Planetary Wave Coupling between the Troposphere and the Middle Atmosphere as a Possible Sun-Weather Mechanism

MARVIN A. GELLER AND JORDAN C. ALPERT

Rosenstiel School of Marine and Atmospheric Sciences, University of Miami, Miami, FL 33149
(Manuscript received 28 November 1979, in final form 11 February 1980)

ABSTRACT

The possibility of planetary wave coupling between the troposphere and solar-induced alterations in the upper atmosphere providing a viable mechanism for giving rise to sun-weather relationships is investigated. Some of the observational evidence for solar-activity-induced effects on levels of the upper atmosphere ranging from the thermosphere down to the lower stratosphere are reviewed. It is concluded that there is evidence for such effects extending down to the middle stratosphere and below. Evidence is also reviewed that these effects are due to changes in solar ultraviolet emission during disturbed solar conditions. A theoretical planetary wave model is then used to see what levels in the upper atmosphere moderate changes in the mean zonal wind state would result in tropospheric changes. It is concluded that changes in the mean zonal flow of ~20% at levels in the vicinity of 35 km or below would give rise to changes in the tropospheric planetary wave pattern that are less than but on the same order as the observed interannual variability in the tropospheric wave pattern at middle and high latitudes. Thus, planetary wave coupling between the troposphere and the upper atmosphere appears to be a plausible mechanism to give a tropospheric response to solar activity. This mechanism is not viable, however, to provide for short-period changes such as the suggested solar sector boundary vorticity index relation, but rather is applicable to changes of longer period such as the 11- or 22-year solar cycles.

1. Introduction

Few subjects in the atmospheric sciences have been the subject of more controversy than has the subject of sun-weather relationships. Recently, the arguments of a "believer" in these relationships have been nicely presented in the review paper of King (1975), while the arguments of a "skeptic" have been nicely presented by Pittock (1978). The purpose of this paper is not to present arguments for or against sun-weather relationships, but instead to discuss a physical mechanism whereby solar disturbance effects in the troposphere may occur by virtue of dynamical coupling to those higher levels in the atmosphere which are known to be affected by solar activity. In this discussion, the evidence, both observational and theoretical, will be reviewed for solar effects on the thermosphere, mesosphere and stratosphere. Then, the possibility that sufficient dynamical coupling exists between the troposphere and upper levels for these solar disturbance effects to significantly affect the troposphere is investigated using a theoretical model.

2. Solar disturbance effects on the upper atmosphere

It is well known that the thermospheric temperature varies greatly with both day-to-day and long-term changes in solar activity. Satellite drag measurements indicate this (Jacchia, 1969), as do incoherent scatter radar measurements (Evans et al., 1979). Volland (1970) has suggested that the increasing thermospheric temperatures with increasing solar activity are due to increases in the solar EUV emission occurring during disturbed conditions. Hicks and Justus (1970) have statistically analyzed 52 rocket vapor trail wind measurements and found that while the winds both above and below 110 km are correlated with geomagnetic variations, the lag correlations of $K_p$ with the winds above this altitude act differently than those with the winds below. This led them to conclude that the solar activity (that initiates geomagnetic disturbances) forces the wind changes above 110 km while the wind changes below this altitude themselves appear to be giving rise to a fraction of the observed geomagnetic changes through dynamo action. The case for solar disturbance effects on the thermosphere is clear. It is non-controversial that the disturbed sun emits increased EUV radiation that significantly affects thermospheric temperatures and winds, leading to the observed association between solar activity and changes in thermospheric parameters.

More controversial are solar disturbance effects on the mesosphere. Ramakrishna and Seshamani
(1973) have analyzed the correlation between mesospheric temperatures, as measured by sounding rockets at Thumba, India, which is located 8° north of the equator, and solar activity as indicated by the $F_{10.7}$ index. They conclude that there is a very significant correlation (at the 99% confidence level) between solar activity and mesospheric temperature at Thumba with a time lag of less than one day, but that the magnitude of the mesospheric temperature response is one order of magnitude less than it is in the thermosphere.

It is interesting that Ramakrishna and Seshamani (1973) hypothesize that this association exists by virtue of increased EUV emission during periods of solar disturbance, but in a later paper (Ramakrishna and Seshamani, 1976) when they found a significant lag-correlation between $K_p$ and mesospheric temperature during daytime hours at Fort Churchill (in the auroral zone), they hypothesized that this was due to auroral electrojet enhancements during disturbed conditions. The situation in the mesosphere is far less clear than that in the thermosphere. There have been studies where a connection between solar and geomagnetic disturbance conditions with mesospheric temperature has been found and physical links have been hypothesized. At the moment, however, our understanding of what physical mechanisms might give rise to such solar control is minimal.

