Winter Monsoonal Surges over East and Southeast Asia¹

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(Manuscript received 19 October 1978, in revised form 22 January 1979)

Abstract

Some of the characteristic features of monsoonal surges during the northern winter were investigated using wind and temperature data over China, the East and South China Seas, and the extratropical and tropical Western North Pacific during the period 16 November 1970 through 15 March 1971. Low-level monsoonal surges are most pronounced over the East China Sea, with dominant spectral peaks around 4.3–4.6 days for meridional winds, temperature, vertical velocity, and heating rate below about 700 mb. For 4.3–4.6 day oscillations in the lower troposphere, the cospectrum between temperature and meridional winds is positive (northward sensible heat transport), while temperature is negatively correlated with vertical p -velocity (local kinetic energy generation). In comparison, the South China Sea is characterized by dominant spectral peaks at 6.0–6.7 days, with secondary maxima occurring around 4.3–4.6 days. The coherence between the East and South China Seas is generally higher than 0.5 in the lower troposphere. Strong monsoonal surges over the South China Sea coincide with above normal ascending motions and consequent condensation heating throughout the troposphere.

Another feature of interest in computed spectra is the existence of marked peaks around 20–30 days, which are clearly separated from short-period (4–6 day) oscillations associated with monsoonal surges. Composite 200 mb maps for 20–30 day filtered geopotential data depict eastward phase propagation ($\sim 4\text{--}10\text{ m s}^{-1}$) in tropical latitudes, which is most distinctly defined over the Bay of Bengal-Malaysia-South China Sea region. This is opposed to near zero phase speed (standing) at higher latitudes around $50^{\circ}\text{--}60^{\circ}\text{N}$. It appears that wet (dry) spells near Malaysia occur in association with the passage of eastward propagating ridge (trough) systems. During wet (dry) spells, the atmospheric temperature becomes warmer (colder) than usual with above (below) normal ascending motions, as confirmed from a large negative cospectrum for 20–30 day oscillations between these two quantities. Phase changes between wet and dry spells over the South China Sea region are closely related to long-period monsoonal fluctuations over the East China Sea and its vicinity. The coherence for T and v between these two regions is substantially large (>0.5) for 20–30 day period oscillations, with the South China Sea lagging behind the East China Sea by about 1–2 days.

1. Introduction

During winter (December–February), rainfall amounts are substantially larger over the Malaysia-Indonesia region as compared to other tropical regions. This region can be identified as a primary energy source region for the winter monsoon circulation. Rainfall is also large off the east coast of Taiwan and the Philippines along about 125°E (refer to Fig. 1 of Murakami, 1978). Un-

doubtedly, this is associated with winter monsoonal surges bursting out of the Siberian high, and occasionally reaching as far south as the equator and possibly beyond. Murakami (1978) showed that a distinct region of 200 mb velocity potential outflow lies centered near 5°S , 130°E , indicating that the updraft portion of the winter monsoon circulation is located over Indonesia. Upper outflow from this region is directed both northward and southward. Thus, a planetary-scale, meridionally oriented, vertical circulation (local Hadley cell) is implied with rising motion in the region of heaviest equatorial rainfall over

¹ Contribution No. 78–7, Department of Meteorology, University of Hawaii.

Fig. 1a Surface maps at 00 GMT (top) and 12 GMT (bottom), 12 December 1970, prepared by the Japan Meteorological Agency. Heavy dashed lines indicate regions where surface wind speeds exceed 20 knots.

the Malaysia-Indonesia region, and downdrafts over Siberia and south Australia. In short, the local Hadley cell over East Asia is characterized by surface monsoonal surges from North China and Siberia, with vigorous upper tropospheric return flow from the equatorial maritime continent.

In general, surface cold surges from the Siberian high are preceded by the development of extratropical cyclones over the East China Sea region (Fig. 1a, top), where air mass modification is quite active and the supply of moisture and sensible heat is abundant over the underlying warm Kuroshio current. AMTEX was a short

Fig. 1b As in Fig. 1a except for 00 GMT, 13 December (top) and 14 December 1970 (bottom).

term, regional observational effort to investigate the role which energy supplied from the sea surface plays in the generation and development of East China Sea disturbances (Ninomiya, 1975; Ninomiya and Akiyama, 1976; Nitta, 1976; Kung, 1977; Soliz, 1977; Henry and Thompson, 1978). Based on a spectral analysis of AMTEX '68

(February, 1968) data, Nitta *et al.* (1973) confirmed the existence of pronounced oscillations (4–6 day period) in the intensity of cold surges over the East China Sea. During AMTEX, detailed measurements of the turbulent fluxes of momentum, heat, and water vapor were made to identify the dynamical and thermal structure

of low-level monsoonal surges over the East China Sea (refer to Collected Scientific Papers of the AMTEX, No. 1 to 3).

When an E.C.S. cyclone interacts with an approaching upper trough, it intensifies into a mature, large-scale, extratropical cyclone (Fig. 1a, bottom). These cyclones then move away east-northeastwards with strong northerly winds bursting out from the cold continent. The monsoonal surge frequently intensifies progressively equatorward and reaches as far south as the South China Sea. In Fig. 1b (top), note the existence of a marked band of strong northerlies (>20 kts) off the east coast of the Asiatic continent, extending from the East China Sea, through Okinawa and Taiwan, to the South China Sea.² Perhaps, this band corresponds to the low-level branch of the local Hadley cell. Association of northerly winds with cold temperatures implies poleward sensible heat transport via the low-level Hadley circulation (monsoon surges).

The majority of occasions of heavy rainfall over Malaysia and the southern South China Sea are due to tropical disturbances developing along the low-level, near equatorial trough around 5° – 10° N. These South China Sea disturbances are possible when the northeast trades are sufficiently deep and broad (Fig. 1b, bottom), surging south-southwestward from north of the South China Sea and eventually merging with the near-equatorial trough and the equatorial westerlies. Rainfall associated with South China Sea disturbances is substantial and occasionally exceeds 200 mm per day. Energy derived from the rains may then act to keep the local Hadley circulation vigorous.

In short, the development of extratropical cyclones over the East China Sea and near-equatorial disturbances over the South China Sea should be viewed as an integral part of the lateral coupling processes (via monsoonal surges) occurring over a fairly extensive region. A comprehension of how regional disturbance development meshes and interacts with the entire winter monsoon system is crucial to the success of the

Winter Monsoon Experiment in December 1978. It is important to investigate the temporal changes and regional differences in low-level monsoon surges to which the development and structure of extratropical cyclones and near-equatorial disturbances are related. This is elaborated on in Sections 3, 4, and 5.

An inspection of surface weather maps [Fig. 1b (bottom), for example] indicates an approximate zonal wavelength of 4,000–5,000 km for eastward propagating midlatitude systems. In Fig. 1b (bottom), we see the presence of a trough extending southwestward from east Siberia to central China. South of this trough is a pronounced high pressure cell centered near 30° N, 120° E, and accompanied by strong (weak) northeast surges over the South (East) China Sea. The cold high cell, previously located over Siberia (Fig. 1a, top), then moves eastward across the western Pacific. This is probably indicative of transporting sensible heat westward from the western Pacific (warm water) to the Siberia-China region (cold continent). We have so far shown that the surface high pressure system over the Asiatic continent is not steady but fluctuates substantially in its location and intensity in association with the passages of eastward propagating disturbances in the midlatitude westerlies. Near Japan, the jet stream is extremely strong with large westerly vertical shear, probably reflecting active local baroclinic instability. In contrast, equatorward of about 20° N over the South China Sea and the western North Pacific, tropospheric winds are predominantly easterly with weak vertical shear, congruent with the small meridional temperature gradient over that region. Charney (1969), and Bennett and Young (1971) showed that wave disturbances could not, theoretically, pass through equatorial regions where the basic zonal winds are easterly relative to the phase speed of incident extratropical waves. On the other hand, Fig. 1 strongly suggests a form of atmospheric teleconnection between midlatitude westerly systems and equatorial easterly regimes. Further study is needed to investigate this important lateral coupling process which occurs during winter.

Despite the fact that violent tropical storms or other vigorous disturbances are extremely rare over the Malaysia-South China Sea region, the winter monsoon circulation undergoes rather drastic changes in association with phase changes between wet and dry spells, which are similar in some respect to “active” and “break” conditions

² Due to the presence of the Himalayas, surges rarely reach west of about 90° E. Surges are much stronger and more frequent east of the Himalayas. Over China, significant pressure rise and temperature decrease occur at the earth's surface. However, surface winds generally remain weak (Fig. 1b, top) possibly due to frictional dissipation. Over Central and South China, monsoonal surges are strongest at 850 mb.

during the summer monsoon. Examination of precipitation data indicates a somewhat longer period than 20 days. When a "dry" monsoon occurs, no near-equatorial disturbances form and rainfall is scanty over the Malaysia-South China Sea region. Also during "dry" spells, the upper subtropical ridge system is ill defined, displaced southward from its normal position near the Philippines, and occasionally disappears from the Malaysian region when upper troughs in the westerlies penetrate southward to equatorial latitudes (Liang, 1974). Upper equatorial winds which should normally be easterly are replaced by strong southwesterlies. It appears that long-period monsoonal fluctuations with "wet" and "dry" phases are a very important part of the winter monsoon circulation regime. However, there remain more questions than answers as to their role in maintaining the winter mean monsoon circulation. Their causes are unknown. This alone indicates the importance of further study in this particular area of the monsoon problem. In Section 6, an attempt is made to detail the structural features of long period (>20 day) monsoonal fluctuations and to investigate their interaction with low-level monsoon surges within a range of 4-6 day periods.

2. Data and computational procedures

The structure of winter monsoonal surges and the character of disturbance activity differ significantly with different regions over East Asia. Fig. 2 shows five regions selected in this study. Upper-air stations used in each region are as follows:

Region 1: 4 stations; Hong Kong (22°N, 114°E), Wuhan (31°N, 114°E), Shanghai (31°N, 122°E), and Shantou (23°N, 117°E).

Region 2: 6 stations; Taoyuan (25°N, 121°E), Shanghai (31°N, 122°E), Mosulpo (33°N, 126°E), Kagoshima (32°N, 131°E), Kadena (26°N, 128°E), and Ishigakijima (24°N, 124°E).

Region 3: 5 stations; Minamidaitojima (26°N, 131°E), Kagoshima (32°N, 131°E), Shionomisaki (34°N, 136°E), Hachijojima (33°N, 140°E), and Chichijima (27°N, 142°E).

Region 4: 5 stations; Saigon (11°N, 107°E), Da Nang (16°N, 108°E), Xishadao (17°N, 112°E), Clark (15°N, 121°E), and Kota Kinabalu (6°N, 116°E).

