind fields there, so that the EEJ strength is modulated at the PW time scales similar to the example presented in Fig. 1. Also to note is the prominent 12–18 days oscillation in the EEJ strength from the last week of September till the end of October that is accompanied by a prominent ~12–20-day oscillation in the zonal wind. We note further that the oscillation in zonal wind precedes by ~12 days that in the EEJ strength, which might suggest that PW of 12–20 days period observed at 88 km could take some time (of the order of 12 days) to propagate to the dynamo region heights (~105 km) in the ionosphere. There are also frequent cases of oscillations present only in one of the two parameters. Examples are the zonal wind periodicities of 5-day in July, 8–12-day in August and ~14–20-day in November–December, and meridional wind periodicities ~14-day in August and ~6–11-day in September. Conversely, the
enhanced 8–11-day oscillation in EEJ is not associated with corresponding mesospheric PW, which might point to the possibility of in situ generation of these oscillations in dynamo region winds. As regards the results for the year 1999, we may note that the EEJ strength presents pronounced oscillations with periodicities near 2-day in the first week of February and near 5–6-day in April which are accompanied, respectively, by significant and less significant oscillations in zonal wind. Some times, dominant oscillations in mesospheric winds are accompanied by less significant or no oscillations in EEJ strength. For example, the dominant ~12-day oscillation in zonal wind in May...
is accompanied by less significant ∼14-day oscillation in EEJ, and the enhanced 14–20-day oscillations in zonal winds during February and March are not accompanied by any signatures in EEJ strength, which might indicate that the waves might be getting damped before penetrating to the dynamo region. It is interesting to note in the results of Fig. 2 that there are more cases of correspondence of EEJ oscillations with the oscillation periods of zonal winds, than with those of meridional winds, which might suggest that the EEJ processes are controlled more by zonal wind than by meridional winds.

Next we consider some evidences on associated PW oscillations in the mesospheric airglow emission and F-layer height in the equatorial region over Brazil. The upper mesosphere and lower thermosphere night airglow intensities in the OI5577 and OH(6,2) bands monitored over Cariri (7.5°S, 36.5°W) and the virtual height of the F-layer monitored by a Digisonde (Reinisch et al., 1989) over Sao Luis (2.6°S, 44.2°W) during the period from March to November 1999 are used in this analysis. It should be pointed out that the airglow data time series has 10-min interval during 6–10 h of a clear sky nights and there are 10–14 nights of observation surrounding each new moon period (see also, Takahashi et al., 2005), whereas the $h'F$ data has (usually) 15-min interval on a continuous basis. For the present analysis we have used $h'F$ values averaged over 21–24 LT of each night. The result of Morlet wavelet power spectral analysis performed on this averaged $h'F$ values are shown in the upper panel of Fig. 3, wherein shaded zones indicate the missing data gaps. Due to the relatively discrete nature of the airglow data its time series was subjected to Lomb–Scargle (LS) spectral analysis for each monthly group, and the results are shown in the lower panels of Fig. 3 (diverging arrows joining the upper and lower panels indicate the intervals of the $h'F$ data series for which the airglow was analyzed). A comparison between the two sets of results brings out some cases of reasonably good coincidences in the oscillation periodicities. A 3–5-day oscillation in the $h'F$ seen during the interval 9–26 March can be easily identified with the airglow oscillation showing an ∼4-day period in both the OI5577 and OH rotational temperature. During the interval of 6–19 June a relatively weaker 2–3-day
oscillation in $h'F$ is associated with strong oscillation of the same period in the airglow emission. In the interval 3–14 September, when the airglow emission presents strong oscillation of 4–5-day period the $h'F$ shows a tendency for 4–8-day oscillation which gets intensified, staying strong during the later half of September. These results do suggest associated oscillations in the equatorial mesosphere and F-region on PW scales.

