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Interannual variations of the mean wind and gravity wave variances in the middle atmosphere over Hawaii

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Abstract

Using simple numerical filters, wind variances with time scales of 0.1–1 h and 1–5 h are estimated from medium frequency radar wind observations at altitudes 70–94 km over Hawaii (22°N, 160°W) during the years 1990–2000. The results show interannual variations of the mean wind and variances. The variances can be attributed to the presence of atmospheric gravity waves (GWs). At altitudes below 80 km a maximum of GW intensity was observed between the years 1992 and 1996. An average seasonal cycle in wind variances is calculated by compositing over the 11 years of data. The monthly residual values, obtained after subtraction of the average seasonal variations show quasi-biennial (∼27 month) and longer period oscillations. Maximum variances are observed in 1993 and 1998–1999 at altitudes above ∼84 km. A correlation is found between GW wind variances and the Southern Oscillation Index, which in turn reflects El Niño activity in the tropical Pacific. However, more analysis with several El Niño events is required to establish the correlation with greater confidence.

Keywords: Gravity waves; Atmosphere; Dynamics; Climatology; El Niño; Medium frequency radar

1. Introduction

Gravity waves (GWs) play an important role in the dynamics of the mesosphere, lower thermosphere and ionosphere (MLTI) region. Propagating from below, they produce turbulence, deposit substantial momentum and energy and influence the general circulation, temperature structure, and composition of the middle and upper atmosphere (Andrews et al., 1987). Hence, GWs have been studied extensively during the last decades (see reviews by Fritts and Alexander, 2003; Hirota, 1997; McLandress, 1998). Considerable information about GWs has been obtained from radar measurements (Vincent, 1984; Fritts and Vincent, 1987; Ebel et al., 1987; Manson and Meek, 1993; Gavrilov et al., 1995, 1996, 1997; Manson et al., 1999; Tsuda et al., 1990; Nakamura et al., 1993a,b, 1996, etc.).

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Since September 1990, a medium frequency (MF) radar at Hawaii (22°N, 160°W) has made continuous measurements of the wind velocity in the altitude range 60–100 km. Gavrilov et al. (2003) studied seasonal variations of the mean wind and wind variances with periods of 0.1–5 h, which can be attributed to GWs.

Several recent studies have been devoted to the interannual variability of the mean winds and wave motions in the low and middle atmosphere (see, for example, Nakamura et al., 1993a,b, 1996; Burrage et al., 1996; Gage et al., 1996; Garcia et al., 1997; Jacobi et al., 1997a,b; Manson et al., 1999; Vincent and Alexander, 2000). Gavrilov et al. (1999, 2001) analyzed the data of multi-year observations with the Middle and Upper (MU) Atmosphere Radar at Shigaraki, Japan (35°N, 136°E), and with the D1 ionospheric drift velocity method at Collm observatory (52°N, 15°E). They found changes in the general circulation and GW intensity, which might be coupled with dynamical processes in the tropo–stratosphere. A number of theoretical studies (see Alexander and Holton, 1997; Sassi and Garcia, 1997; Ray

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et al., 1998; Garcia and Sassi, 1999; Medvedev and Klaassen, 2001) showed that an important contribution to MLTI dynamics may accompany various wave motions generating convective and dynamical processes in tropics.

In this paper we study the seasonal variations of the short-period wind velocity variances attributed to atmospheric GWs and their interannual variations in the middle atmosphere over Hawaii. Two simple numerical frequency filters are used to extract MF radar wind variances with time scales of 0.1–1 h and 1–5 h. This procedure is described in Section 2. Section 3 presents the interannual mean wind components and their variances. Correlations with the Southern Oscillation Index (SOI), a parameter that reflects Pacific El Niño activity, are shown in Section 4 with further discussion of the implications.

2. Data analysis

The Hawaii MF radar provides the horizontal wind velocities with time resolution of about 2 min and height resolution of 2 km. More on the system description and wind analysis can be seen elsewhere (i.e., Fritts and Isler, 1994). Using linear fitting for each altitude, we obtain hourly mean values of zonal and meridional wind velocity components and the corresponding hourly variances. These variances characterize wind perturbations with time scales up to 1 h and will be subsequently referred to as high-frequency (HF) variances. To characterize longer-period wind perturbations we take the differences between consecutive hourly velocity values at given altitudes. This filter passes harmonics with periods of 1–5 h with a maximum at a period of about 2 h. This band of wind perturbations will be referred to subsequently as low frequency (LF). The transmission functions of these filters are shown in Gavrilov et al. (2003).

