Angular Momentum, Length of Day and Monsoonal Low Frequency Mode

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(Manuscript received 29 July 1991, in revised form 7 December 1991)

Abstract

In this paper some global aspects of the intraseasonal oscillations on the time scale of 30 to 50 days are explored. Noting that the variability of zonal flow of the monsoon, the atmospheric angular momentum and the length of day are strongly correlated on this time scale, we have made an effort to examine the global variability using the length of day as a point of reference. The scenario of this cycle is presented starting from a super cloud cluster at the near equatorial latitudes. This seems to be accompanied with an acceleration of zonal flows, an increase of the atmospheric angular momentum and an increase in the length of day. The transfer of westerly angular momentum from the earth to the atmosphere occurs over regions of the surface easterlies to the east of the super cloud clusters resulting in an increase in the length of day. During this transition from a mean length of day to a maximum length of day, an active phase of the Indian summer monsoon is noted. The interesting aspect of the length of day transition occurs on its return cycle when the near equatorial cloud cover eases or moves away from the equator with a decrease in the monsoonal zonal flows and a reduction of this component of atmospheric angular momentum. The length of day does not simply go back to an equilibrium value, but the long term data from the laser ranger shows an overshooting beyond that to a minimum value. This transition is characterized in general by monsoon break-like conditions, counter monsoon flows in the low levels and by a transition from high index to low index conditions in the upper troposphere of the middle latitudes. Phenomenologically, some blocking situations have been noted over the higher middle latitudes during this transition. The reduction of the angular momentum is attributed to the transfer of the westerly angular momentum from the atmosphere to the earth via frictional and mountain torques. These torques exhibit a clear relationship to the changes in the atmospheric angular momentum on this time scale. The behavior of the middle latitude low frequency variability is also in part explained by the meridional wave energy flux. That problem is examined in this context with the full non-linear equations in the frequency domain. It is shown that unlike the linear problems where such fluxes are inhibited beyond the critical latitude, the nonlinear problem permits the temporal oscillations of zonal flows on this time scale. As a consequence, a significant tropical-middle latitude coupling is noted by this process.

A simple mathematical model of the oscillation is also presented. It is a local theory in which ocean and atmosphere interact. Initially, the atmosphere is stably stratified with weak winds at the sea surface and stronger winds aloft; the ocean has a surface mixed layer of temperature T_s lying over deep cold water. Solar heating gradually increases T_s which leads to atmospheric convection with associated transport of horizontal momentum and increased winds at the sea surface. Increased winds lead to deepening of the mixed layer and a drop in T_s because of mixing of deep cold water with surface waters. Convection ceases, winds decay, and the cycle repeats only after solar heating has once more increased T_s . The period of this oscillation is shown to be on the order of 30 days.

1. Introduction

This study addresses the intraseasonal low frequency modes of the atmosphere in the context of atmospheric angular momentum and length of day changes. A number of observational aspects of this problem are reviewed that relate to tropical (mon-

soonal) and middle latitude variability on this time scale. Perhaps the most interesting aspect of this variability is the changes in the length of day. The earth, with a mass roughly 10^6 times larger than that of the atmosphere, swings faster and slower basically in response to atmospheric motions on this time scale. Tropical organized convection appears to be a major factor in the initiation of these mo-

Table 1. List of Acronyms

AEAM	Atmospheric Effective Angular Momentum
ECMWF	European Center for Medium Range Weather Forecasts
FGGE	First GARP Global Experiment
GCM	General Circulation Model
U.S. GOES	U.S. Operational Geostationary Satellite
LOD	Length of Day
MONEX	Monsoon Experiment
NMC	National Meteorological Center
OLR	Outgoing Longwave Radiation
SST	Sea Surface Temperature
VLBI	Very Long Baseline Interferometry
CISK	Conditional Instability of the Second Kind
UTC	Universal Time Coordinate

tions that give rise to an increase of atmospheric angular momentum and an increase in the length of day. The swinging back of the length of day appears to be a response to the evolving intraseasonal global wind systems where transitions from high to low index circulations seem to play a major role. The length of day swings seem thus to be intimately related to the angular momentum exchanges at the lower boundary of atmosphere where the frictional and mountain torques have an important role. The atmospheric teleconnections between the tropics and the middle latitudes appear to be a major issue within the length of day cycle.

Atmospheric variability has been noted over the tropical latitudes particularly in the Indian ocean and the western Pacific ocean. Another region with pronounced variability is over the upper troposphere of the high middle latitudes *i.e.* between 50° and 70° latitudes. The tropical-middle latitude interaction on this time scale is one of the problems we address in this paper.

Atmospheric angular momentum appears to be important for the LOD consideration since it undergoes variations on many time scales. The acronyms are presented in Table 1. Strong easterly trade winds at the earth's surface contribute to a gain of westerly angular momentum from the earth. Similarly over wide belts of strong westerly winds the atmosphere loses westerly angular momentum to the earth through surface friction. In addition, pressure systems anchored near mountains contribute to pressure torques which transfer atmospheric angular momentum between the earth and the atmosphere.

Many of the atmospheric effects which are seasonal produce fluctuation in the length of the day which are of the order of a millisecond.

Most of the length of day changes associated with the annual and seasonal changes in the atmosphere are attributable to the seasonal and annual cycle of the wind systems. One also finds a strong similarity in the atmospheric angular momentum and the

length of the day in the low frequency mode of 30 to 60 days, which is quite large and is the topic of this paper.

Figure 1 from a recent study, Dickey *et al.* (1991) shows a time series of the length of day (LOD) and the atmospheric angular momentum as inferred from the NMC operational analysis. Basically we note an excellent agreement in the fluctuations of the two curves. This data covers a period between 1977 and 1988. The units along the ordinate are milliseconds for these curves. The pronounced feature is the annual cycle; however one also notes smaller amplitude oscillations that are superimposed on these. These include the intraseasonal oscillations. The bottom panel shows the difference. This figure also reveals the changes in the length of day corresponding to the El Niño time scale (*e.g.* Chao, 1989). That is apparent as a peak in the angular momentum around January 1983 when the westerly wind anomalies over the equatorial central Pacific Ocean were quite strong; also seen at around this time is a pronounced maximum in the length of day. The focus of the present study will be limited to the changes on the intraseasonal time scale.

Madden and Julian's (1971, 1972) pioneering analysis clearly demonstrated a global-scale perturbation (zonal wavenumber 1) on the time scale of around 40 days. Using sea level pressure data over the Pacific Ocean, they noted an eastward propagating equatorial wave which had the largest amplitude in the western Pacific Ocean.

A close relationship between the low frequency variability of the atmospheric angular momentum and the length of day has been noted by many authors in recent years. Fig. 2 from a recent study of Dickey *et al.* (1991) provides an excellent illustration. Here the low frequency variability of the total atmospheric angular momentum and of the length of day are both illustrated for the intraseasonal scales.

The following relation, based on Anderson and Rosen (1983), expresses a relationship between

Fig. 1. Time series of the length of the day from 1976 to 1988 as measured (top curve) and also as inferred from the atmospheric angular momentum data of NMC (middle curve) and their differences (bottom curve) all the values are in milliseconds. One year moving average has been removed from both series. (Diagram based on Dickey, J.O., *et al.*, 1991).

changes in the length of day, ΔL , and changes in atmospheric angular momentum ΔM :

$$\Delta L(\text{sec}) = 1.68 \times 10^{-29} \Delta M(\text{kg m}^2 \text{s}^{-1}) \quad (1)$$

Thus for a one millisecond change in the length of day the corresponding change in the angular momentum and a global mean zonal wind respectively are roughly of the order:

$$\Delta M = 1.68 \times 10^{-32} \text{ kg m}^2 \text{s}^{-1}, \quad (2)$$

and

$$\Delta u = 2.25 \text{ ms}^{-1}. \quad (3)$$

Noting that we are dealing with LOD changes of the order of milliseconds and corresponding zonal wind anomalies of the order of several meters/sec this turns out to be an interesting problem on the intraseasonal time scales.

2. Length of the day-observational methods

2.1 Lunar laser ranging

The development of high powered lasers around 1966 and the design of retroreflector arrays and their deployment over the lunar surface has enabled the measurement of the precise distance between the earth and various points on the lunar surface with an accuracy of a few cms. A short pulse of laser light containing approximately 10^{18} photons is fired. It illuminates a few square kilometers around retroreflectors which return sufficient energy to cause a detectable signal at the earth receiver station. This

Fig. 2. Band-pass filtered length of the day time series from 1976 to 1988 in milliseconds (a) as measured by space-geodetic techniques; (b) as inferred from the atmospheric angular momentum data of NMC after J.O. Dickey *et al.*, 1991.

gives the signal propagation time to the lunar surface and return. These observations enable one to find slight deviations from the approximated positions of earth station and lunar retroreflectors. If a sufficient number of these observations are available then accurate earth parameters including length of day can be derived.

2.2 Data used

The length of day data used in this study were from the United States Naval Observatory, Washington and were provided by Bill Benedict of the Naval Post Graduate School, Monterey, California. These data were subjected to a fourth order Butterworth filter to extract the 30 to 50 day modes in the present study.

3. Observational aspects of monsoonal low frequency modes

Observations during the FGGE year and subsequent to that for a recent 8-year period have clearly shown the presence of low frequency motions on time scales of roughly 30 to 50 days. There are several regional and global aspects of these oscillations that have been emphasized in recent literature. The following observational aspects of the low frequency motions are of interest:

Fig. 3. A Hövmoller diagram of the 200 mb velocity potential during the FGGE year. Isopleth interval $1.3 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$.

3.1 Meridionally propagating 30 to 50 day waves in the lower troposphere of the monsoon region

This is a family of trough-ridge systems that can be seen in the streamline-isotach charts of the time filtered motion field at the 850 mb surface. The passage of a trough or a ridge line over central India generally coincides with the occurrence of a wet or a dry spell respectively. The meridional scale of this system is roughly 2000 to 3000 kilometers. The speed of meridional motion is roughly 1° latitude/day. During certain years, the meridional motion and passage of these systems during the summer monsoon season is quite regular while over other years the motion has been somewhat irregular, Krishnamurti and Subrahmanyam (1982), Yasunari (1981), Mehta and Krishnamurti (1988). The reasons for this type of interannual behavior are not quite clear at the present time.

3.2 Zonally propagating planetary scale divergent circulations on this time scale

These seem to have a dominant scale of wave numbers 1 and 2. They traverse the globe, from west to east, in roughly 30 to 50 days Krishnamurti *et al.* (1985), Lorenc (1984). The largest amplitude is in the equatorial latitudes. The divergent circulations have a large meridional extent, the east-west cir-

culations can be seen even as far as 35°N and 30°S . These broad divergent circulations appear to be related to equatorial and monsoonal heat sources and sinks. Figure 3 shows an x-t diagram for 1979 of the velocity potential on the time scale of 30 to 50 days between 5°S and 5°N . This shows a striking behavior of the west to east motion. The interannual variation of these eastward propagating waves has also been studied and one finds that the propagations have been somewhat irregular during several recent years.