2.1. PW scale modulations in the evening zonal electric field and spread F

It is well known that equatorial F-layer height undergoes a large increase at sunset due to the zonal electric field enhancement, PRE (Woodman, 1970) that develops under the action of the F-layer wind dynamo that intensifies at this time. The height of a specific plasma frequency can reach values $>300$ km and, therefore, its rate of change with time is a good/reliable indicator of the vertical plasma drift of the F-layer as discussed in detail by Bittencourt and Abdu (1981) and Abdu et al. (1981). Thus the nighttime F-layer height (at a given plasma frequency) is an approximate representation of the time-integrated vertical drift of the post-sunset F-region (subject to the recombination effects that set in for heights $<300$ km). We have calculated the vertical drift velocity (eastward electric field) over Jicamarca using the F-layer height information available from digisonde data. The result for the entire month of March 2000 (using the almost uninterrupted half-hourly data) is presented in Fig. 4 (the day numbers on the x-axis mark the end of the day in UT, ‘0’ being the beginning of the day ‘1’). The large increase in the vertical drift velocity that falls nearly on the day markings, represents the evening vertical drift velocity enhancement (PRE) that maximizes near 19 LT (note that the local time (LT) over Jicamarca is 5 h behind the UT). We note that the vertical drift velocity peak, Vzp, has relatively large values typical of the seasonal/equinoctial maxima of a solar maximum epoch for this location (Fejer et al., 1991). Large variation in the Vzp amplitude is observed during this period, the lowest and highest values attained being $\sim 20$ and $70$ m/s, respectively. Oscillations of a few days periodicity can be clearly identified in them. The magnetic activity level, in terms of the hourly Dst values plotted in the bottom panel, indicates generally quiet conditions for this period, as confirmed also by the $K_p$ indices that attained slightly higher values only on 1 and 31 March, the corresponding $\sum K_p$ (diurnal sum of the 3-h values) being 28 and 29+, respectively. Thus it appears evident that the large amplitude oscillations in the Vzp are unlikely to have been caused by any disturbance electric field. Consideration of Vzp variations during some specific intervals such as 2–6, 17–21 and 22–29 March, when the Dst remained stable and very low, testifies to the above affirmation. Thus we may attribute these oscillations in the Vzp amplitude to PW-induced effects in the ionosphere similar to the oscillations in the $h'F$ presented in Fig. 3.

The Vzp values of March 2000 (from Fig. 4) are plotted in the top panel of Fig. 5. These were subjected to Morlet wavelet analysis and the result is shown in the lower panel of this figure. The cone of influence due to edge effect in the data series analyzed is indicated by the thick (parabolic) curves. Significant amplitude of 3- and 7-day waves is present during the second week of March and the wave periods evolve to 2–3-day, 4–5-day and 7–8-day towards the third week of the month. These periodicities are in good agreement with the widely observed PW periods discussed before (see, e.g. Forbes, 1996; Forbes and Leveroni 1992; Gurbabaran et al., 2001; Haldoupis et al., 2004; Pancheva et al., 2003, Takahashi et al., 2002). Thus we note that the PW disturbances are capable of modulating
the intensity of the evening equatorial F-region vertical drift velocity (zonal electric field).

The evening F-layer uplift due to the PRE, of which \( V_{zp} \) is the best indicator, is known to be the main cause for the generation of the post-sunset ESF/plasma bubble irregularities (see e.g., Abdu et al., 1983, Fejer et al., 1999). We have plotted in the top two panels of Fig. 6 the ESF intensity as a function of UT and day of the year for the same interval as in Fig. 5 (March 2000). The corresponding \( V_{zp} \) values are plotted in the bottom panel. It should be mentioned here that we have used as a rough indicator of the ESF intensity the spread range of the ESF echoes in the ionograms, one unit of intensity being represented by 100 km of range spreading around the F-layer trace. The middle panel shows the spread F intensity at the lower frequency limit of the ionogram which indicates the first onset of the bottomside spread F. The onset of SF at the higher frequency range that follows in sequence (with some time delay, usually of 15–30 min) marks the evolution towards topside
bubble formations when the intensity (spread range) generally increases, as can be verified from the plots in the top panel. We note that there is significant control of the SF intensity by the Vzp amplitude, higher intensity of the SF, in general, corresponding to larger amplitude of the Vzp. For example, during two of the lowest values of Vzp identified by the vertical lines 2 and 5, corresponding to Vzp \( \approx 20 \text{ m/s} \), the SF was nearly totally absent, and during three of the higher Vzp values identified by the vertical lines 1, 3 and 4, which corresponded to Vzp \( \approx 65–70 \text{ m/s} \), the SF was clearly more intense. Of course a closer examination of this figure would show some important exceptions to a direct relationship between the two parameters, of which we will be discussing later. Independent of the detailed nature of the relationship between Vzp and ESF intensity we may affirm that PW scale oscillations are clearly present in the evening PRE (vertical drift) which in turn modifies/controls the formation and intensity of the ESF.