Region 5: 4 stations; Korrer (7°N, 135°E), Yap (10°N, 138°E), Guam (14°N, 145°E), and Truk (8°N, 152°E).

The data used here consists of twice-daily wind and temperature at standard pressure levels; the surface, 850, 700, 500, 300, and 200 mb during 1970-71 winter (16 November 1970-15 March 1971, totaling 120 days). Spurious values were eliminated by computing the standard deviation of each time series and then discarding those values which fell more than four standard deviations away from the mean. Missing and discarded values were linearly interpolated using the accepted data points. The percentage of missing data increased with decreasing pressure. Temperature data was much less likely to be missing than wind data. Over regions 1, 2, 3, and 5, the percentage of missing wind data was less than 5 percent in the lower troposphere (below 500 mb), but was nearly 10 percent at 200 mb. At Da Nang, Saigon, and Kota Kinabalu, no 12 GMT observations were made during 1970-71 winter. In addition, no surface wind data were available for Saigon, Da Nang, and Xishadao. Thus, data coverage and quality remain a problem over region 4.

We applied a first-order Taylor's series expansion about the gravity center (x_g, y_g) over each region. For example, u can be expressed as follows:³

$$u_i = u_g + \left(\frac{\partial u}{\partial x} \right)_g (x_i - x_g) + \left(\frac{\partial u}{\partial y} \right)_g (y_i - y_g) \quad (1)$$

where (x_i, y_i) designates the locations of the stations used in each region, and the suffix "g" refers to the quantity at the gravity center.

Area-averaged u_g , $(\partial u / \partial x)_g$, and $(\partial u / \partial y)_g$ may be determined from a least square method as follows:

³ A list of symbols is given in the Appendix.

Fig. 2 Regions 1 to 5, to which equations 1 to 10 were applied.

$$E = \sum_i W_i \left[u_i - u_g - \left(\frac{\partial u}{\partial x} \right)_g (x_i - x_g) - \left(\frac{\partial u}{\partial y} \right)_g (y_i - y_g) \right]^2, \quad (2)$$

where

$$W_i = \frac{R^2 - d_i^2}{R^2 + d_i^2}; \quad R = 1500 \text{ km},$$

$$d_i = \sqrt{(x_i - x_g)^2 + (y_i - y_g)^2}.$$

The above computational procedures (1) and (2) were tested by comparing a known function to estimated values at the gravity center. The streamfunction chosen for this test is described by

$$\phi = \cos\left(\frac{2\pi}{X}x\right)\cos^2\left(\frac{2\pi}{Y}y\right),$$

in which $X(Y)$ denotes the zonal (meridional) wavelength. We treated four cases with $X=2,000$ to $5,000$ km. In these four cases, we have let $Y=2,000$ km. The analytical u and v values were evaluated for the latitude and longitude positions of radiosonde stations used in each region. These u and v values were then submitted to computational methods (1) and (2) to estimate u_g and v_g , and their derivatives at the gravity center. The differences between estimated values and corresponding analytical values at the gravity center are due to errors introduced by schemes (1) and (2). Computed results indicate that errors are sufficiently small for $X \geq 3,000$ km over regions 1 to 3. We, therefore, felt that the station distribution and horizontal dimensions of these regions might be adequate to resolve the characteristic features of eastward propagating baroclinic waves with wavelengths of about $4,000$ – $5,000$ km (see Fig. 1). On the other hand, a lack of radiosonde data in the central part of the South China Sea (region 4) appears to inhibit accurate determination of the structural features of near-equatorial disturbances with wavelengths of $2,000$ km or less. Also, a higher degree of horizontal resolution in the station distribution is certainly desirable for the study of monsoonal surges over regions 4 and 5. Thus, the computed results obtained over these regions should be considered preliminary until tested with a more complete data set during the Winter MONEX (December 1978–February 1979).

The mean vertical p -velocity over each region was obtained by

$$\omega = \omega_{p_0} + \int_p^{p_0} \nabla \cdot \mathbf{v} \, dp, \quad (3)$$

where

$$\nabla \cdot \mathbf{v} = \sec \phi \left[\left(\frac{\partial u}{\partial x} \right)_g + \left(\frac{\partial v}{\partial y} \right)_g \right] - \frac{\tan \phi}{a} (v)_g, \quad (4)$$

$$\omega_{p_0} = \left(\frac{\partial p_0}{\partial t} \right)_g + \sec \phi \left[u_g \left(\frac{\partial p_0}{\partial x} \right)_g + v_g \left(\frac{\partial p_0}{\partial y} \right)_g \right]. \quad (5)$$

A lack of surface wind data at Saigon, Da Nang, and Xishadiao precludes the application of the least square technique to estimate surface winds u_g and v_g over region 4. Hence, we assumed $\omega_{p_0} = (\partial p_0 / \partial t)_g$ over that region.

In computing ω , it was assumed that the vertical velocity converges to zero at 100 mb. Divergence estimated from eq. (4) was then modified by subtracting δ , as defined by

$$\delta = \sigma \left[\omega_{p_0} + \int_{100}^{p_0} \nabla \cdot \mathbf{v} \, dp \right] / \int_{100}^{p_0} \sigma \, dp,$$

where σ denotes the standard deviation of divergence at the pressure level P ; 100 mb divergence was extrapolated from neighboring values at 200 mb and 300 mb. Generally, the correction δ becomes larger with decreasing pressure.

Daily T , u , and v data were separated into the time (120 day) mean and departure from it. For example, u was expressed as

$$u = \bar{u} + u'. \quad (6)$$

Thus, u' represents the transient component of u .

The thermodynamic energy equation was expressed in Mercator coordinates as follows:

$$\frac{dT}{dt} = \frac{Q}{c_p}, \quad (7)$$

where

$$\frac{dT}{dt} = \frac{\partial T_g}{\partial t} + \sec \phi \left[u_g \left(\frac{\partial T}{\partial x} \right)_g + v_g \left(\frac{\partial T}{\partial y} \right)_g \right] + \omega_g \left(\frac{\partial T_g}{\partial p} - \frac{\kappa}{p} T_g \right). \quad (8)$$

In the above equation, area-averaged u_g , v_g , ω_g , T_g , $(\partial T / \partial x)_g$, and $(\partial T / \partial y)_g$, determined at the gravity center, are largely related to synoptic-scale disturbances; namely, they are not associated with subdomain eddies. Thus, dT/dt measured from eq. (8) corresponds to what is generally called an "apparent large-scale heating" and conventionally defined as Q_1 by many authors, e.g., Yanai *et al.* (1973) and Nitta (1976). The dT/dt -heating rate computed from eq. (8) was compared with previous estimates by Kubota (1970) and

Ninomiya (1975).

The kinetic energy equation may be written as

$$\frac{dK}{dt} = -\sec \phi \left(u \frac{\partial \Phi}{\partial x} + v \frac{\partial \Phi}{\partial y} \right) + F(k), \quad (9)$$

where $F(k)$ represents the frictional dissipation. The derivative dK/dt is defined according to

$$\begin{aligned} \frac{dK}{dt} = & \frac{\partial K_g}{\partial t} + \sec \phi \left[u_g \left(\frac{\partial K}{\partial x} \right)_g + v_g \left(\frac{\partial K}{\partial y} \right)_g \right] \\ & + \omega_g \frac{\partial K_g}{\partial p}, \end{aligned} \quad (10)$$

The procedure for spectral estimates of time series data as described by Maruyama (1968) was employed in the present study. Before the application of spectrum analysis, the root-mean-square linear trend was removed from twice-daily T , u , and v data of 120 days. A maximum lag of 30 days was used to compute the covariance between two parameters. The cospectrum, quadrature-spectrum, coherence, and phase difference between two variables were also computed following Maruyama's technique. Computed results are presented in Section 5.

3. Season mean fields

In this section, we investigate differences in season mean fields between various regions. Table 1 presents the 120-day season mean T , u , v , ω , dT/dt , and dK/dt , averaged over each region at the surface, 850, 700, 500, 300, and 200 mb, respectively. Also shown are season mean $\overline{u'v'}$, $\overline{v'K'}$, $\overline{v'T'}$, and $\overline{\omega'T'}$. A positive value of $\overline{u'v'}$ indicates northward transport of westerly momentum due to transient eddies. Similarly, positive $\overline{v'K'}$ ($\overline{v'T'}$) represents northward transport of kinetic energy (sensible heat). Negative $\overline{\omega'T'}$ is probably due to an increase in transient eddy kinetic energy at the expense of transient eddy available potential energy over a limited region.

Over region 2, strong westerlies prevail in the upper troposphere, exceeding 55 m s^{-1} at 200 mb. Vertical westerly shear as measured by the difference in u between 200 mb and 850 mb is substantially large (53.5 m s^{-1}), conceivably indicating active baroclinic instability. Region 2 meridional winds are southerly above 500 mb in contrast to the lower troposphere below 700 mb where they are northerly with a maximum of -3.0 m s^{-1} at 850 mb. This confirms the existence of southerly vertical shear, which is congruent with temperature contrasts between the cold Asiatic continent and the warm Pacific

Ocean. Comparable features in the vertical v -distribution are also found over regions 1 and 3.

The vertical velocity ω is positive (descending) in the deep atmospheric layer between 200 mb and 850 mb, over extratropical regions 1 to 3. These regions are located west of a pronounced upper trough which is anchored to the east of Japan in winter. Descending motion is strongest over region 1 (China) with the maximum (>80 units) around 700–500 mb (nondivergent level).

Based on AMTEX '74 data (February 14–28), Ninomiya (1975) showed that heat sources exist in the lower troposphere (below 700 mb) over a region almost identical to our region 2 (East China Sea). His value for 15-day mean heating amounted to 8.3°C per day between 850 mb and the surface. This is approximately twice as large as our season (120 day) mean value of about $4.0 \times 10^{-5}^\circ\text{C s}^{-1}$ for region 2 below 850 mb. In this vicinity, strong low-level northerlies and large temperature contrasts between cold air and underlying warm water result in a large sensible heat flux near the earth's surface. Substantial low-level heating is also encountered over region 3, due not only to a large surface sensible heat flux but also to rather heavy winter precipitation associated with actively developing baroclinic disturbances in the area.

In region 2, season mean dK/dt is positive and very large in the upper troposphere, reaching a maximum (1,553 units) at 300 mb where a strong jet stream is encountered. This is contrasted with dominant negative $\overline{\omega'T'}$ in the lower troposphere between 850 mb and 700 mb, implying large energy generation within the layer. Perhaps, then, excess energy generated most effectively near 850–700 mb, is transported upward to maintain high kinetic energy near jet stream levels (300–200 mb). Interesting, region 3 dK/dt is negative and extremely large above 500 mb. This is due to an overwhelming effect along the eastern boundary where strong westerlies export large amounts of kinetic energy into the central Pacific. Region 3 also exports kinetic energy poleward as indicated by the large $\overline{v'K'}$ flux at 200 mb.