At low altitudes there is a substantial number of random gaps in MF radar data records, particularly during nighttime. The lack of continuous data at these lower altitudes degrades the wind velocity and variance estimates. In this study we reject mean wind and HF variance estimates with less than 8 points in a 1 h interval. With such a criterion, the total relative number of data gaps and rejected data points is 70–90% near an altitude of 70 km. The number of data gaps is smaller in years 1991 and 2001 and it is larger in between. At higher altitudes the number of data gaps decreases and has a minimum of 10–40% for different months at altitudes 85–88 km. Therefore, the proportion of monthly data gaps is 10–60% at altitudes of 80–90 km. There are no noticeable changes in the number of data gaps due to instrumental modifications during the period of observations.

The hourly variances tend to be overestimated when the number of points in the interval is small. Therefore, wind variances below ~76 km should be treated with caution. However, we will include variances at 70–76 km in the figures because they are useful to study *relative* seasonal and interannual changes.

After obtaining hourly mean velocities and the HF and LF perturbations, we calculate monthly mean values taking into account the number of valid data velocity estimates, n_i, during each hour

$$\bar{X} = \sum_i X_i n_i \Big/ \sum_i n_i \, ,$$

where X_i are hourly mean velocities. We use these monthly mean values to study the interannual variations of mean winds and variances at altitudes of 70–94 km.

3. Interannual variations

Fig. 1 represents time–height variations of the mean wind components at Hawaii during 1990–2000, employing the method described in Section 2. Gavrilov et al. (2003) studied seasonal variations of the winds and wind variances over Hawaii. Below ~84 km the monthly mean zonal winds averaged over the years 1990–2000 have a mainly annual variation with an eastward maximum in winter and a westward maximum in summer. Above 84 km, an additional maximum of eastward wind appears in summer and the seasonal variation of the mean zonal wind contains a substantial semiannual component. The seasonal variations of the mean meridional wind in Fig. 1 are more complicated, seemingly a superposition of annual and semiannual harmonics with their phases variable in height.

Figs. 2 and 3 show interannual variations of the HF and LF wind variances as a function of time and height. We assume that the variances in these frequency bands represent the intensities of internal GWs with periods 0.1–1 h and 1–5 h, respectively. Gavrilov et al. (2003) found that the seasonal behavior of the wind variance exhibits a primary maximum in winter and a secondary maximum in summer below altitudes of ~84 km. They also showed that above ~84 km there is a tendency for the maxima of GW activity to shift to the equinoxes. The character of the interannual variations of GW intensity changes with height in Figs. 2 and 3. It should be noted that the magnitudes of HF and LF variances in Figs. 2 and 3 are probably overestimated below 75 km, where there are many data gaps (see Section 2). Therefore, the low altitude variances shown in Figs. 2 and 3 should not be compared directly with those at higher altitudes. However, they yield information about relative interannual changes of the variances at each particular altitude.

To illustrate interannual variations in more detail, Fig. 4 shows variations of the mean wind components at selected heights. Below an altitude of 80 km, Fig. 4 exhibits a winter minimum of eastward winds and a summer maximum of westward winds during 1993, corresponding to similar features observed with the MU radar in Japan (Gavrilov et al., 1999) and with LF D1 ionospheric drift measurements at Collm, Germany (Gavrilov et al., 2001). Fig. 4 also suggests that changes in the amplitude of seasonal variations

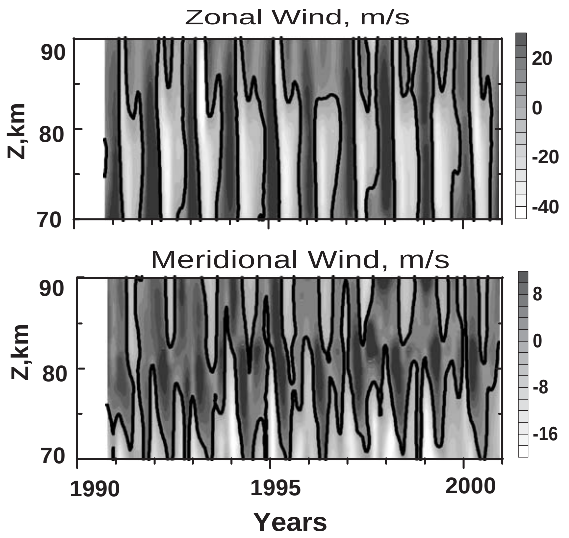

Fig. 1. Interannual variations of monthly mean zonal (top) and meridional (bottom) wind over Hawaii as a function of altitude.