3. Discussion

We have presented observational results in evidence of vertical coupling processes involving PW scale oscillations in the mesosphere E- and F-regions of the equatorial atmosphere–ionosphere system. The results call our attention to two important aspects of the coupling processes underlying these results: (1) Mesosphere E-region dynamic coupling (MEDC) through upward propagation of PW and (2) E- and F-region electrodynamic coupling (EFEC) that produces manifestations at PW oscillation periods.

Both these aspects pose an important question as to the accessibility of the PW to the ionospheric dynamo region and to higher levels. In the case of the MEDC, the penetration of the PW into the E-region could result in dynamo action by the associated winds, leading to the generation of an electric field responsible for the effects observed in the E-regions, such as the PW scale oscillations in \( \Delta H \) observed in Figs. 1 and 2. However, the existing model calculations do not seem to support direct penetration of PW into the E-region capable of inducing wind amplitudes required for a viable dynamo mechanism (Forbes, 1996; Hagan et al., 1993; Forbes et al., 1995). Further, recent analysis results by Pancheva et al. (2003) on MF radar winds seem to show a certain amplitude attenuation in meridional wind with height increase in the 91–103 km region. Several mechanisms have been proposed to explain the PW period oscillations observed in the dynamo region (~100–170 km) and at higher levels of the ionosphere. They include, as explained by Forbes (1996), (a) stratosphere/mesosphere PW modulation of the accessibility of gravity waves to the upper levels, involving mechanisms that may lead to secondary source of PW excitation at higher levels and (b) PW modulation of the upward propagating tides that participate in the dynamo action to generate electric fields at higher levels. The latter mechanism seems to be especially promising and more easily verifiable, in view of the fact that the forcing by upward propagating diurnal and semidiurnal tides is basically responsible for the phenomenology of the quiet time ionosphere, especially over equatorial latitudes. In fact, evidence on the influence of PW modulated atmospheric tides of diurnal and semi-diurnal periodicities on mid-latitude sporadic E-layer formation has been provided recently by Pancheva et al. (2003). Our results in Figs. 1 and 2 showing cases of associated planetary scale oscillations in \( \Delta H \) and mesospheric winds (especially zonal winds) can be examined in the light of this discussion. The \( \Delta H \) represents the diurnal range of the magnetic field \( H \)-component variation due to the EEJ, which is a direct measure of the product of dynamo electric field intensity and electron density/conductivity at E-layer heights. Assuming that the PW oscillations in the E-layer electron density are small (there is no independent evidence of such oscillations so far) we attribute the periodicities observed in \( \Delta H \) to the corresponding oscillations in the dynamo electric field, which is produced dominantly by a diurnal tidal wind dynamo. In view of the observation by Pancheva et al. (2003) that mid-latitude Es-layer formation is affected indirectly by PW through their modulation of the diurnal and semidiurnal tides, we may expect that the equatorial \( \Delta H \) oscillations at PW periods may arise in the dynamo region also through the action of PW modulated tidal winds. This point needs to be checked, however, from further analysis of the data. In any case, a more direct role of PW in the dynamo region cannot be ruled out either, without examining additional concurrent data of winds in the equatorial dynamo region.

As regards the aspect of the EFEC mentioned above, the PW scale oscillations in the F-region parameters that result from the EFEC processes have important implications to our understanding of the causes of the widely observed day-to-day
variability in the equatorial F-region evening vertical drift (PRE) and the related spread F development processes. This problem is complex enough that a detailed discussion is not attempted here. For a brief discussion we may simplify this problem into the two following aspects: (a) PW scale oscillations in the evening PRE/vertical drift and (b) ESF variability arising from the oscillations in the PRE as one of the causes.

