

## Quasi-Biennial and Other Long-Period Variations in the Solar Semidiurnal Barometric Oscillation: Observations, Theory and Possible Application to the Problem of Monitoring Changes in Global Ozone

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### ABSTRACT

A 21-year record of monthly mean determinations of the solar semidiurnal surface pressure oscillation [ $S_2(p)$ ] at Batavia (6.2°S) was analyzed to detect long-period variability. When the  $S_2(p)$  determinations were resolved into components which peak at local midnight (and noon) and 0900 (and 2100) local solar time, considerable evidence was found for a quasi-biennial variation in the 0900 component (but not in the midnight component). It is shown that this is consistent with the expected response of  $S_2(p)$  to the familiar quasi-biennial oscillation of the tropical stratosphere.

Also apparent in the record is a very long term trend in  $S_2(p)$ . It is suggested that this may be an indication of a similar trend in stratospheric ozone, and the possibility of using the surface pressure oscillation in monitoring long-term changes in atmospheric ozone is discussed.

### 1. Introduction

During the past two decades a reasonably complete understanding of the dynamics of thermally forced atmospheric tidal motions has been achieved (Chapman and Lindzen, 1970; Lindzen, 1979; Forbes and Garrett, 1979). In particular, it is well established that the most important excitation for the solar semidiurnal tide is the absorption of solar radiation by ozone in the stratosphere and mesosphere (Butler and Small, 1963; Lindzen, 1968). The response to such forcing can be calculated with considerable accuracy using linear tidal theory. The results of numerical tidal calculations indicate that the direct absorption of solar radiation by ozone and other atmospheric constituents ought to produce a semidiurnal surface pressure oscillation [ $S_2(p)$ ] with an amplitude somewhat greater than 1 mb in the tropics. This feature is in good agreement with observations, but the calculations also suggest that the semidiurnal pressure variation should have maxima just before 0900 (and 2100) local solar time (LST), while the observed phases at tropical stations are generally in the range 0940–1000. The theoretical prediction for the phase of  $S_2(p)$  is not significantly affected by such details as the inclusion of eddy viscosity or mean winds in the tidal calculations (Lindzen and Blake, 1971; Lindzen and Hong, 1974). Fig. 1 summarizes this problem in harmonic dial form. The solid arrow pointing to ~0945 represents the observed  $S_2(p)$ , while the arrow pointing to ~0900 represents the  $S_2(p)$  that is expected to result from the semidiurnal component of direct solar heating of the atmosphere. Lindzen (1978) showed that a tropospheric semidiurnal vari-

ation in latent heat release which peaks at about 0300 (and 1500) LST will produce a contribution to  $S_2(p)$  that will have maximum amplitude at about 0000 (and 1200) LST. If this latent heat excitation has an appropriate magnitude then it can reconcile theory and observations of  $S_2(p)$ , as shown in Fig. 1. Lindzen was able to show that hourly rainfall data at several tropical stations contained a semidiurnal component with the necessary amplitude and phase (assuming that rainfall and latent heat release from condensation are nearly simultaneous). Hamilton (1981a,b) confirmed Lindzen's rainfall results for a much larger number of stations. He also showed that the latent heat forcing mechanism could be used in an explanation of the observed seasonal cycle of  $S_2(p)$  in the subtropical and midlatitude regions.

If the current view of the dynamics of the solar semidiurnal tide is correct then it appears inevitable that  $S_2(p)$  will have a quasi-biennial variation caused by the quasi-biennial oscillation (QBO) of the tropical stratosphere. The well known QBO variations in ozone (Funk and Garnham, 1962; Oltmans and London, 1982), zonal wind (Reed *et al.*, 1961) and temperature (Shah and Godson, 1966) should all affect the tidal barometric oscillation. Observations of the quasi-biennial signal in  $S_2(p)$  would be of interest for at least three reasons. First, such observations might cast some light on the validity of the current theoretical model of the tidal dynamics. In particular, the QBO might be expected to primarily affect the component of  $S_2(p)$  parallel to the 0900 axis of Fig. 1, since this component is believed to be largely excited in the stratosphere. Second, the discovery of a QBO in the years before direct measurements of tropical

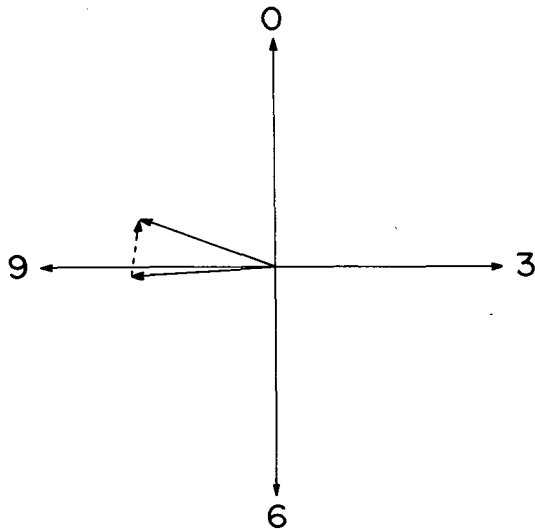


FIG. 1. Harmonic dial representation of the observed  $S_2(p)$  in the tropics (solid arrow pointing towards 0945) and the calculated  $S_2(p)$  resulting from the direct absorption of sunlight in the atmosphere (solid arrow pointing toward 0900). Observations and theory can be reconciled if there is an additional source of excitation for the semidiurnal tide that produces the  $S_2(p)$  response shown by the dashed arrow.

stratospheric winds and temperatures are available (i.e., before about 1950) would provide some evidence that the stratospheric QBO itself must have been present. Finally, a demonstration that biennial variations in the stratosphere can affect  $S_2(p)$  might open the possibility that other kinds of long-term changes in the stratosphere (e.g., ozone variations) could be detected in the record of surface pressure observations.