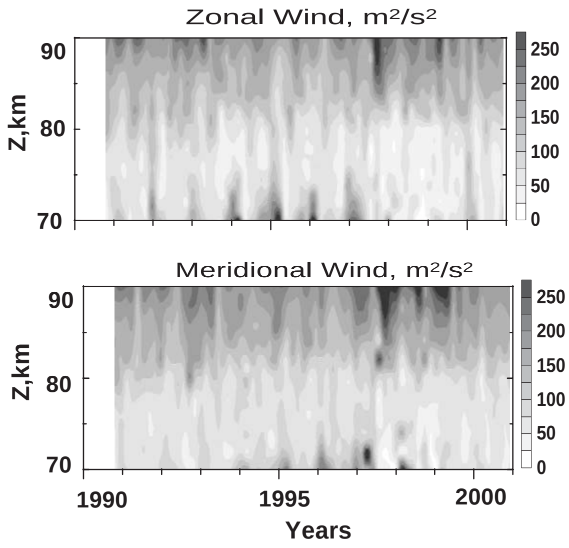

Fig. 3. Same as Fig. 1, but for LF zonal (top) and meridional (bottom) wind variances with periods 1–5 h.

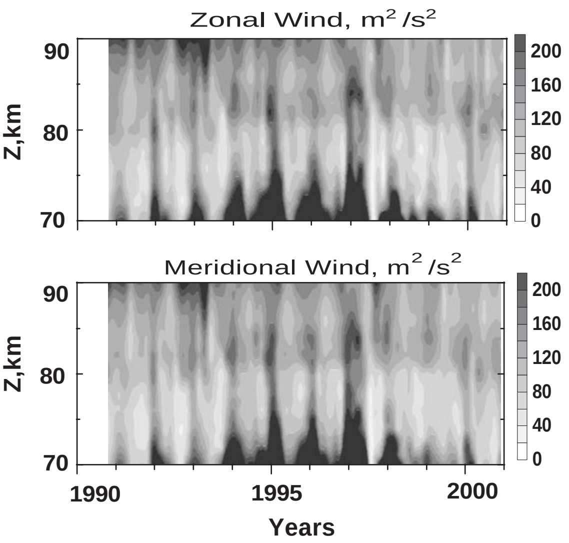

Fig. 2. Same as Fig. 1, but for HF zonal (top) and meridional (bottom) wind variances with periods 0.1–1 h.

of the mean zonal wind at low latitudes appear periodically and may be modulated by the QBO.

Figs. 5 and 6 present interannual changes of the HF and LF variances at selected heights. The character of the interannual trends of GW intensity changes with height. The trend of HF and LF variance is less noticeable near 80 km in Figs. 5 and 6. Interannual changes of zonal and meridional HF and LF variances are similar in Figs. 5 and 6. The reasons for interannual trends of GW variances could be differences in the conditions of wave generation, propagation and filtering in the middle atmosphere in different years. One should keep in mind that some of interannual trends (especially in the lower part of Figs. 5 and 6) may be partly produced by changes in the number of data gaps in different years. Such changes may be made, for example, by changes of electron content during solar cycle.

Figs. 5 and 6 show some evidences of QBO and longer-period variations superimposed on the general interannual trends. The relative amplitudes of these variations are generally larger during years 1994–1998.

To exclude possible influence of seasonal changes, we calculated inclinations of measured monthly mean values from the average seasonal variations, being the sum of annual, terannual, and semiannual harmonics over 11 years of Hawaii radar measurements. Hereafter, we will refer to these inclinations as monthly anomalies. Fig. 7 shows interannual changes of the monthly anomalies for the mean wind components. One can see negative (westward) anomalies of the mean zonal winds in years 1992–1993 at all altitudes. In these years there were weaker winter eastward flows and stronger summer westward flows, as seen in Figs. 1 and 4. Another negative anomaly of the mean zonal wind seen in Fig. 7 occurs during Winter/Spring of 2000 below 90 km. The mean meridional winds in Fig. 7 have a maximum in the northern direction during 1997–1998. One of the reasons for the mean wind variations could be dynamical GW impact. In quasi-steady-state conditions zonally and meridionally propagating GWs influence meridional and zonal mean winds, respectively (Andrews et al., 1987). Figs. 8 and 9