3.3 Air-sea interactions

We have recently examined the oceanic fluxes of sensible and latent heat on this time scale. When fluxes are calculated using the so-called 'surface similarity theory', the basic variables are the SST, the surface wind, the temperature and humidity at the top of a constant flux layer. The similarity fluxes are defined from expressions that invoke the Monin-Obukhov length and a non-linear coupling of the momentum, heat and moisture. Because of this non-linear coupling, one can diagnose the relative importance of the low frequency variations of SST, surface wind, air temperature and humidity and assess their role in the contribution to fluxes on this time scale. A detailed diagnostic study was recently completed by Krishnamurti *et al.* (1988). It was found that the latent heat flux on the time scale of 30 to 50 days can be as large as 10 to 20 watts/m^2 , which was about 5 to 10 % of the total flux over the Indian and Pacific oceans. It was also noted that wind variations on this time scale were very important contributors; next in line were the contributions from SST variations on this time scale. The variations in air temperature and humidity were relatively less important. Although the amplitude of SST anomaly was only of the order of 0.9 to 1°C on this time scale, the SST coupled with wind variations of the order 3 to 5 ms^{-1} contributed to significant latent heat fluxes, *i.e.*, ≈ 10 to 20 watts/m^2 . The sign of these low frequency fluxes are preserved for a couple of weeks, thus their role can become significant. These observations were important in the Design of the Theories presented in Section 9.

3.4 Energetics in the frequency domain

The maintenance of low frequency modes has been addressed via detailed computations of energetics in the frequency domain using daily globally analyzed data sets over many years. These studies are somewhat analogous to the estimates on energetics in the zonal wave number domain. In the latter approach, one speaks of kinetic energy exchanges from zonal flows to eddies of certain scales, and of waves to waves via nonlinear interactions. The other important interactions in the wave number domain are those from potential to kinetic energy for fixed zonal wave numbers. In the frequency domain, anal-

ogous selection rules govern the exchanges of energy. Here, one can visualize a breakdown among long term mean, low frequency and high frequency motions. In a frequency domain, the kinetic to kinetic energy exchanges can occur among long term time mean flows and other frequencies, or among triads of frequencies (analogous to the wave number domain). The potential to kinetic energy exchanges are restricted to occur at the same frequencies.

The results of these energetics, calculations performed by Sheng and Hayashi (1990a, b), show that the kinetic energy of low frequency modes on the time scale of 30 to 50 days are maintained by the following processes:

- i) They receive a substantial amount of kinetic energy from high frequency modes;
 - ii) They lose kinetic energy to the long term time mean flow;
 - iii) The annual cycle is a source of energy for other frequencies;
- and
- iv) They receive a smaller amount of energy from the potential energy on the same frequencies.

The importance of nonlinearity has also been noted by Simmons *et al.* (1983) and by Dickey *et al.* (1991).

3.5 Zonal harmonics of the zonal wind oscillations

We have examined the power spectra of the zonal wind for the 30 to 50 day oscillations in the wave number domain for the 850 and the 200 mb surfaces over the globe. Figures 4a, 4b, 4c and 4d illustrates histograms of percent variance plots as a function of zonal wave number. Basically we note that most of the variance is described by the first 6 or 7 zonal harmonics. In the tropical belt at 200 mb we note a rather pronounced variance for wave number 1 identifying perhaps the global scale monsoon variability on this time scale. In the subtropical latitudes 22°S to 45°S and at 22°N to 45°N there is a larger spread of activity through wave number 7 or 8 with a secondary peak around wave numbers 3 and 4. Low frequency variability on these scales has been noted from the 500 mb data sets by Mechoso *et al.* (1985). In the more polar latitudes between 67°S to 90°S and 67°N to 90°S, most of the variance is described by the first four zonal harmonics. Roughly 70 % of the total variance outside the tropics is described by the first three harmonics. The 850 mb zonal harmonics essentially follow the same pattern as at the 200 mb with a somewhat larger spread of energy to a higher wave number.

It should be noted that the above zonal wave number decomposition does not describe the 30 to 50 day variability for wave number 0. The amplitude for wave number 0, by far, exceeds the contributions of all other zonal wave numbers. The wave number 0

is important in the context of the length of day and the angular momentum issues. Figures 5a and 5b shows a time history of the zonal wind anomaly for wave number 0 for the FGGE year at selected latitude belts. The amplitude of the zonally averaged zonal wind for this time scale are shown for 200 and 850 mb for the several latitude belts in these diagrams. Basically the amplitude varies between 0.5 to 2 ms⁻¹. The strongest winds are found between the equator 22°N during the summer months at 200 mb where the amplitude reaches a value close to 2 ms⁻¹. At 850 mb the amplitude are around 0.5 ms⁻¹. This is the zonally averaged representation and does not describe the local monsoon anomalies which are higher in magnitude locally *i.e.* of the order of 3 to 5 ms⁻¹.

3.6 On the origin of the meridionally propagating wave-trains

In a recent study Comeaux (1991) examined this problem using the 850 mb winds over the monsoon region. An example of these winds is shown in Fig. 6. This data base is saturated with cloud motion vectors from the U.S. GOES satellite which was placed over the Indian ocean during the summer of 1979 *i.e.* the FGGE year. His analysis of the low frequency components on the time scale of 30 to 50 days revealed that these waves originated on the northern edge of the southern equatorial trough over the Indian ocean *i.e.* from roughly at 5°S. Figure 7 illustrates the climatological features of the southern equatorial trough at 850 mb. Pursuing this work further, he compared the passage of the planetary scale divergent wave at 200 mb over this region with the dates of the origin (first appearance) of the meridionally propagating waves at 850 mb. By preparing a two panel film of these two families of waves of the low frequency motions he noted that the meridionally propagating waves seem to be triggered from the southern equatorial trough as the eastward propagating divergent wave passed over it. This careful study suggests that the meridionally propagating wave is of oceanic origin. Figure 8 illustrates a latitude-time diagram of the zonal wind at 850 mb averaged across 70°E to 85°E (*i.e.* across the Indian region). This illustration clearly reveals the origin near the southern equatorial trough. Apparently the 30 to 50 day wave radiates northward and southward from this region. The northward branch has received much attention in the literature, the southward branch of 30 to 50 day wave has only been documented recently. These results seem to be fairly robust since they are based on the vast data collection shown in Fig. 6 which covers the region between 40°S and 20°N.

3.7 Vertical phase of the 30 to 50 day oscillations

Using the year long FGGE data sets Krishnamurti and Gadgil (1985) examined the phase of these os-

Fig. 4a. Histogram illustrating the percent variance for the zonal harmonics of the 850 mb flow on the time scale of 30 to 50 days. These panels include the results for four latitude belts: Equator to 22°N, 22°N to 45°N, 45°N to 67°N and 67°N to 90°N.

Fig. 4b. These are histograms that follows Fig. 4a, but cover the corresponding latitude belts of the southern hemisphere.

Fig. 4c. Histograms illustrating the percent variance for zonal harmonics of the 200 mb flow on the time scale of 30 to 50 days. These panels include the results for four latitude belts: Equator to 22°N, 22°N to 45°N, 45°N to 67°N and 67°N to 90°N.

Fig. 4d. These are histograms that follow Fig. 4c but cover the corresponding latitude belts of the southern hemisphere.

Fig. 5a. The amplitude of the zonal wave number 0 for the 850 mb wind, on the time scale of 30 to 50 days, at several latitude belts are illustrated here. The latitude belts are identified on the top of each panel.

cillations along the vertical coordinate. The zonal flow shows a remarkable change of phase between the lower and the upper troposphere in the tropical convective areas. We selected two grid points where the OLR on this time scale showed minimum values over the tropics—*i.e.* implying cloud cover. At these locations Figs. 9a and 9b we note a reversal of the

phase of the zonal wind. Over nearly all of the extra-tropics we found a barotropic structure *i.e.* no reversal of phase with height. Although, the amplitude of the oscillation increased upward with its maximum value located close to 300 and 200 mb; two typical structures are illustrated in Figs. 9c and 9d. These structures suggest the importance of convection in

Fig. 5b. The amplitude of the wave number 0 for the 200 mb wind, on the time scale of 30 to 50 days, at several latitude belts are shown here. The latitude belts are identified on the top of each panel.

the low latitudes and of the dynamics for the higher latitudes for the excitation of the 30 to 50 day oscillations.

4. Global distribution of the 30 to 50 day oscillation

The global nature of the 30 to 50 day oscillation has been recognized by many authors, Krishnamurti

and Gadgil (1985), Weickman *et al.* (1985), Dickey *et al.* (1991) and several others. Here we shall first present a horizontal and a vertical perspective of this oscillation from our past studies, Krishnamurti and Gadgil (1985). The seasonal distribution of the maximum zonal wind anomaly for the 850 mb surface is illustrated in Figs. 10a, 10b, 10c and 10d. a) Northern Winter: During the winter months, the

Fig. 6. Illustrates the typical data coverage at 850 mb on a typical day during MONEX. All of the data over the ocean, identified by the label 7, denote cloud tracked winds.

Fig. 7. Climatology of the 850 mb streamlines based on FGGE data set for July 1979. The southern equatorial trough is emphasized here as a source region for the meridionally propagating low frequency waves.

largest amplitude of the 30 to 50 day oscillations are found in the middle latitudes 30°N to 70°N . Maximum amplitude of around 6 to 7 ms^{-1} are found over the region of the Aleutian and Icelandic lows and over most of Europe. The curious fact is that in spite of these strong amplitudes for the low frequency motions, the percent variance of the motion field at these high latitudes is quite small because of the stronger high frequency variability. The amplitude of low frequency motion in the tropics is comparatively smaller (≈ 1 to 2 ms^{-1}) although the per-

cent variance is much larger *i.e.* of the order of 10 to 20 %. The monsoonal low frequency mode appears strong in the spring months. The axis of the maximum amplitude extends from the Bay of Bengal to the Western Pacific across North Malaysia and Southern Indochina. Over the Indian Ocean near 10°S , amplitudes as large as 4 ms^{-1} are found over the southern trades. By the summer season, the extent of monsoonal low frequency mode extends over the entire belt of the southwest monsoon from the Arabian Sea to the Western Pacific. The maximum

Fig. 8. A y-t diagram illustrating the origin of the meridionally propagating low frequency waves. The data sets are the 850 mb time filtered (30 to 50 day) zonal wind. Interval of isopleths 1 ms^{-1} , the shaded areas denote positive values.

Fig. 9. Pressure-time analysis of the zonal wind on the time scale of 30 to 50 days for two tropical grid points (a) and (b) and for two higher latitude grid points—(c) and (d) units ms^{-1} . Shaded area denotes negative values. The interval of analysis is shown on the top right of each panel.

amplitude in this belt is roughly 3 to 5 ms^{-1} which is roughly 20 to 25 % of the total variance of the 850 mb flow. During the northern summer, the major middle latitude activity shifts to the southern hemisphere between 50°S and 70°S . Here again we note large amplitudes of the order of 4 to 6 ms^{-1} . During the fall season, the areal extent of the strong monsoonal low frequency mode diminishes covering mostly the Bay of Bengal, Southern Indochina and the far Western Pacific where amplitude reaches intensity of 3 to 5 ms^{-1} . During this period the low frequency activity between 50°S and 70°S continues with comparable amplitudes *i.e.* 3 to 6 ms^{-1} . During most of the year, low frequency activity is

evident in the middle latitudes although the strongest amplitudes are seen over the winter hemisphere. Summarizing the above, we note that low frequency activity appears to be most pronounced in the life cycle of the summer monsoon and over the higher latitudes of the winter hemisphere.