It is worth noting that over region 2, $\overline{v'T'}$ -flux is large and directed northward (positive) near 850 mb. This is clearly the result of a strong association between low-level monsoonal surges and cold temperatures, as has been found in previous studies by Ninomiya (1975) and many others. Also, strong low-level northerly surges are generally correlated with stronger than usual

westerlies, resulting in southward westerly momentum transport ($\overline{u'v'} < 0$) in the lower troposphere. In contrast, momentum transport is directed northward above 300 mb. Here, upper wave disturbances exhibit a distinct NE-SW tilt.

The computed results over region 4 (South China Sea) may be summarized as follows: (1) as in regions 1 to 3, season mean v is northerly below 500 mb as opposed to southerly above 300 mb, (2) in contrast to extratropical regions 1 to 3, the vertical motion is upward throughout the troposphere with a maximum near 500 mb, which is consistent with heavy precipitation due to deep convection in winter, (3) at all levels dT/dt is positive (heating) and substantial (> 2 units), essentially in agreement with Kubota's (1970) finding, and (4) energy generation appears to be most pronounced around 300 mb, as indicated by the large negative $\overline{\omega'T'}$ value.

Compared to region 4, dT/dt over region 5 exhibits only small heating (0.3–0.4 units) below 700 mb. Therefore, it appears that convective activity is much less pronounced over region 5 (equatorial Western Pacific) than over region 4 (South China Sea) during winter. This is reflected by weak ascending or even descending motion above 300 mb in region 5 compared to a relatively strong updraft at all levels over region 4.

4. Changes in regional circulations

The winter monsoon appears to be an extremely complicated sequence of events, rather than a

simple steady motion as shown in Table 1. Massive changes occur in the distribution of weather and circulation systems in association with changes in intensity of low-level monsoonal surges. This is illustrated in Figs. 3 and 4, which depict changes in u and v at 850 mb, and surface pressures p_0 , over regions 2 and 4, respectively. An inspection of Fig. 3 indicates an approximate period of 4 to 6 days in v fluctuations over region 2. Changes in region 4 meridional winds are not as systematic as in region 2. The amplitude of v fluctuations appears to be larger than u variations over region 2 (Fig. 3), while the reverse is true over region 4 (Fig. 4). Over that region, an approximately inverse relationship exists between changes in 850 mb v and p_0 , i.e., $\overline{v'p'_0} < 0$. Perhaps, this implies southward wave-energy flux. In Fig. 4, we also note an approximate in-phase relationship between u and v fluctuations, resulting in northward westerly momentum transport (see Table 1).

To facilitate further study, we have selected two contrasting periods designated as "S" and "W" phases, respectively. In doing so, we first computed the standard deviation σ of 850 mb v over each of the selected regions. Twice-daily 850 mb v data was then compared with the values $\bar{v} \pm \sigma/2$ as follows: "S" phase was designated for 850 mb v data when it deviates more than $\sigma/2$ below the season mean $\bar{v}$, i.e., $v < \bar{v} - \sigma/2$ (see Figs. 3 and 4). Conversely, 850 mb v data was considered to be a "W" phase when $v > \bar{v} + \sigma/2$. The

Fig. 3 Time series of u (top) and v (middle) at 850 mb, and surface pressure p_0 (bottom), over region 2 from 16 November 1970 to 15 March 1971 (120 days).

Fig. 4 As in Fig. 3 except for region 4.

Table 1. Season mean $T(^{\circ}\text{C})$, $u(\text{m}\cdot\text{s}^{-1})$, $v(\text{m}\cdot\text{s}^{-1})$, $\omega(10^{-5}\text{ mb}\cdot\text{s}^{-1})$, $dT/dt(10^{-5}\text{ }^{\circ}\text{C}\cdot\text{s}^{-1})$, $dK/dt(10^{-5}\text{ m}^2\cdot\text{s}^{-3})$, $u'v'(\text{m}^2\cdot\text{s}^{-2})$, $v'K'(\text{m}^3\cdot\text{s}^{-3})$, $v'T'(^{\circ}\text{C}\cdot\text{m}\cdot\text{s}^{-1})$, and $\omega'T'(10^{-5}\text{ mb}\cdot^{\circ}\text{C}\cdot\text{s}^{-1})$, at the surface, 850, 700, 500, 300, and 200 mb, averaged over regions 1 to 5.

	REGION 1						REGION 2					
	P ₀	850	700	500	300	200	P ₀	850	700	500	300	200
T	9.5	4.9	-0.1	-12.4	-33.9	-51.7	10.8	3.4	-2.0	-15.0	-35.7	-51.9
U	-1.6	0.6	9.8	22.4	41.7	48.2	-0.8	1.9	12.0	26.3	48.9	55.4
V	-1.4	-0.7	0.0	0.1	3.8	4.8	-2.7	-3.0	-1.0	0.2	3.7	4.9
W	-1.6	39.7	86.9	84.7	38.7	14.4	-1.9	3.2	10.8	10.5	27.9	21.9
dT/dt	0.9	-0.4	-4.7	-5.0	-6.2	-2.5	4.3	4.0	0.5	-1.5	-2.1	-0.4
dK/dt	2	-11	-51	-264	-530	432	-3	-13	69	447	1553	1140
UV	-0.3	4.4	1.1	-6.3	-1.8	-3.2	-1.9	-3.1	-3.0	-1.5	9.3	23.0
VK	-4	-17	-33	-175	302	546	-10	-71	-60	-51	865	2152
VT	0.5	10.4	6.2	-1.0	3.9	0.3	3.4	13.1	5.7	2.2	1.8	-1.7
WT	1	-32	-92	-52	-88	-6	1	-183	-142	-22	-46	-8
	REGION 3						REGION 4					
	P ₀	850	700	500	300	200	P ₀	850	700	500	300	200
T	12.2	3.2	-2.8	-16.4	-36.5	-50.9	22.1	15.1	8.5	-6.0	-31.4	-53.7
U	0.8	5.8	14.7	29.7	54.2	62.1		-4.3	-3.2	-1.5	-0.8	-2.7
V	-2.0	-2.3	-1.9	0.0	2.1	2.2		-1.9	-0.5	-0.1	1.7	4.9
W	-0.9	4.8	9.6	11.3	8.4	4.4	0	-28.8	-63.6	-65.5	-39.6	-20.7
dT/dt	3.4	2.5	0.3	-1.7	-3.7	-7.1		2.1	3.3	3.2	2.9	2.0
dK/dt	-1	-34	-118	-381	-1637	-2785		-9	-4	-6	9	-3
UV	-1.1	-4.9	0.7	14.8	46.2	35.4		1.3	-1.1	1.5	0.2	-1.3
VK	-3	-54	-73	370	2759	2923		-11	-7	-2	9	86
VT	2.5	7.7	4.0	1.5	10.3	1.7		1.1	-0.1	-0.7	0.0	-0.2
WT	-2	-84	-106	-97	-7	1	0	18	9	-31	-103	-9
	REGION 5						SEASON MEAN					
	P ₀	850	700	500	300	200						
T	27.7	17.8	10.1	-5.0	-30.3	-52.9						
U	-3.6	-7.7	-7.2	-8.0	-6.0	-5.6						
V	-1.2	-0.9	0.0	0.5	0.2	1.9						
W	-0.3	-9.2	-11.5	-2.9	5.9	5.5						
dT/dt	0.3	0.4	0.3	-0.2	-0.5	-0.7						
dK/dt	-1	2	7	9	15	13						
UV	0.3	2.6	5.7	3.7	0.3	-0.6						
VK	-1	-14	-10	-9	-6	17						
VT	-0.1	0.4	0.1	0.2	0.5	0.2						
WT	0	8	17	13	-11	-3						

Table 2. Mean $T(^{\circ}\text{C})$, $u(\text{m}\cdot\text{s}^{-1})$, $v(\text{m}\cdot\text{s}^{-1})$, $\omega(10^{-5}\text{ mb}\cdot\text{s}^{-1})$, $dT/dt(10^{-5}\text{ }^{\circ}\text{C}\cdot\text{s}^{-1})$, and $dK/dt(10^{-5}\text{ m}^2\cdot\text{s}^{-3})$ at standard levels during the “S” and “W” phases, over regions 1 to 3. See text for further information on the definition of “S” and “W”.

REGION 1							S-PHASE (75)							W-PHASE (74)						
p ₀ 850 700 500 300 200														p ₀ 850 700 500 300 200						
T	-1.2	-2.4	-1.8	-0.6	-0.2	0.2	T	1.3	3.0	1.9	0.4	0.1	0.0	T	1.3	3.0	1.9	0.4	0.1	0.0
U	1.4	-0.7	-0.2	1.7	2.8	3.2	U	-1.4	0.9	0.2	-2.1	-3.0	-2.7	U	-1.4	0.9	0.2	-2.1	-3.0	-2.7
V	-0.6	-4.5	-2.9	-1.9	-1.3	0.2	V	0.5	5.1	3.1	2.7	1.6	0.6	V	0.5	5.1	3.1	2.7	1.6	0.6
W	0.1	34.2	68.8	55.4	-0.3	-11.5	W	-0.3	-30.2	-71.1	-93.3	-60.1	-27.4	W	-0.3	-30.2	-71.1	-93.3	-60.1	-27.4
dT/dt	0.7	2.9	-1.9	-1.6	-0.1	0.5	dT/dt	-0.9	-3.2	2.3	4.0	2.3	0.3	dT/dt	-0.9	-3.2	2.3	4.0	2.3	0.3
dK/dt	2	-23	-65	-199	-705	589	dK/dt	-2	22	50	183	43	-332	dK/dt	-2	22	50	183	43	-332

REGION 2							S-PHASE (73)							W-PHASE (75)						
p ₀ 850 700 500 300 200														p ₀ 850 700 500 300 200						
T	-1.9	-3.2	-1.7	-0.9	0.3	0.7	T	2.4	3.5	2.2	1.2	-0.2	-0.4	T	2.4	3.5	2.2	1.2	-0.2	-0.4
U	1.4	1.0	2.6	4.3	3.8	1.8	U	-1.4	-0.6	-2.2	-4.2	-4.6	-3.0	U	-1.4	-0.6	-2.2	-4.2	-4.6	-3.0
V	-1.9	-5.0	-2.9	-1.4	0.2	1.2	V	1.9	5.4	3.1	1.8	-0.6	-2.1	V	1.9	5.4	3.1	1.8	-0.6	-2.1
W	0.0	50.3	77.6	41.6	9.9	9.7	W	-0.3	-65.3	-99.9	-71.8	-25.0	-11.9	W	-0.3	-65.3	-99.9	-71.8	-25.0	-11.9
dT/dt	2.7	3.2	-0.3	0.9	3.0	0.9	dT/dt	-2.6	-2.6	0.2	0.6	-3.3	-0.1	dT/dt	-2.6	-2.6	0.2	0.6	-3.3	-0.1
dK/dt	-2	-59	-19	148	836	32	dK/dt	3	52	25	-30	-706	-198	dK/dt	3	52	25	-30	-706	-198