Evidence that there may actually be a QBO in  $S_2(p)$  is provided in a remarkable unpublished report by Holloway *et al.* (1955). They examined a 21-year (1924–44) record of monthly mean determinations of  $S_2(p)$  at Batavia [now Djakarta (6.2°S, 106.8°E)] and found an indication of a 26-month periodicity in the amplitude. More precisely, they determined that the mean of the squared amplitudes of nine values of  $S_2(p)$  spaced at 26-month intervals starting from December 1925 was 5.9% higher than the mean square amplitude for the entire record. Similarly, the mean value for the ten squared amplitudes at 26-month intervals starting from May 1924 was 4.7% below the mean for the whole record. Thus these results suggest the existence of a 26-month oscillation in the magnitude of  $S_2(p)$  with peak-to-peak amplitude of ~5% of the mean (since the range of the variation in squared amplitude is ~10%). Holloway *et al.* performed their work several years before the discovery of the stratospheric QBO, and they hypothesized that their oscillation might be caused by an oscillation in solar activity. Brier (1966) discussed the findings of Holloway *et al.* as being the possible

result of an approximately 27-month modulation of the lunar-solar gravitational tidal forcing. Today it seems most natural to regard the 26-month oscillation in the Batavia observations (if it is real) as a consequence of the influence of the stratospheric QBO on the excitation and propagation of the semidiurnal tide. In Section 2 of the present paper the time series of the Batavia  $S_2(p)$  data is reexamined in light of the current understanding of tidal dynamics. In Section 3 the expected response of  $S_2(p)$  to the stratospheric QBO is examined theoretically. Section 4 is a discussion of the possible application of observations of  $S_2(p)$  to the problem of monitoring long-term changes in stratospheric composition. The conclusions are summarized in Section 5.

2. Data analysis

Holloway *et al.* (1955) give values of the monthly mean components of  $S_2(p)$  parallel to the 0900 and 0000 axes of Fig. 1 for each of the 252 months between January 1924 and December 1944, inclusive. They were derived from published bi-hourly values of the surface pressure measured at the Dutch Royal Magnetic and Meteorological Observatory at Batavia. The 21-year mean values determined from these monthly averages are 1.267 mb for the 0900 component and 0.469 mb for the 0000 component. This is equivalent to an amplitude of 1.351 mb and a phase of 0941, which compare well with the 40-year mean values (based on data before 1917) of 1.325 mb and 0940 quoted by Haurwitz (1956). The first step in the present analysis of the Batavia  $S_2(p)$  data was the elimination of the mean and annual cycle from both the 0900 and 0000 component records. This was accomplished by subtracting the 21-year mean values for each month of the year from the individual monthly observations. Figs. 2 and 3 show seven-month running means of the resulting anomaly series for the 0900 and 0000 components, respectively. A roughly biennial variation is detectable by inspection of the 0900 series.

Both anomaly series were detrended by removing the least-squares linear fit and were smoothed by taking seven-month running means. The discrete Fourier transforms of the resulting time series were then computed, i.e.,

$$J_k(\omega) = (2/N) \sum_{n=1}^N P_k(n) \exp[i(n-1)\omega],$$

where  $J(\omega)$  is the transform at angular frequency  $\omega$  (expressed in inverse months), the subscript  $k$  labels either the 0900 ( $k = 9$ ) or 0000 ( $k = 0$ ) series,  $P(n)$  is the anomaly at the  $n$ th month of the record, and there are a total of  $N$  months. Fig. 4 shows the squared magnitude of the transform of the 0900 series. There is a great deal of randomness apparent in this figure but there does seem to be a concentration of power

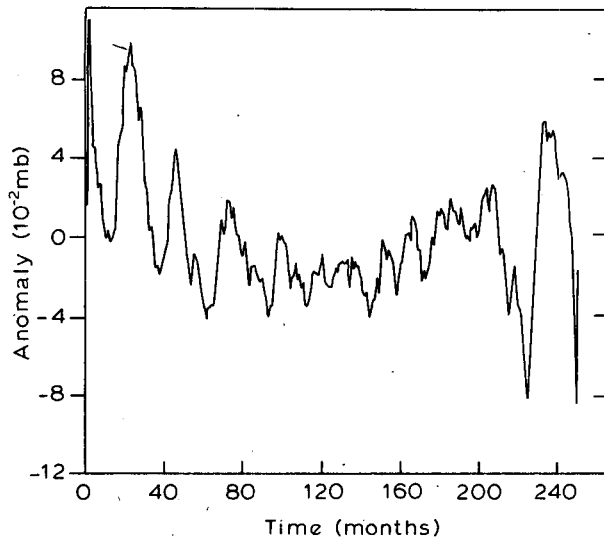


FIG. 2. Seven-month running means of the anomaly in the 0900 component of the Batavia  $S_2(p)$  record during the period 1924-44. See text for details.

at around QBO periods. It should be noted that the arrows in this figure (and in succeeding figures) show the frequencies corresponding to periods of 24 and 30 months; over the last 30 years, periods for the zonal wind QBO have been observed to range between about 22 and 33 months (Quiroz, 1981). This is more clearly seen in Fig. 5 which shows a slightly smoothed version of  $|J_9(\omega)|^2$ . An interesting comparison is provided by Fig. 6 which is the squared magnitude of the Fourier transform of an eight-year (1974-81) time series of monthly mean values of the

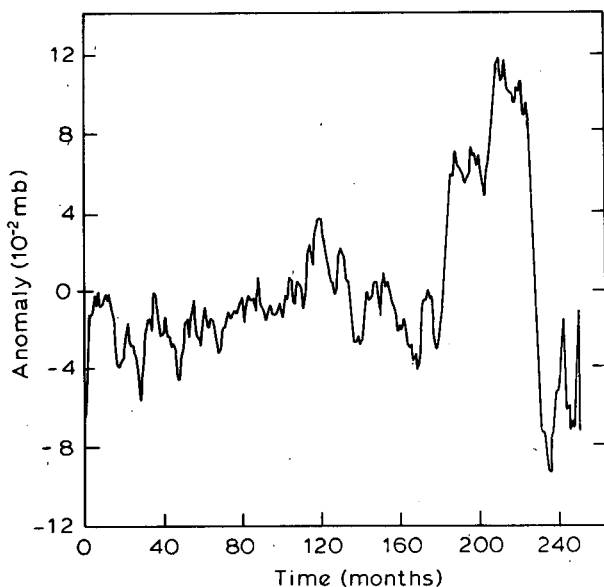


FIG. 3. As in Fig. 2 but for the 0000 component.

