Vol. 5 No. 2 **Advances in Atmospheric Sciences** May 1988

MAINTENANCE AND OSCILLATION MECHANISMS OF SUMMER TROPICAL UPPER-TROPOSPHERIC EASTERLIES

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Received December 18, 1986

ABSTRACT

The mechanisms of the maintenance and oscillation of 1982 summer tropical 200-hPa mean easterly flow and extra-long waves are investigated in terms of the energy equations in wavenumber-frequency space. Calculation results show that the difference in heating between land and sea and the boundary effect serve as the main source of energy; frictional dissipation as the sink; the conversion of available potential energy into kinetic takes place dominantly in the waves of number 1-2; such transformation is accomplished in just a small amount in zonal mean flow and therefore can be ignored because of the value.

In the interaction between wave and zonal mean flow, the latter loses its available potential and gains kinetic energy. The tropical easterly belt over 20°N-5°S is found barotropically stable and that over 10°N-5°S, unstable. The waves of number 2 and 1 manifest themselves a primary source and sink of kinetic energy, respectively, in the interplay between waves and between zonal mean flow and wave.

It is found that zonal mean flow and the waves of number 1-2 have a roughly 40-and 20-day oscillational period of kinetic energy, respectively, whose primary mechanism is the transfer of barotropic energy, the conversion of baroclinic energy, and the boundary effect.

I. INTRODUCTION

The linearity theory accounts satisfactorily for the dampening and developing mechanism of small-amplitude waves in the atmosphere. For waves of finite amplitude, however, the non-linearity theory is of much importance. FjΦtoft (1953) and Lorenz (1960) are concerned with possible mechanisms and dynamical processes of the non-linear interaction of waves: Saltzman (1957) shows that the energy equations of the atmospheric motion can extend to be used in wavenumber space in terms of the alongside-parallel Fourier expansion, and hence the properties of the non-linear effect can be explored with actual meteorological data; subsequently Kao (1968) indicates that the energy equations in wavenumber space can also be extended to apply to wavenumber-frequency space, which makes it possible to investigate the mechanism of varied frequency oscillations, of different-scale waves in the real atmosphere.

The mechanism is explored for the growth and decay of mid-tropospheric waves for the northern winter in wavenumber-frequency space (Kao and Chi, 1978); the mechanism for the energy maintenance at upper-tropospheric 200 hPa over the belt 15°N-15°S is examined for the summer of 1967 and 1972, respectively (Kanamitsu et. al., 1972; Krishnamurti, 1978). Besides, enormous work has been done by Murakami (1977) and even more complete investigation by Depradine (1978). In particular, Lin Hai (1987) indicates that the mechanism for the energy maintenance of tropical zonal mean flow and waves of various scales is fully revealed through the calculation of all the terms of the energy equations with tropical 200-hPa data for 1983. A significant difference, however, exists between their results because of the belts considered. Therefore, more calculation cases are needed for comparison and verification in order to get the fullest possible knowledge of the energetics of the waves in the tropical atmosphere.

II. THE ENERGY EQUATIONS WITH CALCULATIONS

With the aid of the energy equations in wavenumber space on spherical coordinates presented by Saltzman (1957), we have

$$\frac{\partial E_K(K,t)}{\partial t} = N_K(K,t) + M_K(K,t) + A_K(K,t) + A_{KB}(K,t) + N_{KB1}(K,t) + N_{KB2}(K,t) + G_K(K,t) \quad (K \neq 0) \tag{1}$$

$$\frac{\partial E_A(K,t)}{\partial t} = N_A(K,t) + M_A(K,t) + K_A(K,t) + N_{AB1}(K,t) + N_{AB2}(K,t) \\ H_A(K,t) \quad (K \neq 0) \tag{2}$$

where

$$E_K(K,t) = \frac{1}{\sin\varphi_N - \sin\varphi_S}\int_{\varphi_S}^{\varphi_N}\{|u(K,t)|^2 + |v(K,t)|^2\} \tag{3}$$

$$E_A(K,t) = \frac{1}{\sin\varphi_N - \sin\varphi_S}\int_{\varphi_S}^{\varphi_N} C_p\gamma|\theta(K,t)|^2\cos\varphi d\varphi \tag{4}$$

denote kinetic energy (KE) and available potential energy (APE), respectively, of wave K*; φ_N and φ_S the latitude of the northern and southern boundary, respectively (and therefore (1) and (2) represent mean condition of the belt under study); C_p specific heat at constant pressure; $\gamma = ([\bar{T}] - C_pR^{-1}p\partial[\bar{T}]/\partial p)$ where $[\bar{T}]$ denotes the regionally-averaged condition.