The vertical structure of this mode is best brought out in Figs. 11a, 11b, 11c and 11d. Here the power spectral estimates of the meridional and zonal wind are plotted on latitude-height diagrams. The results for 90°E and 140°E for the entire FGGE year are shown here. The salient result here is the large power over the middle latitudes in the upper troposphere. What we saw at 850 mb in the horizontal

Fig. 10. Maximum zonal wind at 850 mb of frequency-filtered motions during individual seasons (units ms^{-1}) (a) winter, (b) spring, (c) summer, (d) fall.

Fig. 11. Vertical distribution of power spectra (power times frequency) analyzed as function of latitude and pressure for the following variables: a) v component at 140°W , b) u component at 140°W , c) v component at 90°E , d) u component at 90°E .

distribution at the latitude belts 50° to 70° appears to amplify with height. The power spectra over the monsoon region 90°E near 10°N exhibits a maximum value near 10°N for the zonal wind.

The large latitudinal separation of the tropical (monsoonal) and the middle latitude low frequency components is not well understood at the present time. Given the tropical easterlies and middle latitude westerlies, the tropical-middle latitude interaction problem, one is faced with the issue of wave energy flux across the critical latitudes. According to Webster and Zhang (1989) longitudinal asymmetry of the zonal flows permits an episodic exchange between the tropics and middle latitudes in certain locations such as the Eastern Pacific Ocean. This problem has mostly been posed as a linear problem where the mean flow $\bar{u}$ is time invariant. We have noted in Sections 3.5 that wave number zero (*i.e.* $\bar{u}$) has most of the variance, among the different wave numbers on the time scale of 30 to 50 days. Thus an understanding of the 30 to 50 wave would not be possible under the assumption of $\bar{u} = \text{constant}$, *i.e.* the assumption of linearity for the wave energy flux and its convergence most perhaps invoke non-linear dynamics. In Section 7 we pose this nonlinear problem in the frequency domain to reassess the tropical-middle interactions.

5. Outgoing longwave radiation, super cloud cluster and zonal flow accelerations

In this section we shall examine the above features in reference to the length of day.

5.1 Outgoing longwave radiation

The low frequency variability of tropical cloud cover (as seen from the OLR distribution) is composited with respect to the selected nodes in the length of day illustrated in Figs. 12a, 12b, 12c and 12d. These panels respectively show the OLR (unshaded area denotes negative OLR values in watts/m^2). This area is supposed to tag regions of high clouds over convective regions in the near equatorial latitudes. The composite cloud cover shows a maximum near 85°E during the minimum length of day. As the length of day increases, the near equatorial zonally oriented axis of cloud cover moves meridionally as well as zonally eastward. During the period of the maxima of the length of day the cloud belt is located along 15°N and illustrates an active monsoon spell. The start of active and inactive spells of the monsoon respectively are apparent in the OLR distributions and these features were noted by Yasunari (1981).

Panels a, b, c and d of Fig. 12 correspond to June 6, June 14, June 23 and July 1 for the year 1979. The strongest westerly zonal flows were noted over the monsoon region, and they appear to be closely coupled to the deep convective activity, over this region a subsequent weakening of near equatorial zonal flows seem to be a response to the meridional motion of cloudy areas. The cloudy areas also exhibit a zonal motion on the time scale of 30 to 50 days. The zonal flow anomalies that follow the eastward propagating cloud clusters are somewhat weaker in amplitude compared to those that follow the meridionally propagating cloud clusters of the monsoon.

Fig. 12. Time filtered outgoing longwave radiation during the length of day cycle. a) June 6, 1979; b) June 14, 1979; c) June 23, 1979 and d) July 1, 1979. Interval of isopleths 10 watts/m²; shaded area denotes positive values.

Fig. 13. Time-longitude section of T_{BB} index (I_{TBB}) integrated between the equator and $5^{\circ}N$ obtained from the 3 hourly GMS IR data from 29 May 00Z to 10 July 21Z, 1980. The letters A, B, C, D denote four of the major super clusters. Contour interval is 10, and shading denotes the region where values are greater than 20.

5.2 Super cloud clusters

Organized near equatorial super cloud clusters have been noted to move eastward at a speed of roughly 10° longitude per day, Lau and Chan (1985). Within these clusters one finds smaller cloud systems associated with tropical disturbances that frequently move westwards at a speed of roughly 5° longitude per day. Figure 13 from a study of Nakazawa (1988) and Lau *et al.* (1989) presents an example of cloud motions on two of these different space-time scales. During episodes of passage of these cloud clusters the near equatorial flows exhibit an enhancement of westerly flows. These have been labeled westerly wind bursts by Nakazawa (1988), Lau *et al.* (1989) and several others. It is recognized that these westerly wind bursts associated with tropical convection contribute to an increase of the angular momentum of the atmospheric flows requiring a slowing down of the earth for the conservation of the

Fig. 14. Time filtered (30 to 50 day) variability during northern summer. a) outgoing longwave radiation, (watts/m^2), b) 850 mb zonal flow over a monsoon domain, units ms^{-1} , and c) length of day, units milliseconds.

total angular momentum. These observations have been noted by numerous authors, Nakazawa (1988).

5.3 Zonal flow accelerations

An example of zonal flow acceleration that follows this evolution of near equatorial convection is shown in Figs. 14a, 14b and 14c. Here the three panels a, b and c correspond to the dates of the minimum

Fig. 15. Time filtered (30–50 days) 850 mb flows during the ascending mode of the length of day cycle. Streamlines (thin solid lines) and isotachs (heavy solid lines) interval 1 ms⁻¹. Heavy west-east oriented line shows a cyclonic circulation of the low frequency mode.

length of day (June 6, 1979), average length of day (June 14) and the maximum length of day (June 23). These dates correspond to the low frequency variation in the length of day and a strong parallel monsoon flow (*i.e.* west to east) at the time of the maximum length of day. Figures 15a, 15b and 15c illustrate the 850 mb low frequency component of the wind field following Krishnamurti and Subramanian (1982). Basically we see a strong counter monsoon

flow at the time of the minimum length of day and a strong parallel monsoon flow (*i.e.* west to east) at the time of the maximum length of day. The zonal flow accelerations follow the evolution of the tropical convection. The amplitude of the isotachs of the low frequency wind is around 3 to 4 ms⁻¹. The heavy dark line identifies an axis of counter-clockwise flow. Since the total atmospheric angular momentum reaches a maximum as the length of day

reaches a maximum, the monsoonal zonal flows appear to exhibit a strong relationship with the total atmospheric angular momentum as well.

Figures 14a, 14b and 14c also includes the low frequency relationship between the outgoing long wave radiation, (OLR) the monsoonal zonal flow and the length of day. Here the zonal flows are averaged over a domain 60°E to 75°E, 5°S to 15°N; the OLR is averaged over a domain 70°E to 90°E and 5°S to 5°N.

A closer look at the phase of the three curves suggests that the OLR and the zonal flow variations follow each other closely, however the length of day lags by a few days. This suggests that the low frequency variation in the length of day which is generally considered to be of atmospheric origin is related to the near equatorial convection, the resulting changes in zonal flows and the change in the atmospheric angular momentum.

5.4 The Pacific trade winds

Thus far we have not addressed the role of the trade winds of the Pacific ocean in this problem. East-west divergent circulations (also called in Walker circulations) follow the easterly trades in the Pacific ocean. In response to the near equatorial convection, the trades and these divergent circulations also exhibit zonal flow accelerations. These easterlies flows in the tropics have a major role in transferring westerly angular momentum from the earth to the atmosphere during the ascending node of the length of day cycle. During this transition from the minimum to the maximum length of day the surface easterlies of the Pacific ocean respond to the organized near equatorial tropical convection. The transfer of westerly angular momentum to the atmosphere, in these surface easterlies, occur via frictional torques. The westerly momentum transfer manifests itself in the strengthening of the overall westerlies and in the increase of the global atmospheric angular momentum. The interaction of the near equatorial easterlies with the ground is thus an important component of this story. A less well understood aspect of the momentum distribution with atmosphere is the vertical flux by cumulus convection.

6. Wave energy flux in the frequency domain

The meridional flux of wave energy is usually cast as a linear problem. The expression for the wave energy flux is obtained from the equation of motion, mass continuity and the thermal equations. Basically this is the meridional analog of the Eliassen and Palm's (1961) work on the vertical wave energy flux problem. The equation for the meridional flux of wave energy is usually expressed by:

$$\overline{\phi v} = -(\bar{U} - C) \left[\overline{uv} - \frac{1}{\sigma_0} \frac{\partial \bar{U}}{\partial p} v \phi_p \right] \quad (4)$$

Here the basic zonal flow $\bar{U}$ is independent of time following the assumption of linearity. The wave energy flux $\overline{\phi v}$ contains a covariance of the Doppler shifted near zonal $(\bar{U} - C)$ and the meridional eddy fluxes of momentum and heat; here C denotes the phase speed for a particular wave. This is a beta plane version of the equation where the basic state zonal flow is a function of y and p only and is independent of x . The basic state also includes specific volume and geopotential on the y, p plane. Furthermore the basic state is in geostrophic balance with respect to the basic state geopotential field and the system is quasistatic. The above equation is obtained from the linear perturbation analysis. Somewhere near the transition zone between the tropical easterlies and the middle latitude Westerlies, i.e. in the vicinity of $\bar{U} = 0$, the Doppler shifted velocity $\bar{U} - C$ also vanishes. This is the critical latitude where the wave energy flux (according to the above linear theory) vanishes. What is evidently wrong here is that $\bar{U}$, based on observations, is not a constant but exhibits rather pronounced intraseasonal variability. The global oscillation of the zonal flow was discussed in Section 4. As stated there, in the wave number domain, the most pronounced intraseasonal oscillations are found for wave number zero. Thus for studies on the wave energy flux around this time scale the assumptions of linearity would be invalid. Here we are interested in deriving an equation for the wave energy flux for the nonlinear problem that permits such variations in $\bar{U}$. We shall cast such an equation in the frequency domain to ask whether the critical latitude permits wave energy fluxes across it in selected frequency bands. We shall next address the nonlinear framework for this problem.

The nonlinear form of the wave energy flux may be obtained from the following equations:

Equation of Motion:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{R \cos \theta \partial \lambda} + v \frac{\partial u}{R \partial \theta} + \omega \frac{\partial u}{\partial p} - \frac{uv \tan \theta}{R} - fv + \frac{\partial \phi}{R \cos \theta \partial \lambda} = 0 \quad (5)$$

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{R \cos \theta \partial \lambda} + v \frac{\partial v}{R \partial \theta} + \omega \frac{\partial v}{\partial p} - \frac{uv \tan \theta}{R} + fu + \frac{\partial \phi}{R \partial \theta} = 0 \quad (6)$$

Mass Continuity Equation:

$$\frac{\partial u}{R \cos \theta \partial \lambda} + \frac{\partial v \cos \theta}{R \cos \theta \partial \theta} + \frac{\partial \omega}{\partial p} = 0 \quad (7)$$

First Law of Thermodynamics:

$$\frac{\partial \phi}{\partial t} + \frac{\partial \phi}{\partial p} + u \frac{\partial \phi}{R \cos \theta \partial \lambda} + v \frac{\partial \phi}{R \partial \theta} + \sigma_0 \omega = 0 \quad (8)$$

Fig. 16. Wave energy flux on the time scale of 30 to 50 days as a function of latitude and days during the FGGE years. Interval of isopleths 0.5×10^8 Joule. s^{-3} .