REGION 3							S-PHASE (75)							W-PHASE (67)						
p ₀ 850 700 500 300 200														p ₀ 850 700 500 300 200						
T	-1.0	-1.8	-1.3	-0.9	0.6	0.8	T	1.9	2.7	1.8	1.3	-0.9	-0.6	T	1.9	2.7	1.8	1.3	-0.9	-0.6
U	1.5	1.3	1.6	2.4	3.0	0.3	U	-1.8	-2.0	-2.7	-4.0	-5.7	-1.9	U	-1.8	-2.0	-2.7	-4.0	-5.7	-1.9
V	-1.1	-4.7	-3.8	-3.3	-2.2	-0.7	V	1.6	5.6	4.0	2.4	0.1	-2.2	V	1.6	5.6	4.0	2.4	0.1	-2.2
W	1.5	30.4	54.2	66.9	34.2	10.3	W	-2.4	-39.8	-64.8	-62.9	-29.0	-9.8	W	-2.4	-39.8	-64.8	-62.9	-29.0	-9.8
dT/dt	1.8	3.4	1.6	0.4	-0.3	-0.1	dT/dt	-2.1	-3.6	-2.2	-0.3	-0.6	-0.4	dT/dt	-2.1	-3.6	-2.2	-0.3	-0.6	-0.4
dK/dt	-6	-70	-86	-122	-510	-265	dK/dt	10	84	104	133	437	-49	dK/dt	10	84	104	133	437	-49

Table 3. As in Table 2 except for regions 4 and 5.

REGION 4							S-PHASE (69)							W-PHASE (60)						
p ₀ 850 700 500 300 200														p ₀ 850 700 500 300 200						
T	-0.1	-0.5	-0.6	0.1	0.1	-0.1	T	0.2	0.7	0.5	-0.2	0.2	0.1	T	0.2	0.7	0.5	-0.2	0.2	0.1
U		-1.0	-0.6	0.2	-1.0	-0.8	U		1.1	0.0	-0.1	1.1	0.7	U		1.1	0.0	-0.1	1.1	0.7
V		-2.3	-0.8	-0.7	0.9	1.2	V		2.3	1.1	0.3	-0.3	-0.5	V		2.3	1.1	0.3	-0.3	-0.5
W	-0.2	-14.8	-24.3	-7.3	-5.0	-3.6	W	0.1	16.7	29.7	22.5	18.8	12.4	W	0.1	16.7	29.7	22.5	18.8	12.4
dT/dt		1.4	1.4	0.4	0.3	0.5	dT/dt		-1.4	-1.5	-1.0	-1.5	-1.3	dT/dt		-1.4	-1.5	-1.0	-1.5	-1.3
dK/dt		-7	1	2	0	20	dK/dt		8	-1	-6	-4	-12	dK/dt		8	-1	-6	-4	-12

REGION 5							S-PHASE (72)							W-PHASE (69)						
p ₀ 850 700 500 300 200														p ₀ 850 700 500 300 200						
T	-0.1	-0.2	-0.1	-0.0	-0.1	0.1	T	-0.2	0.3	-0.1	-0.0	0.1	-0.0	T	-0.2	0.3	-0.1	-0.0	0.1	-0.0
U	-0.3	-1.1	-0.8	-0.5	1.1	1.5	U	0.5	1.3	0.9	0.5	-2.2	-2.3	U	0.5	1.3	0.9	0.5	-2.2	-2.3
V	-0.6	-2.5	-1.5	-0.9	-0.5	0.4	V	1.0	2.5	1.3	1.2	0.4	-0.5	V	1.0	2.5	1.3	1.2	0.4	-0.5
W	-0.1	3.5	11.1	6.1	6.1	6.6	W	0.2	-13.4	-26.2	-27.9	-14.8	-5.6	W	0.2	-13.4	-26.2	-27.9	-14.8	-5.6
dT/dt	0.0	0.1	-0.6	-0.4	-0.4	-0.6	dT/dt	-0.1	0.4	1.2	1.4	0.9	0.4	dT/dt	-0.1	0.4	1.2	1.4	0.9	0.4
dK/dt	-1	2	-2	-2	-8	-14	dK/dt	-0	-0	-4	-8	12	15	dK/dt	-0	-0	-4	-8	12	15

$\sigma/2$ criterion was chosen so that the number of data in each phase was approximately one-third of the total (240) data set. In Table 2 (top), for example, the number of data selected for region 1 is 75 in “S” and 74 during the “W” phase. Averages were then taken to obtain phase means for 850 mb v . Phase means were also computed for all elements, T , u , v , ω , dT/dt , and dK/dt , at all pressure levels with reference to “S” and “W”, which were selected from 850 mb v fluctuations. Utilizing these phase means, it is now possible to compare structural features of different elements in each phase (“S” or “W”), since these phases were prescribed by using only one variable

Table 4. Coherence and phase differences (degrees) for T , u , and v measured at 850 mb (top) and 200 mb (bottom), over regions, 1, 3, 4, and 5, in reference to region 2.

PERIOD		4.3-4.6 DAY				6-6.7 DAY				20-30 DAY			
REGIONS		1	3	4	5	1	3	4	5	1	3	4	5
850 MB	T	0.93 21	0.87 -67	0.59 -43	0.47 -72	0.95 32	0.93 -39	0.64 37	0.45 11	0.97 10	0.92 -9	0.75 -34	0.31 164
	U	0.77 56	0.78 -70	0.53 82	0.68 -137	0.68 21	0.78 -25	0.62 -81	0.54 -150	0.63 -9	0.64 2	0.12 -113	0.50 -138
	V	0.95 43	0.87 -70	0.54 -19	0.52 -154	0.79 32	0.78 -63	0.45 -50	0.41 -158	0.85 5	0.52 -19	0.70 0	0.59 29
200 MB	T	0.63 21	0.84 -28	0.24 -18	0.30 -150	0.92 41	0.87 -38	0.41 175	0.39 -79	0.90 3	0.80 -8	0.51 -58	0.49 -1
	U	0.85 34	0.74 -37	0.47 -143	0.25 168	0.87 11	0.90 -20	0.59 -113	0.41 25	0.94 -4	0.89 -10	0.55 1	0.59 27
	V	0.91 48	0.94 -58	0.52 28	0.41 -119	0.68 44	0.91 -43	0.35 -12	0.32 -29	0.73 2	0.69 -4	0.59 -9	0.43 46

Table 5. Cospectra (multiplied by frequency) between u and v , u and T , v and T , and ω and T , averaged for 20-30 day periods, over regions 1 to 5.

		p_0	850	700	500	300	200
Region 1	$(u'v')$	-0.0	0.4	0.1	-1.4	-1.3	-0.8
	$(u'T')$	-0.3	-0.4	-0.1	-0.7	-0.1	-0.6
	$(v'T')$	0.2	1.6	1.6	-0.2	-0.1	0.3
	$(\omega'T')$	0.0	-13.4	-11.4	10.5	-1.0	-1.5
Region 2	$(u'v')$	-0.2	-0.5	-0.2	-0.3	1.3	5.3
	$(u'T')$	-0.4	-0.6	-1.6	-1.5	-0.0	-0.6
	$(v'T')$	0.5	2.2	0.9	-0.1	-0.6	-0.4
	$(\omega'T')$	0.2	-26.3	-26.9	-1.1	5.6	-0.3
Region 3	$(u'v')$	-0.2	-0.7	0.2	3.6	11.0	9.8
	$(u'T')$	-0.6	-1.9	-1.9	-1.4	2.3	-0.7
	$(v'T')$	0.3	0.7	0.3	-0.2	1.2	-0.5
	$(\omega'T')$	-0.1	-4.9	-11.7	-13.8	10.8	3.7
Region 4	$(u'v')$		0.1	0.4	0.5	0.4	0.1
	$(u'T')$		-0.1	-0.1	-0.1	0.1	-0.0
	$(v'T')$		0.2	0.1	-0.2	0.0	0.0
	$(\omega'T')$	0.0	1.7	3.5	-7.1	-1.8	-0.1
Region 5	$(u'v')$	0.0	0.8	1.5	0.5	-0.3	-0.6
	$(u'T')$	-0.1	-0.0	-0.1	0.2	-0.2	-0.1
	$(v'T')$	0.0	0.1	0.1	-0.1	0.1	0.1
	$(\omega'T')$	0.0	1.1	3.7	2.5	-3.5	-1.1

(meridional wind) at only one level (850 mb), over each region. However, caution must be exercised when comparing conditions in different regions because phases were defined within individual regions.

Phase mean values during “S” and “W” are shown in Table 2 for regions 1 to 3 and Table 3 for regions 4 and 5. For convenience, 120-day season means were removed from phase mean

values. Thus, for example, a negative sign in T represents below season mean temperature. Due to the definition of the two phases, it is only natural to find large negative (positive) values for 850 mb v during “S” (“W”) at all regions.

Table 2 for region 2 shows that during the S-phase, T is slightly warmer than the season mean at 200 mb and 300 mb, while it is colder than average below 500 mb with the coldest

temperature (-3.2°C) at 850 mb. Throughout the troposphere, westerly winds are stronger than usual, accompanied by above season mean descending motion, during "S". Interestingly, dT/dt exhibits large above normal heating between 850 mb and the surface during the S-phase. Henry and Thompson (1978), based on AMTEX '74 data, showed that the rate of heat transfer at the sea surface increases when fresh monsoon surges occur over the East China Sea.

Conditions over region 2 during the W-phase are almost exactly opposite to those in "S". For example, v at 850 mb is 5.4 during "W" as compared to -5.0 during the S-phase. When the season mean 850 mb v of -3.0 m s^{-1} (Table 1) is added, actual meridional winds are southerly (2.4 m s^{-1}) during "W". A comparison between season mean ω and W-phase mean ω reveals that the vertical motion over region 2 becomes upward throughout the troposphere during "W", in contrast to general descending motion for the winter average. In short, significant and systematic changes in circulation regimes occur at all levels in association with the phase changes between "S" and "W", which were defined from 850 mb v -fluctuations. In Table 2, similar systematic changes between "S" and "W" surges are also noted over regions 1 and 3.