Much more work has been done on the extent of solar activity effects on the stratosphere than has been done on the mesosphere. There have been observational studies, theoretical simulation studies, and some observations of the variability of the relevant part of the solar spectrum with solar activity. For example, Schwentek (1971) has indicated that winter stratospheric temperatures over Berlin vary with solar activity but that no clear variation is seen in the summer temperatures. Fig. 1, from Schwentek (1971), illustrates this. It also illustrates, curiously enough, that while an apparent increase in stratospheric temperature with sunspot activity is seen for sunspot numbers below 160, the opposite tendency is seen for higher sunspot numbers. Arranging these same data differently, Schwentek (1971) found that the winter stratospheric temperature was minimum during solar minimum (1964) and maximum during solar maximum (1969), but that the summer stratosphere did not show this behavior. This is shown in Fig. 2. Note that the effect seen by Schwentek is a sizeable one, amounting to $-20^\circ$C at 35 km, $10^\circ$C at 30 km, and no more than $5^\circ$C at 25 km. Of course, Schwentek's study only includes radiosonde data from a single station, Berlin. Recently, Angell and Korshover (1978) have collected Western Hemisphere rocketsonde data for the years 1965–76 for the height layers 26–35, 36–45 and 46–55 km. They segmented these data both by latitude band and by season. Fig. 3 shows Angell and Korshover's (1978) results for the north polar rocketsonde stations in the Western Hemisphere (Fort Greely–Poker Flats and Fort Churchill), the north subtropical stations (Wallops Island, Cape Kennedy, White Sands, Point Magu and Barking Sands), and for the equatorial stations (Ascension Island, Fort Sherman and Kwajalein). Note that the temperatures in the 26–35 and 36–45 km layers in the polar region were observed to be maximum at the time of solar maximum (occurring in 1969) and minimum at the time of solar minimum (occurring in 1964) in qualitative agreement with the results obtained by Zerefos and Mantis (1977) from high-level (24–31 km) radiosonde data; however, this is not seen so clearly in the 46–55 km layer since there is no clear temperature rise from sunspot minimum to sunspot maximum. A similar situation is seen in the subtropical data. In the equatorial region, however, the main feature observed in the data is a general temperature decrease during the period of observation with some indications of a weak temperature rise during the solar maximum. When the data are segmented by season (see Fig. 4) one sees that there is a general temperature decrease in the layer 46–55 km during the spring and summer seasons with temperature maxima at the time of solar maximum during the winter and autumn seasons as well as in the two layers below during all the seasons. Indications are that the winter effect may be strongest. The Angell and Korshover results show the largest amplitude temperature change between solar maximum and solar minimum to be present in the highest layer (46–55 km).

Quiriz (1979) has just published the results of a study that was conducted quite independently from, but is quite similar to, that of Angell and Korshover (1978). Quiriz (1979) examined summer rocketsonde temperature measurements for the time period 1965–77 at several sites which ranged in latitude from 8°S to 64°N. Summer data were chosen since one might expect to be able to identify a statistically significant signal more easily in a summer data set with less meteorological noise than is present during other seasons. Quiriz's results are in general agreement with those of Angell and Korshover, but a greater relation between stratospheric temperature and solar activity is found to take place at 35 km altitude rather than at 50 km altitude. This is shown in Fig. 5 and Table 1 from Quiriz (1979) where the 35 and 50 km temperature trends and sunspot numbers are displayed along with the correlation coefficients. This is at variance with Angell and Korshover's results, but Quiriz attributes this difference to the fact that he used instrumental temperature correc-
tion factors that were sizeable at 50 km, whereas Angell and Korshover did not make such corrections.

Summarizing the stratospheric temperature observations then, it does appear that at least the Western Hemisphere stratosphere between the altitudes of 26 and 50 km in the Northern Hemisphere was warmest in the polar and subtropical regions when the sun was in its most active state during the previous solar cycle. No such clear variation was observed below these altitudes or in the equatorial zone.

There have also been published reports of high coherencies between the 10.7 cm flux of solar radiation and the 10 mb circulation by Ebel and Batz (1977) and between the 10.7 cm solar flux and stratospheric winds over the altitude range of 25–65 km by Nastrom and Belmont (1978). These coherency studies indicate that there exists more of a connection between the 10.7 cm solar flux and the stratospheric flow than if there were independent fluctuations in these parameters with the same period [as Volland and Schaefer (1979) claim is the case for the observed 27-day variability in tropospheric planetary waves and solar activity].

On the theoretical side, there are at least three mechanisms by which solar disturbances may affect stratospheric structure. One of these was originally suggested by Crutzen et al. (1975) who showed that solar proton events should lead to a vastly increased production of NO\textsubscript{x} compounds in the stratosphere. Since polar proton events occur more frequently during disturbed solar conditions, one would expect more NO\textsubscript{x} under such conditions with a resulting increase in the catalytic destruc-