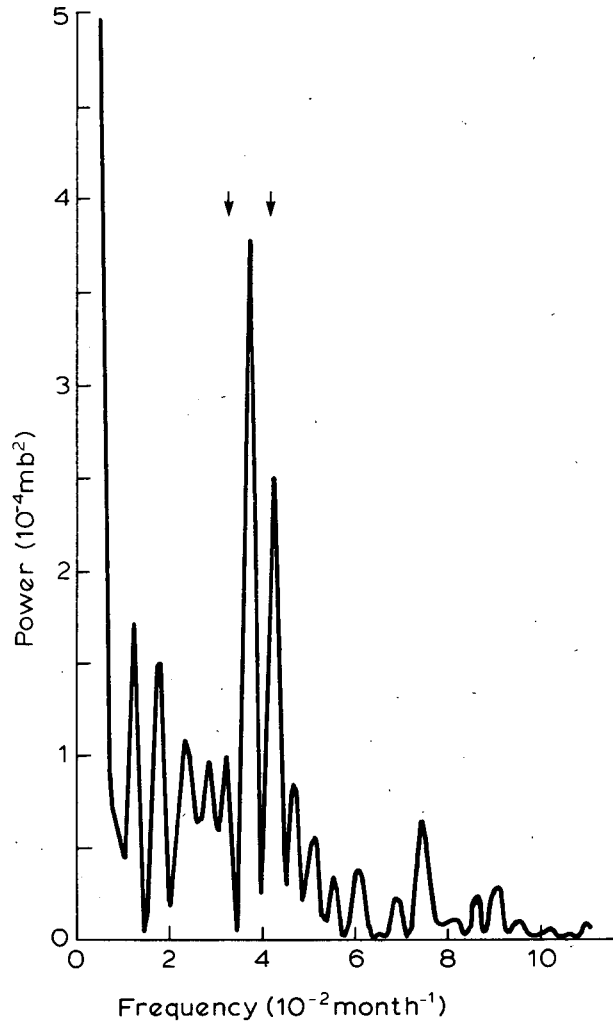


FIG. 4. Squared magnitude of the discrete Fourier transform of the 0900 anomaly series.

zonal wind at Ponape ( $7^\circ\text{N}$ ,  $158.2^\circ\text{E}$ ) where the QBO in the wind is very well defined.

The statistical significance of the QBO peak in Fig. 4 can be judged using Fisher's test (Fisher, 1929; Nowroozi, 1967). Fisher's test statistic is the ratio of the largest value of  $|J(\omega)|^2$  at the Fourier frequencies (i.e., frequencies corresponding to periods which are some integral fraction of the total record length) to the sum of  $|J(\omega)|^2$  at all the Fourier frequencies. Since Fisher assumed that all the elements of the original time series represent independent observations, it is necessary to use the Fourier transform of the original unsmoothed anomaly series (rather than the transform of the slightly smoothed series that is displayed in Fig. 4). In the present case the largest  $|J_9(\omega)|^2$  near QBO frequencies is associated with a ratio of 0.047. From the tables in Nowroozi (1967) one finds that for a time series with  $N = 252$  a test statistic greater than 0.055 is significant at the 90% level. The null

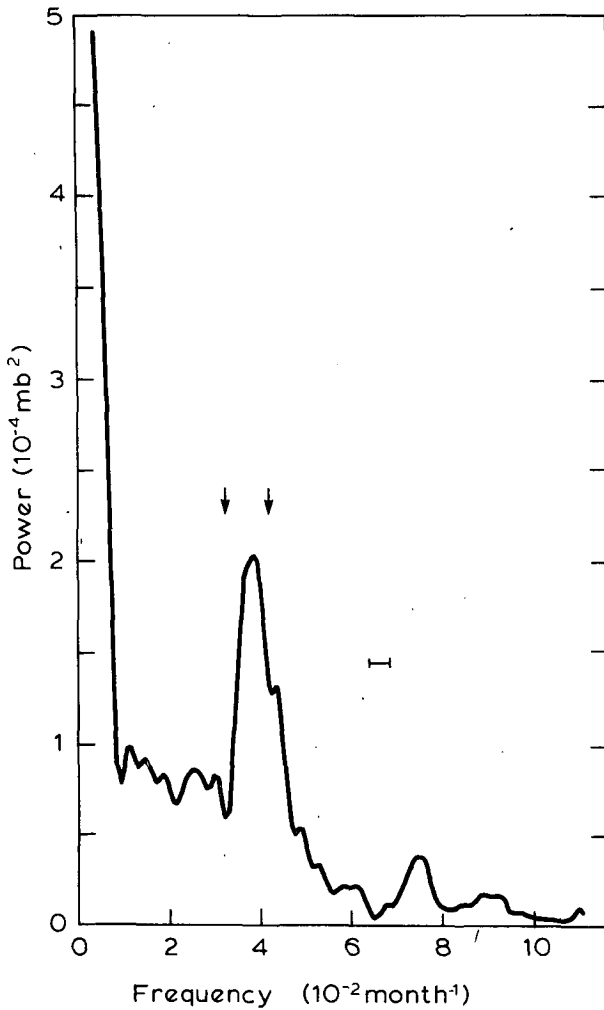


FIG. 5. As in Fig. 4 but smoothed by taking a running mean in frequency. The frequency interval used for the smoothing is indicated by the horizontal bar.

hypothesis for Fisher's test is a series of randomly distributed noise; this is probably not a reasonable hypothesis for the anomaly series which (as one can see from Fig. 2) displays well defined long-term trends. Thus Fisher's test might be more appropriate for a version of the anomaly series that has had some of the very low frequency variability removed. If the least-squares quadratic fit is subtracted from the series before the spectrum is calculated, then the largest test statistic is raised to more than 0.051.

In a formal sense the QBO peak in the spectrum of the 0900 anomaly series appears to fall just short of 90% statistical significance. Given sound theoretical reasons for expecting QBO influence on  $S_2(p)$  one might regard this as support for the reality of a QBO in  $S_2(p)$ . Put in statistical terms, Fisher's test reveals that there is a somewhat greater than 10% chance that some peak in the spectrum of a noise series will be as large as the QBO peak that was actually found in

the Batavia data. The random probability that this large peak should happen to fall within the anticipated period range is much less, of course.

Fig. 7 shows the spectrum of the anomaly series in the 0000 component,  $|J_0(\omega)|^2$ . There is a great deal of low-frequency variability apparent, but there is almost no power whatsoever at QBO frequencies. Thus it seems that the "26-month cycle" discussed by Holloway *et al.* (1955) should probably be regarded not as a variation in  $S_2(p)$  amplitude as such, but rather as an oscillation in the 0900 component of  $S_2(p)$ .

The problem of determining the amplitude of the quasi-biennial variation in the 0900 component of the Batavia  $S_2(p)$  record is necessarily a difficult one, given that the oscillation itself may not be statistically significant. Simple inspection of Fig. 2 suggests that a peak-to-peak amplitude of  $\sim 0.04$  mb might be appropriate. This value is also roughly consistent with the results of the simple analysis of Holloway *et al.* (1955) discussed in Section 1.