The reader is referred to Saltzman (1957) for the specific expressions of the terms in (1) and (2), and here only their physical meanings are indicated, viz.:

$N_K(K,t)$-non-linear effect, which represents the contribution of the interaction between waves (IbWW) to the KE change rate of wave K $(K \neq 0)$;

$M_K(K,t)$-the contribution to the KE change rate of the interaction between wave K $(K \neq 0)$ and zonal mean flow;

$A_K(K,t)$-the conversion rate of APE into KE of wave K;

$A_{KB}(K,t)$-geopotential convergence, i.e., the contribution to the KE change rate of wave K made by the pressure work due to the difference between the boundaries;

$N_{KB}(K,t)$-KE convergence rate caused by the difference in space due to IbWW, which is composed of N_{KB1} and N_{KB2}, denoting convergences, north-south horizontal and vertical, respectively;

$G_K(K,t)$-the residual term, including calculation errors and dissipation of KE of wave K due to Reynolds molecular viscosity stress;

$N_A(K,t)$-the contribution of IbWW to the APE change rate of wave K $(K \neq 0)$;

$M_A(K,t)$-the rate of transport of APE of zonal mean flow to that of wave K $(K \neq 0)$;

$K_A(K,t)$-the rate of conversion opposite to $A_K(K,t)$;

$N_{AB}(K,t)$-the convergence rate of APE resulting from the difference in space caused by the interaction between wave K and other types, which consists of $N_{AB1}(K,t)$

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* The wave of number, say, 1, 2, K, etc. will be represented as wave 1, 2, K, etc. hereafter.

and $N_{AB2}(K,t)$, the convergences in both directions as denoted for N_{KB1} and N_{KB2}, respectively;

$H_A(K,t)$-the contribution of non-adiabatic heating to the APE change rate of wave K with calculation errors involved;

If $K=0$, then (1) and (2) are degenerated into the equations for the KE and APE of zonal mean flow, i.e.,

$$\frac{\partial E_K(0,t)}{\partial t}=M_K(0,t)-A_K(0,t)+A_{KB}(0,t)-G_K(0,t) \tag{5}$$

$$\frac{\partial E_A(0,t)}{\partial t}=M_A(0,t)+K_A(0,t)-H_A(0,t) \tag{6}$$

in which

$$E_K(0,t)=\frac{1}{\sin\varphi_N-\sin\varphi_S}\int_{\varphi_S}^{\varphi_N}\frac{1}{2}(\bar{u}^2+\bar{v}^2)\cos\varphi d\varphi \tag{7}$$

$$E_A(0,t)=\frac{1}{\sin\varphi_N-\sin\varphi_S}\int_{\varphi_S}^{\phi_N}\frac{1}{2}C_p\nu[\bar{T}]'^2\cos\varphi d\varphi \tag{8}$$

represent the KE and APE of zonal mean flow, respectively; $[\bar{\ }]'$ the deviation of zonal averages with respect to regional means; and

$$M_K(0,t)=-\sum_{K=1}^{\infty}M_K(K,t) \tag{9}$$

$$M_A(0,t)=-\sum_{K=1}^{\infty}M_A(K,t) \tag{10}$$

denote the contribution to the KE and APE of zonal mean flow of the interaction between zonal mean flow and wave (IbFW). The other terms of (5) and (6) have the similar interpretation to those for wave K.

In addition, adopted in the present study are the KE equations in wavenumber-freqnency space developed by Kao (1978), namely,

$$\begin{aligned}\frac{\partial[|E_K(K,n)|\sin(nt-\pi/2)]}{\partial t}&=n|E_K(K,n)|\sin nt\\&\quad-|N_K(K,n)|\sin\left[nt+\alpha_{NK}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]\\&\quad+|M_K(K,n)|\sin\left[nt-\alpha_{MK}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]\\&\quad+|A_K(K,n)|\sin\left[nt-\alpha_{AK}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]\\&\quad+|A_{KB}(K,n)|\sin\left[nt+\alpha_{AKB}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]\\&\quad+|N_{KB1}(K,n)|\sin\left[nt+\alpha_{NKB1}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]\end{aligned}$$

$$+|N_{KR2}(K,n)|\sin\left[nt-\alpha_{NKR2}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]$$

$$+|G_K(K,n)|\sin\left[nt-\alpha_{GK}(K,n)-\alpha_{EK}(K,n)-\frac{\pi}{2}\right]\quad (K,n\neq 0)\quad (11)$$

$$\frac{\partial\left[|E_K(0,n)|\sin\left(nt-\frac{\pi}{2}\right)\right]}{\partial t}=n|E_K(0,n)|\sin nt$$

$$=|M_K(0,n)|\sin\left[nt-\alpha_{MK}(0,n)-\alpha_{EK}(0,n)-\frac{\pi}{2}\right]$$

$$+|A_K(0,n)|\sin\left[nt+\alpha_{AK}(0,n)-\alpha_{EK}(0,n)-\frac{\pi}{2}\right]$$

$$+|A_{KR}(0,n)|\sin\left[nt+\alpha_{AKR}(0,n)-\alpha_{EK}(0,n)-\frac{\pi}{2}\right]$$

$$+|G_K(0,n)|\sin\left[nt-\alpha_{GK}(0,n)-\alpha_{EK}(0,n)-\frac{\pi}{2}\right]\quad (n\neq 0)\quad (12)$$

