Intraseasonal Variability in a Dry Atmospheric Model

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(Manuscript received 3 May 2006, in final form 18 October 2006)

ABSTRACT

A long integration of a primitive equation dry atmospheric model with time-independent forcing under boreal winter conditions is analyzed. A variety of techniques such as time filtering, space–time spectral analysis, and lag regressions are used to identify tropical waves. It is evident that oscillations with intraseasonal time scales and a Kelvin wave structure exist in the model tropical atmosphere. Coherent eastward propagations in the 250-hPa velocity potential and zonal wind are found, with a speed of about 15 m s⁻¹. The oscillation is stronger in the Eastern Hemisphere than in the Western Hemisphere.

Interactions between the tropical and extratropical flows are found to be responsible for the simulated intraseasonal variability. Wave activity flux analysis reveals that a tropical influence occurs in the North Pacific region where a northeastward wave activity flux is found associated with the tropical divergent flow in the western and central Pacific. In the North Atlantic sector, on the other hand, a strong extratropical influence is observed with a southward wave activity flux into the Tropics. The extratropical low-frequency variability develops by extracting kinetic energy from the basic mean flow and through interactions with synoptic-scale transient eddies. Linear experiments show that the tropical atmospheric response to the extratropical forcing in the North Atlantic leads to an eastward-propagating wave in the tropical easterly mean flow of the Eastern Hemisphere.

1. Introduction

The Madden–Julian oscillation (MJO) is the dominant mode of intraseasonal variability in the Tropics. It organizes convection and precipitation, thus has a great impact on the weather in the Tropics. It has a significant influence on the extratropical atmospheric variability, possibly through Rossby wave propagation (e.g., Ferranti et al. 1990; Hsu 1996), and thus could provide an important signal source for the extratropical weather forecasts on intraseasonal time scales. The MJO is also observed to be associated with changes in the atmospheric angular momentum (e.g., Anderson and Rosen 1983; Gutzler and Ponte 1990), the length of day (e.g., Madden 1987), and the earth’s electric and magnetic fields (e.g., Anyamba et al. 2000).

Since its first detection by Madden and Julian (1971), the MJO has been under intensive study. Numerous papers have been published trying to depict its behavior and understand its physical mechanisms. Progress has been made as reviewed by Madden and Julian (1994) and Zhang (2005). The MJO, a complex phenomenon, is not a regular oscillation. Instead, it is episodic and varies from case to case and from year to year (e.g., Salby and Hendon 1994; Hsu et al. 1990). The common features of the MJO are obtained in observational studies through techniques such as composite and lag regression analyses (e.g., Knutson and Weickmann 1987; Hendon and Salby 1994; Sperber 2003). In general, the MJO is a tropical large-scale oscillation that is dominated by periods of 30–60 days and zonal wavenumber-1 propagating eastward. The eastward-propagating disturbance is seen in the zonal wind, the 200-hPa velocity potential, and the outgoing longwave radiation (OLR). It was suggested that the MJO takes the form of a zonal-vertical cell that propagates eastward along the equator, associated with eastward-propagating con-
vective activity (Madden and Julian 1972). Another important feature of the MJO is its strong seasonal dependence. The amplitude is largest during Northern Hemisphere winter months and smallest during the summer (e.g., Madden 1986).

Several mechanisms have been proposed for the MJO to explain its main features and answer the question of where the energy comes from to maintain a large-scale oscillation against damping. Because the MJO is similar to a Kelvin wave in many aspects, such as an eastward propagation, a large amplitude on the equator, and a dominant signal in the zonal wind, most theoretical studies have been based on equatorial Kelvin wave dynamics. The first baroclinic mode of a Kelvin wave in a dry atmosphere, however, circles the equator in about 10 days, much faster than the MJO. In addition to finding an explanation for this speed difference, one also has to identify the energy source that sustains the Kelvin wave. In the wave–conditional instability of the second kind (CISK) mechanism, the energy comes from latent heat release that results from the convection forced by low-level moisture convergence (e.g., Lau and Peng 1987; Chang and Lim 1988). The low-level convergence itself is part of the propagating Kelvin wave. The slow eastward propagation may be caused by excitation of different vertical modes by the wave–CISK heating. Considering boundary layer frictions, Wang (1988) and Salby et al. (1994) proposed a frictional wave–CISK, where boundary damping interacts with the Kelvin wave–induced convergence leading to preferred planetary scales for the unstable waves. In a different theory, the wind-induced surface heat exchange (WISHE) or evaporation–wind feedback (EWF; e.g., Emanuel 1987; Neelin et al. 1987), interactions between surface evaporation and the surface wind component of planetary-scale Kelvin waves provide a source of instability for the MJO. When interacting with convections, radiation may be essential in generating MJO signals (e.g., Hu and Randall 1994; Raymond 2001). There are also studies suggesting that feedback from sea surface temperature (SST) is important in the MJO dynamics (e.g., Flatau et al. 1997; Waliser et al. 1999; Inness and Slingo 2003).

Almost all of the instability mechanisms discussed above are aimed at explaining the intraseasonal variability in the Tropics, and thus have an energy source generated within the Tropics. As a matter of fact, intraseasonal variability is also an important feature in the extratropical atmosphere (e.g., Barnston and Livezey 1987). The extratropical low-frequency variability interacts with that in the Tropics. For example, Lau and Phillips (1986) found coherent fluctuations between the extratropical circulation and tropical convection on the intraseasonal time scale. Ferranti et al. (1990) suggested a coupled normal mode of low-frequency variability between the Tropics and midlatitudes that may be forced by convection anomalies in the tropical western Pacific associated with the MJO. The intraseasonal fluctuation of tropical convection in the western and central Pacific was found to influence the East Asian winter jet, which in turn affects wave propagation and the Pacific–North American (PNA) pattern (Schubert and Park 1991). In a numerical study Matthews et al. (2004) were able to reproduce some major features of the extratropical low-frequency variability by specifying a time-dependent tropical forcing that mimics the convective heating of the MJO.