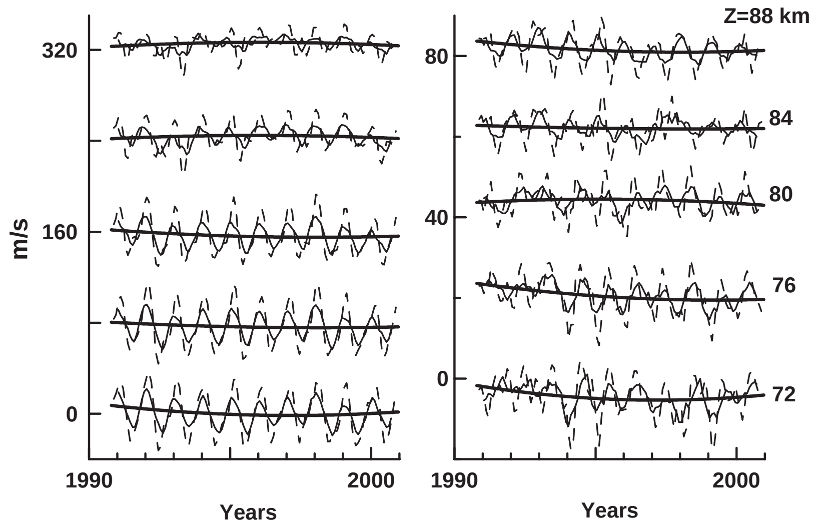

Fig. 4. Monthly mean zonal (left) and meridional (right) winds at different altitudes (right numbers) over Hawaii smoothed running averages of 3-month (dashes) and 6-month (thin solid lines). Thick parabolas show quadratic trends. An offset of 80 m s^{-1} is added to each consecutive curve.

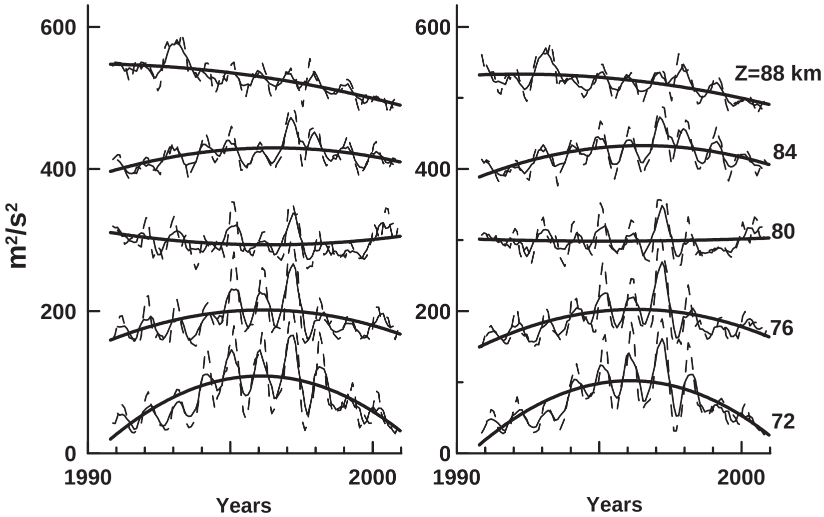

Fig. 5. Same as Fig. 4, but for HF zonal (left) and meridional (right) wind variances with periods 0.1–1 h. An offset of 100 m^2 s^{-2} is added to each consecutive curve.

exhibit maxima of GW variances in years 1992–1993 and 1997–1998, which could produce corresponding changes in the mean winds observed in Fig. 7 (see also discussion below).

Fig. 8 presents changes of the monthly anomalies of HF wind variances with periods of 0.1–1 h. One can see that general trends are different at different altitudes in Fig. 8. At altitudes of 92 and 88 km we observe a general decrease in both zonal and meridional variances that is less obvious at lower altitudes. Local maxima of the variances are superimposed on the general trends in Fig. 8. They may have quasi-bienniel and longer periods. One can see possible examples of QBO influences in Fig. 8 at altitudes 80–84 km with maximum wind variances in winters during the years 1993, 1995, and 1997. At the same time, other causes of interannual changes of wind variance almost surely exist. They may enhance or produce additional maxima, like the 1993 maximum at altitudes 88–92 km or a shift from the 1997 maximum (seen at altitude 80 km in Fig. 8) to the maximum in the year 1998 at higher altitudes.

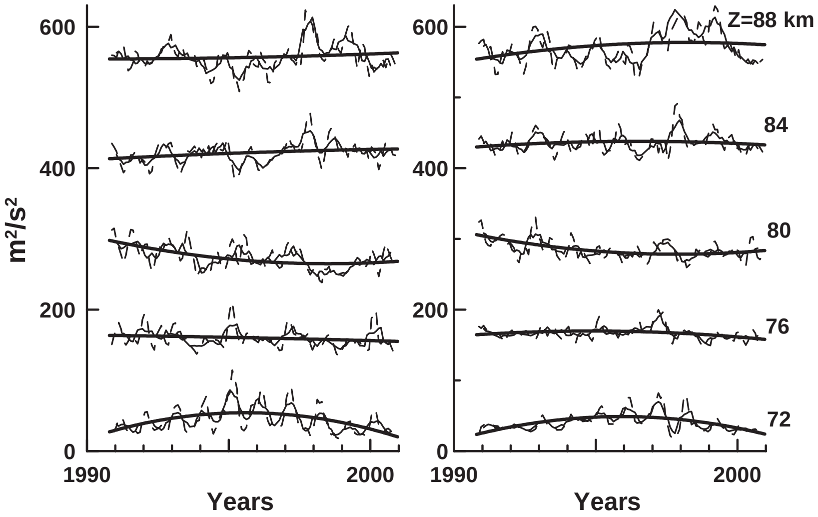

Fig. 6. Same as Fig. 4, but for LF zonal (left) and meridional (right) wind variances with periods 1–5 h. An offset of 100 m^2 s^{-2} is added to each consecutive curve.