---

**Fig. 1.** Median values of summer (left) and winter (right) stratospheric temperature \( T \) at four different heights (25, 30, 35 and 40 km) obtained from daily measurements by means of radiosonde launchings at Berlin plotted against Zurich sunspot number \( R \). For definite groups of subspot number, i.e., \( R = 0–20, 10–30, 20–40, \) etc., the median as well as the upper and lower quartile values of \( T \) have been determined, separately for summer and winter (from Schwentek, 1971).
tion of ozone which, in turn, would alter the radiative balance of that portion of the atmosphere affected. Such a depletion of ozone during a solar proton event has been observed to take place by Heath et al. (1977); however, Schoeberl and Strobel (1978a) have illustrated that sizeable as this effect is locally, its geographical extent is too limited to appreciably affect global stratosphere dynamics. Ruderman and Chamberlain (1975) and Chamberlain (1977) have hypothesized that the known modulation of cosmic rays by solar activity leads to a modulation in stratospheric NOx which results in a modulation of ozone. It should be mentioned in this context that it is far from being settled whether total ozone, in fact, is observed to vary with solar activity, (see, e.g., London and Reber, 1979). However, it is thought to be likely that ozone concentrations in the upper stratosphere show such a variation. A third mechanism by which solar activity may affect the structure of the stratosphere was motivated by some observations by Heath (1973) in which he claims that the solar output of ultraviolet radiation in the wavelength range of $0.175 < \lambda < 0.310 \mu m$ varies by some tens of percent. These observations are thought to be somewhat controversial (e.g., see Smith and Gottlieb, 1974). Photochemical models by Callis and Nealy (1978) and Penner and Chang (1978) show that sizeable temperature changes would result from variable solar ultraviolet radiation as observed by Heath (1973), with the Callis and Nealy (1978) results appearing to indicate much larger stratospheric temperature changes with solar activity than were observed by Angell and Korshover (1978). The Penner and Chang
(1978) results show some qualitative agreement with the observations, but there also are areas of significant disagreement between these observations and their theory.

3. Dynamical coupling between the troposphere and the upper atmosphere

The most difficult obstacle to overcome in constructing a theory for how solar activity may give rise to significant tropospheric changes is to have extremely small changes in solar energy output bring about changes in tropospheric energetics that are orders of magnitude larger (see Willis, 1976). For instance, according to Livingston (1978) the solar constant does not vary by more than 0.1% on the short term, and Volland (1977) has demonstrated that if the solar output did vary by 0.1% with its 27-day rotation cycle this would generate tropospheric planetary waves with amplitudes of no more than 1 gpm.

Hines (1974) has suggested a possible mechanism that might be operative despite this energy mismatch between solar input and tropospheric energetics. The basis of this mechanism is that the surface airflow over topography and the global distribution of diabatic heating in the troposphere force planetary-scale disturbances that propagate their energy upward. Stratospheric and mesospheric winds play a dominant role in determining the "refractive index" for these waves (e.g., see Charney and Drazin, 1961; Matsuno, 1970; and Schoeberl and Geller, 1976) which will, in turn, determine the transmission-reflection properties of these waves. Thus, changes in the middle atmosphere flow might lead to changes in the tropospheric amplitudes and phases of planetary waves that propagate to this level. The energetics
of these changes are such that relatively small amounts of energy may give rise to significant effects in the upper atmosphere where the density is low, and these upper atmosphere effects merely act to modulate the effect of fixed energy sources in the troposphere. Some of the relevant dynamical model studies that relate to the viability of this mechanism, that was suggested by Hines (1974), are those of Bates (1977), Schoeberl and Strobel (1978a) and Mahlman et al. (1978). Since the results of these studies show significant differences, the relevant results from each one will be reviewed here briefly.

Bates (1977) developed a set of scaled equations for steady-state planetary wave motions assuming that both the wave variables and the basic-state parameters have zonal and meridional

---

out, a number of simplifying assumptions were made to allow him to obtain an analytic solution to this problem, and one must be careful to apply his formulation only to geophysical situations where these assumptions are satisfied over a range of latitudes.

Schoeberl and Strobel (1978a) have used a numerical quasi-geostrophic model to look at the effect of ozone reductions on the zonally averaged circulation of the middle atmosphere. They calculated the response to the August 1972 solar proton event, to halocarbon pollution, to uniform ozone density reductions, and to changes in the solar constant. They also calculated the resulting effects of their obtained changes in the zonal mean state on stationary planetary waves using essentially the planetary wave model of Schoeberl and Geller (1977). They found that no significant effect on the planetary wave structure should accompany an event of the nature of the August 1972 solar proton event. They calculated the alteration in middle atmosphere planetary wave structure to halocarbon pollution and found changes near the 50 km level that were no more than those of normal winter variability. They also found that the mean zonal state of the stratosphere would change so little in response to small variations in the solar constant that the planetary wave structure would be negligibly affected.

Mahlman et al. (1978) used a primitive equation general circulation model with 40 vertical levels between the earth's surface and 80 km altitude for their studies. They performed an experiment in which they ran a "control" case and compared it with a case in which ozone amounts were halved. This led to temperature decreases in excess of 20°C at the tropical stratopause and in excess of 10°C at the tropical tropopause. The winds were seen to decrease by ~10–15 m s⁻¹ in the region of the middle atmospheric jet decreasing downward to generally <5 m s⁻¹ at ~40 km. Mahlman et al. (1978) noted that although there were differences in the modeled troposphere for their two cases, that further analysis was required to assess their significance.