Table 3 (top) summarizes computed results over region 4. Compared to extratropical regions 1 to 3, temperature changes between "S" and "W" are much smaller over region 4. At 850 mb where monsoon surges are generally most pronounced, T increases only 1.2°C from "S" to "W". Of particular interest is large above normal ascending motion during "S", with an associated dT/dt -heating rate also well above season mean. Thus, region 4 (South China Sea) is characterized by strong ascending motion and large heating (rainfall) when monsoonal surges are fresh (above average) at 850 mb.

Table 3 (bottom) for region 5 shows a different character in the changes in ω and dT/dt with reference to phase changes from "S" to "W". Here, negative 850 mb v is correlated with positive ω and negative dT/dt at most levels during "S", while positive 850 mb v is associated with negative ω and positive dT/dt above 850 mb during the W-phase. These features are similar, in some respects, to the structural features of so-called "easterly" waves. As observed by Chang *et al.* (1970), these waves are associated with southerlies, ascending motion and wet weather at the rear of troughs, in contrast to northerlies,

descending motion, and dry weather ahead of troughs.

Burpee (1975) and Reed *et al.* (1977) have previously published African wave composites for the GATE period. The same compositing technique was employed in this study to examine phase relationships between different variables, at different levels, over each of the selected regions. Prior to compositing, we selected several cases with systematic 850 mb v fluctuations. For instance, in Fig. 3, strong northerly winds occurred at day 8, followed by southerlies around day 10 and a refreshing of strong northerlies at day 13. Thus, the period from day 8 to day 13 constitutes one cycle of v -fluctuations at 850 mb. The number of cycles chosen for this study was 18 for region 2 and 11 for region 4, as shown by arrows in Figs. 3 and 4. Each twice-daily observation in one cycle was assigned an appropriate category (numbered 1 to 9) relative to nine different 850 mb v phases. Category 1 is centered on the time of strongest northerly wind at 850 mb, while category 5 is assigned to the time of strongest southerlies or weakest northerlies at 850 mb. Categories 2 to 4 and 6 to 8 occupy intermediate 850 mb v phases; category 9 is assumed to be identical to category 1. Averages were then taken over all cycles selected to compute category means for 850 mb v . Category means were also computed for all variables at all levels between the surface and 200 mb over each region. This allowed us to use category means for different quantities to examine the phase relationships between them, since categories were defined using one variable (meridional winds) only. Likewise, vertical phase tilt can be detected from the vertical dependency of category means, since categories were prescribed at only one level (850 mb).

The variables treated in this manner were T , u , v , ω , p_0 , and dT/dt . Before compositing, prefiltering was applied to original twice-daily data. Smoothed time series data at i -time were obtained by

$$\begin{aligned} \bar{A}_i = & \sum_{k=i-N(1)}^{i+N(1)} W_k(1) A_k / \sum_k W_k(1) \\ & - \sum_{k=i-N(2)}^{i+N(2)} W_k(2) A_k / \sum_k W_k(2), \end{aligned} \quad (11)$$

where

$$W_k(n) = [2\pi\sigma(n)^2]^{-1/2} \exp[-t^2/2\sigma(n)^2],$$

$n=1$ and 2 ,

$$t = k\Delta t, \quad (k=0, \pm 1, \pm 2, \dots), \quad \Delta t = 1/2 \text{ day}.$$

We have let $\sigma(1)=0.5$ day and $\sigma(2)=2.0$ days. The response curve (not shown) indicates satisfactory reduction of A -fluctuations with periods shorter than 3 days and longer than 7 days, while retaining about 76 percent of the original amplitude in a period range of 4 to 6 days. In eq. (11), $N(i)$ was assigned in proportion to $\sigma(i)$, $i=1$ and 2.

Fig. 5 presents composite vertical structures of u , v , T , and ω perturbations with an average period of 5 days over region 2. The coordinates used are phase category as the abscissa and pressure as the ordinate, with category 1 and 850 mb as the origin of the coordinate system. The abscissa may be regarded as an approximately west-to-east oriented axis along which a category width represents about 650 km. As will be shown later, the average phase speed of eastward propa-

gating, 5-day period waves is about 12 m s^{-1} .

In Fig. 5, the meridional wind component has its largest amplitude (exceeding 4 m s^{-1}) at 850 mb, with secondary maxima occurring near 300 mb. The v -axis slopes slightly eastward with height below 850 mb and westward above. These features agree well with those obtained by Nitta *et al.* (1973) and Ninomiya (1975) over the AMTEX region.

The associated temperature pattern is shown in Fig. 5 (top-right). A maximum temperature decrease of about -2.2°C is encountered at 850 mb between category 1 and 2, *i.e.*, about one-half category lag from the strongest northerly wind at 850 mb. The T -axis is nearly vertical below 850 mb and slopes westward above. Regions of warm and cold T -perturbations correlate well with regions of southerly and northerly v -perturbations, respectively, an indication that disturbances transport sensible heat northward. Thus, sensible heat transport is down the $\partial \bar{T} / \partial y$ gradient.

Strongest downward motion with a peak value of about 8 units occurs slightly above 700 mb (nondivergent level) at category 2. An approximately inverse relationship exists between ω - and T -perturbations, implying $\overline{\omega'T'} < 0$ at most levels (Table 1).

Fig. 6 illustrates phase relationships between different variables at 850 mb over region 2. Also shown are phase changes in surface pressure, p_0 . Interestingly, dT/dt reaches its minimum at category 4 and maximum around categories 8 and 9. Thus, it follows that maximum heating occurs slightly in advance of the strongest northerly surges. Another feature of interest in Fig. 6 is an approximate 90 degree (2 category widths) out-of-phase relationship between p_0 -perturbations and 850 mb v -perturbations. Consequently, $\overline{v'p_0'}$ is near zero. Presumably, then, 850 mb v and p_0 perturbations are in approximate geostrophic balance with no meridional wave-energy flux.

Fig. 7 is prepared to investigate zonal phase propagation of v -perturbations at 200 and 850 mb over regions 1 to 3. Compositing was made with reference to region 2 v at 850 mb and category 1. Thus, over region 2, 850 mb v -perturbations lead corresponding perturbations at 200 mb by about two categories, as was already noted in Fig. 5. At 200 mb, there exists about a 3-category phase difference in v -perturbations between regions 1 and 3. From this phase difference and distance (2,000 km) between regions 1 and 3, one

Fig. 5 Vertical cross sections of 4-6 day filtered u (m s^{-1} ; interval 1 unit), v (m s^{-1} ; interval 1 unit), T ($^\circ\text{C}$; interval a half unit), and ω ($10^{-4} \text{ mb s}^{-1}$; interval 5 units), over region 2. Categories 1 to 9 were determined with reference to v at 850 mb. See text for further information.

Fig. 6 Changes in 4–6 day filtered T , dT/dt (heating rate), u , v , and ω with respect to categories 1 to 9, at 850 mb over region 2. Also shown are the surface pressure, p_0 , changes with category.

Fig. 7 Changes in 4–6 day filtered v at 200 mb (top) and 850 mb (bottom) with respect to category, over regions 1 to 3.

can estimate a zonal wave-length and phase speed of 200 mb v -perturbations. Estimated values are about 5,200 km and 12 m s^{-1} (eastward), respectively. In Fig. 7 (top), it is also worth noting that the amplitude of 200 mb v -perturbations is largest (smallest) over region 3 (1). In fact, region

3 amplitude amounts to about 6 m s^{-1} , which is approximately three times as large as its region 1 counterpart ($\sim 2 \text{ m s}^{-1}$). Perhaps, this is indicative of wave amplification due to baroclinic instability with an approximate growth rate of 0.6 day^{-1} .

At 850 mb (Fig. 7, bottom), v -amplitude is largest over region 2 as compared to regions 1 and 3. The strongest northerly v -surge in region 2 exceeds -4 m s^{-1} at category 1 while the corresponding v -surge in region 1 amounts to only about -2 m s^{-1} at categories 7 to 8 (about one day earlier than the occurrence of region 2 surges). Similarly, surges in region 3 are weak ($\sim 3 \text{ m s}^{-1}$) and lag about one day behind the strongest surges over region 2. In short, low-level monsoon surges propagate eastward from China (possibly further west), through the East China Sea where they become strongest, and across Japan to the Western Pacific.

The composite technique was also applied over region 4 (South China Sea) by selecting 11 cases from 850 mb v -fluctuations (refer to Fig. 4). Of these 11 cases, 7 were selected from the early winter period (15 November to 31 December 1970). Over this region, disturbance activity is generally more pronounced in early rather than late winter. As in region 2, compositing was made by prescribing the time of strongest northerly surges at 850 mb as a reference point (category 1). Computed results are summarized in Fig. 8. The most striking feature is an inverse relationship in v -perturbations between the upper and lower troposphere. At category 1, northerly surges below 500 mb are capped by southerly (return) flows aloft, while the reverse is true for category 5. In Fig. 8, upward motions tend to be correlated with dT/dt heating and downward motions with dT/dt cooling. A numerical examination (not shown) reveals that the diabatic dT/dt warming (cooling) nearly balances the adiabatic cooling (warming) due to prevailing upward (downward) motions. Such an approximate balance is quite common in equatorial tropospheric motions (Holton, 1971; Murakami, 1972).

In Fig. 8 (top-right), dT/dt heating reaches a maximum near 500 mb at category 2 (lagging about 15–20 hours behind the time of maximum northerly surges at 850 mb). Conceivably, latent heat release in convection can account for this maximum heating that occurs in the midtroposphere near 500 mb. An inspection of satellite photographs indicates frequent development of disturbances over the South China Sea when low-

Fig. 8 Vertical cross sections of 4–6 day filtered u (m s^{-1} ; interval a half unit), v (m s^{-1} ; interval 1/2 unit), dT/dt ($10^{-5} \text{ }^{\circ}\text{C s}^{-1}$; interval 1/2 unit) and ω ($10^{-4} \text{ mb s}^{-1}$; interval a half unit), over region 4. Categories 1 to 9 were determined with reference to v at 850 mb over region 4.

level monsoon surges are deep and broad. These disturbances, although weak in their circulation intensity, are generally associated with deep convection concentrated over the northern and northwestern sectors. South China Sea disturbances develop in the vicinity of the Philippines or North Borneo, move very slowly westward across the South China Sea along about 5° – 10°N , and rapidly decay when approaching the Indochina coast and/or the Malaysian peninsula. In other words, development, movement, and dissipation of these disturbances occur near or even beyond the southern boundary of our region 4 (see Fig. 2). However, rains associated with S.C.S. disturbances are widespread over their northern and northwestern sectors, as mentioned earlier. This is, perhaps, reflected in the calculation of the dT/dt -heating rate over region 4.