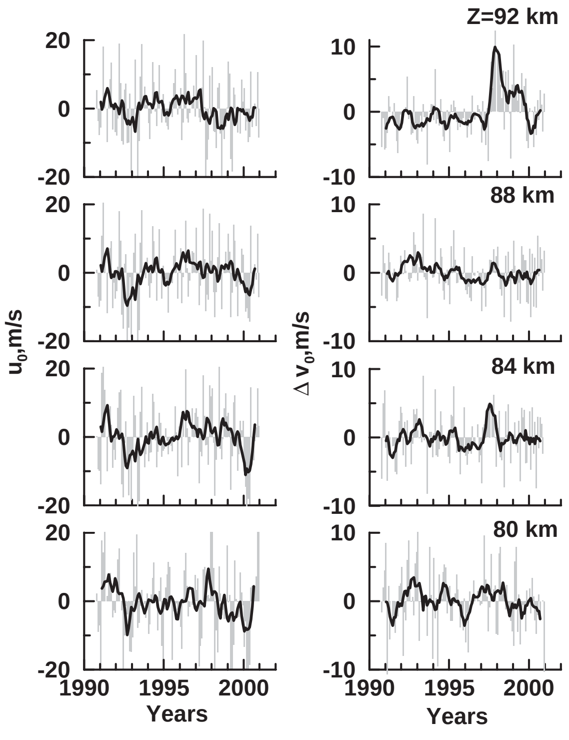

Fig. 7. Inclinations (gray vertical bars) of monthly mean zonal (left) and meridional (right) winds from their average seasonal variations at different altitudes over Hawaii. Solid lines show 6-month running averages.

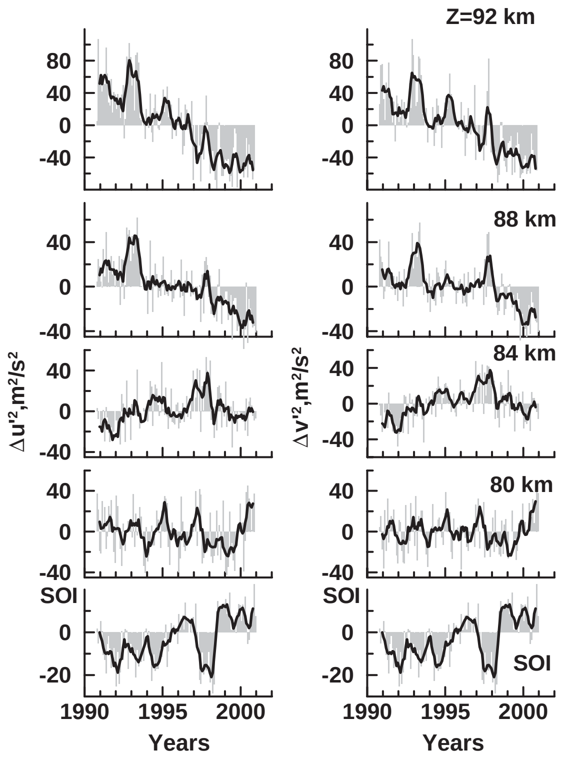

Fig. 8. Same as Fig. 7, but for HF zonal (left) and meridional (right) wind variances with periods of 0.1–1 h. Bottom plots show the SOI.