In assessing the confidence to be placed in these and other studies of dynamical coupling between the troposphere and the upper atmosphere, it is very important to take into account properly the role of both radiative and mechanical dissipation. Obviously, the planetary wave energy must be able to penetrate upward to the level where the refraction-transmission properties are altered as well as to return down to the region of forcing after reflection in order for the Hines (1974) mechanism to be viable. The three main dissipation mechanisms operative in the stratosphere and mesosphere are radiative damping acting through carbon dioxide and ozone, mechanical dissipation which is presumed to act through the turbulent mixing that results from the "breaking" of waves and

---

**Table 1.** Coefficients of correlation between summertime temperature departure from long-period mean and mean annual sunspot number, 1965–77 (from Quiroz, 1979). Period measured at Kwajalein is 1970–77.

<table>
<thead>
<tr>
<th>Station</th>
<th>Longitude</th>
<th>35 km A*</th>
<th>35 km B**</th>
<th>50 km A*</th>
<th>50 km B**</th>
</tr>
</thead>
<tbody>
<tr>
<td>Poker Flat</td>
<td>84°N</td>
<td>0.82</td>
<td>0.93</td>
<td>0.14</td>
<td>0.24</td>
</tr>
<tr>
<td>Churchill</td>
<td>59°N</td>
<td>0.55</td>
<td>0.83</td>
<td>0.63</td>
<td>0.89</td>
</tr>
<tr>
<td>Point Mugu</td>
<td>34°N</td>
<td>0.71</td>
<td>0.89</td>
<td>0.79</td>
<td>0.94</td>
</tr>
<tr>
<td>White Sands</td>
<td>32°N</td>
<td>0.76</td>
<td>0.89</td>
<td>0.49</td>
<td>0.84</td>
</tr>
<tr>
<td>Cape Kennedy</td>
<td>28°N</td>
<td>0.73</td>
<td>0.83</td>
<td>0.66</td>
<td>0.89</td>
</tr>
<tr>
<td>Kwajalein</td>
<td>9°N</td>
<td>0.75</td>
<td>0.90</td>
<td>0.75</td>
<td>0.87</td>
</tr>
<tr>
<td>Ascension Island</td>
<td>8°S</td>
<td>0.47</td>
<td>0.72</td>
<td>0.17</td>
<td>0.32</td>
</tr>
<tr>
<td>All stations</td>
<td></td>
<td>0.77</td>
<td>0.89</td>
<td>0.67</td>
<td>0.89</td>
</tr>
</tbody>
</table>

* Based on smoothed yearly data.
** Based on 3-year running means (1-2-1 smoothing) of temperature data, 1966–1976.
tides, and the possible absorption of wave momentum at critical levels. We will remark on each of these processes in the following.

Dickinson (1973) has developed a Newtonian cooling approximation to the infrared cooling due to CO$_2$ and O$_2$ using the temperature structure of the 1962 U.S. Standard Atmosphere. Dickinson's formulation has been used by such investigators as Schoebel and Strobel (1978b) and Holton and Wehrbein (1979). Dickinson (1973) found the minimum dissipation time scale due to infrared processes to be about five days in the stratosphere. Blake and Lindzen (1973) have developed a Newtonian cooling approximation for infrared cooling plus photochemical acceleration (the apparent cooling that arises from the photochemical destruction of O$_3$ that takes place as the ambient temperature increases). They found the minimum dissipative time scale to be about two days also at the stratosphere. Schoebel and Strobel (1978b) have indicated that they believe the Blake and Lindzen (1973) values to be inapplicable to global vertical large-scale motion, however, for reasons given in their paper. A recent analysis of simultaneous O$_3$ and temperature satellite observations of the upper stratosphere by Ghaizi et al. (1979) suggests, however, that the total rate of radiative damping may be twice the damping due to infrared radiation alone in approximate agreement with the Blake and Lindzen (1973) results albeit for different physical reasons. Ramanathan and Grose (1978) have indicated that although the Newtonian cooling approximation cannot be used for detailed simulation of stratospheric climate, it is probably sufficient for mechanistic studies of the type being discussed here.

The situation is even less understood for mechanical dissipation. Both Schoebel and Strobel (1978b) and Holton and Wehrbein (1979) have found it necessary to introduce a parameterization for a marked increase in mechanical damping above the stratosphere to achieve reasonable looking simulations of the zonal mean state of the middle atmosphere. Mahlman et al. (1978) also have included a parameterization for enhanced mechanical dissipation in their general circulation model. Comparing the Schoebel and Strobel (1978b) and the Holton and Wehrbein (1979) parameterizations for mechanical dissipation, it appears that they both have dissipation time scales of about 10 days at 65 km with longer time scales below (Schoebel and Strobel's time scales being shorter there) and shorter time scales above (Holton and Wehrbein's time scales being shorter there).