Here R denotes the radius of the earth, θ is the latitude, ϕ is the geopotential and σ_0 is a dry static stability parameter $= -\frac{\alpha}{\theta^*} \frac{\partial \theta^*}{\partial p}$, where α is the specific volume and θ^* denotes the potential temperature. After some detailed algebraic manipulations of the zonal equation of motion and the first law requiring an elimination of the vertical velocity w , one can obtain the following expression for the wave energy flux.

$$\overline{\phi v} = \left[\frac{1}{f_0 - \bar{\Gamma}} \right] \left[\overline{\phi \beta} + \left[\frac{u^2}{2} - \frac{1}{\sigma_0} \frac{\partial \phi}{\partial p} \frac{\partial u^2/2}{\partial p} \right] \gamma \right] \quad (9)$$

where

$$f_0 = 2\Omega \sin \theta_0 \quad (10)$$

$$\bar{\Gamma} = \frac{\tan \theta_0}{R} u - \frac{\partial u}{R \partial \theta} - \frac{1}{\sigma_0} \frac{\partial}{R \partial \theta} \left\{ \left[\frac{\partial \phi}{\partial p} \right] \frac{\partial u}{R \partial \theta} \right\} \quad (11)$$

$$\beta = \frac{\partial u}{\partial t} - \frac{1}{\sigma_0} \frac{\partial}{\partial p} \left[\frac{\partial \phi}{\partial p} \right] \frac{\partial u}{\partial p} + \frac{\partial}{R \cos \theta \partial \lambda} \left[\frac{\partial u^2/2}{\partial p} \right] \frac{\partial \phi}{\partial p} \quad (12)$$

$$\gamma = \left[u \frac{\tan \theta}{R} - \frac{\partial u}{R \partial \theta} + f \right] v + \frac{v}{\sigma_0} \frac{\partial}{R \partial \theta} \left[\frac{\partial \phi}{\partial p} \right] \frac{\partial u}{R \partial \theta} \quad (13)$$

The computational procedure for the estimation of $(\phi v)_n$ where n is the frequency follows the method proposed by Hayashi (1980). Ideally the problem of wave energy flux should be examined in the wave number-frequency domain. For the present we are presenting only a breakdown in the frequency domain.

We shall next show an illustration on the wave energy flux. Figure 16 shows the total contribution for the 30 to 50 day time scale on a latitude-time diagram. The daily values shown here are averaged around the globe and from the bottom to the top of the atmosphere and shown for the entire FGGE year. This zonal-time section of the wave energy

flux is particularly interesting in the northern summer monsoon months, *i.e.* May through September, where one can clearly see meridional wave energy flux emanating from the near equatorial latitude to almost 60°N. The critical latitude separating the tropical easterlies from the middle latitude westerlies is located somewhere near 20°N during the northern summer. Thus it appears that in the frequency domain the wave energy passes through this region. Our interest in this problem comes from the observations of the large variance of the low frequency modes in the middle latitude upper troposphere, these were discussed in Section (5) and Fig. 11. During the summer months the slope of the shaded and the unshaded lines clearly shows a positive slope in the y, t diagrams. This slope implies that information (*i.e.* wave energy) travels from the tropics to roughly 60°N. Around that latitude the wave energy flux convergence appears to be quite large since the sloping lines appear to terminate near 60°N. The alternation in sign suggests that positive and negative wave energy fluxes on the intraseasonal time scale emanate from the tropics to account for the middle latitude variability. We confirmed further by plotting the time variability of the 300 mb low frequency component of the zonal wind at 60°N and compared that with the convergence of wave energy flux at that latitude. This reference point was selected from the diagram illustrating the maxima of the middle latitude power spectra *i.e.* Fig. 11. We noted a very strong coherence in the phase of the oscillations of the middle latitude zonal wind and the wave energy flux for the summer months. This suggests that the middle latitude variability is in part connected to the wave energy fluxes from the tropics. It appears from the calculation presented here that the wave energy from the summer monsoon latitudes can easily make it to these high latitudes when the fluxes are viewed in the frequency domain. The strong connection to the summer monsoon is revealed by the large amplitude of this wave train during the months when the monsoon is most active.

We have also looked at the wave energy fluxes

Fig. 17. Difference in zonal wind (ms^{-1}) between a minimum LOD and the preceding maximum LOD at 300 mb. a) January 11 to January 29, 1979 b) July 25 to August 8, 1979.

at higher frequencies *i.e.* on the time scale of 1 to 10 days. This produces a busy y-t diagram since there are many alternations of sign in 365 days. We shall not present that diagram here except to state that it contains several instances of the southward propagation of wave energy flux from the middle latitudes to the tropics. There are however a few occasions when the reverse also occurs *i.e.* tropical wave energy propagation is directed poleward. The latter seems more episodic.

The overall maintenance of the middle latitude upper tropospheric low frequency component may partly be attributed to the tropical events, as shown here. It is also possible that part of such an accumulation of energy can occur via vertical flux convergence (related to the angular momentum problem) in the middle latitudes. The role of the mountain and frictional torques in the low frequency component of the angular momentum budget are discussed in Section 8.

7. On the transition from the maximum to the minimum length of day

Here we shall illustrate two examples during the transition for the maximum to the minimum length of day. With the reduction in the angular momentum during the transition, we expect to see a weakening of the zonal flows in the atmosphere *i.e.* a possible change from a high to a low index situation. We had noted that during this transition the largest changes in the zonal flows are found in the upper troposphere of the middle latitudes. The two cases we selected cover the following periods:

- a) (Northern winter) maximum length of day on: January 11, 1979. Minimum length of day on: January 29, 1979.

- b) (Southern winter) maximum length of day on: July 25, 1979. Minimum length of day on: August 8, 1979.

First we calculated a difference in the zonal flows between the date of the minimum and the maximum length of day at the 300 mb surface. Figures 17a and 17b illustrates these zonal flow differences from the maximum to the minimum length of day. The northern winter 300 mb differences show a belt of positive differences with maximum values as large as 60 ms^{-1} . That shows a large collapse of the zonal flow during the transition. The corresponding 300 mb geopotential height charts are shown in Fig. 18. They illustrate a transition from a high to a low index situation culminating in a major Omega-block over the northern Atlantic Ocean. It was also curious to note that the previous length of day cycle from the maximum to the minimum length of day occurred between December 15, 1978 to December 30, 1978 where similar features were also noted. These were described as the period of two of the major blocking events of the FGGE winter, Kung and Baker (1986). Figure 19 from their study provides a Hovmöller diagram of the ridging events of that period. It is interesting to note that these two major blocks (shown by heavy horizontal lines) were separated by a period close to that of the intraseasonal oscillation. These two blocking events lasted for roughly two weeks covering almost the entire period between the maximum and the minimum length of day. Backtracking these respective individual cycles to the period between the minimum to the maximum length of day we were able to identify maximum convection and zonal flow acceleration in the region of the winter monsoon. Those were the onset and next active spell during the winter mon-

Fig. 18a. A sequence of 300 mb contour charts during the LOD cycle covering the period from January 11, 1979 to January 19, 1979. Maps are shown at interval of every 2 days.

soon of 1978–1979. The westerly bursts of the north-east monsoon have been addressed by numerous authors. The dates of those westerly bursts correspond roughly to the dates of maxima in the length of day. Figure 20a shows a curve of the 850 mb zonal flow (averaged from 110E to 150E, 5N to 20N) during the period December 1978 through February 1979. The westerly bursts can be seen around January 11 and February 14th, they correspond to the two spells of the winter monsoon. We also present two curves showing time filtered zonal flows on Fig. 20b and the length of day is presented in Fig. 21. The zonal flows covers the period December 1, 1978 through February 28, 1979 and the length of day curves illustrates two of the cycles between January 1 though February 28, 1979. During the ascending node of the length of day cycle the winter monsoon activ-

ity increases and decreases during the descending node. This tropical-middle latitude teleconnection appears to be linked to the angular momentum cycle of the earth and atmosphere. A sequence of 300 mb charts for the northern summer case is presented in Fig. 22. This sequence also goes through a major change in the zonal flow and the zonal index during the transition from the maximum to the minimum of the length of day. In this instance a major blocking event is seen south of Australia around August 8. This is not surprising since changes in the atmospheric angular momentum follows the length of day cycle. It is interesting that the mountain and frictional torques provide a major drain of atmospheric westerly angular momentum to the earth during this transition. This entire scenario appears to start with the tropical convection as the length of day increases

Fig. 18b. Same as Fig. 18a for the period January 21 to January 29, 1979.

from a minimum value and culminates in a low index cycle over the upper troposphere of the middle latitudes.

7.1 Vertical cross-section of the zonal wind anomaly

As one proceeds from a maximum length of day to a minimum length of day the atmospheric angular momentum decreases and equivalent zonal wind anomalies should appear somewhere in the atmosphere. We have noted such anomalies over the middle and polar latitudes in the upper troposphere near the 300 mb. Figure 23a shows a vertical cross section of such a composite for the northern summer season. Here we show the difference in the zonal wind of the 30 to 50 day oscillation for the maximum to a minimum length of day composited over many cycles of the length of day. The largest changes in the zonal wind (for wave number 0) occur poleward

of 45° latitude during this period. The amplitude of this change is around 2.8 ms^{-1} at 300 mb *i.e.* easterlies build up as the atmospheric angular momentum decreases. This is a composite for the northern summer when we find the strongest response in the southern hemisphere. The maximum easterly wind response over the northern hemisphere is around 1.8 ms^{-1} near 55°N. This seasonal low frequency activity is consistent with the results shown in Section 5. A stronger westerly wind anomaly of 2.5 ms^{-1} near 80°N occupies a smaller area near the north polar cap. The decrease of atmospheric angular momentum between the maximum and minimum length of day is largely accounted for by the build up of easterlies near of 67°S and 55°N. The latitudinal plot of the 300 mb zonal wind change is shown in Fig. 23b. This illustration portrays that these large wind speed changes seem to occur in the highest middle

Fig. 19. Longitude-time diagram of significant ridges during the FGGE winter as identified in the daily 0000 GMT 500 mb charts. Bold lines are for cases of observed blockings.

Fig. 20. The top panel (a) shows the amplitude of trade wind easterlies in the monsoon region during the winter of 1979. The bottom panel (b) shows the time filtered zonal wind for panel (a), units ms^{-1} .

Fig. 21. Typical intraseasonal length of day cycle during the winter of 1979.

latitudes. We have also constructed vertical cross-section of the zonal wind anomaly for the northern winter season and noted similar structures primarily over the upper troposphere of the northern hemisphere. These are not shown here since these results are clearly evident from the northern summer maps leading to the blocking event.