An influence of the midlatitudes on the tropical low-frequency variability has also been observed. Liebmann and Hartmann (1984) reported that the change of tropical convection lags that of the extratropical 500-hPa height, suggesting that the midlatitude flow drives the Tropics. By analyzing the intraseasonal oscillation in the 1985/86 northern winter, Hsu et al. (1990) suggested that convection in the Indian Ocean could be triggered by a subtropical Rossby wave train, and the cold surge in south China could also lead to increased convection in the tropical Indonesian region.

Despite the above indications of extratropical forcing, it is still unclear how the MJO is excited and maintained. Using a linear two-level model with a realistic three-dimensional basic-state flow and a cumulus convection parameterization, Frederiksen and Frederiksen (1997, hereafter FF) found that one of the unstable modes couples the extratropics with a tropical 40–60-day disturbance, which is similar to the MJO. The connection of the MJO to the extratropical instability was also suggested by Straus and Lindzen (2000).

In the present study, the output of a long integration of a dry atmospheric model with a time-independent forcing is analyzed. There are no processes related to moisture, convection, feedbacks with radiation, or SST in this model. As will be seen, such a dry model produces a localized region of intraseasonal Kelvin wave activity in the Eastern Hemisphere, which may be a source for the MJO. The importance of interactions between the tropical and extratropical flows will be emphasized, leading to a global view of the intraseasonal variability.

In section 2 the model and analysis procedures are briefly described. The model climatology is shown in section 3. A spectral analysis and the time evolution of the equatorial flow are presented in section 4. In section 5 the extratropical signature of the intraseasonal vari-
ability is analyzed. The time evolution of the wave activity flux associated with the intraseasonal variability is discussed in section 6. In section 7 mechanisms responsible for the extratropical intraseasonal variability are investigated. The tropical response to an extratropical forcing is analyzed using a linear model in section 8. Section 9 gives a summary and discussion.

2. The model, data, and analysis procedures

The simple general circulation model (SGCM) as described in Hall (2000) is used in this study. It is based on the primitive equation model of Hoskins and Simmons (1975). It is a global spectral model with no moisture representation. The resolution used in this study is T31 with 10 vertical levels. The time step for integration is 22.5 min. A detailed description of the model parameters can be found in Hall and Derome (2000). A scale-selective dissipation of the form of $v^6$ with a time scale of 12 h at the smallest scale is used in the vorticity, divergence, and temperature equations. A level-dependent linear damping is also imposed on temperature and momentum. For the upper levels, the time scale of this damping is 30 days for momentum and 10 days for temperature. At the two lowest levels of the model, the momentum damping has time scales of 2 ($\sigma = 0.85$) and 0.67 days ($\sigma = 0.95$), while the temperature damping has time scales of 4 and 1.34 days at these two levels, respectively.

An important feature of this model is that a time-independent forcing is used that is calculated from observed daily data. This forcing is obtained as a residual for each time tendency equation by computing the dynamical terms of the model, together with the dissipation, with daily global analyses and averaging in time. All the processes that are not resolved by the model’s dynamics are thus included in the forcing. No topography is prescribed in the model, but its time-mean effect is accounted for by the forcing. It is necessary to keep in mind that the MJO, which is a transient phenomenon, can contribute to the time-independent forcing, especially in the Indian and western Pacific area. The time-mean effect of the MJO, if any, is thus included in the forcing to maintain a realistic climatology.

The daily data of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses (Kalnay et al. 1996) are used to calculate the forcing fields. The forcing fields are to represent the climatological conditions for the Northern Hemisphere winter season. Therefore, data of 30 winters from 1969/70 to 1998/99 are utilized, where the winter is defined as the 90-day period starting on 1 December. Forcing fields are calculated separately for each winter, and an average of the 30 forcing fields is obtained as the time-independent climatological forcing. The NCEP–NCAR reanalyses are also used to validate the model climatology.

With the time-independent climatological forcing, a perpetual Northern Hemisphere winter integration of 3660 days is conducted starting from an observed initial condition. Daily output is saved for analysis. The period of the first 40 days is taken as a spinup of the model, and is not used. This long integration will also be called the control run in the following discussions.

Space–time spectra are calculated following Hayashi (1982) to identify the dominant spatial and temporal scales of the tropical perturbations. This technique also allows us to separate the eastward- and the westward-traveling waves. To extract the variabilities of the intraseasonal time scales, a 20–100-day bandpass filter is used which is a 121-point Lanczos filter (Duchon 1979).

3. Model climatology

To assess the overall quality of the model, some basic climatological features of the model are presented in this section and compared with the observations. It is important to have a well-simulated time-mean flow, since the behavior of the intraseasonal variability is influenced by the basic flow. For example, wave propagations in the mid- and high latitudes are largely determined by the structure of the mean zonal flow (e.g., Webster and Holton 1982). One possible source for the extratropical low-frequency variability is the barotropic energy transfer from the mean flow (e.g., Simmons et al. 1983). The importance of the mean background state to the tropical intraseasonal variability has also been suggested (e.g., Hendon 2000; Inness and Slingo 2003).

The general climatological features of the SGCM were reported in Hall (2000) for the T21 version. It was found that this model not only reproduces remarkably well the stationary planetary waves but also the broad climatological characteristics of the transients. Here the upper-tropospheric model climate is shown for the current long integration of the T31 version of the SGCM (Fig. 1).

The simulated climatological zonal wind at 200 hPa (Fig. 1b) agrees well with the winter [December–February (DJF)] 200-hPa zonal wind climatology calculated from 30 yr (1969/70–1998/99) of the NCEP–NCAR reanalyses (Fig. 1a). The position and strength of the wintertime subtropical jet in the Northern Hemisphere with maxima near the east coasts of Asia and North America are reproduced. In the Tropics, the extent and strength of the mean easterlies and the west-
erly duct over the eastern Pacific are reasonably well simulated, though the model easterlies over South America extend eastward across the Atlantic to join the easterlies over Africa, while in the observations there is a weak mean westerly wind in the central Atlantic. The maximum of easterlies with a magnitude over 10 m s

The time-mean eddy streamfunctions at 200 hPa for the observations and the model are shown in Figs. 1c,d, respectively. Good agreement between the model and the observations can be found, especially in the Northern Hemisphere. The strength, position, and shape of the stationary waves in the Northern Hemisphere are well simulated. Discrepancies between the observed and the simulated climates are found in the Southern Hemisphere south of 30°S.