Finally, we come to the role of critical levels as a dissipation process for planetary wave energy. Until recently, based on the work of Dickinson (1968a, 1970) it was believed that stationary planetary waves were absorbed at locations of zero mean zonal wind. This concept was used in the design of a number of mechanistic planetary wave models (e.g., Matsuno, 1970; Schoebel and Geller, 1977). Tung (1979) has recently argued that nonlinearities probably dominate over dissipation in the vicinity of critical levels, and has shown that most of the planetary wave energy is reflected rather than absorbed there if this is the case. This becomes important in the context of dynamical coupling between the troposphere and the upper atmosphere since if the singular wind lines reflect rather than absorb planetary waves, the only dissipation mechanisms are radiative damping and mechanical dissipation in the free atmosphere and Ekman friction in the planetary boundary layer. Since the first two of these mechanisms grow with altitude, significant tropospheric change due to variations in the upper atmosphere are more likely to occur if such changes occur at lower levels in the upper atmosphere. For instance, Tung and Lindzen (1979) argue that the troposphere will be affected much more greatly if the middle atmosphere jet is lowered below the middle stratosphere than if its structure is changed at higher levels.

4. Model calculations

The model we will use in this section is a quasi-geostrophic model of stationary planetary waves that extends in the vertical from the ground to 100 km. This model is essentially the same as that used by Schoebel and Geller (1977) except it has been altered by S. K. Avery, M. A. Geller and J. C. Alpert to include tropospheric forcing by the surface airflow over the appropriate zonal harmonic of the Northern Hemisphere topography [as given by Berkofsky and Bertoni (1955)] and planetary wave forcing by diabatic heating (by using the appropriate zonal harmonic of the lower tropospheric diabatic heating that was calculated for the Northern Hemisphere by Geller and Avery (1978)].

The model equations are as follows:

$$\sin^2 \theta \left( \frac{\theta}{\phi} \frac{\partial}{\theta} \left( \frac{\partial \psi_m}{\sin^2 \theta} \frac{\partial \psi_m}{\theta} \right) \right) + Q_m \psi_m + \left( \frac{\bar{u}m - ia_0}{\bar{u}m - iB_R} \right) \sin^2 \theta \left( \frac{\partial \psi_m}{\partial z} \right)^2$$

$$- \frac{i \sin \theta}{\bar{u}m - iB_R} \frac{\partial \psi_m}{\partial z} \frac{\partial \psi_m}{\partial z} = -iG_m, \quad (1)$$

where

$$Q_m = \frac{m \frac{\partial q}{\theta}}{\left( \frac{\bar{u}m - iB_R}{\cos \theta} \right)} + \frac{\bar{u}m - ia_0}{\bar{u}m - iB_R} \frac{\sin^2 \theta}{S} Y$$

$$- \frac{i \sin \theta}{\bar{u}m - iB_R} \frac{\partial \psi_m}{\partial z} \left[ \frac{1}{2S} \left( 1 + \frac{1}{S} \frac{\partial S}{\partial z} \right) \right], \quad (2)$$
\[
\frac{\partial \tilde{u}}{\partial \theta} = \cos^2 \theta \left[ 2(\Omega + \tilde{u}) + 3 \tan \theta \frac{\partial \tilde{u}}{\partial \theta} - \frac{\partial^2 \tilde{u}}{\partial \theta^2} \right] - \sin^2 \theta \frac{\partial}{\partial z} \left( \frac{e^{-\tilde{u}^2}}{S} \frac{\partial \tilde{u}}{\partial z} \right),
\]
(3)

\[
G_m = \frac{\sin^2 \theta}{S^{1/2}} \frac{\partial}{\partial z} \left( \frac{e^{-\tilde{u}^2}}{S} R J_m \right),
\]
(4)

\[
Y = -\left( \frac{3}{4} \frac{1}{S^{1/2}} \frac{\partial S}{\partial z} \right)^2 - \frac{1}{2S} \left( \frac{\partial S}{\partial z} - \frac{\partial^2 S}{\partial z^2} \right) - \frac{1}{4}.
\]
(5)

In writing Eqs. (1)-(5), the stationary planetary wave geopotential, with zonal wavenumber equal to \( m \), is given by \( \phi = e^{i2\pi S_{1/2} \psi_0 e^{im\lambda}} \). Here \( \lambda \) is longitude; \( \tilde{u} = \tilde{u}/\cos \theta \), where \( \tilde{u} \) is the mean zonal wind, \( a \) the earth's mean radius and \( \theta \) latitude; \( a_0 \) is the Newtonian cooling coefficient;

\[
S = \frac{R}{(2\Omega)^2} (RT\theta/c_p + \partial T/\partial z),
\]

where \( R \) is the gas constant for dry air, \( \Omega \) the earth's rotation rate, \( T \theta \) the horizontally averaged temperature profile, \( c_p \) the specific heat of dry air at constant pressure and \( z = -\ln(p_0/p) \), \( p \) being pressure and \( p_0 \) the surface pressure (taken to be 1000 mb). \( \beta \) is the Rayleigh friction coefficient. \( J_m \) is the \( m \)th Fourier coefficient of the diabatic heating, i.e., \( J = \sum J_m e^{im\lambda} \), where \( i = \sqrt{-1} \).