Regarding the aspect (a), it is known that the equatorial evening electric field enhancement arises from the F-layer wind dynamo, which involves electrical and electrodynamical coupling of the E- and F-regions of the equatorial ionosphere (Rishbeth, 1971; Heelis et al., 1974). Briefly, the thermospheric zonal wind (eastward in the evening hours) in the presence of the strong longitudinal/LT gradient in the E-layer conductivity (caused by the conductivity decay towards night side of the terminator) produces by dynamo action vertical (downward) electric field increasing towards the night side, as can be verified with the help of the following equation:

\[ E_z = U_y \times B_0 \left[ \frac{\sum F}{\left( \sum F + \sum E \right)} \right], \]  

where \( U_y \) is thermospheric zonal wind and \( B_0 \) is the Earth magnetic field intensity, and \( \sum F \) and \( \sum E \) are the integrated conductivities, respectively, of the E- and F-regions (Abdu et al., 2003). Due to the faster decay of the E-layer conductivity, \( \sum E \), as compared to \( \sum F \), in the post-sunset hours, \( E_z \) tends to increase towards the nightside. The application of a curl-free condition to such an electric field could lead to the enhanced zonal (eastward followed by westward) electric field as proposed originally by Rishbeth (1971) and modeled by Eccles (1998) (for other different approaches used in modeling this problem see e.g., Farley et al., 1986; Batista et al., 1986; Haerendel et al., 1992; Crain et al., 1993, Fesen et al., 2000). Thus it appears evident that the magnitudes of the zonal wind as well as that of the longitudinal conductivity gradient across the terminator could control the intensity of the PRE. Therefore, the PW oscillations observed in the PRE (Figs. 4 and 5) could be caused by such oscillations either in the intensity of the thermospheric zonal wind or in the degree of the conductivity longitudinal gradient. From the discussion on the vertical penetration of the PW presented above it looks less likely that significant oscillations in the neutral winds at thermospheric height could be directly induced by the PW. There is a possibility of PW modulation of the solar diurnal thermal tide that is mainly responsible for the winds at higher levels of the thermosphere, which is difficult to verify however. A more likely source of these oscillations may be sought in the possible dependence of the E-layer conductivity longitude/LT gradient on the lower thermospheric zonal wind as explained below. Fig. 7 taken from Abdu et al. (2003) shows the E-layer integrated Pedersen conductivity as a function of time calculated for diurnal tidal winds of different amplitudes ranging from 0 to 100 m/s. The time \( t = 0 \) is adjusted to coincide with 18 LT, close to local sunset (for details of the calculation see Abdu et al., 2003). We note that the LT gradient (which is equivalent to longitudinal gradient), positive towards dayside, increases significantly with increase of the zonal wind amplitude in the E-region (90–140 km). On the basis of this dependence we may expect that the PRE amplitude which depends upon the longitudinal gradient in the E-layer integrated conductivity (as a dominant controlling factor) could undergo fluctuations due to corresponding oscillations in the amplitude of the E-layer zonal wind of diurnal or semi-diurnal tidal mode. The amplitude of the tidal wind itself could be modulated by PW through nonlinear interaction, as mentioned before. Here again we cannot rule out the possibility of a direct role of the PW in causing the zonal wind fluctuation of the required magnitude at E-layer heights.

The aspect (b) concerns the day-to-day variability in the ESF occurrence and intensity due to oscillations in the PRE amplitudes as one of its causes. It has been known from previous studies
that ESF can develop and increase in intensity for
Vzp amplitude of ~20 m/s and higher (see for example, Abdu et al., 1983; Fejer et al., 1999). This relationship is evident also in the plots of these two parameters shown in Fig. 6. The ESF development and intensity depend also on other conditions besides a direct dependence on the PRE amplitude. To understand its variability we need to examine the linear growth rate, γ, of the ESF/plasma bubble instability by the generalized Rayleigh–Taylor mechanism that uses the background ionospheric parameters expressed in terms of their values integrated along the magnetic flux tube in which the instability begins to develop at the bottom-side gradient region of a rapidly rising evening F-layer. Based on flux tube integrated quantities, as proposed by Haerendel (1973, unpublished report) (also, Haerendel et al., 1992), a generalized form of the linear growth rate derived by Sultan et al. (1996) is given by

\[ \gamma_{\text{FT}} = \frac{\sum_{E}^{F} \left( \frac{E}{B} - U_{\text{FT}}^{P} + \frac{g}{v_{\text{eff}}} \right)}{\sum_{P}^{E} + \sum_{F}^{P}} \frac{1}{L_{\text{FT}}} - \beta_{\text{FT}}. \]  