As for $n=0$, it indicates the time-averaged condition. From (1), (2), (5) and (6) we derive the balance equations for the maintenance mechanism governing the KE and APE of both wave and zonal mean flow, that is,

$$N_K(K,0)-M_K(K,0)+A_K(K,0)+A_{KR}(K,0)+N_{KR1}(K,0)-N_{KR2}(K,0)+G_K(K,0)=0\quad (K\neq 0)\quad (13)$$

$$N_A(K,0)-M_A(K,0)-K_A(K,0)+N_{AR1}(K,0)-N_{AR2}(K,0)+H_A(K,0)=0\quad (K\neq 0)\quad (14)$$

$$M_K(0,0)-A_K(0,0)+A_{KR}(0,0)+G_K(0,0)=0\quad (15)$$

$$M_A(0,0)+K_A(0,0)+H_A(0,0)=0\quad (16)$$

Computation is made of the ECWMF/WMO data at 1200 GMT between July 1 and September 2 (64 days altogether), 1982 in terms of the above equations. Since our main interest lies in planetary-scale waves, interpolation is performed of the primitive data at 64 points alongside the parallels with the grid of 5° × 5.625° instead of 2.5° × 2.5° in oredr to facilitate fast Fourier transform (FFT). The manipulation is indicated in Lin Hai (1987).

III. PROPERTIES OF ZONAL MEAN FLOW

As shown in Fig. 1, a great depth of easterly flow is observed above 700 hPa between 20°N–5°S for the summer of 1982, with the strongest easterly at over 16 m s^{-1} at 100 hPa and the 200-hPa one near 5°N. It can also be seen that the extent of the easterly flow is narrowest in the vicinity of 200 hPa, widening downward and below 700 hPa a small region of westerly wind appears between 2.5°N–12.5°N, which is the reflection of the equatorial westerly flow. As compared with the summer conditions of other years, especially those of 1983, it is obvious that both the upper easterly and lower westerly winds are a bit stronger in 1982. This may be associated with the event of ENSO of that year, which caused the weakening of the Walker circulation near the equatorial Pacific, thus leading to the reinforcement of both the winds at lower latitudes. Yet in the year of

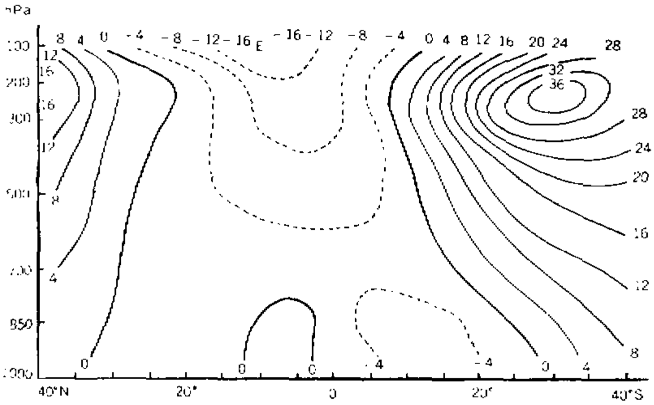


Fig. 1. The meridional vertical cross-section of zonal mean wind over the period of July 1 to September 2. 1982.

1972, when a similar event occurred, the upper mean easterly flow there was weak, which remains to be studied.

It should be noted here that the 64-day zonal mean flow is diverse from the zonal winds at different meridions and times. As far as the 200-hPa tropical easterly belt under consideration is concerned, there is a break in northern South America and the Caribbean Sea, and sometimes in the middle of the Pacific. And even though no break occurs the easterly belt is quite narrow over these areas. Some contribution of westerly flow, therefore, would be indicated on the map of zonal mean flow, which is obviously small in magnitude and hence would not distort the characteristics of the flow and present no difficulties for our discussion.

IV. MAINTENANCE OF KE OF BOTH WAVE AND ZONAL MEAN FLOW

Since our attention is focused on the low-latitude upper-tropospheric easterly flow, the region in question will be confined to 20°N–5°S at the 200-hPa level.