The large-scale divergent component of the upper-tropospheric flow is expressed by the velocity potential and is shown in Figs. 1e,f for the observations and model, respectively. Again, the model does a good job, though small differences in details can be observed.

4. Simulated intraseasonal variability

a. Spectral analysis

Space–time power spectrum analyses are performed on the 250-hPa zonal wind and velocity potential averaged between 5°S and 5°N. The daily unfiltered data from the perpetual winter integration are used. Fourier power spectra are calculated separately for each of fifteen 240-day periods. The spectra are averaged to obtain the mean power spectrum. The results are shown in Fig. 2.

The spectra with positive and negative frequencies correspond to eastward- and westward-propagating waves, respectively. As can be seen, eastward-propagating waves dominate in both the zonal wind and velocity potential fields. Both exhibit large spectral values between frequencies 0.02 and 0.04 day−1 (corre-
sponding to periods of 25–50 days) for wavenumber 1. The straight line represents the nondispersive Kelvin wave phase speed of 20 m s$^{-1}$, indicating a dominating Kelvin wave signature.

b. Zonal propagation

Figures 3a,b show the time–longitude distribution of the unfiltered 250-hPa velocity potential perturbation averaged between 10°S and 10°N for two 200-day periods of the perpetual winter integration. The eastward propagation of the perturbations can clearly be seen. The time–longitude diagrams of the 20–100-day filtered 250-hPa velocity potential for the same two periods are shown in Figs. 3c,d. This time, the eastward-propagating signal is even more obvious. On average, the amplitude of the oscillation is about 4–6 × 10$^6$ m$^2$ s$^{-1}$, which is comparable to that of the observed MJO (e.g., Slingo et al. 1996). The oscillation moves around the equator in about 30 days, corresponding to a phase speed of about 15 m s$^{-1}$, which is faster than the observed MJO (e.g., Hendon and Salby 1994) but comparable to that in many GCMs (Slingo et al. 1996). We will refer to this simulated variability as a tropical intraseasonal variability (TIV).

The low-frequency wave seems to reach its maximum amplitude from the Indian Ocean to the date line. The amplitude is reduced in the longitudes from 180°W. This is clear in Fig. 3c. This feature is confirmed in Fig. 4, which shows the variance of the 20–100-day filtered 250-hPa velocity potential as a function of longitude along the equator. A similar longitudinal dependence of the observed MJO was found in many studies (e.g., Knutson and Weickmann 1987). It is often attributed to the strong convective activity in the Indian Ocean and western Pacific. Since there is no convection in the SGCM, here another mechanism must be responsible for the longitudinal distribution of the intraseasonal variability. We attribute this to the tropical mean zonal wind distribution as will be discussed in section 8. The signal, however, is not ubiquitous throughout the whole integration. For example, at the beginning of the second period (Fig. 3d), the amplitude is weak and the eastward propagation is less clear. Also the period of oscillation in the second period is a little longer than that at the beginning of the first period. As reported in Salby and Hendon (1994), the observed MJO is episodic and irregular.

c. Spatial structure

To identify the spatial structure of the simulated TIV, an empirical orthogonal function (EOF) decomposition is applied to the daily data of the 20–100-day filtered 250-hPa velocity potential. The area of analysis is the global band from 60°S to 60°N. The effect of unequal areas represented by different grid points is taken into account by multiplying the daily data by the square root of the cosine of the latitude at that grid point. Figure 5 shows the distribution of the two leading EOFs, presented as the linear regression patterns of 250-hPa velocity potential to the respective principal component (PC) time series. They account for 34% and 30% of the 20–100-day variance, respectively. The combination of these two modes represents a dominant part of variability on this time scale (EOF3 and EOF4 only account for 6% and 4% of the variance, respectively). Both EOF1 and EOF2 have a mainly zonal wavenumber-1 structure. EOF1 has a negative center over the equatorial western Pacific, corresponding to an anomalous upper-tropospheric divergence. Over the equatorial Africa, a positive center of velocity potential is ob-
served, which corresponds to an upper convergence. The structure of EOF2 is almost in quadrature with that of EOF1, with centers shifted westward, such that the upper-divergence center is over the equatorial Indian Ocean. The time-lag correlation of PC1 and PC2 is presented in Fig. 6. As can be seen, the maximum positive (negative) correlation ($-0.70$) occurs when PC2 leads (lags) PC1 for 8 days. This indicates that a linear combination of EOF1 and EOF2 represents an eastward wavenumber-1 propagation with a period of about 32 days. Thus, a TIV index is constructed using PC1 and PC2 as

$$\text{index}(t) = \frac{PC2(t-1) + PC2(t) + PC2(t+1)}{3} + \frac{PC1(t+7) + PC1(t+8) + PC1(t+9)}{3},$$

where $t$ is the time in days.
As can be seen, the TIV index is defined as in phase with PC2. This TIV index is used as the basis to construct the life cycle of the simulated TIV. Note that either PC1 or PC2 can be used as a TIV index. Except for a time lag, a different choice of the index makes little difference to the results.

d. Lag regression

Time-lag regressions were calculated between the TIV index and some 20–100-day filtered 250-hPa variables averaged for the band 10°S–10°N. Figure 7a shows the lag regression for the velocity potential. This gives a composite view for the 20–100-day oscillation as seen in Fig. 3. An eastward-traveling wavenumber 1 is clearly shown, which crosses the equator in about 30 days, although the amplitude is reduced in the longitudes from 180° to 60°W. The lag regression for the 250-hPa zonal wind is shown in Fig. 7c. Corresponding to one standard deviation of the TIV index, the amplitude of the equatorial TIV zonal wind reaches about 2 m s$^{-1}$, which is comparable to that in the observations (e.g., Hendon and Salby 1994). The eastward propagation for the zonal wind is obvious in the Eastern Hemisphere, with maximum amplitude near 30° and 120°E. In the Western Hemisphere, however, the oscillation looks more like a standing wave. Figure 7c illustrates the lag regression for the geopotential height at 250 hPa. In the Eastern Hemisphere, again clear eastward propagation can be observed. The geopotential height anomaly in this region seems to be in phase with the zonal wind anomaly, implying a Kelvin wave signal. In the Western Hemisphere, the propagating feature of the geopotential height anomaly is complex. A separation of the anomaly occurs near 30°W, where the anomaly changes sign. It behaves as if a forcing has been applied at this longitude.