The vertical velocity that results from the zonally averaged surface airflow over the \( m \)th zonal harmonic of the surface topography is used as the lower boundary condition, and we use a radiation boundary condition to assure upward energy flux at the upper boundary of our computations (\( z = 14 \) or \( \sim 100 \) km). The north-south gridpoints are separated by 5° in latitude and the vertical grid points are separated by \( \Delta z = 0.2 \). For lateral boundary conditions, we take \( \psi_0 \) to vanish at both the pole and equator [for details see Schoeberl and Geller (1977)]. Solution of Eq. (1) is accomplished by the method described in Lindzen and Kuo (1969). The Newtonian cooling coefficient is taken to be equal to that computed by Dickinson (1973) between 30 and 70 km and is taken to go smoothly to zero at the ground and at 100 km altitude, and the Rayleigh friction is taken to have a small background value of \( 5 \times 10^{-3} \) s\(^{-1} \), which implies a dissipation time scale of about 23 days, with larger values at zero wind lines to provide for critical level absorption (see Schoeberl and Geller, 1977). This point will be returned to later in the discussion of the previous section. The static stability profile \( S(z) \) is that which was calculated for January by Geller (1970). The effect of dissipation in the planetary boundary layer is included in our computations through an Ekman pumping component to the surface vertical velocity. The structure of this model will be described in more detail in a forthcoming publication.

Now, given the evidence for solar disturbance effects altering the state of the thermosphere, the mesosphere and the stratosphere, we inquire in this section to what extent, if any, the planetary waves in the troposphere will respond to changes in the mean zonal state at various levels of the upper atmosphere. Since Schoeberl and Geller (1977) have shown that the structure of planetary waves appears to be much more sensitive to changes in the zonal wind structure \( \tilde{u} \) than to changes in \( S \), the static stability profile, we will be restricting ourselves to examining the response to changing the mean zonal wind state. Our "control" mean zonal wind state was derived from the January mean zonal wind state in Oort and Rasmussen (1971)\(^2\) below 50 mb and by utilizing the CIRA (1972) atmosphere above. This is shown in Fig. 6. Our planetary wave model is then run for the control case as well as for cases where the control case mean zonal winds are decreased by 20% at various levels ranging from \( z_0 = 12 \) (\( \sim 85 \) km) to \( z_0 = 2 \) (\( \sim 17 \) km). The choice of a uniform 20% reduction is arbitrary and is motivated by our desire to explore the effects of moderate changes in the mean zonal flow which could result through the thermal wind relation from latitude dependent solar activity induced temperature changes. In each case the mean zonal wind values are essentially unchanged from their control values over the height regions \( z \leq z_0 - 1 \) and \( z \geq z_0 + 1 \). For instance, Fig. 7 shows the basic wind state when \( z_0 = 6 \) (\( \sim 42 \) km) and Fig. 8 shows the mean zonal wind values that are given when the wind values shown in Fig. 7 are subtracted from the wind state that is shown in Fig. 6.

What is generally seen in these computations is that when the mean zonal flow is changed at a given level in the atmosphere by 20% over a restricted altitude range as previously indicated, the effects of these changes on the planetary wave structure are seen at all levels above the level where the wind change has taken place; however, only those levels higher than about \( z_0 - 3 \) show any measurable change in planetary wave structure. Several test comparisons will be shown to illustrate these points. Fig. 9 shows a comparison of the amplitude and phase structure for the computed planetary wave with zonal wavenumber 1 between the control and the \( z_0 = 10 \) (\( \sim 73 \) km) mean zonal wind structures for the latitudes of 30, 50 and 65°N. Note that no perceptible change in either the phase or the amplitude of wavenumber 1 is seen below \( z = 7 \) (\( \sim 50 \) km) at any of the latitudes. There is also an indication that the change penetrates furthest down-

---

ward at the highest latitude. Similar calculations for planetary waves 2 and 3 for these two wind states give qualitatively similar results for the change in wave structure due to mean zonal flow alterations around $z_0 = 10$ although, of course, the wave structures for the different wavenumbers are quite different. When the wind structure is changed below the middle stratosphere, however, significant changes in the tropospheric planetary wave structure become

Fig. 6. Control case mean zonal wind state (m s$^{-1}$). Positive values indicate westerly flow and negative values indicate easterly flow.

Fig. 7. Mean zonal wind state when the wind magnitudes in the vicinity of $z = 6$ are reduced by 20%.
evident. For instance, Fig. 10 shows a comparison of the amplitude and phase structure for the computed planetary wave with zonal wavenumber 1 between the control and the $z_0 = 5$ (~35 km) mean zonal wind structures for the latitudes of 30, 50 and 65°N. Note that while no perceptable change in either the phase or the amplitude is seen below $z = 2$ (~17 km) at the two lower latitudes, a significant change in the phase of wavenumber 1 is seen at 65°N that penetrates all the way down to the earth’s surface.

Changes in the wind structure at levels lower than 35 km give alterations in the planetary wave structure that penetrate both further down into the troposphere and to lower latitudes. For instance, Fig. 11 shows the comparison of the amplitude and phase structure for the computed planetary wave with zonal wavenumber 1 between the control and the $z_0 = 2$ (~17 km) mean zonal wind structures for the latitudes 30, 50 and 65°N.