Table 1 summarizes the calculation results of all terms in (13) and (15) with only waves 0–16 considered owing to the sharp decrease in the order of magnitude of the energy spectrum with the increasing wavenumber. It is clear that the acquisition of KE by zonal mean flow is attributed chiefly to: i) the work done by the lateral boundary pressure (A_{KP}) and ii) the IbFW (M_K). For the former case the acquisition is due to the fact that the action of the 200-hPa South Asian and Mexican highs on the north side of the belt results in the divergence flow crossing the isobars in the direction of gradients from the centers, and the zonal flow is thus accelerated on the south side of the systems and for the latter zonal mean flow gains KE from waves, with the maximum contribution made by wave 2, through barotropical transfer (M_K). As for the transformation of APE into KE (A_K), it can be neglected because of its minute and negative magnitude as a result of the Hadley circulation quite weak at that time. On the other hand, the Reynolds molecular viscosity stress (G_K) brings about the dissipation of KE of zonal mean flow, thus maintaining the easterly flow.

Table 1. Calculation Results of the Terms in (13) and (15) for 20°N-5°S. Unit: $m^2\ s^{-2}\ day^{-1}$

Wavenumber K	N_K	M_K	A_K	N_{KB1}	N_{KB2}	A_{KB}	G_K
0	0.0	1.51	0.09	0.0	0.0	1.03	−2.48
1	2.88	1.18	4.92	−3.10	0.62	2.72	−9.22
2	−1.98	−1.12	0.89	−1.13	0.19	3.31	−0.46
3	0.05	0.28	0.22	1.77	0.08	2.20	0.51
4	0.92	0.06	0.11	−0.26	−0.01	1.18	−0.98
5	1.13	0.18	0.23	−0.34	0.34	−0.65	−0.64
6	1.21	−0.22	0.22	−0.59	0.00	−0.29	−0.33
7	0.04	−0.07	0.02	−0.07	0.06	−0.23	0.25
8	0.16	0.05	0.13	−0.01	0.01	0.19	0.05
9	0.19	0.14	0.10	0.01	−0.04	0.27	0.29
10	0.43	0.21	0.09	−0.16	0.01	0.08	−0.21
11	−0.04	−0.07	0.38	−0.15	0.04	0.06	0.08
12	−0.10	−0.08	0.10	0.11	−0.00	−0.01	0.28
13	−0.43	−0.08	−0.03	0.23	0.02	0.13	0.16
14	0.08	−0.08	−0.02	0.05	0.00	−0.14	0.11
15	0.43	−0.11	0.04	0.24	0.02	0.04	0.20
16	−0.11	−0.07	0.01	0.04	0.03	0.05	0.11

As regards waves, they obtain KE from various sources, usually the conversion of APE being one of the important ones for most waves . extremely significant for waves 1–2, almost by one order of magnitude greater than in other types. This occurs owing to the discrepancy in the east-west direct thermal circulation resulting from the land-sea thermal difference. The pressure work at the lateral boundary (A_{KB}) serves as a main source of KE for waves 1-4 and 8-11 alike, and deprives 5-7 of their KE. Waves 12–16 are not considered here because of their order of magnitude.

The lateral boundary convergence in the non-linear IbWW (N_{KB1}) plays a negative role in the increase of KE for waves 9–11, with wave 9 having a tiny positive value, and it has substantial effect on mere waves 1-3. This indicates that the non-linear IbWW among these extra-long waves in the tropical easterlies causes their KE to diverge to both sides, i.e., to be transported to westerlies. The convergence in vertical (N_{KB2}) plays a considerable role in making KE of wave 1-2 grow in contrast with others. On the whole, however, the boundary convergence results in the dissipation of KE of the waves considered.

Since the barotropical transfer of KE is quite complex in IbWW and IbFW, the exchange is schematically shown in Fig. 2 for discussion. It is apparent that in IbFW (M_K) all waves except 1, 5 and 8 lose KE, with wave 2 losing most and wave 1 gaining most. On the whole, waves 1–16 lose their KE to zonal mean flow. Hence it can be assumed that easterlies are barotropically stable, a result identical with the calculation for the 1983 summer (Lin Hai, 1987) and different from that derived by some researchers (see, for instance, Kanamitsu et al., 1972). Also, Kanamitsu et al. (1972) indicates an enormous exchange of energy between wave 11 and zonal mean flow, the former losing to the latter. Yet waves 4–15 gain KE, and the ultimate result is that waves acquire KE from the flow, suggesting a barotropically unstable condition. This may be due to the