8. Angular momentum

Using the year long FGGE data sets we have calculated the following components of the angular momentum:

- mountain torques
 - frictional torques
 - meridional flux of total angular momentum
- and
- vertical flux of total angular momentum

Our interest is in their contribution to the low frequency variability. The calculations in the frequency domain require that special care be exercised in handling of the nonlinear terms. Strategies for the computations of products of two or more time series of variables have been developed by Hayashi (1980), Krishnamurti *et al.* (1988) and several others.

8.1 Review of earlier work on angular momentum

In a classical study on the flow of angular momentum in the atmosphere, Widger (1948) studied the generation and transport of angular momentum. Equations for the rate of change of momentum in terms of mountain, friction torques and momentum transport were used in his study.

Using data sets for sea level (1230 UTC) and 500 mb (0400 UTC) from Northern Hemisphere Histor-

Fig. 22. A sequence of 300 mb charts during the LOD cycle covering the period from July 25, 1979 to August 8, 1979. Maps are shown at interval of every 2 days.

Fig. 23a. Difference of the zonally averaged zonal velocity between a minimum LOD and the preceding LOD during the northern summer. Results are composited over several cycles of LOD. Units $\pm 0.5 \text{ ms}^{-1}$.

Fig. 23b. This illustrates a meridional profile of the difference of zonal velocity; illustrated in Fig. 23a, for 300 mb units ms^{-1} .

ical Weather Maps for January 1946 and at 700 mb (0400 GCT) level from the analyzed charts of U.S. Air Force, he evaluated geostrophic winds for a 5 degree latitude long grid. Budget of various terms were calculated for the region 10N to 80N and below the 7.5 km level for the entire month of January 1946. Some of his results were (in units of $10^{25} \text{ kg M}^2 \text{ Sec}^{-1}$) the following:

- (a) Net northward transport of relative angular momentum, at 35°N for the month.

$$R \cos \phi \iint \rho v dxdz = 3.16$$

- (b) Net change of relative angular momentum for latitude belt 25

$$\frac{\partial}{\partial t} \iiint \rho R u \cos \phi dxdydz = 0.72 \quad (14)$$

- (c) Net transport of omega angular momentum at 10°N for the month.

$$\omega R^2 \cos^2 \phi \iint \rho v dxdz = 2.21 \quad (15)$$

- (d) Surface frictional torque for the latitude belts, integrated effect for the month.

$$\iint \kappa \rho u (u^2 + v^2)^{1/2} R^3 \cos^2 \phi d\phi dx$$

for $10^\circ - 15^\circ\text{N} = -10.2$
for $40^\circ - 45^\circ\text{N} = 1.8 \quad (16)$

- (e) Mountain torque for $40^\circ - 45^\circ\text{N}$ latitude belt, integrated effect for the month.

$$\int p r d\sigma = 1.55 \quad (17)$$

where R is the radius of the earth and other symbols have usual measuring.

The conclusions of this study were that the mountain and the frictional torques were of the same order as those for the earth's and the uv transport terms.

Newton (1971a, b) synthesized observations and the then available information in the global angular

momentum balance. He provided estimates for the mountain torque and the atmospheric fluxes. His study arrives at providing a consistent picture of the annual (and the seasonal) cycle of the angular momentum. He notes that the frictional source of momentum in the southern hemisphere is on the annual average almost twice as large as in the southern hemisphere. The largest frictional contribution occurs in the winter season. He also notes that in the annual mean the mountain torque parallels the frictional torque in sign *i.e.*, they vary in the same sense in the two hemispheres. This was also alluded to by Peixoto and Oort (1984) in a more recent study. Both of these studies refer to the difficulties in obtaining reliable estimates of the frictional and the mountain torques.

The problems on the former stem from the use of neutral stability drag coefficients and the problems of the latter are largely due to the formulation of the pressure at the ground surface or the difference of this large terms in pressure gradient force in sigma coordinates.

Some 15 years later Peixoto and Oort (1984) re-examined this problem in considerably more detail. Using 15 years of a more complete data set over the entire globe they evaluated the global budgets. In their study the horizontal and vertical transports were roughly 50 % larger than Widger's estimates. They used Newton's estimates of mountain torques and evaluated frictional torques using the Bulk aerodynamic approach. The stronger drain of westerly angular momentum at around $\pm 40^\circ$ latitude and the net gain in the tropics (except for the annual cycle of the monsoon *i.e.*, the summer monsoon) were the salient features. Their estimates of the annual mean mountain torques show a substantial drain of westerly momentum around 40°N . The effect of the Andes was comparatively smaller in the latitude belt 40° to 60°S . Their horizontal and vertical transport terms were of comparable magnitude. Their annual transport of momentum was illustrated from the isopleths of a transport streamfunction. This is an interesting and useful climatological diagram. In the context our study is focused on the 30 to 50 day oscillation of the momentum transport streamfunction. With the variation in the length of day and in the atmospheric angular momentum we can expect some interesting fluctuations of this diagram on the intraseasonal time scales which can be revealing on the tropical-extratropical teleconnections, this we shall present in Section 9.6.

Following a somewhat different method Sakellariades (1989) calculated what he calls Atmospheric Effective Angular Momentum (AEAM) functions. These are robust functions that do not suffer from the computational problems facing the evaluation of the pressure gradient force *i.e.* the small difference of two large terms. The excitation functions provide

changes in the atmospheric angular momentum by invoking the dynamics of earth's rotation. By keeping track of these functions based on ECMWF analysis and from the ECMWF forecasts an assessment of the model's ability to handle the atmospheric angular momentum changes was made. The angular momentum function was calculated from ensembles of days 0, 5 and 10 day forecasts. The results of these calculations showed that the model's results at day 10 of forecasts, especially over the southern hemisphere were quite degraded.

The tropical angular momentum transfer cycle has been successfully described by Madden (1987). Given an eastward propagating tropical super cloud cluster, the Kelvin wave response to its east is described by a vertical east-west circulation cell *i.e.* the Walker circulation. To its west is a Rossby-Gravity wave response *i.e.* another vertical circulation cell. The ascending branch of the two cells lie within the super cloud cluster. The surface winds to the west are usually labelled as the westerly monsoon circulations. The surface winds to the east are the easterly trades.

Madden has studied the contribution to the length of day cycle to the surface frictional torques. During the ascending mode as the length of day increases the slowing down in the earth's rotation is attributed to a gain of westerly angular momentum over the regions of the easterly trades. He calculated the oceanic frictional torques using the Bulk Aerodynamic drag laws. He was able to demonstrate a full cycle of the passage of the planetary scale divergent cell east of the super cloud cluster. The torques show the importance of the drag exerted by the trade winds on the length of day. Figure 24 illustrates a schematic cloud passage with the vertical circulations from the study of Madden (1987). This also includes the estimates of surface torques from his study, which shows an interesting intraseasonal oscillation in its magnitude over the Pacific ocean.

8.2 Computational methodology for the mountain torque

The mountain torque which contributes to the east-west angular momentum transfer is the net east-west component of pressure force across the opposite surfaces of the mountains. It is obvious that the actual pressure values at the ground surface over the mountains is required to be known. In the height and pressure coordinate systems in the vertical, the earth surface is not a coordinate surface, therefore, data are to be specially interpolated at the earth surface. In such coordinate systems the data may also be artificially generated below the ground surface. It is for this reason that sigma, the normalized pressure coordinate system has been adopted for use. In the sigma coordinate system the lowest surface with sigma equal to 1.0 coincides with the earth's surface.

Fig. 24. Estimated $\bar{\tau}_x$ bottom. Time-mean circulation cells are taken from Streten and Zillman [1984, p. 356]. Regions of large scale convection are indicated schematically over Africa, the Indonesian region, and South America. The upper eight panels show τ_x for phases of Fig. 1. Anomalies in the mean circulation associated with the 40–50 day oscillation are taken from Madden (1987). Units are N m^{-2} .

Since for the computation of the pressure torques the pressure values are required at the ground surface, the data in the sigma coordinate system are therefore, the most suitable for the calculation of pressure torque.

In the sigma coordinate system the pressure term takes the form

$$\nabla_{\sigma} \phi^* - \frac{\sigma}{p_s} \frac{\partial \phi^*}{\partial \sigma} \nabla p_s \quad (18)$$

It is also convenient to have data in a staggered manner particularly in the vertical, where temperatures and vertical velocity refer to an intermediate level in the vertical. Therefore, an averaging process is used to get the data suitably configured at the appropriate place for the calculation of the pressure torque.

The mountain torque as used by Swinbank is expressed as

$$\left(\frac{R^2}{g} \right) \cos \phi \Delta \phi \sum_i p_s \sum_k \left(\delta_{\lambda} \bar{\phi}^{*\lambda} + \overline{RT^{\lambda} \delta_{\lambda} \ln p_s^{\lambda}} \right)_i \Delta \sigma_i \quad (18a)$$

Here ' ϕ ' is the latitude, ' ϕ^* ' the geopotential ' R ' denotes the radius of the earth; $\Delta \phi$ is the width of the latitude belt used for calculations. Subscript ' s ' refers to surface data. The $\Delta \sigma$ refers to vertical spacing of sigma levels. The operators

$$(\delta_{\lambda} x)_i = [x_{i+1/2} - x_{i-1/2}]; \quad \bar{x}_i^{\lambda} = [x_{i+1/2} + x_{i-1/2}] / 2 \quad (19)$$

and $\sum_i$ denotes sum over all points in the latitude belt and $\sum_k$ a sum over all model layers.

In the present study the data from a 12 level T42 global spectral model has been used. The expression for the mountain term is used exactly in the form as given in Eq. (18a) above. Following Swinbank (1985) the convention used in this paper is that positive mountain and friction torque refer to angular momentum transfer to the earth.

8.3 Formulation of frictional torques

The formulation of the frictional torque is based entirely on the surface flux estimates from the FSU global spectral model (T42) following Pasch (1983). The daily computation of these torques was carried out over the entire globe.

8.4 Horizontal and vertical fluxes of atmospheric angular momentum

Absolute angular momentum at any point in the atmosphere is expressed by the sum of the omega and relative angular momentum,

$$\text{i.e. } M = \Omega R^2 \cos^2 \phi + u R \cos \phi \quad (20)$$

the flux across a vertical wall following a latitude circle is given by

$$\iint \left[-\frac{R^3}{g} \cos^3 \phi p_s v \Omega \right] d\lambda d\sigma + \iint \left[-\frac{R^2}{g} \cos^2 \phi p_s u v \right] d\lambda d\sigma \quad (21)$$

where λ is the longitudinal and σ is the vertical coordinate. Other symbols have the usual meaning. These terms are calculated for each one of the grid points and then averaged along latitudes, to finally get meridional profile of fluxes.

The vertical flux of the absolute angular momentum is expressed by

$$\int_{\phi_1}^{\phi_2} \oint \Omega \frac{R^4}{g} \cos^3 \phi \omega d\lambda d\phi + \int_{\phi_1}^{\phi_2} \oint \frac{R^3}{g} \cos^2 \phi \omega u d\lambda d\phi \quad (22)$$

It should be noted that although the vertical velocity ω appears here, these calculations are performed on the sigma surfaces.