In order to give a better idea of the magnitude of the changes in tropospheric circulation that occur as a result of altering the middle atmospheric transmission-reflection properties for stationary planetary waves, Fig. 12 shows a comparison of the 500 mb planetary wave patterns that result from our model with the control mean zonal winds and with the 20% reduction case at $z = 4$ (~30 km). These patterns are generated by adding our modeled January wavenumbers 1, 2 and 3 together for both of the wind state cases. Our choice of including wavenumbers 1, 2 and 3 was made on the basis of van Loon et al.’s (1973) finding that these wavenumbers account for 96.2% of the deviations from the zonal mean values of geopotential and temperature at 50°N in January at 500 mb. Also shown in this figure is the planetary wave pattern obtained by adding the January wavenumbers 1, 2 and 3 at 500 mb from van Loon et al. (1973) together. In order to get an idea of the level of relevance of our model to the actual atmosphere, we compare the observed pattern (from van Loon et al.’s analysis) to our control case results. A ridging in excess of ~14 dam is seen over the British Isles in the observations. This is compared to the modeled ridging of ~18 dam in roughly the same location. Over northeastern North America a troughing of about 14 dam is seen in the observations. A similar magnitude troughing is seen in the model results but shifted somewhat to the northwest. A ridging is seen along the western coast of North America reaching a maximum of about 11 dam in the Gulf of Alaska. The model results show a general area of ridging centered in the Pacific Ocean off the west coast of the North America of roughly the same magnitude. The observations show a large troughing area centered in the vicinity of Japan with a magnitude of ~26 dam. The model results show a large troughing area centered about 30° to the west of Japan with a magnitude of only ~11 dam. We only compare our model results to observations at middle latitudes since good comparison between the observations and the model at very high and very low latitudes, if it
Fig. 9. Comparison between the modeled planetary wave with zonal wavenumber one for the control mean zonal wind state and that when the wind magnitudes are decreased by 20% in the vicinity of \( z = 10 \). Left-top: amplitude in units of \( \text{cm}^2 \text{s}^{-2} \) (1 dam in geopotential height corresponds to \( \sim 10^8 \text{cm}^2 \text{s}^{-2} \)); left-bottom: west longitude of ridge; middle: same at 50°N; right: same at 65°N.
Fig. 10. As in Fig. 9 except when 20% reduction is made at z = 5.
FIG. 11. As in Fig. 9 except when 20% reduction is made at z = 2.
Fig. 12a. Observed mean Northern Hemisphere January 500 mb planetary wave height pattern (dam) due to zonal harmonics 1, 2 and 3, using the results of van Loon et al. (1973).

occurred, would be considered fortuitous given the limitations of both the observations and the model. Given that the observational results from van Loon et al. (1973) are for a 7-year average and that a substantial year-to-year variability is noted by these authors (see Fig. 13), we believe that our model agrees reasonably well with the observation. In order to compare the control case with the 20% reduction case at z_0 = 4 most easily, Fig. 14 shows the z_0 = 4 pattern subtracted from the control pattern. This figure shows that in addition to the differences in the magnitude of the ridging and troughing that were seen in Fig. 12 (for instance, a difference of 0.4 dam to the west of the British Isles), there are larger differences produced at higher latitudes due to phase differences. For instance, in response to the changes in the zonal winds around z_0 = 4, the 500 mb heights over northeastern Asia have increased by ~1.9 dam; the 500 mb heights over northeastern Europe have decreased by ~1.4 dam; the 500 mb heights off the northeastern coast of North America have increased by ~1.3 dam; and the 500 mb heights over north central Canada have decreased by ~1.5 dam. These changes are put into perspective by comparing them with the observed interannual variability as shown in Fig. 13. The difference field in Fig. 14 amounts to a substantial fraction of the observed interannual variability in the amplitude of the individual planetary waves. We have also pointed out previously that larger changes in the mean zonal flow at lower levels would produce larger effects. What we have shown then is that regional changes in the mean winter 500 mb height field of ~20 m in response to solar disturbance effects are quite possible on theoretical grounds and that this change is not negligible when compared to observations of the interannual variability in the mean January planetary wave field. Thus, the upper atmosphere appears to vary in response to solar activity down to sufficiently low altitudes so that changes in tropospheric planetary wave structure occur as a result of the altered planetary wave propagation through the middle atmosphere. One candidate mechanism for giving rise to changes in the temperature structure at these stratospheric levels is changes in the sun's ultraviolet emission with changing solar activity. Therefore, it appears that in light of our best present knowledge that dynamical coupling of the troposphere to those regions of the upper atmosphere that seem to be directly affected by solar activity is a viable mechanism for solar activity to make itself felt on the tropospheric circulation. This mechanism, however, is incapable of explaining any near instantaneous tropospheric response to solar activity such as that reported by Wilcox et al. (1974). This is because of the finite time that it takes for planetary waves to propagate their energy upward to reflection levels, be reflected, and return down to the forcing re-

Fig. 12b. Modeled Northern Hemisphere January 500 mb planetary wave height pattern (dam) due to zonal harmonics 1, 2 and 3 using "control" mean zonal wind state.
Fig. 12c. As in Fig. 12b except for 20% reduction of the mean zonal winds in the vicinity of $z_o = 4$.