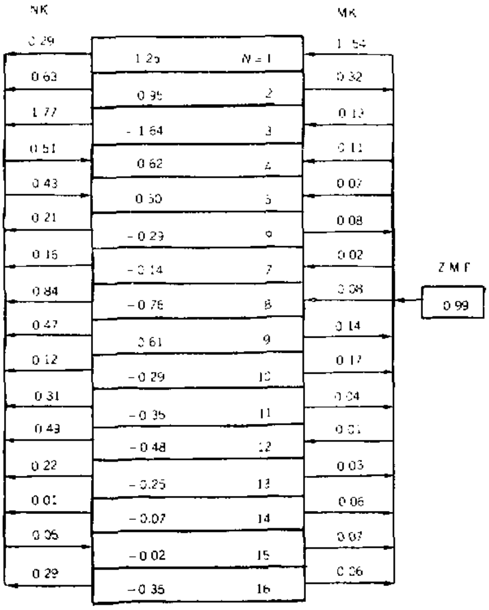


Fig. 2. The exchange of KE between 20°N–5°S at 200 hPa. The number in the middle indicates the net loss or gain of energy and those on the left and right show the loss and gain due to IbWW and IbFW, respectively. Unit: $m^2\ s^{-2}\ day^{-1}$.

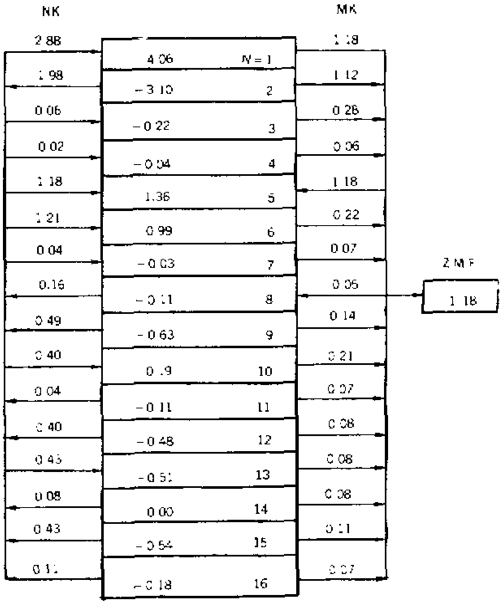
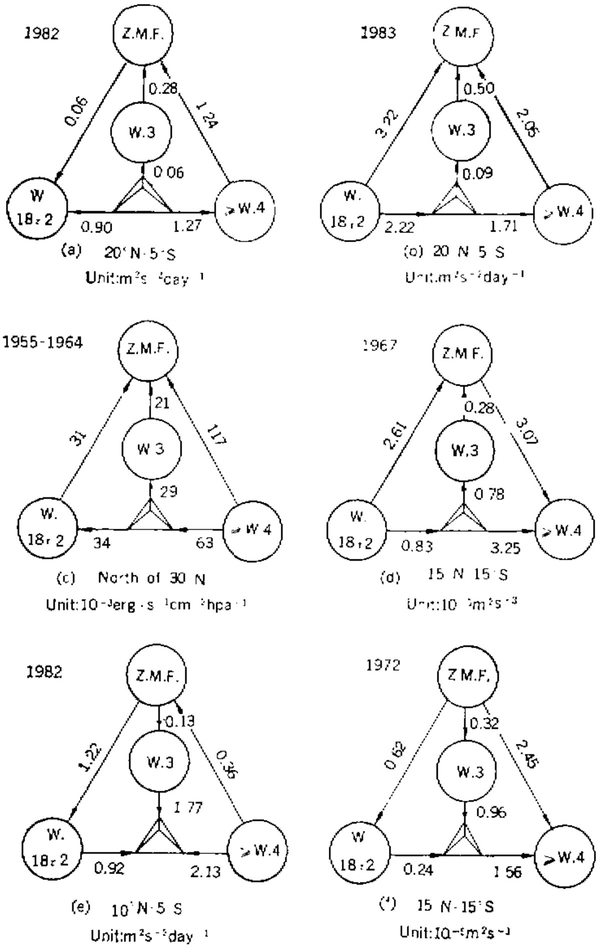


Fig. 3. as in Fig. 2 except for 10°N–5°S.

regions under investigation. Kanamitsu et al. (1972) takes for calculation the area 15°N–15°S including the SH subtropical westerly jet stream, thus leading to a fact that the flow averaged over the zonal belt is westerly. On account of the effect of the inclination of the trough and ridge, mean westerly (easterly) flow is barotropically unstable (stable). However, it is not the case in the vicinity of the equator. Figure 3 is a chart showing the exchange of barotropical energy within 10°N–5°S for the period of July–early September, 1982. It is evident from Fig. 3 that such waves as 1, 3–5 and 7–8 obtain from and waves 2, 6 and 9–16 (but 12) lose KE to zonal mean flow with the result that these waves acquire energy out of the flow, indicating that the zonal belt is of barotropical instability. This is probably attributable to the non-inclination of the trough and ridge near the equator. It can , therefore, be assumed that within 20°N–5°S there is a subbelt 20°–10°N of barotropical stability of intense easterly flow, which conceals the barotropical instability