In the next section it will be argued that the quasi-biennial variation in  $S_2(p)$  found in the Batavia data is consistent with the expected influence of the fa-

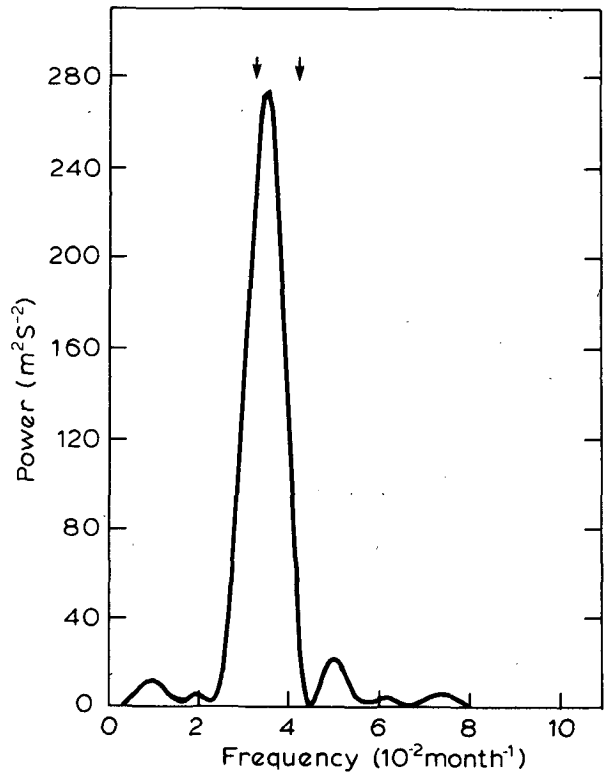


FIG. 6. Squared magnitude of the Fourier transform of the series of monthly mean zonal wind observations for January 1974–December 1981 at Ponape. The data were taken from *Monthly Climatic Data for the World* published by the National Climate Center. The mean, annual cycle and linear least-squares trend were removed from the time series before the transform was calculated.

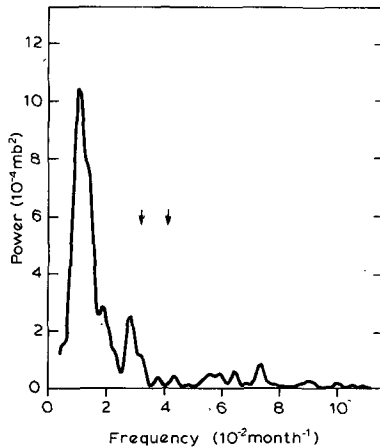


FIG. 7. As in Fig. 4 but for the 0000 component.

miliar stratospheric QBO on the semidiurnal tide. However, it is worthwhile to emphasize that there is as yet no direct observational evidence to link the stratospheric QBO and the QBO in  $S_2(p)$ . The most important extension of the present observational study would be to include an  $S_2(p)$  record extending well past 1950, so that the correlation between the zonal wind QBO and  $S_2(p)$  could be directly examined.

### 3. Theoretical considerations

#### a. Review of classical tidal theory applied to $S_2(p)$

Atmospheric tides below the mesopause ought to be reasonably well described by the so-called classical tidal theory. The classical theory assumes that the tidal fields can be treated as linear perturbations about a basic state which is characterized by an absence of mean winds and by a temperature structure that is a function only of height. The earth's spherical geometry is considered but topography is ignored. The infrared cooling associated with the tidal temperature variations can be included within a Newtonian cooling approximation, but this has only a very small effect on the surface pressure oscillation (Lindzen and McKenzie, 1967). No other dissipative processes are considered. When all these approximations are made the equations governing the height and latitude dependence of the tidal fields can be separated. The solution for  $S_2(p)$  at the surface can then be written as a linear combination of Hough functions; in the standard notation this is

$$S_2(p) = \sum_{n=2} p_n \Theta_{2,n}(\theta) \exp(i\sigma t_{loc}),$$

where the Hough functions  $\Theta_{2,n}$  are solutions of the homogeneous Laplace tidal equation,  $\theta$  is the latitude,  $\sigma$  is the semidiurnal frequency,  $t_{loc}$  is LST, and the

$p_n$  must be determined by solving a vertical structure equation which includes the projection of the semidiurnal component of the heating onto  $\Theta_{2,n}$  as an inhomogeneous term [see Chapman and Lindzen (1970) for a more complete description].

The most striking aspect of realistic tidal theory solutions for the semidiurnal tide is the dominance of the gravest Hough function  $\Theta_{2,2}$  in the surface pressure response [at the equator  $\Theta_{2,2}$  can account for  $\sim 98\%$  of the total calculated  $S_2(p)$ ]. Fig. 8 shows  $\Theta_{2,2}$  as a function of latitude. This particular Hough function is so important for two reasons. Except near the time of the solstices the thermal excitation will peak in the tropics and thus much of this forcing will project onto the similarly shaped  $\Theta_{2,2}(\theta)$ . Even more significant is the fact that the vertical wavelength in the vertical structure equation associated with  $\Theta_{2,2}$  is extremely long ( $>200$  km) and thus this mode is very efficiently excited by a thermal forcing of broad vertical extent such as the ozone heating (e.g., Chapman and Lindzen, 1970).

In order to assess the possible impact of the ozone QBO on  $S_2(p)$ , classical tidal theory calculations were performed with thermal excitations appropriate for the extrema in the ozone oscillation. The mean ozone distribution and the radiation model used for calculating solar heating rates are discussed in the next subsection. The theoretical results for the variation of  $S_2(p)$  resulting from the ozone QBO are then considered in Section 3c.

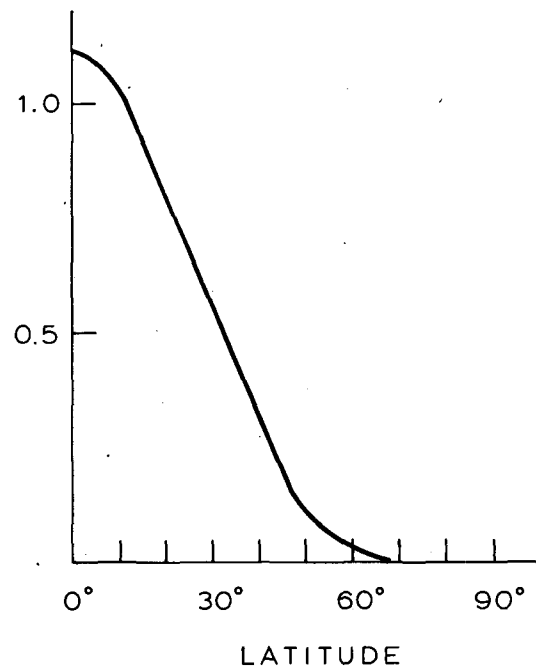


FIG. 8. The latitude structure of the gravest solar semidiurnal Hough function. The function is symmetric with respect to the equator.