Regression maps were calculated for the 250-hPa velocity potential field at different lags with respect to the TIV index (not shown). An eastward propagation of a perturbation with wavenumber 1 centered along the equator was clearly seen. The eastward propagation was quite continuous from the Greenwich longitude to about the date line. In the Western Hemisphere from the date line to western Africa, however, the propagation was found to be less continuous. As the divergence center moved close to the date line, it was seen to stall while its amplitude decreased. At the same time, another divergence center developed in the tropical Atlantic. The Atlantic center then was seen to increase in amplitude, and eventually dominated the upper divergence. Therefore, the eastward propagation of the TIV in the Western Hemisphere is a manifestation of the process of a decay of the oscillation near the date line and a new development in the tropical Atlantic, implying that the source for the simulated TIV lies in the Atlantic sector. As will be discussed in the next section, the development of the TIV in the Atlantic is a result of an extratropical influence. Though with a weaker signal, lag regression maps of the 850-hPa velocity potential (not shown) have a similar structure and evolution to its 250-hPa counterparts, but with the opposite sign, indicating a baroclinic structure in the Tropics, consistent with the observations.

In previous observational studies (e.g., Knutson and Weickmann 1987; Hsu 1996), it was found that the MJO-related OLR, which represents tropical convec-
tion, moves eastward from the Indian Ocean to the central Pacific, but little evidence of eastward propagation was found east of the date line. Hsu (1996) also found that the convective activity in Central and South America and the tropical Atlantic develops in situ. The time evolution of the TIV in the SGCM simulation shown above is thus similar in these respects to the observed MJO.

In Fig. 8 regression maps are shown for the wind and geopotential height anomalies at 250 hPa. At day −8 (Fig. 8a), westerly wind anomalies appear over the tropical Indian Ocean, accompanied by positive geopotential height anomalies. The wind distribution constitutes an upper convergence over the western Pacific. From day −8 to day 0, the area of westerly wind anomalies decreases and moves eastward. At day −4 (Fig. 8c), easterly wind anomalies start to appear near the Greenwich longitude and propagate eastward afterward. At day 0, significant upper divergence develops over the Indian Ocean with westerlies to the east and easterlies to the west. The easterly anomalies then propagate eastward across the Indian Ocean (from day −4 to +6). During this period, positive (negative) geopotential height anomalies are paired with westerly (easterly) wind anomalies, implying a Kelvin wave structure. Another important feature is observed in the subtropical Atlantic region, where a strong anticyclonic circulation centered near 20°N, 45°W is present from day −4 to +6. As will be seen in the next section, the generation of this flow pattern is linked to extratropical activities. Its intensification in the subtropical Atlantic provides a forcing to the tropical flow. This is also reflected in the propagating feature of the geopotential height anomaly in the tropical Atlantic (Fig. 7c). The generation of tropical low-frequency waves by subtropical forcing will be discussed in section 8. An obvious disagreement with the observed MJO is the lack of the Rossby wave gyre that is observed to be dragged along with the divergence in the Eastern Hemisphere (e.g., Hendon and Salby 1994).

The vertical structure of the TIV is shown in Fig. 9, which shows equatorial cross sections of zonal wind
(Fig. 9a), vertical wind (Fig. 9b), and temperature (Fig. 9c). They are calculated as lag regressions to the TIV index at day +8, which corresponds to an upper-level divergence centered in the western Pacific as shown in EOF 1 (Fig. 5a). The zonal wind has a baroclinic distribution, with a sign reversal from the upper to lower troposphere. The Kelvin waves appear to be confined mainly to the upper troposphere and thus have a vertical structure different from the first internal mode, which would explain the slower phase speed than that of the first internal mode Kelvin wave ($\sim 40$ m s$^{-1}$). The upper-tropospheric Kelvin wave is likely a result of the extratropical forcing in the upper troposphere as will be discussed in the next sections.

5. Extratropical signature of the simulated intraseasonal variability

As described in the last section, a significant intraseasonal variability is generated in the long SGCM integration. As discussed in Sardeshmukh and Hoskins (1988), a tropical divergent flow results in a Rossby wave source in the subtropics. In the extratropical westerlies, the wave energy propagates eastward to generate a wave pattern downstream. The Rossby wave source associated with the tropical divergent flow is not the only energy source of intraseasonal variability in the mid- and high latitudes. As discussed in Simmons et al. (1983), the low-frequency circulation anomalies can extract energy from the mean flow by barotropic conversion. Moreover, the transient eddy forcing anomaly associated with a modified storm track feeds back to the low-frequency flow (e.g., Lau 1988). In this section we take a look at the time evolution of the extratropical intraseasonal variability associated with the simulated TIV. Discussions of contributions to the low-frequency circulation anomalies by different mechanisms will be presented in the next sections.

Figure 10 shows global lag regression maps of the...
250-hPa streamfunction field with respect to the TIV index at a 2-day interval from day $-8$ to day $0$. The shaded areas represent regressions with a significance level of 0.05 according to a two-tailed Student’s $t$ test, where 120 degrees of freedom are used (the number of time series points is 3600) considering that the data is autocorrelated. A global-scale signal in the 250-hPa circulation anomalies can be found associated with the TIV. The low-frequency signal in the mid- and high latitudes occurs mainly in the Northern Hemisphere. The apparent reason for that is the stronger westerly wind and stationary waves in the Northern Hemisphere than those in the Southern Hemisphere in DJF. The westerly wind and stationary waves are important for Rossby wave propagation, barotropic energy conversion, and transient eddies, which generate and maintain the low-frequency flow anomalies.