Fig. 9 is similar to Fig. 6, but for region 4. Here, cold T -perturbations tend to be correlated

Fig. 9 As in Fig. 6 except for region 4.

with northerly winds and warm T -perturbations with southerlies. Consequent northward sensible heat flux is down the weak $\partial \bar{T} / \partial y$ gradient over region 4. The close association of u - and v -perturbations indicates northward transport of westerly momentum. Of particular interest is an approximately inverse correlation between 850 v - and p_0 -perturbations. Thus, southward transport of wave-energy flux, $\overline{v'p_0'} < 0$, is implied. This southward wave-energy flux is probably accomplished by low-level lateral coupling with extratropical regions (*e.g.*, region 2) and can account, at least partly, for the maintenance of low-level monsoon circulations over region 4 and possibly further south.

5. Results of spectral analysis

Our primary interest in this section is to illustrate the power spectral density for several important elements of the winter monsoon system. We shall follow the representation of Zangil (1975) in our depiction of the power spectral density; namely, the product of the power spectral density and frequency is plotted along the ordinate, and the log of the frequency along the abscissa. This method of presentation has various advantages, as pointed out by Zangil (1975). We also computed the cospectrum, quadrature spectrum, coherence, and phase difference between two series of data, using techniques described by Maruyama (1968).

Fig. 10 depicts the meridional wind spectrum, computed at all standard levels over region 2.

Fig. 10 Power spectrum for meridional winds at surface (p_0), 850, 700, 500, 300, and 200 mb over region 2. See text for detailed information on the method of spectral depiction.

The salient v -oscillation is around 4.3–4.6 days and is prominent at all levels.⁴ Another peak may be noted above 500 mb at about 7 days and is most pronounced at 200 mb. In the lower troposphere between the surface and 700 mb, shorter-period (~ 2.5 day) oscillations are conspicuous. Nitta *et al.* (1973) confirmed the existence of short-period (2–3 days), medium-scale ($\sim 2,000$ km wavelength) disturbances in the lower troposphere above the AMTEX region. As mentioned in Section 2, our observational network over region 2 is not adequate to detail disturbances with wavelengths of less than 2,000 km. For this reason, no further discussion concerning these short-period, medium-scale disturbances will

⁴ Prescaling was applied to the original spectrum power computed at each level. In doing so, we first computed the mean of products of the original power spectrum and frequency, over the entire frequency range treated. The mean determined for each level is shown at the top of each diagram in Fig. 10. Finally, at each level, the original power spectra for all frequencies were divided by the mean at that level. This facilitates the detection of spectral peaks in the lower troposphere, where the original spectrum power is generally small compared to the upper troposphere.

Fig. 11 As in Fig. 10 except for zonal wind component u .

be given.

As in the v -spectrum, a 4.3–4.6 day peak is dominant in the u -spectrum at the surface (Fig. 11). Similar peaks, although somewhat ill defined, are also encountered at 850 and 700 mb. At these levels, the most pronounced peak is in a period range of between 6.0 and 6.7 days. Here, a question arises as to whether or not the 6.0–6.7 day spectrum is independent of the 4.3–4.6 day peak. Choosing 30 days as a maximum time lag, a resolvable bandwidth in the present study turns out to be about 0.044 cycle per day. Thus, the two spectral peaks in question may be regarded, at least statistically, as independent of each other.

Fig. 12 (left) shows the power spectrum of surface pressure, p_0 , and its derivatives $\partial p_0 / \partial x$ and $\partial p_0 / \partial y$, respectively. The spectra for p_0 and $\partial p_0 / \partial y$ show marked peaks at 6.0–6.7 days, while the $\partial p_0 / \partial x$ spectrum has a maximum at around 4.6 days. Interestingly, this $\partial p_0 / \partial x$ period is nearly coincident with the dominant period (4.3–4.6 days) of v -oscillations at the surface, perhaps implying an approximate geostrophic balance between the two quantities. In Fig. 12 (right), note the presence of well defined peaks at about 4.3–4.6 days in the spectra of T , dT/dt , and ω at 850 mb.

Fig. 12 Left: Power spectrum for surface pressure, p_0 , and its derivatives $\partial p_0/\partial x$ and $\partial p_0/\partial y$, over region 2. Right: Power spectrum for 850 mb T , dT/dt , and ω over region 2.

For the convenience of the reader, the characteristic spectral features obtained over region 2 (East China Sea) are summarized as follows:

4.3–4.6 day period: Conspicuously large v spectra at all levels, dominant peaks for T , dT/dt , and ω spectra below 700 mb, and clear maxima of u and $\partial p_0/\partial x$ spectra at the surface.

6.0–6.7 day period: Ill defined v spectrum below 500 mb, marked peaks in u spectra between 850 and 500 mb, and pronounced maxima for p_0 and $\partial p_0/\partial y$ spectra, with no prominent peaks in T , dT/dt , and ω at all levels.

The summary above, leads us to speculate that low-level monsoonal surges over region 2 are primarily associated with 4.3–4.6 day period oscillations in u , v , T , ω , dT/dt , and p_0 , while 6.0–6.7 day variations are of secondary importance. This is further supported by computing the cospectrum between ω' and T' for 4.3–4.6 day and 6.0–6.7 day periods, respectively. A comparison between the two computed results (Fig. 13, left) suggests more effective generation of local kinetic energy by 4.3–4.6 day period oscillations. Also, the cospectrum between v' and T' (Fig. 13, right) indicates larger northward sensible heat transport by 4.3–4.6 day oscillations compared to 6.0–6.7 day variations in the

Fig. 13 Cospectra (multiplied by frequency) between ω and T (left) and between v and T (right), for 4.3–4.6 and 6.0–6.7 day oscillations over region 2.

Fig. 14 Power spectrum for u at 850 and 300 mb (top), v at 850 and 300 mb (middle), and 700 mb ω and surface pressure, p_0 (bottom), over region 4.

lower troposphere.

Over region 4 (Fig. 14), both u and v spectra at 850 mb exhibit dominant peaks around 6.0–6.7 days with secondary maxima occurring at 4.3–4.6 days. The significant p_0 -oscillation is also around 6.7 days (Fig. 14, bottom-right). Thus,

low-level monsoonal surges over region 4 tend to be associated more with the 6.0–6.7 day period rather than the 4.3–4.6 day period.

The relative importance of 6.0–6.7 day period oscillations, as opposed to 4.3–4.6 day fluctuations, was further confirmed by computing the cospectra between v and K at all levels, and between 850 mb v and surface pressure, p_0 , over region 4 (not shown). The (v', K') cospectrum indicates larger negative values for the 6.0–6.7 day period around 700–850 mb than it does for 4.3–4.6 days. The same is true for the (850 mb v' , p_0') cospectrum with larger contributions from 6.0–6.7 day periods. These are indications of southward transport of kinetic energy and wave-energy, possibly due to lateral coupling with extratropical regions. This could be one mechanism through which low-level monsoon circulations and associated disturbances are generated and maintained over region 4 (South China Sea). Additional contributions to regional energetics appear to be accomplished by the processes of ascending warm air and descending cold air, as confirmed by the large negative (ω', T') cospectrum above 500 mb for the 6.0–6.7 day period range.

Coherence and phase differences between region 2 and each of the other regions were computed for T , u , and v data series at the same pressure level. Table 4 presents results computed at 850

mb and 200 mb, respectively. The coherences between regions 2 and 1, and between 2 and 3, for 850 mb T , u , and v at 4.3–4.6 and 6.0–6.7 period ranges, are all greater than 0.7 which is well above the 5 percent significant level. Region 1 elements lead those in region 2, while region 3 variables are lagging behind their region 2 counterparts, an indication of eastward propagation with an average phase speed of about 10 m s^{-1} (refer also to Fig. 7). Similar eastward movements are also noted at 200 mb.

Of particular interest is a generally high coherence between regions 2 and 4 for both 4.3–4.6 day and 6.0–6.7 day periods, at 850 mb. The corresponding coherence at 200 mb is relatively low, implying that atmospheric teleconnections between regions 2 and 4, with the 4.3–4.6 day and 6.0–6.7 day periods, are more likely to occur via the lower rather than the upper troposphere.⁵ In comparison, for the 20–30 day period range, upper tropospheric teleconnection tends to be equally or even more important than its low-level counterpart. This will be elaborated on in Section 6.

Compared to region 4, the v -spectrum at 850 mb over region 5 (Fig. 15) is more complex and less systematic with a number of minor peaks having periods of less than 10 days. In addition, there exists an isolated, somewhat ill defined, v -peak with a 20-day period at 850 mb. The u -spectrum, by comparison, is more systematic with a major peak near 20–30 days and a secondary maximum around 4.6 days. In Fig. 15 (bottom-right), the p_0 -spectrum possesses a distinct and very sharp peak near 20 days, which is clearly separated from a secondary peak around 4.3–4.6 days. Therefore, we feel that the activity of low-level monsoonal surges with shorter (4.3–6.7 day) periods is less pronounced over region 5 as compared to region 4. The coherence of 850 mb T , u , and v between regions 2 and 5 is generally lower than that for regions 2 and 4 (Table 4 for

Fig. 15 As in Fig. 14 except for region 5.

⁵ The 850 mb- v phase difference between regions 2 and 4 (East and South China Seas) is only -19° (~ 6 hours) for 4.3–4.6 day period, while it is -50° (slightly less than one day) for the 6.0–6.7 day period range. As shown in Figs. 1a (bottom) and 1b (top), northerly surges over the East (South) China Sea are associated with a strong E-W (N-S) pressure gradient to the east (south) of an anticyclonic cell centering over China. The increase in northerly winds over region 5 occurs about 2 to 3 days later when the anticyclonic cell propagates eastward to the east of Japan (Fig. 1a, top).

4.3–4.6 and 6.0–6.7 day periods).

The most significant feature in the region 5 spectra is the predominance of 20–30 day peaks. This is clearly shown in the spectrum distributions for u at all levels, v in the lower troposphere, and surface pressure, p_0 (Fig. 15). The coherences for these quantities between regions 2 and 5 are relatively high as they exceed 0.5 (Table 4 for 20–30 day period).

The predominance of 20–30 day spectral peaks is also characteristic of many other regions. This is clearly illustrated in Fig. 16, which depicts 200 mb u -spectra over regions 1 to 5, respectively. With the exception of region 3, the existence of a dominant 20-day peak is clearly confirmed. In Table 4, it should also be noted that there is a high (>0.5) coherence in 20–30 day u -oscillations at 200 mb between region 2 and regions 1, 3, 4, and 5. We are aware that the emphasis of such long-period oscillations is subject to some criticism, since low-frequency spectral estimates are not as reliable, at least statistically, as those in the middle range of the entire frequency domain under consideration. However, the prominent spectral peaks for various important elements and high coherences between region 2 and all other regions, as confirmed in Fig. 16 and Table 4, probably indicates that the existence of 20–30 day oscillations is real and requires further study.