*b. Radiation model for calculating the semidiurnal thermotidal excitation*

The heating rate resulting from the direct absorption of solar radiation by ozone was computed at 1 km intervals between the ground and 70 km for each half-hour between sunrise and sunset. The calculations were performed for 10° latitude intervals and only equinoctial conditions were considered. The heating rates were determined using the parameterization of Lacis and Hansen (1974). Only their formulation for cloudless skies was employed, but the effect of clouds was crudely incorporated by adopting a large effective "ground" albedo of 0.24 (this should have only a very small effect on the stratospheric and mesospheric heating rates). The ozone distribution without any QBO perturbation was assumed to be independent of latitude and is shown in Fig. 9. It was chosen to be typical of tropical conditions and peaks somewhat higher than some distributions based on midlatitude observations (e.g., Lacis and Hansen, 1974). The value for the solar constant was taken from Thekaekara (1976).

Once the heating rates were determined as a function of time of day, height and latitude, they were Fourier analyzed to determine the semidiurnal com-

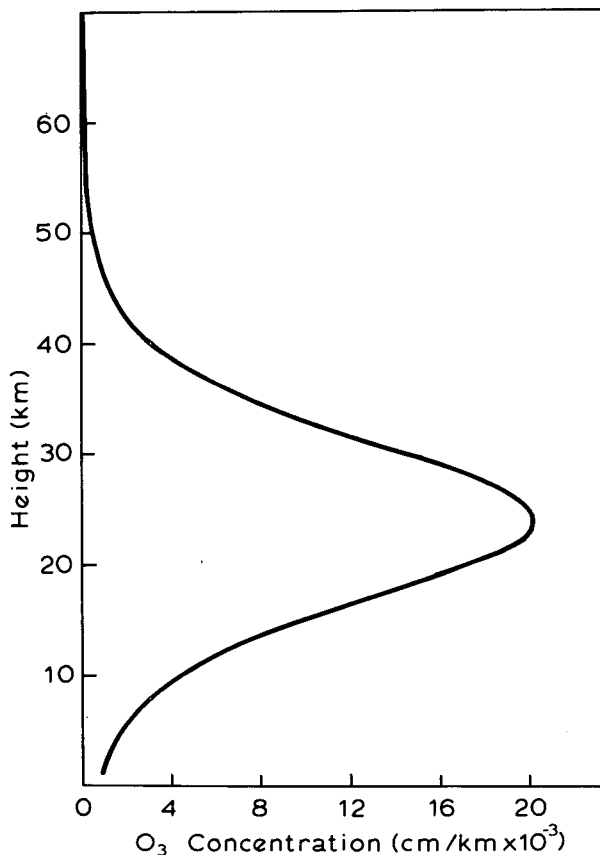


FIG. 9. The ozone distribution adopted in the present study.

ponent. Then this component was resolved into the three gravest symmetric semidiurnal Hough functions  $\Theta_{2,2}$ ,  $\Theta_{2,4}$ ,  $\Theta_{2,6}$ . The result for  $\Theta_{2,2}$  is displayed in Fig. 10 in terms of the amplitude of the semidiurnal temperature oscillation that would result from this heating in the absence of a dynamical response. The thermal excitation is strongly peaked in the upper stratosphere, in agreement with the recent calculations of Forbes and Garrett (1978) and Walterscheid *et al.* (1980).

*c. The effects of the ozone QBO*

Observations of the QBO in the total ozone column have been discussed by Angell and Korshover (1973, 1978) and Oltmans and London (1982) among others. These investigators conclude that there is a very clearly defined quasi-biennial variation of total ozone in the tropics which is well correlated with the stratospheric zonal wind QBO. The peak-to-peak amplitude of this oscillation is ~4% of the mean total column, with the maximum coinciding approximately with the maximum westerly phase of the zonal wind QBO at 50 mb. Outside the tropics (say poleward of about 25°) there is still a great deal of ozone variability at roughly biennial timescales but this is apparently not significantly correlated with the tropical ozone or wind QBOs. The vertical structure of the quasi-biennial ozone variability is not so well understood. One study by Wilcox *et al.* (1977) employed a series of balloonborne ozonesonde measurements from several stations in North and Central America to deduce a latitude-height section of the QBO up to altitudes of ~32 km. Their results indicate that the amplitude of the tropical ozone QBO (when expressed in absolute concentrations, e.g., molecules  $m^{-3}$ ) peaks at ~25 km and that it drops off rapidly with height away from the peak. There is also a downward phase progression which (near the peak amplitude) is about 1 km per month.

There apparently has been no observational study of the vertical structure of the tropical ozone QBO above 32 km. However, it is well known that the chemical lifetime of ozone above 30 km is very much shorter than the QBO timescale, and hence for present purposes the ozone concentrations in this region of the atmosphere may be considered to be controlled exclusively by photochemistry. Thus the only way the dynamical QBO could affect the ozone above ~30 km is through temperature changes (excluding more exotic possibilities such as a QBO related transport of long-lived species involved in the ozone photochemistry). It is believed that the QBO wind and temperature oscillations are strongly coupled through the thermal wind relation (Reed, 1962). Thus it seems reasonable to expect that above 30 km there will be a significant ozone QBO only where there is a wind QBO. The work of Hamilton (1981c) indicates that