At day $-8$ (Fig. 10a), in the tropical Eastern Hemisphere a cyclonic circulation is observed on each side of the equator. This cyclonic couplet straddling the equator with anomalous westerly wind along the equator is located between the upper convergence over the tropical western Pacific and divergence over western tropical Africa. In the northern extratropical region, a wave train of circulation anomalies can be observed over the PNA region, which is reminiscent of the PNA pattern. From day $-6$ to 0, the tropical cyclonic couplet moves eastward and decreases in extent. This process is consistent with the eastward propagation of the upper-divergent flow and wind field (Figs. 8b–e). During this period, the Pacific wave train weakens and gradually disappears. By day $+2$, the tropical cyclonic couplet only has a weak signal over the western Pacific, and the PNA has only a weak anticyclonic circulation anomaly left over northwestern Canada.

During the same time, low-frequency anomalies develop vigorously in the Atlantic region. An anticyclonic circulation anomaly appears over the subtropical Atlantic on day $-8$, downstream of the PNA-like wave train. This anomaly increases in amplitude afterward. A cyclonic anomaly in the North Atlantic starts to develop from day $-4$. Together with the subtropical anticyclonic circulation anomaly, a dipole structure of anomalies in the North Atlantic is constructed, which reaches its maximum strength around day $+2$. The dipole then weakens gradually in amplitude.

When the Atlantic signal is strong, a pair of anticyclonic anomalies on each side of the equator appear near the Greenwich longitude, to the west of the upper-divergence center over the tropical Indian Ocean. This couplet of anticyclonic circulation anomalies moves eastward together with the upper-divergent flow. By day $+4$ when the upper-divergence center reaches the western Pacific, another PNA-like wave train starts to develop. The subtropical cyclonic circulation anomaly in the western Pacific that is left over from the last half cycle of the TIV serves as the southern center for the PNA-like wave train. By day $+10$, the pattern is fully developed, similar to that of day $-8$ with a reversed sign.

The above analysis indicates that the intraseasonal variability in the Tropics is coupled with that in the extratropics. An EOF analysis was conducted for the 20–100-day band-filtered 250-hPa streamfunction for the region of 60°S–85°N, 0°–360°. The results (not shown) confirm this coupling. EOF 1, which accounts for...
for 17% of the total variance, has a global distribution that is very similar to the day-0 regression of 250-hPa streamfunction to the TIV index (Fig. 10e); whereas EOF2, which explains 13% of the total variance, is very similar to the day −8 regression map in both the Tropics and extratropics (Fig. 10a).

Lag regression maps of the 850-hPa streamfunction (not shown) show that in the Tropics the anomalies tend to be out of phase with those at 250 hPa, implying a baroclinic structure. In the extratropics poleward of about 30°, the 850-hPa streamfunction anomalies tend to be in phase with those at 250 hPa, implying an equivalent barotropic structure. In summary, a global low-frequency signal in the rotational circulation is associated with the simulated TIV. In the Eastern Hemisphere, couplets of cyclonic
and anticyclonic circulation anomalies straddling the equator propagate eastward up to west of the date line. In the extratropical region, the pattern seems develop in situ with little evidence of phase propagation. In the extratropical PNA region, the signal communicates with the downstream anomalies in the form of a wave train. In the North Atlantic area, intensification of anomalies with a dipole structure is followed by the development of another tropical couplet of circulation anomalies near the Greenwich longitude.

Somewhat similar coherent fluctuations between the MJO and the extratropical circulation were found in previous observational studies (e.g., Lau and Phillips 1986; Knutson and Weickmann 1987; Ferranti et al. 1990; Hsu 1996). The most robust signal is the PNA-like wave train that is associated with the MJO-related anomalous tropical convective activity in the western-central Pacific. Besides the PNA-like wave train, Lau and Phillips (1986) found another wave train that originates from the North Atlantic and is associated with the MJO. Ferranti et al. (1990) found that the North Atlantic Oscillation (NAO) is linked to a phase of the MJO. Hints of development of North Atlantic circulation anomalies can also be found in the lag correlation of the global streamfunction to the MJO in the observational study of Hsu (1996).

6. Extratropical wave activity flux

In the last section, it was observed that a coherent variation exists between the circulation anomalies in the mid- and high latitudes and the TIV. Here we look at the time evolution of the wave activity flux associated with the extratropical circulation anomalies.

The wave activity flux vector (W vector) proposed by Takaya and Nakamura (2001) is analyzed, which is based on the conservation of wave activity pseudomomentum. This formulation of wave activity flux vector is applicable to both stationary and migratory quasigeostrophic eddies on a zonally varying basic flow. The phase independence of the W vector allows snapshots of wave dispersion to be taken in each stage of the evolution of the circulation anomalies. As seen in the last section, in the extratropical region the pattern develops in situ, changes amplitude in time but has a negligible phase speed. Thus, the horizontal components of the W vector can be given by

\[
W = \frac{1}{2|U|} \left[ \begin{array}{c} U(\psi_u^2 - \psi_u \psi_{uu}) + V(\psi_u \psi_v - \psi_{uv}) \\ U(\psi_u \psi_v - \psi_u \psi_{uv}) + V(\psi_v^2 - \psi_{vv}) \end{array} \right]
\]

where \( \psi \) is the perturbation streamfunction and the subscripts represent partial derivatives. Here \( U = (U, V) \) is the two-dimensional time mean flow of the long integration. The wave activity flux vectors are calculated based on lag regression maps of the 250-hPa streamfunction with respect to the TIV. The amplitude corresponds to one standard deviation of the TIV index.

Figure 11 depicts the wave activity flux with respect to the TIV index at a 2-day interval from day \(-8\) to \(+8\). Also shown on the maps as contours is the potential vorticity (PV) anomaly, which is calculated as the lag regression of the PV onto the TIV index as well. The square of the PV anomaly is proportional to the wave activity. At day \(-8\) (Fig. 11a), a wave train with four alternating positive and negative PV anomaly centers is seen arching across the PNA regions. This wave train is similar to the PNA pattern, in agreement with the streamfunction anomaly (Fig. 10a). Along the path of the wave train, a wave activity flux is clearly seen originating from the subtropical Pacific and propagating northeastward. From day \(-6\) to \(0\), the Pacific wave train weakens and gradually disappears. By day \(+2\), almost no trace of the PNA can be found (Fig. 11f).