Fig. 9 exhibits monthly anomalies of LF wind variances having time scales 1–5 h. One can see variations similar to Fig. 8 with QBO maxima in the years 1993, 1995 and 1997 at altitudes near 80 km with larger amplitudes in the zonal component. This maximum for the longer period component in Fig. 9 is broader than the corresponding maximum

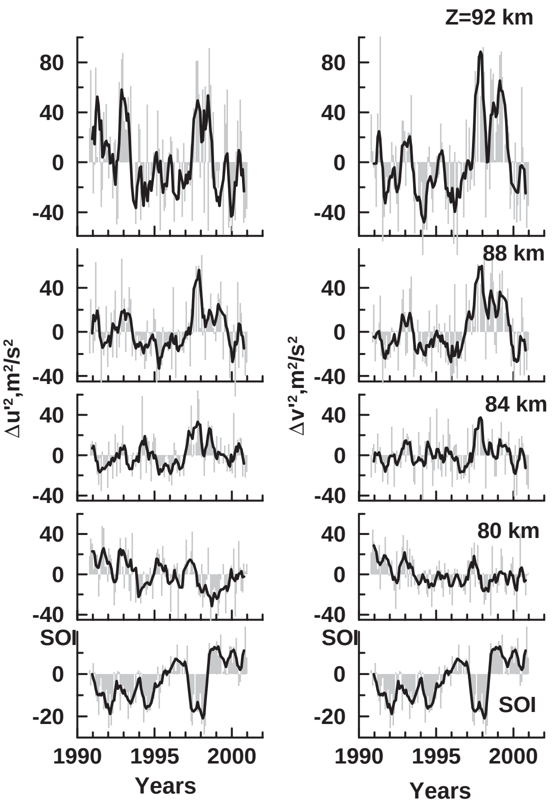

Fig. 9. Same as Fig. 7, but for LF zonal (left) and meridional (right) wind variances with periods of 1–5 h. Bottom plots show the SOI.

of the shorter-period component in Fig. 8. The reason could be different dispersion of propagation of longer-period GW energy from lower atmospheric sources. Therefore, LF variances in Fig. 9 may contain superpositions of GWs coming from much wider source regions than the HF variances in Fig. 8.

A comparison of interannual changes of the mean wind shown in Fig. 7 with the variances in Figs. 8 and 9 reveals a correlation of the 1993 GW intensity maximum with a decrease in the mean zonal wind. During the 1998–1999 GW maxima in Figs. 8 and 9, there was an increase in the northward meridional wind. Thus, variations of the mean wind and GW intensity appear correlated and may be suggestive of wave-mean flow interactions.

4. Discussion

Gavrilov et al. (2003) analyzed Hawaii MF radar observations and found semiannual changes of wind variances with periods of 0.1–5 h having maxima at equinoxes at altitudes above 85 km that appear correlated with the semiannual changes observed in the tropical tropo–stratosphere by Ray et al. (1998). The latter authors analyzed gravity wave climatologies in the tropics from the United Kingdom Meteorological Office (UKMO) global assimilated model meteorological data and found the maxima of gravity wave zonal forcing in March–April and August–October at pressure level 2.1 mbar at a latitude of 22°N. Therefore, GWs may propagate to the MLTI region from tropo–stratospheric sources. At low altitudes their amplitudes are small, but they grow approximately exponentially versus height. Near altitudes of 85 km and above, the amplitudes may become larger than background atmospheric variations and the waves become prominent in Hawaii radar data, suggesting semiannual variations of their tropo–stratospheric sources.

Numerical modeling by Gavrilov et al. (2003) reproduced several of the main features of the observed seasonal variations of GW activity in the MLTI region over Hawaii, including their changes as a function of altitude observed with the MF radar. In the numerical model, seasonal variations were caused by changes in the background atmospheric fields and in the strength of GW hydrodynamic sources, which depend on the mean wind. Numerical modeling showed that the contribution of GW sources located in the atmosphere below an altitude of 20 km may be the most important for GW wind variances at altitudes of 70–90 km.

The results of the mean wind observations with the Japanese MU radar at altitudes of 65–80 km in 1983–1999 by Gavrilov et al. (1999) show a decrease in the winter eastward velocities during 1990–1994, which corresponds to a maximum of short-period wind variance attributed to atmospheric GWs. Our results with Hawaii radar data reveal a similar decrease in the winters of 1992 and 1993 at all altitudes in Fig. 7. Minimum eastward velocities were observed in the winters of 1992 and 1993 and maximum westward velocity in Summer 1993. Analysis of the mean zonal wind measured with the HRDI instrument aboard the UARS satellite also showed smaller eastward velocity in Winter 1993 and very high westward velocity in Summer 1993 in the tropical MLTI (Burrage et al., 1996; Garcia et al., 1997). During 1992–1993 one can see maxima of wind variances in Figs. 8 and 9, especially in winter above altitude of an 85 km.

Consideration of these figures show that the 1992–1993 minima of winter westerlies and the wind variances appeared during the longest El Niño event in history (negative SOI), that occurred from 1990 to 1994 with the largest maximum in 1992. The Pinatubo volcano eruption in June 1991 might also play a role in causing changes in the thermal regime and circulation of the middle atmosphere.