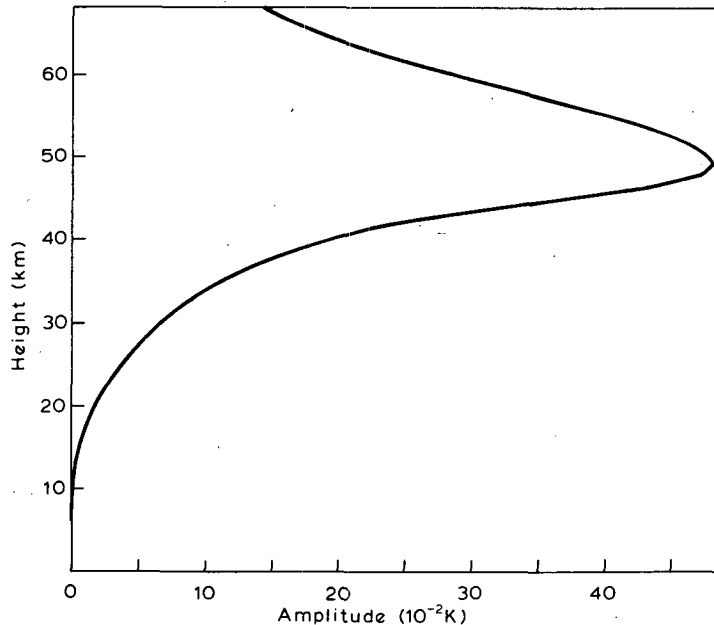


FIG. 10. The calculated amplitude of the semidiurnal component of heating by direct absorption of solar radiation by ozone, when projected onto  $\Theta_{2,2}$ . The result is given in terms of the magnitude of the temperature oscillation that this heating would produce in the absence of a dynamical response. The atmospheric densities required for this calculation were obtained using the temperature profile shown in Fig. 11 along with a surface pressure of 1000 mb.

the zonal wind QBO drops off rapidly in amplitude above 35 km and that the QBO may not exist above  $\sim 40$  km.

All these considerations led to the adoption of the following form for the ozone QBO (in absolute concentrations):

$$\delta O_3 = \begin{cases} A \cos[2\pi(z - z_0)/26] \exp[-(z - 25)^2/64], & z < 40 \text{ and } |\theta| < 25^\circ \\ 0, & z > 40 \text{ or } |\theta| > 25^\circ. \end{cases}$$

Here  $z$  is in kilometers and the choice of  $z_0$  determines the phase. The maximum perturbation in total ozone is obtained with  $z_0 = 25$  and the value of  $A$  was chosen so that equatorward of  $25^\circ$  a total column perturbation of 2% of the mean was obtained ( $A = 0.00144$  cm STP  $\text{km}^{-1}$ ; note that 1 cm STP  $\text{km}^{-1}$  is equivalent to  $2.7 \times 10^{20}$  molecules  $\text{m}^{-3}$ ).

The shortwave heating model described in the previous section was applied to the calculation of the thermotidal excitations associated with the three gravest symmetric semidiurnal Hough functions for various phases of the ozone QBO. These excitations were employed in classical tidal theory calculations of  $S_2(p)$ . In these calculations standard numerical methods were used for the solution of the vertical structure equation (see Chapman and Lindzen, 1970). The mean temperature profile employed is shown in Fig.

11 and was meant to be characteristic of the tropical atmosphere (appropriate since the tidal response is dominated by an equatorially trapped Hough mode and since the data considered in this work come from low latitudes). The calculations indicated that the largest perturbations in the equatorial  $S_2(p)$  induced by the ozone QBO were for  $z_0 = 29$  (i.e., four months before the maximum in the total column) and  $z_0 = 16$  (four months before the minimum in the total column). The altitude dependence of the ozone perturbation for  $z_0 = 29$  is shown in Fig. 12 (the  $z_0 = 16$  perturbation is just the negative of the one with  $z_0 = 29$ ). The equatorial results for these extreme cases are shown in Table 1. The calculated  $S_2(p)$  response is so strongly dominated by  $\Theta_{2,2}$  that the results away from the equator (at least equatorward of  $\sim 40^\circ$ ) can be determined quite accurately by simply modulating the values in Table 1 by the function displayed in Fig. 8.

As can be seen from Table 1, the theoretical response of  $S_2(p)$  to the ozone QBO involves primarily the 0900 component, in agreement with the observed quasi-biennial variations in the Batavia data. However, the amplitude of the response in the 0900 component is only about one-tenth of that tentatively identified in the Batavia time series. There is, of course, much uncertainty in the vertical structure of the ozone QBO and the form adopted in the present

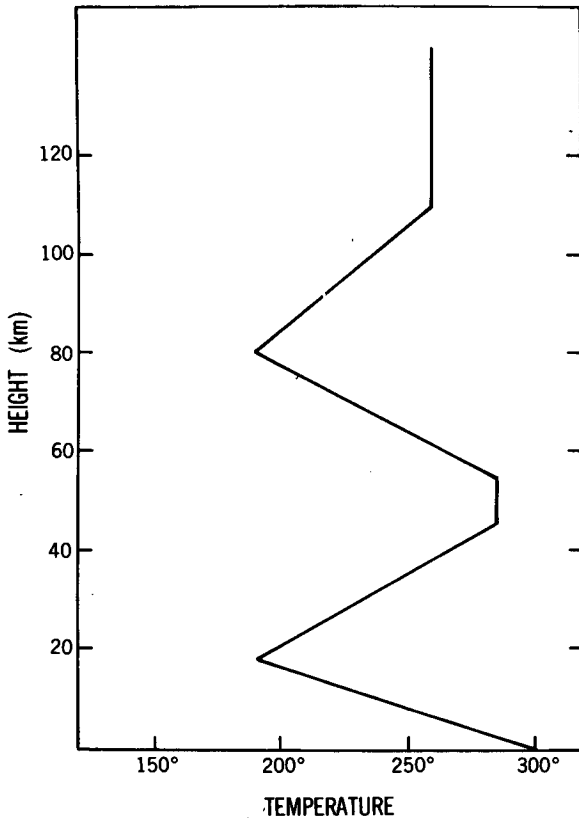


FIG. 11. The mean temperature profile used in the classical tidal theory calculations described in the text.

calculations is undoubtedly a great idealization of reality. It would be possible to obtain a larger response in the calculated  $S_2(p)$  if the ozone oscillation were concentrated higher in the atmosphere. However, the reasons given above for believing that the amplitude of the ozone QBO should drop off rapidly above 35 km appear to be sound. Thus it is more likely that the discrepancy between theory and observations is an indication that the wind and temperature QBOs must be significantly affecting  $S_2(p)$ . This issue is discussed briefly in the following section.

d. The effects of the dynamical QBO

The phase speed of the semidiurnal tide is much larger than the magnitude of the mean zonal winds in the atmosphere and thus the classical tidal theory is satisfactory for most purposes (Chapman and Lindzen, 1970). It is known, however, that the inclusion of realistic mean winds and meridional temperature gradients in tidal calculations can change the  $S_2(p)$  results by several percent (Lindzen and Hong, 1974; Walterscheid *et al.*, 1980). The calculations of Walterscheid *et al.* (WDV) are of particular interest. WDV computed the  $S_2(p)$  resulting from direct ab-