During the same period, PV anomalies and wave activity fluxes develop vigorously in the Atlantic region. A dipole PV anomaly structure is formed in the North Atlantic that is composed of a negative PV anomaly center near 30°N and a positive PV anomaly center near 50°N. A trace of this structure appears as early as day \(-6\). It increases in strength and reaches its maximum amplitude around day \(+2\) (Fig. 11f). Accompanied with the PV anomalies, two branches of wave activity fluxes are observed starting from the midlatitude Atlantic. The northern branch has a group of vectors pointing to the east, along the path of a zonal wave train up to central Russia. A similar wave train was observed in the observational study of Lau and Phillips (1986). The southern branch has a group of W vectors pointing to the south, indicating wave activity flux into the Tropics. Interestingly, the south branch has a much stronger magnitude than the eastward flux of the northern branch. Strong divergence of the wave activity flux (not shown, but can be inferred from the flux vectors) is located in the midlatitude Atlantic near 40°, implying a source in this area. Equatorward wave activity flux was also found in previous studies in the North Atlantic (e.g., Schubert and Park 1991), and even for the zonal average (e.g., Magnusdottir and Haynes 1996). The dipole structure then weakens gradually in amplitude. Another PNA-like wave train and northeastward wave activity flux starts to be established in the Pacific sector from around day \(+6\) (Fig. 11h). By day \(+8\), the distribution looks similar to that of day \(-8\) except for a reversal of sign in PV anomalies (Fig. 11i).

In summary, a clear pattern of wave activity flux
emerges in the extratropics associated with the TIV. The wave activity propagates from the Tropics to the mid- and high latitudes in the Pacific region, implying a tropical influence in this sector. On the other hand, in the Atlantic sector, there is a strong wave activity flux from the midlatitudes southward into the Tropics, implying an extratropical influence. These fluxes of wave activity are coupled with the divergent flow in the Tropics. Compared with the 250-hPa velocity potential field, it is seen that the northward wave activity flux occurs when the tropical convergent (divergent) center moves into the western Pacific, while the development of the tropical Atlantic divergent flow occurs when strong southward wave activity flux makes its way into the

![Fig. 11. Time evolution of the wave activity flux (m² s⁻¹) associated with the TIV. The arrows are the horizontal (W vectors), and the contours the potential vorticity anomalies. The amplitude corresponds to 1 std dev of the TIV index. The CI for the PV anomalies is 0.3 PVU, where 1 PV unit (PVU) = 1.0 × 10⁻⁶ m² s⁻¹ K kg⁻¹. The zero line is omitted. Contours in dash and solid are of negative and positive values, respectively. The shaded areas represent PV regression anomalies with a significant level of 0.05 according to a two-tailed Student’s t test. Scaling for arrows is given below each panel. Wave activity flux with magnitude smaller than 0.1 m² s⁻² is not plotted.](image)
Tropics (Fig. 11e). The largest northward wave activity flux in the Pacific is followed by the largest southward wave activity flux in the Atlantic about 8 days later. The southward wave activity flux in the Atlantic sector is much stronger than the northward wave activity flux in the Pacific region. It appears that there is a source in the extratropics that supplies energy to the global intraseasonal variability. In the next section, we will discuss mechanisms responsible for the generation of kinetic energy for the low-frequency variability in the extratropics.

7. Mechanisms of extratropical intraseasonal variability

a. Barotropic energy exchange with the basic flow

One energy source for the extratropical intraseasonal variability is the barotropic energy exchange with the background mean flow. Following Simmons et al. (1983), the barotropic conversion of kinetic energy from the mean flow to a disturbance, to a good approximation, can be written as

$$ C = \mathbf{E} \cdot \nabla \pi, $$

where \( \mathbf{E} = -(u_i^2 - v_i^2, u_i, u_i) \) is the extended Eliassen–Palm flux, which has been discussed in detail by Hoskins et al. (1983). The overbar represents the climatological mean and the subscript \( l \) refers to the low-frequency perturbation. The calculations of \( \mathbf{E} \) and \( C \) are done using the lag regression of 250-hPa wind with respect to the TIV index.

Figure 12a shows the \( \mathbf{E} \) vector at day +2, when the North Atlantic anomaly dipole is at its peak strength (Fig. 10f) and the southward wave activity flux reaches its maximum (Fig. 11f). Superimposed on the map as contours is the climatological zonal wind \( \bar{u} \). The vectors
are largest in the North Atlantic jet exit region and point westward in the direction of the gradient of the climatological zonal wind, suggesting that the intraseasonal variability is extracting kinetic energy from the basic state. Figure 12c depicts the energy conversion (C) from the basic state to the TIV-related flow at day $+2$. Strong conversions are seen in the central North Atlantic.

At day $+10$, when the Pacific PNA pattern reaches its maximum amplitude (Fig. 10j), largest $E$ vectors are found in the North Pacific jet exit region and they also point westward in the direction of the gradient of the climatological zonal wind (Fig. 12b). The intraseasonal perturbation extracts kinetic energy from the basic mean flow in the North Pacific (Fig. 12d). The energy conversion in the North Pacific is, however, a little weaker than that in the North Atlantic about 8 days earlier.

It was found in previous observational studies that the PNA pattern grows by extracting kinetic energy from the mean flow (e.g., Schubert and Park 1991; Hsu 1996). Here it is found that the mean flow provides kinetic energy to low-frequency variabilities in the jet exit areas in both the Pacific and Atlantic. The conversion in the North Atlantic is even stronger.

b. Interaction with transient eddies

Changes in circulation patterns such as those associated with the TIV modify the path and strength of the storm track and synoptic-scale transients. The modified transient activity, in turn, feeds back onto the low-frequency pattern. This provides another source for the low-frequency variability. To estimate this effect, the geopotential height tendency of the time-mean flow is calculated for each of the 5 days. The height tendency caused by the convergence of the transient eddy vorticity flux can be written as

$$\frac{\partial \zeta}{\partial t} = \frac{f}{g} \nabla^{-2} \left[ \nabla \cdot \left( \zeta \nabla^2 \right) \right],$$

where $\zeta$ is the relative vorticity, and $\mathbf{V}$ is the horizontal velocity vector. The overbar represents the 5-day average, and the prime is the departure from the 5-day average. A similar formulation of geopotential height tendency by transient eddies was used in previous studies (e.g., Lau 1988; Lin and Derome 1997). The daily tendency is calculated using the 250-hPa winds as for the 5 days centered on the target day. Then lag regressions are calculated between the daily tendency and the TIV index.