(a) 20°N-5°S
Unit: $m^2s^{-2}day^{-1}$

(b) 20°N-5°S
Unit: $m^2s^{-2}day^{-1}$

(c) North of 30°N
Unit: $10^{-3}erg\cdot s^{-1}cm^{-2}hpa^{-1}$

(d) 15°N-15°S
Unit: $10^{-5}m^2s^{-3}$

(e) 10°N-5°S
Unit: $m^2s^{-2}day^{-1}$

(f) 15°N-15°S
Unit: $10^{-5}m^2s^{-3}$

Fig. 4. The loss and gain of barotropical energy among waves of 4 different scales.

of its adjacent subbelt 10°N-5°S. From the chart in Fig. 2 it can also be seen that *in IbWW waves 1, 3–7, 10 and 14 gain KE from all the others*, wave 1 obtaining most and wave 2 losing most. However, as shown in Fig. 3, the IbWW feature over the belt 10°N-5°S is quite different, indicating that waves 4-5 gain and all the others (except wave 15) lose KE, with waves 3 and 8 deprived more.

If we take a look at the characteristics shown in Figs. 2 and 3, we can see that between 20°N-5°S wave 2(1) is the principal source (sink) of KE. It is of interest that Saltzman et al. (1964) shows a similar result out of the calculations of the decade-averaged mid-latitude KE exchange for 1955–1964, yet indicating that wave 1 serves as the sink of energy that comes mainly from the other waves via IbWW.

A noticeable fact is found *in* wave 3. As indicated *in* Fig. 4a (*denoting* 20°N-5°S), it gains KE via IbWW and loses through IbFW and the reversal is true for 10°N-5°S. The results shown in Fig. 4b, d, (for low) and c (for middle latitudes) are identical with that in 4a and the 1972 result is similar to the 1982 calculation for 10°N-5°S (see Fig. 4c).

From Fig. 4 it can also be seen that, for the summer of 1967 and 1972, waves >4 *gain* enormous KE by virtue of both IbWW and IbFW; in the belt 20°N-5°S for 1982 and 1983 they obtain substantial KE only through IbWW and lose it via IbFW. They lose the energy in either of the interactions over 10°N-5°S for 1982, a result identical with that by Saltzman *et al. (1964).*

C_N as the sum of calculation error and Reynolds molecular viscosity stress will not be taken into account at wave$>$7 in view of the considerable growth of the error. Yet the stress plays a significant part in the dissipation of energy for waves 1–6, particularly for wave 1.

V. MAINTENANCE OF APE OF ZONAL MEAN FLOW AND WAVES

Table 2 indicates that zonal mean flow acquires APE by means of non-adiabatic heating and conversion of KE of its own, *while in IbFW APE is transferred from the flow to waves.* It is worth noting that all terms in the APE equations for zonal mean flow have minute values, all smaller by nearly 1–2 orders than those for wave 1 except the IbFW term.

Also, it is evident that for practically all the waves non-adiabatic heating makes their APE *grow; the IbFW gives rise to a small increase in APE of the waves and a greater* growth in waves 1–2; the IbWW causes waves 1 and 3 to lose their APE, with very little effect on wave 2; almost all the waves have their APE transformed into KE of their own, more important being waves 1–6.

The N–S directed component of the lateral boundary convergence in IbWW (N_{AB1}) has a significant influence upon waves 1–3, and through the convergence waves 1 and 3 attain and wave 2 loses APE while the vertical component results in the increase of APE of waves 1 and 2, with the growth much higher in the former than in the latter. This indicates that waves *1 and 2 in the 200-hPa easterlies* possess a mechanism for getting APE from the layers over-and underlying and the lateral boundary convergence does not make significant contribution to APE of waves of smaller scale.

VI. CYCLE OF ENERGY

Following the above analyses, we construct a diagram showing the cycle of KE and APE (Fig. 5). The processes for the cycle can be summarized as follows:

1) The non-adiabatic heating of various scales serves as the main source of energy for

Table 2. As in Table 1, except for in (14) and (16)