8.5 Results of computations

Following the intraseasonal variability of the length of day a number of interesting questions on the role of the mountains, friction and transports can be asked with respect to the atmospheric phenomenology. We shall first present some of the diagrams showing the contribution of these terms of the angular momentum equation during a typical cycle of the length of day. First we shall examine the meridional variability of the zonally averaged total torques. This will be followed by estimates on the time scale of 30 to 50 days. The latter are calculated in the frequency domain essentially following methods shown in Krishnamurti *et al.* (1988). We have also examined the temporal variability of these intraseasonal fluxes. The horizontal and vertical fluxes are examined via a transport streamfunction in the meridional-vertical plane. We shall discuss these results in sequence.

(a) Total frictional torques as a function of latitude

The meridional profiles of the frictional torque for minimum and the maximum of the LOD *i.e.*, for May 1 and 17, 1979 respectively are in Fig. 25a. Similarly for the following cycle of the minimum and the maximum of the LOD *i.e.* for June 6 and 23, 1979 respectively are given in Fig. 25b. The following features of the frictional torque profiles are evident in these illustrations.

Atmospheric angular momentum is transferred to the earth by the surface frictional torque in the southern hemisphere roughly between 30° to 60°S. The maximum amount of angular momentum transfer being $5 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ for every 2.8 deg latitude belt, these are some of the largest values *i.e.* over the belt of the strongest surface winds.

During the maximum of the LOD, the atmosphere loses more angular momentum than during the minimum of LOD as the difference of friction torque is $1-2 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ per 2.8 deg latitude belt. This feature is more evident in the southern hemisphere where the change appears to be larger.

Fig. 25. Latitudinal profile of total mountain torque for the a) maximum LOD, May 17, 1979 and minimum LOD, May 1, 1979. b) maximum LOD, June 23, 1979 and minimum LOD, June 6, 1979. Units $10^{18} \text{ kg m}^2 \text{ s}^{-2}$.

Over the tropical region, on both sides of the equator, the atmosphere gains angular momentum as value of friction torques is about $-1 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ for every 2.8 deg latitude belt.

In the middle latitudes of the northern hemisphere the westerlies lose angular momentum to earth due to frictional torque at the rate of about $1.5 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ for every 2.8 degrees belt.

There is a net global transfer of the atmospheric angular momentum to the solid earth by the frictional torques. The results shown here are consistent with the surface wind variability.

(b) Total mountain torque as a function of latitude

The meridional profiles of the mountain torque for the minimum and the maximum of the LOD on May 1 and 17, 1979 respectively are presented in Fig. 26a. The results for the next cycle of the minimum and maximum of the LOD *i.e.* for June 6 and 23, 1979, are presented in Fig. 26b.

In the southern hemisphere, the mountain torques are generally small as was noted by Newton (1971a, b). Atmosphere gains angular momentum in southern tropics and loses in the northern tropics at the rate of about $2 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ for 2.8 degree latitude belts. In the northern hemisphere, from about 15°N to about 55°N, large transfers of angular momentum to the atmosphere by the mountains are noted. The amounts of these range from about $8 \times 10^{18} \text{ kg m}^2 \text{ sec}^{-2}$ for a 2.8 degrees latitude belt. The variations of mountain torques are less marked

Fig. 26. Latitudinal profile of total mountain torque for the a) maximum LOD, May 17, 1979 and minimum LOD, May 1, 1979. b) maximum LOD, June 23, 1979 and minimum LOD, June 6, 1979. Units $10^{18} \text{ kg m}^2 \text{ s}^{-2}$.

during the ascending and descending nodes of the LOD cycle.

During this period of May 1 to end of August 1979 there is a net gain of the angular momentum by the atmosphere due to mountain torques. The mountain torque is generally larger than the frictional torques. The mountains continually transfer angular momentum to the atmosphere, the atmospheric relative angular momentum remains positive. The frictional torques have a major role in the maintenance of the atmospheric and the earth's angular momentum balance.

(c) *Latitudinal variation of low frequency variability of the frictional torque*

In Fig. 27a three curves are provided. They respectively denote the meridional profiles of frictional torques for the minimum length of day on May 1, the average length of the day on May 9, and the maximum length of the day on May 17, 1979 *i.e.* during the ascending node. Similarly the meridional profiles of the low frequency variability of the frictional

Fig. 27. Latitudinal profile of the low frequency component of frictional torque a) minimum of LOD May 1, 1979 and maximum of LOD May 17, 1979. Average of LOD May 9, 1979 b) maximum of LOD May 17, 1979 and minimum of LOD June 6, 1979. Average of LOD May 27, 1979. Units $\times 10^{18} \text{ kg M}^2 \text{ Sec}^{-2}$.

torques for the descending node of the LOD are presented on Fig. 27b. These curves refer to May 17, May 27 and June 6, 1979 respectively. They denote the torques for the maximum of the length of the day, the average length of the day and for the minimum of the length of the day.

In Fig. 27 we note that during the minimum of the LOD *i.e.* on May 1, 1979 a transfer of angular momentum from the atmosphere to the earth occurs from 49S to 60S, the rate of the transfer is around $2 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ for 2.8 degree latitude; to the south of this region the rate of transfer of angular momentum is from the earth to the atmosphere with a magnitude around $1 \times 10^{18} \text{ Kg M}^2 \text{ Sec}^{-2}$ in every 2.8 degree latitude belt. Over the tropics, on either side of the equator a transfer of angular momentum from the atmosphere to the earth is noted. In the northern hemisphere westerlies the transfer of angular momentum on the intraseasonal time scale is weak. The meridional profile of the intraseasonal momentum transfer reverses during the LOD maxima. In the following cycles as we proceed from the maximum of the length of the day to the minimum LOD and from the minimum LOD to the maximum LOD the same essential pattern repeats.

(d) *Latitudinal variation of the low frequency variability of the mountain torque*

In Figs. 28a and 28b each of the Figs. 28a and 28b illustrate three curves for the variability of the low

Fig. 28. Latitudinal profile of the low frequency component of mountain torque a) minimum of LOD May 1, 1979 and maximum of LOD May 17, 1979. Average of LOD May 9, 1979. b) maximum of LOD May 17, 1979 and minimum of LOD June 6, 1979. Average of LOD May 27, 1979. Units $\times 10^{18} \text{ kg M}^2 \text{ Sec}^{-2}$.

frequency modes of mountain torques. This set of the three profiles for the ascending node respectively refer to the maximum of the length of the day, the average of the length of the day and the minimum length of day.

In Fig. 28a the profiles on May 1 pertaining to the minimum of the LOD has generally negative values except from 35°S to 4°S suggesting a net transfer of westerly angular momentum from the earth to the atmosphere. During the maximum of the LOD on May 17, 1979, the shape of mountain torque profile reverses at most of the latitudes. It implies a transfer of angular momentum from the atmosphere to the earth at most of the latitudes except between 18°S and 35°S and 43°N to 60°N . The profile for May 9, 1979, for the average LOD, however, shows maxima in the momentum transfer from the earth to the atmosphere south of 38°S , from 29°S to 12°S and from 26°N to 43°N .

Figure 28(b) covers the period from the maximum of the LOD on May 17 to the minimum LOD on June 6, 1979, here the changes in the angular momentum transfers do not appear to be significant. During the average LOD a large latitudinal belt exhibits a momentum transfer from the earth to the atmosphere over the southern hemisphere; on the other hand, the atmospheric angular momentum is being transferred to the earth over the northern hemisphere.

A similar pattern of momentum transfer was also noted in the following cycle.

The significant finding here is that the largest angular momentum transfers occur during the average length of day. The average length of day is defined as a period halfway between the minimum and maximum LOD.

8.6 Angular momentum transport stream function

An illustration following the work of Widger (1948) and Peixoto and Oort (1984) can be constructed to illustrate the horizontal and vertical flow of angular momentum. Such a streamfunction satisfies a continuity equation *i.e.*, however it is somewhat artificial since it assumes no sources or sinks by satisfying a continuity equation.

Given the mean meridional circulation (V, ω), the streamfunction for the angular momentum transport is given by the equations:

$$\frac{\partial \Psi}{\partial p} = \frac{+2\pi R^2}{g} \cos^2 \phi \cdot UV \quad (23)$$

$$\frac{\partial \Psi}{\partial y} = \frac{-2\pi R^2 \cos^2 \phi}{g} \omega U \quad (24)$$

here R is the radius of the earth, and ϕ is the latitude.

One can construct the geometry of Ψ from the meridional wind V alone by imposing a boundary value for Ψ at the top of the atmosphere. Furthermore since this equation is linear, hence the transient transport stream function can be easily constructed; *i.e.*, for the 30 to 50 day oscillations using the time filtered values of V and U .

Here we shall first show the time mean angular momentum transport (for the unfiltered winds) and the transport by the transients for the different reference points typifying the length of day variations. These are illustrated in Figs. 29a, 29b, 29c and 29d.

The composite mean total transport of relative angular momentum for the four legs along the length of day cycle shows a pronounced cell with an ascending branch near 25°S and 50°S and a descending lobe near 75°S . This depiction is for the northern summer months of the year 1979. The cell over the winter hemisphere is stronger than that over the summer hemisphere. The annual mean picture of Peixoto and Oort (1984) exhibits nearly the same intensity for the cells in both hemispheres.

The *transient* transport streamfunction is shown in the four panels starting from the minimum length of day. The four panels show a pronounced gyre around 30°S and 70°S . During the length of day cycle we do not see any reversal of circulations for these transient gyres. The reason for that being the U and V are strongly correlated and they both exhibit a change in sign together. A strong source is evident at the earth's surface near 40°S . The transient

Fig. 29. Angular momentum transport streamfunction during different phase of length of day cycle a) minimum length of day b) average length of day c) maximum length of day d) average length of day Units: $10^{18} \text{ kg m}^2 \text{ s}^{-2}$.

sources are located near 15° and 75°S in all of these panels. These transient momentum transport cells south of the equator seem to be in phase with the seasonal mean transport stream function between 40°S to 70°S . The magnitude of these transient transports are roughly 2 orders of magnitude smaller than the mean transport.

Over the southern hemisphere winter the strongest contribution to the transient fluxes at the earth's surface came from the frictional torques in these latitudes of strong surface wind variability. The mountain torques are also large but not comparable to those of the northern hemisphere. It appears that these transients play a significant role in

transferring angular momentum for the earth to the upper troposphere in the high latitudes of the southern hemisphere, where the anomalous values of angular momentum are seen as zonal wind anomalies.

In summary, the role of the easterly trades appears to provide a significant contribution to the increase in the length of day via frictional torques. During the ascending nodes westerly angular momentum is transferred from the earth to the atmosphere. The mountain and frictional torques over the middle latitude appear to have a major role in the transfer of westerly angular momentum from the atmosphere to the earth during the descending node *i.e.* during the decrease in the length of day. This is in line with the work of Madden (1987) who has addressed the surface frictional torques.

9. Theoretical considerations

a) Brief review of other work

A linear model for intraseasonal oscillation was presented by Emanuel (1987). The emphasis of this paper was on the air sea flux of latent heat and its rapid redistribution in the vertical by cumulus convection. This linear model successfully predicts an eastward wave propagating number 1 line disturbance in the equatorial latitudes. This study supports the findings of Lau and Chan (1985) *i.e.*, a lower tropospheric heat source for the excitation of these Kelvin waves.