(2)

Here, \( \sum_{P}^{E,F} \) is the field-line integrated conductivities for the \( E \)- and \( F \)-region segments of a field line; \( U_{\text{FT}}^{P} \) is the conductivity weighted flux tube integrated vertical wind; \( \beta \) is the recombination loss rate; \( v_{\text{eff}} \) is the effective ion-neutral collision frequency. The subscript FT in Eq. (2) stands for flux tube integrated quantity. The influence of the different terms in the ESF variability has been discussed briefly in Abdu (2001). We note from Eq. (2), especially, that larger flux tube integrated conductivity (the denominator of the conductivity term) reduces the instability growth rate. Even if \( \gamma_{\text{FT}} \) is sufficiently positive for the initiation of the instability the ensuing nonlinear growth of the instability into developed topside bubble structures can be retarded or even totally inhibited depending upon the magnitude of the flux tube integrated conductivity. It has been shown by Maruyama (1988) based on model calculations that trans-equatorial winds (that blow from summer to winter hemisphere) can cause significant increase of the flux tube integrated conductivity. Its effect will be to reduce instability growth rate as well as to inhibit the polarization electric field responsible for the nonlinear development of the plasma bubbles. Thus possible day-to-day variability in the intensity of trans-equatorial winds could be a significant source of ESF day-to-day variability although this effect has not been adequately established/investigated so far (see e.g. Mendillo et al., 2001; Abdu, 2001). Another source of variability for the field line integrated conductivity seems to reside in sporadic E-layers when present at the extremities of the flux tubes that participate in the ESF development processes which include also the development of the PRE. Model calculations have shown that increase of field line integrated conductivities of sufficient magnitude to affect the instability growth rate can occur due to the presence of \( E \)-layers of significant intensity (Stephan et al., 2002). However, observational evidence on the effect of \( E \)-layers on ESF development so far available lead to diverging conclusions. For example, while an isolated case of ESF inhibition in the presence of sporadic \( E \)-layer over the equatorial station Jicamarca has been reported by Stephan et al. (2002) an exactly opposite effect from the statistical data set of ESF enhancement in the presence of sporadic \( E \)-layers over an equatorial anomaly (low latitude) station, Chung-Li, has been shown by Bowman and Mortimer (2003). Such apparently ambiguous/contradictory manifestations of the \( E \)-layer–ESF relationship can be understood in the light of a direct relationship between the PRE development and the evening \( E \)-layer formation conditions that has been recently demonstrated by Abdu et al. (2003). It was shown based on analysis of extensive data sets that \( E \)-layer formation in the evening hours over a sub-equatorial station (such as Fortaleza, Brazil) can be disrupted during the development of the PRE of significant amplitude. An example of this result (taken from Abdu et al., 2003) is presented in Fig. 8 in which the right panel shows the interruption of an ongoing \( E \)-layer at the time of the maximum \( F \)-layer vertical drift (zonal electric field) near sunset, the \( E \)-layer formation resuming after 2–3 h. When the vertical drift is relatively less intense such disruption of the \( E \)-layers does not occur, as can be seen in the results plotted in the left panel. This was explained (by Abdu et al., 2003) in terms of the effect of a vertical electric field in causing vertical plasma transport at \( E \)-layer heights that can act, depending upon the \( E \)-field direction, to retard or enhance the vertical ion convergence driven by a wind/wind shear mechanism that is basically responsible for the \( E \)-layer formation. An upward-directed vertical electric field can disrupt the vertical ion convergence while a downward electric field can enhance it (see Abdu et al., 2003, for further details). The vertical
structure of the evening vertical electric field over the equator, which is associated with the PRE, as modeled by Haerendel et al. (1992) consists of an upward directed field in the height region \( \sim 90-300 \) km which reverses to a downward directed field above \( \sim 300 \) km (see Haerendel et al., 1992, for further details). When this electric field is field-line mapped to off-equatorial latitudes we have an upward-directed electric field over Fortaleza (3.9°S, 38.45°W; dip angle: \(-9^\circ\)) and a downward-directed electric field at a location farther away in latitude, such as Cachoeira Paulista (22.6°S, 315°E; dip angle: \(-28^\circ\)). As explained by Abdu et al. (2003) this situation leads to disruption of Es-layer formation over Fortaleza while uninterrupted, or even enhanced, Es-layer formation over Cachoeira Paulista during the PRE development. Results in support of this explanation are presented in Fig. 9, which shows, as an example, a few days of observations during December 1988 of simultaneous Es-layer variations over Fortaleza and Cachoeira Paulista, together with the F-layer height and the associated vertical drift variations over the former station. It is clear that Es-layer disruption over Fortaleza is accompanied by its formation over Cachoeira Paulista when the PRE development was at its peak over the former station, which is in complete agreement with the vertical electric field control of the Es-layer just explained above. We have verified that ESF was present over Fortaleza on all these nights as expected for Vzp amplitudes > 20 m/s, in agreement with the results of Fig. 6 as well. Thus we note that the nature of the association between the ESF and Es-layer occurrences to be expected in observational data would depend upon the latitude region for which the data analysis is performed. Further, it appears that a possible effect on ESF development and intensity due to an
enhanced field line-integrated conductivity arising from the mere presence of Es-layers is a challenging question warranting experimental/observational verification. In any attempt to verify this point it is necessary, as a minimum requirement, that the Es-layer, to be associated with ESF occurrence/intensity, be located in common areas in the conjugate E-layers at the feet of a potentially unstable flux tube. Further, the direct association (in an interactive manner) that exists between the