In Fig. 13 we present the height tendency caused by transient eddies at day $+2$ and $+10$. As can be seen, at day $+2$ there is a negative height tendency (Fig. 13a) in the North Atlantic. This negative height tendency is collocated with the positive PV anomaly in the North Atlantic as can be seen from Fig. 11f. It is also in phase with the north center of the North Atlantic dipole anomaly (Fig. 10f). This indicates that the transient eddies are reinforcing the height anomalies, thus acting as an energy source.

At day $+10$, however, when the North Atlantic dipole anomaly disappears and the PNA-like patterns reaches its maximum amplitude, a noticeable contribution can be found from the transient eddies to reinforce the North Pacific pattern (Fig. 13b). If the extratropical activities have a positive contri-

![Fig. 13. Lag regressions of 250-hPa geopotential height tendency caused by transient vorticity flux convergence with respect to the TIV index for day (a) $+2$ and (b) $+10$. The CI is $0.8 \times 10^{-8}$ m s$^{-1}$. Contours in dash and solid are of negative and positive values, respectively, and the zero line is omitted.](image-url)
8. Tropical response to extratropical forcing

From the lag regression maps of the 250-hPa velocity potential (not shown), it was seen that from about day −8 to about day +8 a divergence center in the central Pacific decays and a new divergence center develops in the tropical Atlantic. During this period, a series of extratropical processes occur: a wave train accompanied by northward wave activity flux in the North Pacific; development of low-frequency anomalies by extracting kinetic energy from the mean flow and transient eddies, especially the dipole in the North Atlantic; and then a strong southward wave activity flux in the Atlantic. It can be concluded that in the Western Hemisphere the eastward propagation of the TIV is connected with the extratropical processes.

In the Eastern Hemisphere, continuous eastward propagation along the equator was found. The disturbance takes about 15 days to travel across the Eastern Hemisphere. As seen from the wave activity flux analysis, the development of the divergent field near equatorial Africa is accompanied by a strong equatorward wave activity flux from the midlatitude North Atlantic. A possible mechanism for the tropical disturbance is the extratropical forcing.

To study the tropical response to an extratropical forcing, a linear perturbation model is used that is based on the SGCM, with the approach as described in Hall and Derome (2000). The basic state is chosen as the model winter climate. A forcing is applied to maintain this basic state, which is calculated as a residual in the model equations with only time mean quantities. A perturbation forcing is then added in the form of a heat source. The heating perturbation results in vertical motion and upper-level divergence, which simulates a vorticity source, and is defined as a stationary transient heating:

\[
f(x, y, \sigma, t) = A(x, y, \sigma) \sin \left( \frac{2\pi t}{T} \right) ,
\]

where \(A(x, y, \sigma)\) is the amplitude, and takes an elliptical form in the horizontal, with major and minor axes of 40° longitude and 20° latitude. The magnitude is proportional to the squared cosine of the distance from the center. The heating amplitude has a vertical profile of \((1 - \sigma) \sin [\pi (1 - \sigma)]\), which peaks at \(\sigma = 0.35\) with a vertically averaged heating rate at the center of 1 K day\(^{-1}\). The linear model is integrated for 30 days starting from the winter climate. To ensure a small-amplitude linear response, the forcing amplitude is scaled by \(10^{-3}\) times the anomalous forcing, and the solutions are scaled back for display purposes.

Three experiments are conducted: 1) the heating source centered at 30°N, 30°W and with a period \(T = 10\) days (Exp1); 2) the heating source centered at 30°N, 30°W and with a period \(T = 30\) days (Exp2); and 3) the heating source centered at 30°N, the date line and with a period \(T = 30\) days (Exp3). As discussed in Jin and Hoskins (1995) and Hall and Derome (2000), the free tropical Kelvin wave in the same model travels across the equator in about 10 days. Exp1 is designed to ana-
analyze the tropical response to a stationary extratropical forcing with the same frequency as the free Kelvin wave. Exp2 is used to mimic the North Atlantic forcing associated with the TIV. Exp3 is designed to test the sensitivity of the result to the longitudinal location of the forcing.

Time–longitude distributions of the 250-hPa velocity potential response at the equator are shown in Fig. 15 for the three experiments. As can be seen, the forcing with a period of 10 days generates an eastward-propagating wave (Fig. 15a). Like the free Kelvin wave, the forced Kelvin wave travels across the equator in about 10 days. This is not surprising, since energy is supplied periodically to the equatorial Kelvin wave to maintain its eastward propagation. It can be noted also that the amplitude tends to be larger in the tropical mean easterly basic flow from 0° to 180° (Fig. 1a) than that in the mean westerly flow.

With the Atlantic TIV forcing (T = 30 days), slow eastward propagation occurs in the Eastern Hemisphere (Fig. 15b). It takes about 15 days for the disturbance to travel from the Greenwich longitude to the date line. The slightly different behavior comparing to the full model (Fig. 7b) may be due to the deep heating.

Fig. 15. Power spectra of the height tendency caused by transient activity at 50°N, 45°W.

Day 9 response

Fig. 16. Exp2 response of wind vectors and geopotential height at 250 hPa. The CI is 5 m. Winds with a speed less than 0.2 m s⁻¹ are not plotted. Scaling for arrows is given below the panel. Areas with negative height anomalies are shaded.
profile used in the linear experiment. Figure 16 depicts the horizontal distribution of the 250-hPa wind and geopotential height response at day 9 for Exp2, about 2 days after the maximum forcing. Right over the extratropical forcing, an anticyclonic flow response is dominant. To the south of the forcing, a strong southwestward cross-equator flow is found. The northeastward-returning flow from the Southern Hemisphere, together with the anticyclonic flow response, forms a cyclonic circulation centered near 20°N, 20°E. Farther downstream along the climatological subtropical westerly jet, a wave train is generated that is discernible up to east of Japan. The above feature is an indication of a Rossby wave. Over the equator, on the other hand, a continuous band of eastward flow response can be found from the middle Atlantic up to the middle Indian Ocean, reminiscent of a Kelvin wave signal.