Region as is required for this mechanism to be operative. Muench (1965) has observed that periodic amplifications of planetary waves with zonal wave-numbers 1 and 2 appear to propagate upward at a rate of about 6 km day$^{-1}$. Taking this to be an indication of the planetary wave group velocity implies a two-way transit time between the troposphere and 35 km of $\sim$10 days. So, there should be at least a time lag of this order between a solar disturbance affecting the atmosphere at this altitude and its effect being felt on the tropospheric planetary wave structure. In fact, this planetary wave reflection mechanism is probably most applicable to climatic time scales (on the order of a solar cycle, say) rather than on the scales appropriate to changes in the weather (about a few days).

It should be mentioned once again, however, that in view of the importance of dissipation effects for the planetary wave transmission-reflection problem, what such effects are included in this model. Dickinson’s (1973) values of Newtonian

Fig. 13. Monthly mean wave structure of zonal harmonics 1, 2 and 3 at a constant latitude during January for seven different years from 850 to 200 mb (from van Loon et al., 1973).
cooling were used between 30 and 70 km altitude with smaller values above and below. A small background value for Rayleigh friction $5 \times 10^{-7} \text{s}^{-1}$ was used with enhanced values for the Rayleigh friction being used in the vicinity of zero wind lines in an effort to totally absorb planetary waves there. No enhanced mechanical dissipation was used at high altitudes which almost certainly accounts for the continuous growth of the wavenumber 1 planetary wave above $\sim 60$ km in our model, which is not observed. Investigations into the effects of altered dissipation in our model are underway with particular emphasis on the treatment of singular wind lines.

5. Discussion

We have seen in previous sections that observational evidence indicates that the atmosphere from the thermosphere down to an altitude of $\sim 25$ km varies in a manner that is consistent with a response to solar activity. Of course, since in most cases observations are only available for one 11-year solar cycle, no firm statistical inference of such an effect can be made (Pittock, 1978). The results of our modeling study indicate that changes in the mean zonal flow at levels of about 35 km and below should alter the propagation of stationary planetary waves such that the winter tropospheric circulation is significantly affected in the manner that was hypothesized by Hines (1974). Changes in the mean zonal flow above this level as well as changes in the mean zonal flow during the summer and equinox seasons would not appear to affect the tropospheric circulation to such an extent. Our modeling study also shows that such alterations in the tropospheric flow should take place mainly at high latitudes.

Our interpretation for this is as follows. By looking at Figs. 1 and 2 in Schoeberl and Geller (1977), it can be seen that the higher meridional mode eigenfunctions to the planetary wave equation for the idealized case of an isothermal atmosphere in constant rotation have their energy concentrated at high latitudes, and that these higher meridional wave modes are reflected by relatively low values of the mean zonal wind. For instance, the (1,5) mode is seen to be reflected at levels where the mean zonal flow is $\sim 30$ m s$^{-1}$. This is consistent with the results shown in Schoeberl and Geller’s (1977) Fig. 13 which indicates that for a basic state wind model that closely resembles that shown in Fig. 6 the (1,5) mode changes from being vertically propagating at 17 km to evanescent at 34 km. We believe that when the winds in the vicinity of $z = 5 \sim 35$ km) or lower were decreased by 20%, this had the effect of raising the reflection height of a high-latitude wave mode, most probably (1,5), which led to our change in the computed phase structure of wavenumber 1 at high latitudes.

Putting these results into perspective, we know that the inter-annual variation in stationary planetary wave patterns are related to very significant changes in regional climate (Namias, 1959, 1966). Our modeling study indicates that if solar disturbance effects lead to changes in the mean zonal middle stratospheric flow of $\sim 20\%$, then changes in the strengths of the high-latitude quasi-stationary troughs and ridges amounting to $\sim 20\%$ of the observed range of variability may occur. These are not negligible changes, but other physical effects appear to be more important in producing changes in this stationary trough-ridge pattern. Furthermore, by the simplified nature of this model (linear and steady-state), the estimate of the tropospheric effects are most probably an overestimate of such effects. Nonetheless, we do believe that we have demonstrated the viability of the Hines (1974) mechanism for solar effects on the troposphere and have given an idea of the magnitudes of the effects to be anticipated if the stratospheric mean zonal winds vary with solar cycle below an altitude of 35 km. We reemphasize at this point that we are not claiming a variation in the mean zonal wind of 20% its mean value though a solar cycle, but have merely tried to make a case that observational studies indicate some solar activity influence of winds and temperatures in this altitude range. Furthermore, our results indicate that the tropospheric planetary wave response to solar-
induced changes in the zonal mean state of the stratosphere occurs regionally, that is to say, they may be much more evident at some longitudes and latitudes than at other locations where they may be absent altogether.

Acknowledgments. The authors wish to acknowledge the support of the Climate Research Section of the National Science Foundation under Grants ATM 7718679 and ATM 7908352. We wish to thank Drs. Mark Schoeberl and Darrell Strobel for their valuable comments on an earlier version of this manuscript. We also wish to thank Dr. M. H. Davis and USRA (the University Space Research Association) for commissioning a study which helped to motivate the writing of this paper (through NASA Contract NAS 8-32482 Task 21).

REFERENCES
Geller, M. A., 1970: An investigation of the lunar semi-