Based on geopotential height and temperature data, operationally determined by the National Meteorological Center, Washington, D.C., McGuirk and Reiter (1976) observed a strong, persistent and significant oscillation of about 24-day periodicity in hemispheric-scale energy parameters over the Northern Hemisphere extratropics during the winter season. A set of independent data (Webster and Keller, 1975) revealed a cycle of about the same periodicity in the extratropical kinetic energy of the Southern Hemisphere as manifested by Eole balloon data.

In the next section, we examine a number of interesting problems which relate to these oscillations occurring over the Northern Hemisphere extratropical and tropical regions. In particular, their detailed structural features and their role in maintaining the winter mean circulation system are investigated.

6. 20–30 day oscillations

From original twice-daily data sets over each region, we extracted 20–30 day filtered u , v , and T data by applying eq. (11) with prescribed values of $\sigma_1=3.5$ days and $\sigma_2=6.0$ days. The response curve of this filtering operator shows that about 38 (30) percent of the original data is retained in a 20 (30) day period, while greatly reducing short period oscillations (for example,

Fig. 16 Power spectra for 200 mb u over regions 1 to 5.

Fig. 17 Time-series of 20–30 day filtered $T(^{\circ}\text{C};$ interval $1/2$ unit), $v(\text{m s}^{-1};$ interval 1 unit), and $u(\text{m s}^{-1};$ interval 1 unit), over region 2.

only 9 percent for a 10-day period). Thus, 20–30 day oscillations contribute most to the filtered time series data.

Fig. 17 illustrates time-height sections of 20–30 day filtered T , v , and u over region 2. At 850 mb, filtered v is northerly and substantial during the periods of days 45 to 56 and 67 to 85. Almost simultaneously, filtered 850 mb T is negative (colder than usual) during these two periods.⁶ An inspection of original time series v data at 850 mb, shown in Fig. 3, reveals that a sequence of three extremely strong, partially overlapping, monsoonal surges occurred between days 45 and 56. Similarly, between days 67 and 85, we note the occurrence of several (5 or 6) time-clustering monsoon surges. On the other hand, between days 57 to 66 and 86 to 100, twice-daily 850 mb v over region 2 never exceeds -10 m s^{-1} and is even positive (southerly) occasionally. This is reflected in 20–30 day filtered 850 mb v data which exhibits southerly anomalies during the corresponding two periods; namely, these

periods are characterized by weaker than usual short-period monsoon surges.

The relationship mentioned above, between long-period v - or T -oscillations and individual monsoon surges in a 4–6 day period range during winter, is very similar to that observed in summer during “active” and “break” monsoons. When the summer monsoon is “active,” several partially overlapping monsoon disturbances develop and rainfall is widespread over the entire monsoon region with precipitation around disturbances much higher than normal. Conversely, no monsoon disturbances form during a monsoon “break” and rainfall activity over most of the monsoon region becomes very weak. The period associated with “active” and “break” monsoons is about 15–20 days, which is much longer than the life cycle (3–7 days) of individual summer monsoon disturbances (refer to Murakami and Frydrych, 1974). Also, the characteristics of “active” and “break” monsoons, on the horizontal scale, are much larger than that of individual monsoon disturbances. Although “active” and “break” monsoons are an important part of the total summer monsoon system, their structural features have never been completely documented. Their causes and role in the maintenance of the summer mean monsoon circulation are also unknown, which indicates the importance of further study in this particular area of the summer monsoon problem. We feel that similar studies, using 20–30 day filtered data, are required for long-period, large-scale monsoon fluctuations during winter. For brevity, throughout the subsequent discussion, the term “filtered” is rarely used unless absolutely necessary.

Let us again look at Fig. 17 for region 2. We at once note the near simultaneous occurrence of negative (positive) T with negative (positive) v around 850 mb. This, together with a large positive (v' , T') cospectrum at 850 mb (Table 5), indicates northward sensible heat transport due to 20–30 day oscillations over region 2. Likewise, the close association of negative (positive) T with positive (negative) u in the lower troposphere (Fig. 17) and a large negative (u' , T') cospectrum in that vicinity (Table 5), imply westward sensible heat transport. This, in turn, indicates that 20–30 day oscillations play an important role in the reduction of heat contrasts between the cold Asiatic continent and warm Pacific Ocean during winter. Another feature of interest in Table 5 for region 2, is a large negative (ω' , T') cospectrum between 850 and

Fig. 18 Time-series of 20–30 day filtered $T(^{\circ}\text{C})$; interval $1/2$ unit) over regions 1, 3, and 4, respectively.

⁶ The approximate amplitudes of filtered v - and T -oscillations at 850 mb are 1 m s^{-1} and 1°C , respectively. When taking into account their reduction rate of about 30–38 percent, these values are actually comparable to or even greater than those associated with low-level monsoonal surges of about 4–6 days (see Fig. 5).

700 mb, an indication of local kinetic energy generation due to 20–30 day oscillations. The predominance of a negative (ω', T') cospectrum in the lower troposphere below 500 mb is also noted in regions 1 and 3. Over tropical regions 4 and 5, on the other hand, the (ω', T') cospectrum tends to be negative in the upper troposphere above about 500 mb, perhaps reflecting the changes in rainfall (convective activity) in association with 20–30 day oscillations. This point will be examined later.

A comparison of Figs. 17 (top) and 18 (top) reveals that T -oscillations over region 1 occur slightly in advance of region 2 T -fluctuations. The T -phase difference between these two regions is 10° (3°) at 850 (200) mb, with an extremely high coherence of more than 0.9 (Table 4, right). High coherence is also noted in T -oscillations between regions 2 and 3, with region 2 leading by phase differences of 9° (8°) at 850 (200) mb. From these phase differences, one can estimate the phase speed of T -perturbations to be about $2\text{--}3\text{ m s}^{-1}$ (eastward) near region 2. In comparison, the eastward phase speed, estimated from 850 mb v -phase differences between region 2 and regions 1 and 3, is approximately 4 m s^{-1} .

A glance at Figs. 17 (top) and 18 (bottom) shows that 850 mb T -oscillations over region 4 lag behind corresponding T changes over region 2. As shown in Table 4, coherence and phase difference between these regions are 0.75 and -34° (about 2–3 days lag), respectively, for T at 850 mb. This table also reveals a fairly high coherence (>0.5) at 200 mb for T , u , and v between regions 2 and 4. Coherence at 850 mb, by comparison, is higher (>0.7) for T and v but lower (~ 0.1) for u between these two regions. Presumably, then, lateral coupling between regions 2 and 4, associated with 20–30 day fluctuations, occurs more favorably via the upper troposphere rather than through the lower atmosphere.

Fig. 18 (bottom) exhibits a large vertical phase difference for region 4 T -oscillations between the lower (below 850 mb) and upper (above 500 mb) troposphere, with rapid phase changes between 850 mb and 500 mb. Consequently, large T -oscillations near 850 mb are inversely correlated with secondary maxima around 500 mb. This is in sharp contrast to what was observed in extratropical regions 1 to 3, where T -oscillations are nearly in phase (vertical) within the deep atmospheric layer from the surface to 500 mb and possibly even higher.

Next, an attempt was made to further examine

the relationship in 20–30 day oscillations between the extratropical and tropical regions. To accomplish this, we used daily geopotential data, prepared by the Computer Center, Japan Meteorological Agency (JMA), at 850 and 200 mb over an extensive region covering India to the west, the Central Pacific to the east, and from the equator to about 60°N , during the period 16 November 1970 through 15 March 1971. From this data, we extracted 20–30 day filtered geopotential fields by applying eq. (11) with $\sigma_1=3.5$ days and $\sigma_2=6.0$ days over the entire JMA domain. The composite geopotential maps were then constructed by assigning an appropriate category (numbered 1 to 9) to these 20–30 day filtered data. The category was determined in reference to the changes in 20–30 day filtered v at 850 mb over region 2; 850 mb v changes were most pronounced and systematic over region 2 as compared to regions 1, 3, 4, and 5. Category 1 was centered on the time of strongest northerly winds or weakest southerlies at the reference point (850 mb, region 2), *i.e.*, days 20, 32, 51, and 81 (see Fig. 17, middle). Conversely, category 5 was assigned to the time of strongest southerlies; namely, days 27, 38, 63, and 89. Categories 3 and 7 corresponded to the times when 850 mb filtered v was zero at region 2. Finally, category means were computed for 20–30 day filtered geopotential data at 850 and 200 mb over the entire JMA domain. Composite geopotential maps were then used to examine phase relationships between different latitudinal zones, and between 850 and 200 mb. This was possible since categories were determined from only one region (2) and one level (850 mb).

Figs. 19a and b depict 850 mb composite geopotential fields at categories 1, 3, 5, and 7, respectively. At category 1, northerly winds prevail over the East China Sea region, as they should despite our category definition using region 2 data (not JMA data). Conversely, southerlies are dominant at category 5 over the East China Sea. In fact, category 5 geopotential configurations, with an opposite sign, are surprisingly similar to category 1 patterns over the entire JMA domain. This may confirm the contention that the use of two different data sets (region 2 data for category definition and JMA data for composite maps) does not cause any serious errors and is adequate for investigating changes in long-period (20–30 day), large-scale circulations during winter.

In Figs. 19a and b, a rather systematic westward propagation of 850 mb geopotential fields

Fig. 19a Composite 20–30 day filtered geopotential (m; interval 5 units) maps at 850 mb, for categories 1 (top) and 3 (bottom). Categories were determined in reference to 20–30 day filtered v at 850 mb over region 2 (Fig. 17, middle). Regions of negative geopotential anomalies are shaded.

Fig. 19b As in Fig. 19a except for categories 5 and 7.

poleward of about 30°N is immediately evident. At category 1, a distinct zone of positive geopotential extends southward from 65°N to 30°N along about the dateline. Between categories 1 and 7, this positive zone (ridge) shifts westward while increasing in intensity, and reaches Central Asia near 90°E at category 7. The average westward phase speed is about -4 m s^{-1} at 60°N . A close examination of Fig. 19 reveals that the

westward movement of this ridge system is not uniform but somewhat intermittent with higher speeds ($\sim -7\text{ m s}^{-1}$) between categories 3 and 5. Equatorward of about 30°N , on the other hand, 850 mb geopotential fields exhibit near zero phase speed (standing oscillations). This is most clearly seen over the South China Sea where 850 mb geopotential is high during categories 1 and 3, while it is low between categories 5 and 7. Here, it should be mentioned that the phase propagation discussed above is an indication of relative speed with reference to v -perturbations at 850 mb over region 2; at this reference point, the estimated v -phase speed is approximately 4 m s^{-1} (eastward), as mentioned earlier. Thus, it follows that the actual phase propagation is eastward ($\sim 4\text{ m s}^{-1}$) in the equatorial regions while it is near zero (standing) in the higher latitudes around 60°N .