Using an ocean covered earth, Hayashi and Sumi (1986), simulated the equatorial 30 to 50 waves in a general circulation model. Their results show a reasonable agreement to observations of the divergent planetary scale waves. Emanuel (1987) interprets their results by stating that neither land-sea contrast nor zonal asymmetry are necessary to explain the basic mechanism of the oscillation. That appears to be true for the zonally propagating equatorial waves but they do not explain many of other observed features we have reviewed in Section 1.

Webster (1983) addressed the meridionally propagating monsoonal low frequency waves. He showed that the temperature contrast between the warm land area to the north of the cloud lines provides a differential heating, which if modeled with a detailed surface hydrology, can provide realistic phase propagation. However, it appears from his results that he was able to describe a 10 to 20 day oscillation which is about twice as fast as the observed time scale for the meridionally propagating modes.

The issue of the vertical distribution of heating for these intraseasonal oscillations remains unresolved at the present time. When one looks at the vertical distribution of the apparent heat source (*i.e.* the parameter Q_1 made well known by Professor Yanai) one notes that the maximum value of Q_1 is found near 400 mbs and not near 700 mbs. Those values of Q_1 were reported by Luo and Yanai (1984)

from their examination of the Indian monsoon data sets. It should be stated that their measures of Q_1 includes the contributions of both high and low frequencies and furthermore their measures of Q_1 include the contributions of non-convective heating as well. It is not clear whether the vertical distribution of convective heating on the time scale of 30 to 50 days would have a maximum near 700 mb as is suggested in the studies of Lau and Peng (1987). In a full GCM, the low and the high frequency motions all coexist and the heating in the model should include contributions from all frequencies.

b) Air sea interaction theory for the 30–50 day oscillation

The earth's atmosphere obviously has a large number of degrees of freedom. Thus there may be many different excitation mechanisms that through interaction among these degrees of freedom, result in oscillations, particularly when the period is broad band as it is in the 30 to 60 day oscillation. Thus different mechanisms may be operating in different realizations, leading to different explanations of the phenomenon.

In the following we present one possible mechanism for producing such a low frequency oscillation. It involves ocean-atmosphere interaction in which atmospheric convection transports horizontal momentum from aloft to the sea-surface, thereby enhancing the wind-inducing mixing of the surface layer of the ocean. Mixing with the deeper cold water produces a drop in surface temperature, followed by a drop in convective activity, until solar heating restores the sea-surface temperature and convection starts once again. This is shown schematically in Fig. 30. We will show that this leads to an oscillation of approximately 50 days.

In the simplest for of this model, we take the sea-surface to be warmed by solar heating and cooled by wind-mixing as follows:

$$\rho_w c_w h \frac{dT_s}{dt} = S_0 - \rho_w c_w h \left[\frac{T_w^\infty}{h} \frac{dh}{dt} \right] \quad (25)$$

where

T_s = sea-surface temperature

ρ_w = sea-water density

c_w = specific heat of sea-water

h = depth of the surface mixed layer of the ocean

S_0 = solar heating, taken to be 0.1 of the solar constant

T_w = deep ocean temperature

t = time

For the rate of deepening of the surface layer we use the relation given by Turner and Kraus (1966).

$$\frac{dh}{dt} = k (u_s)^3 \quad (26)$$

Fig. 30. Three phases in one cycle of the low frequency oscillation a) stably stratified atmosphere over an ocean with warm mix layer at temperature T_s and depth h , lying over cold deep water of temperature T_w^∞ . Winds at the sea surface are weak. b) solar heating increases T_s , leading to atmospheric convection with momentum transport and increased surface wind u_s . c) Increased u_s deepens the mixed layer and lowered T_s by mixing with deep cold water. Solar heating of the sea surface will cause the cycle to repeat.

where u_s is the surface wind speed and $k = 2.75 \times 10^{-10} \text{ sec}^2 \text{ cm}^{-2}$, for u_s in cm sec^{-1} , and h in cm. In a more complete model we include other cooling mechanism as well and a mechanism for restoring h after it has been deepened.

The surface wind speed is taken to be damped by friction and increased by convective transport of high momentum from aloft. The formulation is based on laboratory results, (Ingersoll, 1966, Krishnamurti and Zhu, 1991), where the momentum flux M is found to be proportional to the horizontal speed gradient and to the cube root of the Rayleigh number R .

$$M = c\rho v \frac{\Delta u}{d} \left[\frac{R}{R_c} \right]^{1/3} \quad (27)$$

Here $R = \frac{g\alpha}{kv} \Delta T d^3$, where g denotes the acceleration of gravity, α the thermal expansion coefficient, k the thermal diffusivity, v the kinematic viscosity, ΔT the temperature difference across the layer of depth d . R_c is the critical Rayleigh number which is of order 10^3 . ρ is the fluid density, Δu the difference in horizontal speed from bottom to top, and c is a dimensionless number of order unity.

Although these results refer to statistically steady state experiments, it seems that on a time scale of tens of days (of the low frequency mode) quasi-steadiness should hold for the small scale and short-lived convective plumes. The momentum flux divergence introduces a vertical length scale which finally leads to the $2/3$ power of the Rayleigh number entering the horizontal momentum equation as follows:

$$\frac{du_s}{dt} = -k'u_s + k''(u_{a\infty} - u_s)(T_s - T_{a\infty})^{2/3} \quad (28)$$

where

$$\begin{aligned} u_a^\infty &= \text{wind speed aloft} \\ T_a^\infty &= \text{air potential temperature aloft} \\ k' &= v/d^2 \\ k'' &= c^2 \left[\frac{g\alpha}{kv} \right]^{2/3} [R_c]^{-2/3} \end{aligned}$$

k' and k'' are estimated as follows. The eddy viscosity v of the air is taken to be $10^5 \text{ cm}^2 \text{ s}^{-1}$ (Krishnamurti, 1975), d the depth of the atmosphere involved. For k'' we use $c=1$, $\alpha=(1/300)(^\circ\text{C})^{-1}$, $k=v=10^5 \text{ cm}^2 \text{ s}^{-1}$, $R_c=10^3$.

This is a two-component system since dh/dt can be eliminated. The fixed point occurs when $T_s - T_s^*$ and $u_s = u_s^*$ given by

$$\begin{aligned} u_s^* &= (S_0/\rho_w c_w k T_w^\infty)^{1/3}, \\ (T_s^* - T_a^\infty)^{2/3} &= \frac{k' u_s^*}{k''(u_a^\infty - u_s^*)} \end{aligned} \quad (29)$$

Perturbing the system and linearizing around the fixed point leads to the following equations governing the perturbation temperature θ and perturbation wind speed μ .

$$\frac{d\theta}{dt} = -A\mu \quad (30)$$

$$\frac{d\mu}{dt} = B\theta - K\mu \quad (31)$$

with damped oscillatory solutions θ, μ of the form $e^{-\frac{K}{2}t \pm i\omega t}$

where

$$K = k' \left[1 + \frac{u_s^*}{u_a^\infty - u_s^*} \right] \quad (32)$$

Table 2.

d	u_s^* (ms^{-1})	T_s^* ($^{\circ}\text{C}$)	Decay time τ_d (days)	Oscillation period τ_0 (days)
5	0.85	25.84	29	43
10	0.85	25.10	116	30

$$\omega = \sqrt{AB - \frac{K^2}{4}} \quad (33)$$

$$AB = \frac{1}{\sqrt{k'}} \left[\frac{2kT_w^\infty}{h} \right] [k''u_s^*(u_a^\infty - u_s^*)]^{3/2} \quad (34)$$

The free parameters are T_w^∞ , T_a^∞ , u_a^∞ , and h . These we choose as follows:

$$\begin{aligned} T_w^\infty &= 20^{\circ}\text{C} \\ T_a^\infty &= 25^{\circ}\text{C} \\ u_a^\infty &= 2u_s^* = 1.69 \times 10^2 \text{ cm s}^{-1} \\ h &= 10 \text{ m} \end{aligned}$$

For d specified, k' is specified. For atmospheric convection of depth $d=5$ km and 10 km, we shows in Table 2 the corresponding fixed point surface wind speed U_s^* , fixed point surface temperature T_s^* , the decay time τ_d of the perturbed solution ($\tau_d = \frac{2}{K}$), and the oscillation period $\tau_0 = \frac{2\pi}{\omega}$.

10. Conclusions

We have attempted to present a review and to provide new results on the intraseasonal low frequency motions of the earth's atmosphere. This problem is of global context because one notes length of day variations correlate very closely to the changes in the total atmospheric angular momentum. The newly acquired lunar laser ranger data sets are of good quality. The recent advances in four dimensional data assimilation have also improved the quality of the daily global wind analysis. The improvements have led to the recognition that millisecond to fraction of millisecond changes in the length of day are indeed strongly related to intraseasonal oscillations of the atmosphere.

The scenario portrayed here is roughly as follows:

- Organized tropical convection in the near equatorial latitudes (particularly over the Indian and the Western Pacific Ocean) appears to be the start of a number of events.
- In response to the convection surface easterlies to its east and monsoon westerlies to the west amplify.
- The surface easterlies in the tropics contribute to a transfer of westerly angular momentum from the earth's surface to the atmosphere.
- The westerly angular momentum is evidently transferred to the upper troposphere of the tropics via deep cumulus convection. This aspect is not covered in the paper.

- The atmospheric angular momentum increases and reaches a maximum during this period.
- The length of day reaches a maximum during this period.
- The angular momentum estimates show the contributions of the frictional and mountain torques on the intraseasonal time scale during the different phases of the length of day cycle.
- A local theory for the 30 to 50 day oscillation has been proposed involving air-sea interaction. Solar heating enhances atmospheric convection which leads to increased surface winds, which then leads to deepening and cooling of the oceanic mixed layer from entrainment of cold deep water. Convection ceases and surface winds decay. Solar heating warms the mixed layer for a repetition of the cycle. This along with the Kelvin wave theory (reviewed here) presents a mechanism for initial excitation of this cycle.
- Mean zonal flows on the intraseasonal time scales exhibit a strong oscillation. Thus the assumption of linearity for the calculations of wave energy flux appears invalid. The nonlinear problem is cast in the frequency domain and we have shown a tropical-middle latitude interaction by a flux of wave energy from the tropics to the high middle latitudes especially during the summer monsoon months.
- The descending mode of the length of day is especially interesting in several aspects: the total atmospheric angular momentum decreases during this period. The frictional and mountain torques on the low frequencies provide of drain of atmospheric angular momentum by transferring it to the earth. The flow field in the upper troposphere exhibits a transition from a high to a low index which often culminates in major blocking events.

The implication to extended range forecasting lie in our ability to correctly model the changes in the length of day. That requires a better description of all of the components of the atmosphere-earth angular momentum cycle. The most difficult areas seem to be the better parameterization of tropical convection, the meridional fluxes of wave energy, the better representation of frictional and mountain torques. Monitoring of the length of day would be most important for the isolation of areas of deficiency in weather and climate models.