Wavenumber K	N_A	M_A	K_A	N_{AB_1}	N_{AB_2}	H_A
0	0.0	−0.13	0.09	0.0	0.0	0.01
1	−2.23	0.18	−4.92	0.47	1.43	5.07
2	1.01	0.07	−0.89	−0.52	0.26	1.07
3	−1.14	0.01	−0.22	0.71	0.32	0.59
4	−0.07	−0.00	−0.11	−0.05	0.02	0.21
5	0.18	−0.02	−0.23	0.00	−0.03	0.10
6	0.01	0.00	−0.22	−0.08	−0.05	0.34
7	0.19	−0.00	−0.02	−0.05	−0.12	0.10
8	0.02	−0.01	0.13	−0.00	0.01	−0.15
9	−0.15	−0.01	−0.10	0.02	0.06	0.18
10	−0.04	−0.01	−0.09	0.00	−0.01	0.15
11	−0.02	−0.00	−0.08	−0.00	0.01	0.09
12	−0.00	−0.00	−0.10	−0.03	0.00	0.13
13	0.02	−0.00	0.03	−0.03	0.01	−0.03
14	0.09	−0.02	0.02	−0.04	−0.03	−0.02
15	0.10	−0.01	−0.04	−0.04	−0.02	0.01
16	0.02	−0.00	−0.00	−0.04	0.01	0.01

the upper-tropospheric easterlies, with the thermal difference between land and sea making predominant contribution to APE of waves 1 and 2. A further source is the boundary effect including the boundary-pressure work and IbWW lateral boundary convergence, and

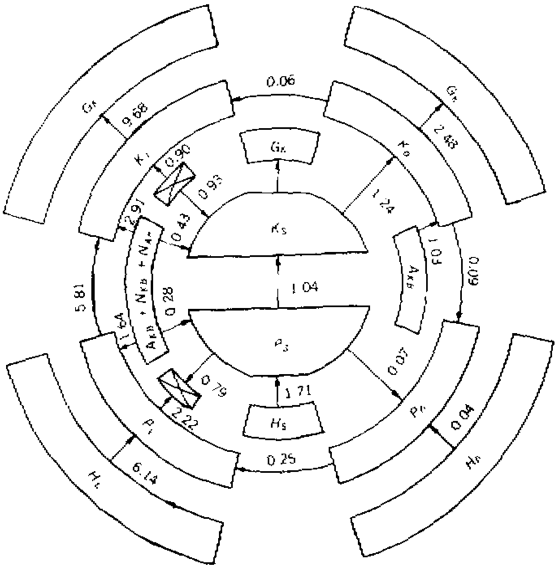


Fig. 5. Maintenance of energy at 200 hPa from July 1 to September 2 (64 days altogether), 1982 over 20°N–5°S. Unit: m^2s^{-2} day^{-1}.

⊠ represents non-linear effect and the subscripts L, s, o indicate waves 1–2, 3–16 and zonal average, respectively.

the sink is represented by the dissipation of KE of diverse-scale waves caused by Reynolds molecular viscosity stress;

2) For the mutual transformation of APE and KE of waves, the conversion into KE is predominant, especially in waves 1 and 2, whereas the change of KE to APE is accomplished only in a very small amount in zonal mean flow;

3) In IbFW the flow loses its APE but gains KE and in IbWW waves 1–16 lose APE and gain KE;

4) An unbalance in the energy exchange due to IbWW occurs in the transfer of barotropical energy obtained by Kanamitsu et al. (1972), Krishnamurti et al. (1978) and Lin Hai (1987) (See Fig. 4). In our calculation, however, waves 1–16 lose APE and gain KE with no exception. We cannot help taking interest in the course that APE and KE follow in their transfer and conversion. The following is possible: through IbWW waves 1–16 give their APE to waves >16, which is then converted into KE of their own and at length the KE is transferred back to the original waves.

VII. LOW-FREQUENCY OSCILLATION MECHANISM OF KE OF WAVES

In order to determine what distinctive periods exist in the oscillation of KE of waves of different scales, the time-sequence power spectrum is computed of KE of zonal waves $K=0$–8. Results indicate the principal periods as follows (figure omitted): roughly 40 and 20 days for the zonal mean flow: 14–20 for wave 1; 20 and 7 or so for wave 2; about 10 for waves 3 and 4; 6–8 for waves 5 and 6; around 20 for wave 7; approximately 10 for wave 8. In view of the fact that tropical flow has in itself such conspicuous periods of oscillation as ca. 40, 20 and 14 days, it can be proved that these periods are intimately related to those found for the KE oscillation in the present study.

Since the KE possessed by zonal mean flow and extra-long waves takes a substantial proportion in the total KE of upper-tropospheric easterlies, special attention is directed towards the oscillation mechanisms of the 40-day period for zonal mean flow and the 20-day one for waves 1 and 2.