Composite geopotential maps for 200 mb are shown in Figs. 20a and b, respectively. They closely resemble their 850 mb counterparts poleward of about 30°N , which implies that there is very little (if any) vertical tilt in ridge or trough systems between the two levels. This, in turn, means that at higher latitudes (around 60°N), actual phase propagation of 200 mb ridge and trough systems is near zero (standing), as it was at 850 mb.

In the equatorial regions, the character of 200 mb geopotential waves is considerably different from those in higher latitudes. Let us first examine their relative phase propagation, as illustrated in Figs. 20a and b. A high cell located over the Malaysian peninsula at category 1 moves intermittently eastward across the South China Sea and the Philippines between categories 3 and 5, and reaches as far as the western Pacific around 140°E , 20°N at category 7. Similarly, a trough extending equatorward from Burma to the eastern Bay of Bengal in category 3, shifts eastward and, after crossing the Malaysian peninsula at around category 5, reaches the central South China Sea at category 7. An average relative phase speed over the Bay of Bengal-Malaysia-South China Sea region is about 3 to 6 m s^{-1} (eastward), corresponding to an actual eastward phase speed of 7 – 10 m s^{-1} .

Liang (1974) investigated the changes in weather over the Malaysian region in association with eastward propagating upper troughs. He showed that weather associated with the equatorial part of the upper trough was clear due to the presence of dry air in and around the trough. In Fig. 20b (top), a 200 mb trough is located near

Fig. 20a As in Fig. 19a except for 20–30 day filtered 200 mb geopotential height (m). Isopleths are for 10 m intervals. Dashed lines are for 5 units.

Fig. 20b As in Fig. 20a except for categories 5 and 7.

the Malaysian peninsula, an indication that category 5 may correspond to dry spells over that region. This 200 mb trough lies directly above an 850 mb ridge (Fig. 19b, top) near the Malaysian peninsula. Thus, category 5 is characterized by negative temperature anomalies (colder than usual) between 850 and 200 mb. Conversely, positive temperature anomalies are encountered between 850 mb and 200 mb over Malaysia at category 1 when an 850 mb trough is capped by a 200 mb high cell, which leads us to speculate that category 1 may correspond to wet spells

over the Malaysian region. This contention was formulated in reference to a dominant negative correlation between ω and T above 500 mb over region 4 (Tables 1 and 5); namely, the association of warm temperature with ascending motion (wet weather) and cold temperature with descending motion (dry weather).

Recall that category 1 was defined as the time when 850 mb northerly winds are strongest and accompanied by the coldest temperature anomalies over region 2 (Fig. 17, top and middle). Also recall that during category 1, a sequence of exceptionally strong, time-clustering, low-level monsoon surges occurred over region 2 [compare Fig. 3 (middle) with Fig. 17 (middle)]. Referring to the above, our tentative hypothesis in this study is as follows: wet spells over the Malaysia-South China Sea region are associated with unusually pronounced and prolonged activity of monsoon surges over the East China Sea. An extensive study is now in progress at the University of Hawaii to examine the above hypothesis using 20–30 day filtered divergence and velocity potential, among other quantities, over both hemispheres during the 1970–71 winter. The results of this study will be reported elsewhere by Lyons and Murakami.

7. Concluding remarks

It is evident from this study that the character of low-level monsoonal surges differs considerably with region during the northern winter. To summarize, the main results of this study are:

(1) Over the East China Sea (region 2, Fig. 2), low-level monsoonal surges are primarily associated with 4.3–4.6 day period oscillations in u , v , T , ω , dT/dt (heating rate), and surface pressure, p_0 . When northerly surges are stronger than usual, T is colder than the season mean and is accompanied by large above normal heating below 850 mb. Similar features are noted over China (region 1) and the Western North Pacific (region 3).

(2) In extratropical regions 1 to 3, the axes of u -, v -, T -, and ω -perturbations tilt westward with height above 850 mb. Strongest downward motion lags about a half day behind the strongest northerlies and coldest temperatures. The co-spectrum relating to 4.3–4.6 day oscillations between v and T (ω and T) is positive (negative) in the lower troposphere, indicating northward sensible heat transport (local kinetic energy generation). It appears that low-level monsoonal surges occur in association with the eastward

propagation of baroclinic wave perturbations from China (possibly further west). These surges then move through the East China Sea where they become strongest, and cross Japan to the Western Pacific, as revealed by extremely high coherences and systematic phase differences between these regions in a period range of 4.3–4.6 days.

(3) Over the South China Sea (region 4), the spectra for 850 mb u , v , T , and surface pressure, p_0 , exhibit dominant peaks around 6.0–6.7 days with secondary maxima occurring at 4.3–4.6 days. In this region, strong low-level northerly surges are accompanied by above normal ascending motion throughout the troposphere with an associated heating rate (rainfall) also well above the season mean. The most striking feature is an approximately inverse relationship in v -perturbations between the lower and upper troposphere; namely, northerly surges below 500 mb are capped by southerly (return) flows aloft.

(4) Coherence is high (≥ 0.5) at 850 mb between regions 2 (East China Sea) and 4 (South China Sea) for both 4.3–4.6 day and 6.0–6.7 day periods. Over region 4, the cospectrum between meridional winds and kinetic energy at 850 mb is negative (southward kinetic energy flux), as is the cospectrum between 850 mb v and p_0 (southward wave-energy flux). The lateral couplings associated with monsoonal surges are more likely to occur via the lower, rather than the upper troposphere.

It is fair to say that the results, summarized in (3) and (4) for the South China Sea region, are somewhat speculative and should be regarded as preliminary. A serious weakness in this study is poor data coverage and questionable accuracy at certain stations over region 4 (Fig. 2), which precludes an accurate measurement of vertical velocity and heating rate associated with monsoonal surges. Perhaps, this problem can best be investigated when the Winter Monsoon Experiment data set, including data from three U.S.S.R. ships and two dedicated U.S.A. research aircraft over the South China Sea, becomes available.

(5) Spectral analysis revealed the existence of dominant and isolated peaks around 20–30 days for many important elements over most regions. Using 20–30 day filtered u , v , and T data, detailed structural features and their relationship with low-level monsoonal surges (4–6 day periods) were investigated. During periods when 20–30 day filtered v at 850 mb was northerly over the East China Sea, a sequence of extremely strong, partially overlapping, monsoon surges occurred.

In contrast, periods of southerly anomalies in 20–30 day filtered v , were characterized by weaker than usual short-period (4–6 days) monsoonal surges with below normal heating (sensible heat supply) near the surface. These features are similar, in some respects, to “active” and “break” monsoon conditions during summer.

(6) The coherence for T and v between regions 2 and 4 is substantially large (>0.5) for 20–30 day period oscillations, with region 4 lagging behind region 2 by about 1–3 days. Phase relationships between extratropical and tropical latitudes were further investigated using Northern Hemisphere geopotential data during the 1970–71 winter, prepared by the Japan Meteorological Agency. Composite geopotential maps for 20–30 day period perturbations revealed eastward phase propagation ($\sim 4\text{--}10\text{ m s}^{-1}$) in the equatorial regions covering India, the Bay of Bengal, the South China Sea, and the Western Pacific, as opposed to near zero phase speed (standing) in higher latitudes around $50^\circ\text{--}60^\circ\text{N}$. When eastward propagating 200 mb ridges (troughs) approach the Malaysia-South China Sea region, wet (dry) spells appear to commence over that region with above normal ascending (descending) motions and above (below) normal temperatures, which result in a large negative cospectrum between ω and T for 20–30 day oscillations. It also appears that wet (dry) spells over the Malaysia-South China Sea region approximately coincide with unusually strong (weak) activity in low-level monsoonal surges over the East China Sea.

So far, we have emphasized the importance of long-period (20–30 days) oscillations as a part of the entire winter monsoon system. These long-period oscillations appear to occur over an extensive area, covering at least East and Southeast Asia and the North Pacific, as indicated in Figs. 19 and 20. However, their causes and associated lateral coupling mechanisms still remain unknown. During Winter MONEX (December 1978–February 1979), one of the major scientific objectives is to investigate this problem in relation to heavy rainfall and dry spells over the Malaysian region.

Acknowledgements

The author is indebted to Dr. Y. Ookochi, the Computer Center, the Japan Meteorological Agency, for providing the Northern Hemisphere geopotential data for this study. He also thanks Mr. L. Ho and Mrs. C. Fenander for programming assistance and Mrs. S. Arita for typing the manuscript.

This research has been supported by the Global Atmospheric Research Program, Climate Dynamics Research Section, National Science Foundation, under Grant No. ATM77-28358.

Appendix: List of symbols

- λ, ϕ : longitude and latitude
 x, y : eastward and northward Mercator coordinate axes
 p : pressure
 t : time
 u, v : zonal and meridional winds
 ω : vertical p -velocity
 p_0 : surface pressure
 ψ : streamfunction
 Φ : geopotential height
 dT/dt : heating rate defined in eq. (8)
 Q : diabatic heating rate per unit mass
 a : radius of earth
 κ : R/c_p ; R gas constant of air and c_p heat capacity of air at constant pressure
 K : kinetic energy
 dK/dt : rate of change of kinetic energy defined in eq. (10)
 $(\bar{\quad})$: time average
 $(\quad)'$: departure from time average

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東アジアおよび東南アジアにおける寒波の吹出しについて

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冬期（120日間）における風と温度の資料を用いて中国、東支那海、南支那海、西太平洋域における吹出しの性質と構造をしらべた。吹出しは東支那海でもっとも強く、風、温度、上昇流、加熱量に顕著な4.3–4.6日周期があらわれる。吹出しは偏西風波動の一環としておこり、吹出しにともない下降気流が強くなり、下層における加熱（海上における顕熱輸送）が増加する。

南支那海では6.0–6.7周期が卓越し、4.3–4.6日周期はあまり顕著でない。吹出しが強くなると上昇気流が強くなり、対流圏全域で加熱（凝結）が大きくなる。南支那海と東支那海の吹出しの間には強いコヒーレンス（ >0.5 ）がみられる。

すべての地域で20–30日周期の変動が卓越する。4–6日周期の吹出しとは深い関係がある。700 mb と 200 mb の等圧面コンボジット天気図によると、20–30日周期の変動は高緯度では定常性であるが、30度以南の熱帯域では東進波動と関係している。東進波のトラフ（リッジ）がマレーシア、南支那海に達すると少（多）雨期となる。この地域の天気変動と、東支那海附近の吹出しの変動とは密接な関係がある。