Finally it is worth mentioning that many areas on the angular momentum cycle are not covered by this study. For instance, in an elegant study, Joussaume *et al.* (1986), examined the water cycle along the trajectories of a global model. Basically he asked the question where did the water originate that rained somewhere later. He tagged the source and the sinks of the water along its pathways. Such studies are generally quite difficult to carry out if analysis of real data are used, since the construction of four dimensional trajectories require a good observational coverage. In the context of the monsoon it is however apparent that a 30 to 50 day fluctuation in rainfall is quite pronounced, Hartman and Michelson (1989). The source region for roughly 50 % of this monsoon rainfall resides nearer the equator. The moisture is canoeyed from the equatorial latitude to roughly 25N where the heavy rainfall along the eastern foothills of the Himalayas occurs. The rainfall on a given episode such as the onset or revival of monsoon can deposit several tons of water on the ground. The eventual return of this water via runoff and rivers back to lower latitude or origin is a slower process. This northward transport of water within the atmosphere and 30 to 60 day oscillations of rainfall suggest yet another possible contributor to the change of earth's angular momentum, *i.e.* that from a sudden deposition (time scale 3 to 4 days from the time of evaporation) of water. This can perhaps also contribute to small changes in the length of day. In modeling studies, where good account of the water budget in all of its facets is carefully considered can perhaps lead to further understanding of low frequency variability of the angular momentum.

Acknowledgements

The research reported here was supported by the following grants: NSF ATM-8812053, ONR N0014-89J01662 and NOAA NA88AA-D-AC049. Computations for this work were carried out on the CRAY Y-MP at the National Center for Atmospheric Research. We acknowledge the assistance provided by Sarah Meador and Rosemarie Raymond. Mr. Vivek Hardiker assisted in the coding of part of this problem.

References

- Anderson, J.R. and R.D. Rosen, 1983: The latitude height structure of 40–50 day variations in atmospheric angular momentum, *J. Atmos. Sci.*, **40**, 1584–1591.
- Chao, F., 1989: Length of day variation caused by El Niño-southern oscillation and quasi-biennial oscillation, *Science*, **243**, 223–243.
- Comeaux, J., 1991: Origin and structure of the low frequency modes, M.S. Thesis available from Department of Meteorology Florida State University, Tallahassee, FL 32306.
- Dickey, J.O., M. Ghil, S.L. Marcus, 1991: Extratropical aspects of the 30–60 day oscillation in length-of-day an atmospheric angular momentum, in press.
- Eliassen, A. and E. Palm, 1961: On the transfer of energy in stationary mountain waves, *Geof. Pub.*, **No. 3**, 1–23.
- Emanuel, K.A., 1987: An air-sea interaction model of intraseasonal oscillations in the tropics. *J. Atmos. Sci.*, **44**, 2324–2340.
- Gill, A.E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**, 447–462.
- Hartman, D.L. and M.L. Michels, 1989: Intraseasonal periodicities in Indian rainfall. *J. Atmos. Sci.*, **46**, 2838–2862.
- Hayashi, Y.-Y. and A. Sumi, 1986: The 30–40 day oscillations simulated in an “aqua planet” model. *J. Meteor. Soc. Japan*, **64**, 451–467.
- Hayashi, Y., 1980: Estimation of nonlinear energy transfer spectra by the cross-spectral method, *J. Atmos. Sci.*, **37**, 299–307.
- Ingersoll, A.P., 1966: Thermal convection with shear at high Rayleigh number. *J. Fluid Mech.*, **25**, 209–228.
- Joussaume, S., R. Sadorny and C. Vignal, 1986: Origin of precipitating water in a numerical simulation of the July climate, *Ocean-Air Interactions*, **1**, 43–56.
- Krishnamurti, R. and Y. Zhu, 1991: Heat and momentum fluxes in sheared turbulent convection. (to appear in) *J. Fluid Mech.*
- Krishnamurti, R., 1975: On cellular cloud patterns. Part 3: Applicability of mathematical and laboratory models. *J. Atmos. Sci.*, **32**, 7, 1373–1383.
- Krishnamurti, T.N., D. Oosterhof and A. Mehta, 1988: Air sea interaction on the time scale of 30 to 50 days during FGGE, *J. Atmos. Sci.*, **45**, 1304–1322.
- Krishnamurti, T.N. and S. Gadgil, 1985: ON the structure of the 30 to 50 day mode over the globe during FGGE, *Tellus*, **37A**, 336–360.
- Krishnamurti, T.N., P.K. Jayakumar, J. Sheng, N. Surgi and A. Kumar, 1985: Divergent circulation on the 30 to 50 day time scale, *J. Atmos. Sci.*, **42**, 364–375.
- Krishnamurti, T.N. and D. Subhrmanyam, 1982: The 30–50 day mode at 850 mb during MONEX, *J. Atmos. Sci.*, **39**, 2088–2095.
- Kung, E.C. and W.E. Baker, 1986: Spectral Energetics of the observed and simulated northern hemisphere general circulation during blocking episodes, *J. Atmos. Sci.*, **43**, 2792–2812.
- Lau, K.M., Li Peng, C.H. Sui and T. Nakazawa, 1989: Dynamics of super cloud clusters, westerly wind bursts, 30–60 day oscillations and ENSO: A unified view, *J. Meteor. Soc. Japan*, **67**, 205–219.
- Lau, K.M. and P.H. Chan, 1985: Aspects of the 40–50 day oscillation during the northern winter as inferred from outgoing longwave radiation, *Mon. Wea. Rev.*, **113**, 1889–1909.
- Lorenc, A., 1984: The evolution of planetary scale 200 mb divergence during the FGGE year. *Quart. J. Roy. Meteor. Soc.*, **110**, 427–442.
- Luo, H. and M. Yanai, 1984: The large-scale circulation and heat sources over the Tibetan Plateau and surrounding areas during the early summer of 1979.

- Part II: Heat and moisture budgets. *Mon. Wea. Rev.*, **112**, 966–989.
- Madden, R.A., 1987: Relationship between changes in the length of day and the 40 to 50 day oscillation in the tropics, *J. Geophys. Res.*, **92**, 8391–8399.
- Madden, R.A. and P.R. Julian, 1971: Detection of a 40–50 day oscillation in the zonal wind in the tropical Pacific, *J. Atmos. Sci.*, **28**, 702–708.
- Madden, R.A. and P.R. Julian, 1972: Description of global-scale circulation cells in the tropics with a 40–50 day period, *J. Atmos. Sci.*, **29**, 1109–1123.
- Mechoso, C.R., K. Yamazaki, A. Kitoh and A. Arakawa, 1985: Numerical forecasts of stratospheric warming events during the winter of 1979. *Mon. Wea. Rev.*, **113**, 1015–1029.
- Mehta, A.V. and T.N. Krishnamurti, 1988: Interannual variability of the 30–50 day wave motion, *J. Meteor. Soc. Japan*, **66**, No. 4, 538–548.
- Nakazawa, T., 1988: Tropical super clusters within intraseasonal variations over the western Pacific. *J. Meteor. Soc. Japan*, **66**, 823–839.
- Newton, C.W., 1971a: Mountain torque in the global angular momentum balance. *J. Atmos. Sci.*, **28**, 1329–1341.
- Newton, C.W., 1971b: Global angular momentum balance, earth torque and atmospheric fluxes, *J. Atmos. Sci.*, **28**, 1329–1341.
- Pasch, R.L., 1983: On the onset of the planetary scale monsoon FSU rep. no 83.91, 197 pp. (Available from the Department of Meteorology, Florida State University, Tallahassee, FL 32306).
- Peixoto, J.P. and A.H. Oort, 1984: Physics of climate, *Reviews of modern Physics* **56**, 365–429.
- Sakellarides, G., 1989: Atmospheric effective angular momentum fluctuations for 1986–87; research department technical report no. 62, Feb. 1989, European Centre for Medium Range Weather Forecast, Shienfield Park, Reading.
- Sheng, J. and Y. Hayashi, 1990a: Estimation of atmospheric energetics in the frequency domain during the FGGE year, *J. Atmos. Sci.*, **47**, 1255–1268.
- Sheng, J. and Y. Hayashi, 1990b: Observed and simulated energy cycles in the frequency domain, *J. Atmos. Sci.*, **47**, 1243–1254.
- Simmons, A.J., J.M. Wallace and G.W. Branstator, 1983: Barotropic wave propagation and instability and atmospheric teleconnection pattern, *J. Atmos. Sci.*, **40**, 1363–1392.
- Streten, N.A. and Z.W. Zillman, 1984: *Climate of the south Pacific ocean, in World Survey of Climatology*, Vol. 15, Ed. H. Van Loon, p 716, Elsevier, New York.
- Swinbank, R., 1985: The global atmospheric angular momentum balance inferred from analyses made during the FGGE, *Quart. J. Roy. Meteor. Soc.*, **111**, 977–992.
- Turner, J.S. and E.B. Kraus, 1966: A one-dimensional model of the seasonal thermocline, I.A. laboratory experiment and its interpretation, II. The general theory and its consequences. *Tellus*, **19**, (1), 88–106.
- Webster, P.J. and C. Zhang, 1989: Effects of zonal flows on equatorial trapped waves, *J. Atmos. Sci.*, **46**, 3632–3652.
- Webster, P.J., 1983: Mechanisms of monsoon low-frequency variability: surface hydrological effects, *J. Atmos. Sci.*, **40**, 2110–2124.
- Weickmann, K.M., G.R. Lussy and J.E. Kutzbach, 1985: Intraseasonal (30–60) fluctuations in outgoing longwave radiation and 250 mb streamfunction during northern winter, *Mon. Wea. Rev.*, **113**, 941–961.
- Widger, W.K. Jr., 1948: A study of the flow of angular momentum balance of the atmosphere, *J. Atmos. Soc.*, **5**, 291–299.
- Yasunari, T., 1981: Structure of an Indian summer monsoon system with around 40 day period, *J. Meteor. Soc. Japan*, **59**, 336–354.

角運動量、1日の長さ、および季節内低周期現象

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この論文では、30日から50日の時間スケールの現象が扱われている。モンスーンの東西風、角運動量そして自転周期の長さとはその時間スケールでよい相関がみられる。その関係は、赤道付近の積雲群の活動から始まるシナリオで考えられる。つまり、東西風が加速され、角運動量が増加し、日の長さが延びる。西向き角運動量は地上の東風の地域で地球に移り、自転周期を遅らせる。その長さが最長になりつつあるときにインドモンスーンが活発化であり、それが最長から転じる時期には、赤道付近での積雲群の活動がおとろえる。現象的には、そのころにブロッキングが中高緯度に見られる。角運動量の減少は西向き角運動量が地表摩擦や山岳トルクを通じて地球に移ると説明できる。そのトルクが、考えている時間スケールにおいて大気角運動量の変化に関係していることは明らかである。中緯度の低周期現象の特徴

もある部分において波動による子午面方向のエネルギー輸送で考えられる。非線形で検討すると、線形では存在した臨海緯度の影響がなく、波動を通じての熱帯・中緯度の結合が示される。

振動に関する簡単な数字モデルを提示する。大気・海洋相互作用の局所理論である。つまり、最初、大気は安定で海面では弱い風、上層ほど強い風とする。海洋では混合層があり、下層ほど冷たい。そこに日射が当たると海面温度が上がり、大気中に積雲対流が生じ、海上の風速を強める。そのため海洋混合層が深くなり、海面温度が下がる。その結果、積雲対流はおとろえ、風も弱まる。そして、再び日射による昇温が繰り返される。この周期が 30 日のオーダーになる。