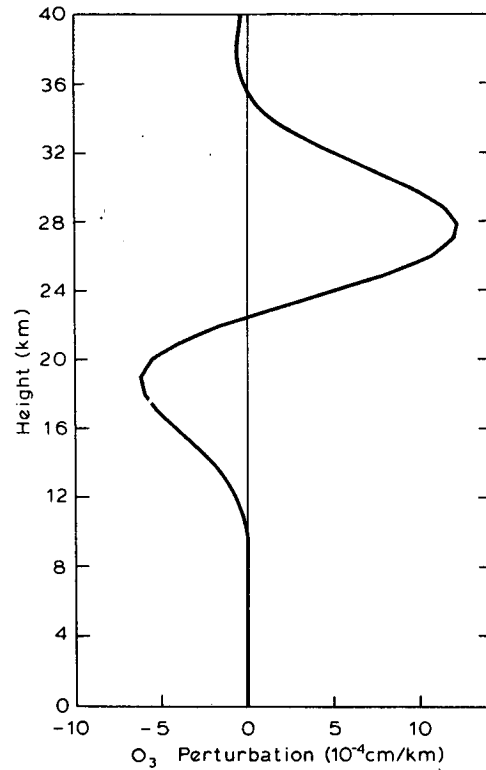


FIG. 12. The ozone perturbation that causes the maximum calculated response in the equatorial  $S_2(p)$ .

sorption of sunlight by ozone and water vapor. They considered two cases: 1) an atmosphere with no mean winds and a global mean  $T_0(z)$ , and 2) an atmosphere with the solstitial zonal wind field of Murgatroyd (1965, 1969). In this second case the mean temperature field  $T_0(\theta, z)$  was obtained by adding the meridional temperature variations required by the thermal wind relation to the global mean  $T_0(z)$  of case 1). The WDV results for the equatorial  $S_2(p)$  are summarized in Table 2. The change from no winds to a complete solstitial wind field alters the 0900 component by 0.123 mb but the 0000 component by only 0.012 mb. The less dramatic changes in the wind and temperature fields associated with the QBO would presumably cause smaller changes in  $S_2(p)$ . It is quite

TABLE 1. The results of the classical tidal theory calculations of the equatorial  $S_2(p)$  near the extreme phases of the ozone quasi-biennial oscillation. Results are given in millibars.

	0900 component	0000 component
Near maximum total ozone	0.6489	-0.1165
Near minimum total ozone	0.6449	-0.1154
Difference	0.0040	-0.0011



TABLE 2. Results for linear tidal theory calculations of the equatorial  $S_2(p)$  in an atmosphere with no winds and in an atmosphere with the climatological zonal mean winds (and associated meridional temperature gradients). These values have been computed on the basis of the results shown in Table 4 of Walterscheid *et al.* (1980).

	0900 component	0000 component
Solstitial winds	1.000	-0.022
No winds	0.877	-0.010
Difference	0.123	-0.012

plausible that the effects of the dynamical QBO could account for a significant fraction of the  $\sim 0.04$  mb amplitude of the observed QBO in  $S_2(p)$ . The dominance of the 0900 component in the response to mean wind changes is a consequence of the very long vertical wavelength associated with  $\Theta_{2,2}$  and would almost certainly also be found in the calculated  $S_2(p)$  response to the zonal wind and temperature QBO.

It appears that the QBO in  $S_2(p)$  evident in the Batavia data might be adequately explained by the effects of the familiar ozone and wind QBOs on the excitation and propagation of the semidiurnal tide. Further detailed non-classical tidal calculations including the stratospheric wind QBO would clearly be needed before a definitive statement on the agreement between theory and observations could be made. Such calculations ought to include a tropospheric semidiurnal latent heat forcing since the wind QBO could conceivably influence the 0000 component of the surface pressure oscillation through downward reflections of the part of the tide thought to be generated in the troposphere.

#### 4. The possible use of observations of $S_2(p)$ in monitoring ozone variability

The effect of human activities on the chemistry of atmospheric ozone remains one of the most important environmental issues of the day. In particular, theoretical predictions suggest that the halocarbon compounds that have been released into the atmosphere since World War II ought to decrease the global inventory of ozone by  $\sim 3\%$  by 1990 (e.g., Borucki *et al.*, 1980). Such a decrease would have deleterious effects on human health (e.g., Mole, 1980) and might also produce a change in the climate at the ground (e.g., Ramanathan and Dickinson, 1979). However, studies of the time evolution of total ozone measured using ground-based spectrophotometers appear to indicate that over the last two decades there has actually been an increase in global ozone (e.g., Angell and Korshover, 1973, 1978). Unfortunately, such observational estimates of global ozone trends are subject to a number of significant uncertainties. Instrument calibrations can drift by several percent

over a period of years (Komhyr and Grass, 1972). To make matters worse, during the 1960's some stations changed the wavelength pairs used by their Dobson instruments in determining total ozone (Angell and Korshover, 1973). In addition, spurious trends in total ozone measurements can be introduced by long-term changes in tropospheric air quality at a particular station (Komhyr and Evans, 1980). Finally, it is important to note that the network of stations with reliable total ozone observations over a significantly long period is not really adequate for the evaluation of global means (Moxim and Mahlman, 1980).

Any changes in ozone must, in principle, affect  $S_2(p)$ , and this suggests the possibility of using observations of  $S_2(p)$  to monitor changes in the ozone. This method avoids the shortcomings of direct observations since 1) surface pressure is a quantity that is very easy to measure reliably and it is most unlikely that instrumental effects could introduce spurious trends, and 2) the  $S_2(p)$  at a single station is influenced by the distribution of ozone over much of the globe rather than by the ozone profile directly over the station. This second point may require some explanation. Atmospheric tides can be regarded as internal gravity waves that are somewhat modified by rotation. The internal gravity wave response to an isolated excitation in a plane non-rotating atmosphere is concentrated along conical surfaces radiating away from the source region. The slope of these surfaces is given by the ratio of the angular frequency of the waves to the Brunt-Väisälä frequency. For the semidiurnal tide this slope is  $\sim 5 \times 10^{-3}$  in the stratosphere and  $10^{-2}$  in the troposphere. Thus tidal excitation at 30 km height should influence the surface  $S_2(p)$  signal in a region with horizontal dimensions of the order of  $10^4$  km. The main effect of the earth's rotation and sphericity is to inhibit the vertical propagation of waves at high latitudes. Thus the influence of a perturbation of the stratospheric heating on the surface  $S_2(p)$  depends largely on the projection of the perturbation onto  $\Theta_{2,2}$ .