One of the possible reasons for the interannual variations of the mean wind and wind variances at low latitudes could be variations of the atmospheric circulation caused by periodic warming (El Niño) and cooling (La Niña) of the tropical central Pacific. According to recent knowledge, the El Niño and La Nina events may produce changes in circulation and the strength of wave sources in the tropical tropo–stratosphere. Such variations might also produce interannual

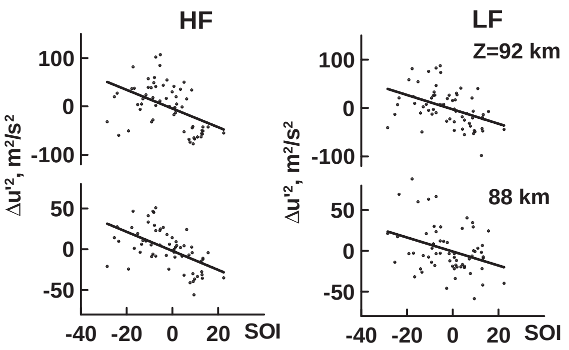

Fig. 10. Anomalies of monthly variances of zonal wind from their average seasonal variations with time scales of 0.1–1 h (left) and 1–5 h (right) plotted versus the SOI in winter at different altitudes over Hawaii. Straight lines show linear regressions to the measured points.

changes in the mean winds and wave activity in the middle atmosphere of tropical and middle latitudes. An indicator of El Niño effects in the atmosphere is the SOI reflecting the surface pressure difference between the tropical observation points of Tahiti and Darwin. Positive SOI values correspond to La Nina (cool central Pacific) and negative SOI to El Niño (warm central Pacific) events. Zonal pressure gradients imply eastward or westward wind directions of the lower atmosphere near the equator, where the Coriolis force tends to zero. Variations of SOI during analyzed Hawaii MF radar observations are shown in the bottom plots of Figs. 8 and 9. Possible reasons for correlations between wind variances in the MLTI and SOI could be changes in the conditions of wave generation and propagation due to changes in atmospheric dynamics caused by variations of Pacific surface temperatures.

Monthly anomalies of HF and LF zonal wind variances about average seasonal values versus monthly mean SOI values are shown in Fig. 10. One can see a regression with larger GW intensity during El Niño (negative SOI). Similar regressions for the meridional wind component for summer, and for lower altitudes are weaker than those shown in Fig. 10. Straight lines in Fig. 10 show least-squares fits to the experimental points with a linear function $\Delta u'^2 = \alpha \mathrm{SOI} + \beta$, where α and β are constants. The values of α and β for all plots in Fig. 10 are shown in Table 1. According to Hald (1952a), a random value $t = \alpha/\delta_\alpha$ (where δ_α is the standard deviation of α) has a t-distribution of probability with $f = M - 2$ degrees of freedom, M being the total number of monthly values. In our study, for the winter season, $M = 60$. Calculating the values of t, and using statistical tables (Hald, 1952b; Lowry, 2003), one can evaluate the probabilities P_α of the hypothesis that the regression coefficients α have nonzero values. The probabilities (P_α) obtained are given in Table 1 and have high values, especially for HF GWs.

Table 1
Parameters of the regression lines

Height (km)	α	β	α/δ_α	P_α	r	P_r
HF GWs						
88	−1.15	−0.87	5.0	> 0.999	−0.55	> 0.999
92	−1.95	−6.76	4.2	> 0.999	−0.47	> 0.999
LF GWs						
88	−0.83	2.19	2.7	> 0.99	−0.32	> 0.98
92	−1.50	−4.32	3.3	> 0.99	−0.42	> 0.99

$\Delta u'^2 = \alpha \mathrm{SOI} + \beta$ from Fig. 10, correlation coefficients, r, and nondirectional probabilities, P_α and P_r, of statistical hypotheses of nonzero values of α and r, respectively.

Another indication of the connection between GW variances and the SOI arises from the consideration of correlation coefficients, r. These are presented in Table 1. A significance test (see Lowry, 2003) estimates the probabilities, P_r of nonzero values of the correlation coefficients shown in Table 1. The probabilities have high values similar to P_α. Both probabilities P_α and P_r in Table 1 show a high reliability of possible correlation between observed GW variances and SOI at altitudes above 85 km.

Table 1 shows a better correlation between GW intensity and SOI for HF than for LF components. The reason may be connected with differences in the angles of GW energy propagation from tropospheric sources to the MLTI region. LF wave energy should propagate at smaller angles to the horizon and reach the upper atmosphere at longer distances from a lower atmosphere wave source than the HF GW energy. Therefore, LF variances contain superpositions of GWs coming from more widely distant areas (which may include higher latitudes) than the HF variances in Figs. 8–10 and Table 1.