When the forcing is placed near the Pacific jet exit (Exp3; Fig. 15c), the response in Tropics is weaker than Exp2 and almost no eastward propagation is found in the Eastern Hemisphere.

Significant tropical responses to extratropical forcings were also found in previous studies. For example, Zhang and Webster (1992) investigated the equatorial response to an extratropical stationary transient forcing using a simple linear model. They also found that the Kelvin wave is stronger in the mean easterlies than in the mean westerlies. Their forcing to generate a significant Kelvin wave signal, however, has a very short time scale (≈1 day). Zhang (1993) and Hoskins and Yang (2000) studied the tropical response to an extratropical forcing that has an eastward or westward phase speed. Again, it was found that the equatorial Kelvin wave response is stronger in an easterly mean flow than that in a westerly mean flow, which they attributed to the Doppler-shift effect on the forcing frequency. Our result is in agreement with the previous studies that a significant Kelvin wave–like response is generated in the equatorial easterly mean flow region (Eastern Hemisphere), though we are dealing with an extratropical forcing with a much lower frequency.

9. Summary and discussion

In the present study, the output from a long integration of a primitive equation dry atmospheric model was analyzed. It was found that significant intraseasonal variability can be generated in the absence of moisture and tropical convections. The model-generated TIV has a Kelvin wave structure.

Through analyses of lag regressions, extratropical wave activity flux, barotropic energy conversion from the mean flow, and feedback from the transient eddies, a global picture of the simulated TIV emerges. When the tropical divergence (convergence) center reaches the western Pacific, a PNA-like wave train is generated in the extratropical PNA region, accompanied by a northeastward wave activity flux. The extratropical low-frequency flow develops by extracting kinetic energy from the basic mean flow in the Pacific and Atlantic jet exit regions. In the North Atlantic, the transient eddy anomalies associated with a modified storm track also help to enhance the low-frequency flow. This leads to a strong equatorward wave activity flux in the subtropical Atlantic, which acts as a forcing to the tropical disturbance. In the tropical Eastern Hemisphere, the TIV continuously propagates eastward. When it reaches the western and central Pacific, another PNA-like wave train propagates eastward and the above process repeats.

The ultimate energy source of the simulated TIV is likely located in the extratropical atmosphere. A part of the energy comes from the mean flow though barotropic conversion in the westerly jet exits. Transient eddy activity that is caused by baroclinic instability also contributes to the TIV, especially in the North Atlantic area, where the vorticity flux convergence by transients reinforces the low-frequency flow.

There are many aspects that the simulated TIV and the observed MJO are in common. These include the longitudinal dependence and very large scale of the tropical divergence field, the intraseasonal time scale, and the eastward propagation. It is possible that in nature the eastward-propagating tropical intraseasonal variability is originally created by the dry dynamics of the extratropics–Tropics as in the model, and modified by tropical convective processes into the observed MJO.

One important aspect of the observed MJO is its seasonal dependence. As observed by Madden (1986), the MJO has its largest amplitude during the boreal winter months. From the point of view that the energy source of the TIV is in the extratropics, this seasonal dependence can possibly be explained. The Northern Hemispheric westerly jets as well as the baroclinic instability are strongest in northern winter, and so is the kinetic energy supply from the mean flow and transients.

The atmospheric dynamics in the upper troposphere is very important to the TIV, considering that strong interactions occur at the extratropical jet level. On the one hand, the extratropical barotropic energy conversion and transient eddy feedback are largest near the jet level. On the other hand, the upper-level easterly winds in the Eastern Hemisphere allow an equatorial Kelvin wave response to an extratropical forcing to be largest there. It is possible that the dynamically generated
tropical upper divergence (convergence) creates a condition to favor (suppress) the deep convection in the Indian Ocean in the real atmosphere.

Our result indicates that a realistic 3D climatology supports intraseasonal variabilities in the Tropics that have some important characteristics of relevance to the MJO. In a sense, this provides support for the instability theory of FF, who suggest that the MJO is one of the unstable modes in a realistic 3D varying basic state. Another point in common between FF and this study is that the CISK mechanism that was proposed in many earlier studies is excluded. In our case a dry model is utilized, while in FF the relevant moist static stability parameter is set positive everywhere, so that wave-CISK instability is not possible. There are also significant differences between these two studies. While FF is built on instability analysis in a linear framework, the present study is constructed from a long integration of a nonlinear model. In FF the effect of moisture and convection is included in the equation set by a parameterization, while in the present study a dry model is used that has no interactions with moisture and convection.

Many previous studies have viewed moisture and convection as a central part of the MJO. The fact that some of the key features of the MJO are missing in the dry model simulation also indicates that tropical deep convective processes are crucial for the observed MJO. The latent heat release of tropical deep convection, especially in the Indian Ocean, provides energy to fuel the MJO. The slower than simulated eastward-propagating speed that is observed to be about 5 m s⁻¹ in the Eastern Hemisphere is likely to be related to moisture and convection. FF also suggested that latent heat release is important in determining the vertical structure of the tropical disturbances. The difference in vertical structure of the TIV in the present study compared to that of the observed MJO is likely related to the lack of convection. This may also be responsible for the absence of the Rossby wave gyre to the west of the tropical upper divergence. In the real atmosphere, the equatorial Rossby wave gyre may be a response to an active diabatic heating associated with latent heat release. The relative importance of the upper-level dynamics and that of the deep convection needs to be further investigated for a more complete understanding of the MJO.

**Acknowledgments.** This research was made possible by funds provided to the Canadian CLIVAR Research Network by the Natural Sciences and Engineering Research Council of Canada (NSERC) and the Canadian Foundation for Climate and Atmospheric Sciences (CFCAS). We thank Dr. Harry Hendon and two anonymous reviewers whose comments and suggestions helped to improve our paper.

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