While avoiding some of the difficulties inherent in direct observations, the use of  $S_2(p)$  in monitoring changes in stratospheric composition has its own problems. First of all the effect to be observed is likely to be very small. In the previous section it was shown that a 4% change in total ozone distributed as in Fig. 12 should change the 0900 component of the equatorial  $S_2(p)$  by only  $\sim 0.004$  mb. The anticipated changes in ozone concentrations due to the effects of halocarbon compounds extend considerably higher than the QBO perturbation of Fig. 12 (e.g., Vuppuri, 1979; Pyle, 1980) and so they will produce much more alteration in  $S_2(p)$  for the same total column variation. That small changes in  $S_2(p)$  might be detectable is suggested by the Batavia time series. Fig. 13 shows 25-month running means of the 0900 com-

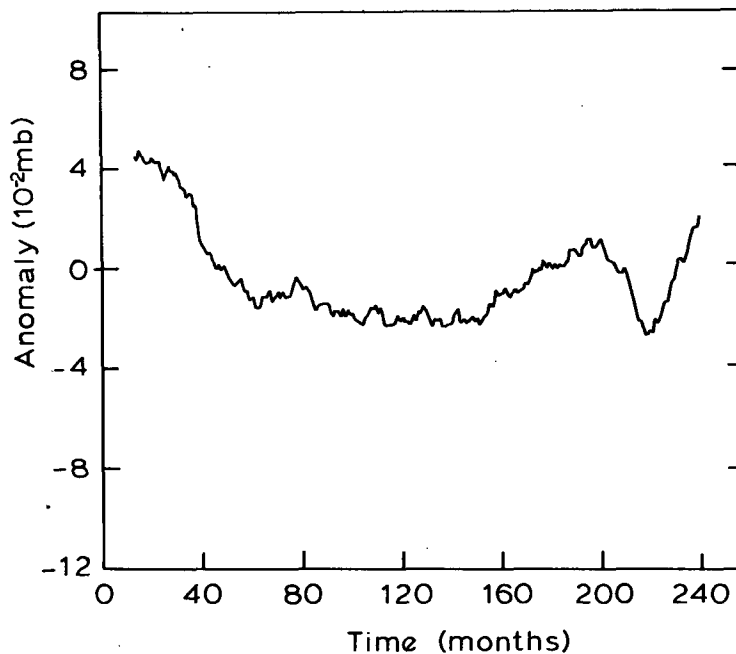


FIG. 13. As in Fig. 2 but for 25-month running means.

ponent of the anomaly in  $S_2(p)$ . Once the variability at biennial and shorter time scales has been eliminated the noise is so strongly reduced that trends as small as 0.02 mb per decade apparently can be detected by simple inspection. Presumably the signal-to-noise ratio could be further enhanced by using observations averaged over several stations.

A more serious shortcoming of the proposed approach to ozone monitoring is the ambiguity inherent in the interpretation of an observed trend in  $S_2(p)$ . For example, the decrease in the 0900 component during the first decade in Fig. 13 might be due to any or all of the following causes: 1) a decrease in total ozone, 2) a rearrangement of the mean ozone profile resulting in a net downward displacement without any change in the total column, 3) a decrease in the concentration of some other radiatively active atmospheric constituent (e.g., water vapor or clouds), 4) a decrease in the solar constant in the ultraviolet, or 5) a change in the phase of the latent heat excitation (see Fig. 1). In regard to these possibilities one should note that rather large changes in global water vapor concentrations or cloud amounts would be required to significantly alter  $S_2(p)$ , and it is possible that such changes would be detectable by other independent observations. Similarly one might hope that (in the near future at least) the part of any  $S_2(p)$  variations caused by changes in the solar flux might be accounted for by employing satellite observations of the solar spectrum. It is interesting to note in this connection that Haurwitz *et al.* (1957) and Cooper

(1982) find little evidence for solar cycle related variations in  $S_2(p)$ .

The question of whether one should expect long-term phase variations in the latent heat tidal excitation cannot be answered definitively. The best one can say is that the diurnal variability of latent heat release is probably under the control of local radiative processes (e.g., Gray and Jacobson, 1977), and thus there is no particular reason to suppose that there should be any systematic, global, long-period variability in the average time of maximum rainfall, for example. However, the current understanding of the diurnal variability of cloudiness and precipitation is so incomplete that one cannot positively rule out such a possibility.

Despite the possible ambiguity in the interpretation of the results, the analysis of  $S_2(p)$  data may have some role to play in corroborating direct observations of ozone trends. Thus the recent claims of increases in global ozone (e.g., Angell and Korshover, 1973) would be more credible if they could be supported by evidence of a similar trend in the 0900 component of  $S_2(p)$ .

## 5. Conclusions

Any conclusions drawn from the present work on the QBO in  $S_2(p)$  must necessarily be somewhat tentative, given the rather limited data record examined and the absence of definitive tidal theory calculations. However, there does appear to be a QBO in the Ba-

tavia  $S_2(p)$  record, and the features of this oscillation (in particular the absence of any biennial variation in the 0000 component) are consistent with what would be expected on the basis of tidal theory. Thus the present work can be regarded as evidence supporting the validity of the model of the solar semi-diurnal tide advanced by Lindzen (1978) and as an indication that the stratospheric QBO existed in something like its present form during the period 1924-44. Much more work on this subject clearly remains to be done. Particularly exciting is the possibility of using surface pressure records to attempt to verify the existence of the QBO as far back as the mid-nineteenth century [data from Singapore exist at least as far back as 1841; the Batavia data extend back to 1866; see Chapman and Lindzen (1970)].

Attention has been focused on the semidiurnal tide in this paper. The solar diurnal tide should also be affected by the stratospheric QBO. Unfortunately, the diurnal surface pressure oscillation  $S_1(p)$  is only about one-third to one-half as large as  $S_2(p)$ , and  $S_1(p)$  is believed to be strongly influenced by some purely local effects that are unrelated to the stratospheric forcing (Chapman and Lindzen, 1970). Thus it is unlikely that the effects of the stratospheric QBO would be detectable in the  $S_1(p)$  record. [Incidentally, Holloway *et al.* (1955) give data only for  $S_2(p)$ .]

The apparent ability of the  $S_2(p)$  record to resolve such a miniscule phenomenon as the biennial variation naturally leads to speculation that surface pressure observations might be employed to monitor changes in stratospheric composition. Due to the difficulty in interpreting observed variations in  $S_2(p)$  this approach can be used only to corroborate other more direct measurements.

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