Filtering is carried out by using the band-pass filtering technique (see Murakami, 1979). For zonal mean flow the KE band width is taken as 35–45 and for waves 1 and 2 as 15–25 days, and a filtering curve is thus obtained (figure omitted). Then with the aid of the curve, each cycle of the KE oscillation is divided into 4 phases: phase I for low-valued KE, where $E_K=(E_K)_{\min}\times 85\%$; phase II for its gorwth; phase III for its high values, where $E_K-(E_K)_{\max}\times 85\%$; phase IV for its decrease. Finally, the averages of the terms in (11) and (12) for each phase are found through calculation and composited averaging is done of these cycles. It is noted that during the 64 days for waves 1 and 2 three cycles are obtained and for zonal mean flow only one observed because of the length of the oscillation period taken. The results are summarized in tables 3, 4 and 5. It is evident that $\partial E_K/\partial t$ is positive at phase II and negative at IV, with very small absolute values for the other phases. It can also be seen that the whole process of the KE oscillation can not be accounted for by any individual or any set of terms because such oscillation is the result of the combined effect of all these terms. On the other hand, for the different phase some of the terms make greater contribution. For the 40-day period of zonal mean flow, IbFW (frictional dissipation) is the main factor for increasing (decreasing) KE, and therefore this period depends chiefly on the non-linear barotropical effect within easterlies. For waves 1 and 2, however, barotropical transfer and baroclinic and boundary effect come into play. As far as the 20-day period of wave 1 is concerned, A_{KE}, A_K and N_K (G_K and N_{KB1}) are the major factors for increasing (decreasing) KE and for wave 2, A_{KB} and A_K (N_{KB1} and G_K) make KE grow (decay). Therefore, in any dynamic model dealing with low-frequency oscillation, it is necessary to take into account non-linear barotropical and baroclinic effect, and mid-and low-latitude IbFW and IbWW.

Table 3. The Mechanism of the 40-day KE Oscillation Period of Zonal Mean Flow. Unit: $m^2\ s^{-2}\ day^{-1}$

Phase	I	II	III	IV
Period (day)	13-19	20-31	32-38	39-51
trend	min.	upward	max.	downward
$\partial E/\partial t$	−0.02	0.83	0.03	−0.80
M_K	1.94	1.31*	2.12	0.33
A_K	−0.79	0.46	0.59	0.05
A_{KB}	1.97	0.51	2.60	2.27
G_K	−3.14	−1.45	−5.28	−3.90*

* Values of the main factors (the same below).

Table 4. The Mechanism of the 20-day KE Oscillation Period of Wave 1, with the Unit as Indicated in Table 3

Phase	I	II	III	IV
Period	10-12	13-18	19-21	22-27
(day)	28-31	32-39	40-44	45-50
trend	min.	upward	max.	downward
$\partial E/\partial t$	0.05	1.91	−0.10	−2.77
N_K	3.11	3.34*	3.08	3.54
M_K	0.40	0.79	1.24	1.69
A_K	5.99	3.45*	5.20	6.97
N_{KB1}	−3.50	−2.67	−1.92	−5.74*
N_{KB2}	0.82	1.02	1.06	−0.03
A_{KB}	1.78	4.34*	2.21	2.06
G_K	−8.55	−8.38	−10.97	−11.26*

Table 5. As in Table 4 except for Wave 2

Phase	I	II	III	IV
Period	10-12	13-18	19-21	22-27
(day)	28-31	32-39	40-44	45-50
trend	min.	upward	max.	downward
$\partial E/\partial t$	−0.02	0.41	0.01	−0.46
N_K	−1.37	−0.60	−6.63	−0.64
M_K	−0.05	−1.02	−1.96	−0.84
A_K	0.94	1.03*	1.95	1.25
N_{KB1}	−0.61	−0.96	−0.09	−3.98*
N_{KB2}	0.29	0.30	0.64	0.43
A_{KB}	1.43	2.73*	5.75	5.30
G_K	−0.65	−1.07	0.35	−1.98*

VIII. CONCLUDING REMARKS

From the above calculations and analyses we have arrived at the following conclusions:

i) The principal source of energy for maintaining tropical upper-tropospheric easterlies is the E-W land-sea thermal difference, which produces APE for extra-long waves that is then converted into KE for maintaining the easterlies by means of E-W direct thermal circulation. On the other hand, boundary pressure work is also a substantial contribution to the maintenance of KE of easterlies. And the main sink is frictional dissipation and non-linear transport of KE to westerlies.

ii) Calculations of the non-linear transport of barotropical energy indicate that for waves 1-16, in general, easterlies are barotropically stable but a barotropically unstable belt is observed near the equator.

iii) Zonal mean easterly flow and waves 1-2 have the major KE oscillation periods of 40 and 20 days, respectively, and the primary physical mechanism for oscillation is the transfer of barotropical energy; conversion of baroclinic energy, and boundary effect. Therefore, interaction between mid-and low-latitude should be considered in addition to baroclinic effect in any dynamical model for quasi-periodic oscillation phenomena in the atmosphere.

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