Fig. 9. F-layer height \( h'F \), vertical drift velocity \( V_z \), and Es-layer parameters, \( f_tEs \) and \( h'Es \), over Fortaleza, plotted in the top four panels and the Es-layer parameters over Cachoeira Paulista in the bottom two panels.
PRE and the Es-layer, as shown by Abdu et al. (2003), needs to be isolated in order to establish any possible effect on the ESF arising from the mere presence of the Es-layers. These considerations show that while the PW scale oscillations in the PRE amplitude is an important factor contributing to the day-to-day variability in the ESF, other factors such as the meridional/ trans-equatorial winds, field line integrated conductivities, etc., mentioned before, as well as the Es-layer connections, also contribute to its variability in ways difficult to evaluate from limited data bases.

4. Conclusions

The following conclusions may be noted from the present study: PW scale oscillations of different periodicities with episodic nature occur simultaneously in the mesospheric winds and EEJ intensity. It is not clear from the present analysis if the EEJ effects are due to direct penetration of the PW up to the dynamo region or indirectly by a mechanism of tidal wave modulation by PW interaction at lower height as found to be operative in the case of mid-latitude Es-layer responses. Nighttime F-layer height oscillations at PW periodicities are found to be associated with such oscillations in the mesospheric airglow intensity and OH rotational temperature. In this case the F-region oscillations represent the time-integrated effect of the evening vertical drift (zonal electric field) due to F-region dynamo. For the first time we have provided evidence for oscillations at PW periods in the equatorial evening zonal electric field believed to result from the combined action of thermospheric zonal wind and the evening longitudinal gradient in the E-layer Pedersen conductivity. Considerations on the relevant electrodynamic processes suggest that PW oscillations in zonal winds at E-layer height, rather than such oscillations in the zonal winds at F-region height, could be a more likely cause of the observed PW scale oscillations in the evening zonal electric field /F-region vertical plasma drift. A direct consequence of the PW scale oscillations in the evening electric field is its role in the quiet time day-to-day variability of the ESF/plasma bubble occurrence and intensity. However, other factors, such as the relationship between the PRE field and sporadic E-layer formation in the evening hours, to mention as an example, should be considered in any investigations aiming for a better understanding and evaluation of the causes of the widely observed day-to-day variability in the ESF phenomenon.

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