One may assume that westward-propagating waves generated over the tropical and subtropical Pacific during El Niño may propagate to the mesosphere at middle latitudes, producing larger than normal westward wave drag, and likely modify the general circulation of the middle atmosphere. Studies of GW interannual variations in the tropical lower stratosphere from radiosonde data by Vincent and Alexander (2000) reveal the largest westward momentum fluxes in the winters of 1992 and 1993 at altitudes 18–25 km during the entire period of their analysis from 1992 to 1998. Gage et al. (1996) illustrated the mechanism of wave drag with radar observations of the stratospheric circulation at Christmas Island. They found the smallest zonal wind seasonal variations in 1991 and 1992. Our observations of a similar minimum of eastward wind over Hawaii in the winter of 1992–1993 and a corresponding maximum of westward wind in Summer 1993 in Fig. 7 allows us to hypothesize the existence of a time lag in the response of the tropical stratosphere and subtropical MLT to Pacific surface temperature changes.

Larger responses of the zonal mean wind to this El Niño event (see Fig. 7) may be connected with co-phase maxima of the wave accelerations and QBO wind. Analysis of HRDI and Christmas Island MF radar data by Burrage et al. (1996) showed that during 1992–1993 there was a "negative" phase of QBO oscillation at altitudes 82–90 km with a preference for westward winds in all seasons. One can also see a negative wind minimum in 1993 at altitudes near 80 km also in Fig. 7. This minimum is simultaneous with the 1993 GW variance maxima in Figs. 8 and 9. Probably, a superposition of very long El Niño event, the Pinatubo eruption and the negative phase of the quasi-biennial oscillation induced a larger response of the MLT circulation in Winter 1992–1993 compared to 1997–1998.

The better correlation between HF and LF wind variances and the SOI obtained in winter than in summer appears to confirm our assumption about an increased generation of westward GWs during El Niño. Penetrating to the MLT over Hawaii from below in winter, westward GWs propagate opposite to the eastward mean wind at all altitudes. They generally do not encounter critical levels and may grow to large amplitudes before saturation. In summer, the vertical structure of the zonal circulation in the middle atmosphere is more complicated with a strato–mesospheric westward jet stream. In this flow, westward propagating GWs will likely dissipate near critical levels and may be constrained by saturation to smaller amplitudes. As a result, westward propagating GWs will enter the summer MLT more reduced in amplitude, exhibiting a smaller correlation with the SOI than observed in winter.

There are a number of theoretical studies of the wave origin of the QBO and semiannual wind oscillations in the low-latitude atmosphere (for example, Dunkerton, 1982; Mengel et al., 1995; Alexander and Holton, 1997; Sassi and Garcia, 1997; Ray et al.,1998; Garcia and Sassi, 1999; Medvedev and Klaassen, 2001). The main concept in these theories is the production of long-period oscillations of the zonal mean atmospheric circulation due to the differential mean flow interaction of eastward and westward propagating waves. Previously, Gavrilov et al. (2003) showed that the transition from annual to semiannual variations of the zonal mean wind over Hawaii occurs at altitudes above 83–85 km, where we observe correlations of HF and LF wind variances with the SOI in Fig. 10. Fig. 7 shows increasing GW intensities above 83–85 km. At lower altitudes their amplitudes are smaller than existing atmospheric noise. This may explain the reduction in the correlation of GW variances to the SOI at lower altitudes.

The interval of MF radar observations at Hawaii spans only about 11 years, and only two El Niño events occurred during this period. Therefore, our apparent correlations of GW intensity and the SOI are preliminary and further study is needed to establish the correlations with greater confidence.

5. Conclusions

The mean zonal wind has a largely annual variation with maximum eastward wind in winter and westward wind in summer below 82–85 km and a largely semiannual variation with an additional maximum of eastward wind in summer above. Seasonal variations of the mean meridional wind exhibit a superposition of annual and semiannual harmonics with their phases variable in altitude. Variances of the GWs with periods of 0.1–5 h exhibit seasonal changes with the primary maxima in winter and secondary maxima in summer. The latter is more noticeable for the HF component. Variances of GWs with periods 0.1–5 h reveal a transition from seasonal changes with the main maxima in winter and secondary maxima in summer below altitudes of 82–85 km.

The amplitudes of annual variations of the mean zonal and meridional winds are different in different years and are modulated by the QBO and other multi-year variability. Interannual changes in HF and LF variances have maxima between the years 1994–1998 at altitudes below 75 km. The character of interannual variations of GW intensity changes with altitude and is modulated by oscillations with periods of 2–7 years. Correlations of the wind variances with the SOI suggest increased activity of atmospheric wave sources during El Niño.

Acknowledgements

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