The Internal Gravity Wave Observed in the Typhoon T8124 (Gay)

By Seiichi Matsumoto* and Hirofumi Okamura

Meteorological Research Institute, Nagamine, Yatabe, Ibaraki-ken, 305 Japan

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Abstract

An extremely sharp pressure dip accompanied with a mesoscale disturbance was observed in the northwestern quadrant of Typhoon 8124 (Gay) passing off the coast of the Boso peninsula. The amount of pressure drop was so large that the minimum pressure value associated with the disturbance (967.4 mb) was almost comparable to the central pressure of this typhoon (965 mb). The steering velocity and the lifetime of the disturbance was estimated to be a little faster than the typhoon movement and more than five hours respectively. The structure was analyzed based on the data from dense network of surface meteorological observations and a doppler radar installed in MRI, and it was shown that the surface wind field and the cloudiness distribution had internal gravity wave characteristics. The disturbance was accompanied with a rainband system, and the surface pressure drop-jump phenomena seems to be closely related to the finer structure of systematic convective activity with the east-west wavelength of 90 km embedded in the rainband.

1. Introduction

Internal Gravity Waves (hereafter denoted by IGW) appearing in the stratified lower atmosphere have sometimes been observed. IGW characteristics are often revealed in *Mesoscale Disturbances* (hereafter denoted by MD) analyzed within severe snowstorm or heavy rainfall regions (Matsumoto & Ninomiya, 1965; Matsumoto & Akiyama, 1969; Uccellini, 1975; and others). Bosart & Cussen (1973) analyzed an IGW-related *Pressure Drop-jump* (hereafter denoted by PD) line of 500 km length which moved north-eastwards in the Eastern United States and showed, like many others, a divergence area in front of this line and a convergence area in the rear of it. Eom (1975) investigated the relationship among surface pressure, surface wind, cloudiness field and their changes around a PD line within a MD which moved north-eastwards over the Midwest of USA, and explained their rela-

tionship by the IGW theory derived from a simple two-layer model with vertical wind shear.

A force that causes the up and down oscillation of air particles in IGW is *buoyancy* by the density difference between stratified layers, and it is ultimately due to the action of gravity (Brunt, 1927; Thorpe, 1968a). The results of theoretical, laboratory experimental and observational studies indicate that the Richardson number less than 0.25 is a necessary condition for large-amplitude IGW developments (Miles, 1961; Thorpe, 1968b; Gedzelman & Rilling, 1978). Therefore the existence of strong vertical wind shear in stratified layers is important for such a development.

There have been some reports describing such a PD-like disturbance within a typhoon. Fujita (1952) analyzed a pressure disturbance within T4902 (Della), and named it *pressure dip*. Tatehira (1961) investigated a *pressure dip* phenomena within a rainband of T5821 (Helen). Mitsuta *et al.* (1979) and Nakajima *et al.* (1980) reported similar anomalous pres-

* Present affiliation: Japan Weather Association, 4-5 Kojimachi, Chiyoda-ku, Tokyo, Japan.

Fig. 1 Locations of the stations with barograph records. a: Hamamatsu, b: Irozaki, c: Tateyama, d: Tokyo, e: Chiba, f: Tateno (Tsukuba, MRI), g: Onahama, z: Mt. Fuji PD was not recorded at stations from st. 21 to st. 26.

sure field related to T7818 (Irma) and T7916 (Owen) respectively.

A typhoon named T8124 (Gay) moved north-eastwards off the coast of the Japan Islands on 22 October 1981. A sharp PD line, which had north to south 200 km length at its mature stage, was generated (about at 1200GMT) in the northwest quadrant of this typhoon, and tracked along with the typhoon for more than 5 hours. This PD phenomena at its mature stage was quite fortunately caught on every trace of five barographs arrayed within 25 km from north to south of Mt. Tsukuba, which was specially designed and operated by the orographic rain research group in Meteorological Research Institute (MRI). These traces showed 9 mb rapid drop within 20 min and 7 mb rapid jump within 20 min right after re-

spectively. The amount of pressure deficit or its pressure gradient is the greatest one that we have ever experienced. On the meantime the meteorological tower (213 m height) observation at MRI showed clearly that the maximum gust of 37 m/s with a veering wind from NNW to NE and then backing to NW corresponded to the pressure disturbance. Gradually weakening heavy rainfalls associated with the typhoon once stopped at the time of its passage, and a very weak rainfall followed for a while. On the other hand, any notable changes were not recognized in the temperature and the dew point temperature records. Such remarkable changes in the meteorological elements as mentioned above were observed at many stations from Hamamatsu (st. a) at 1210 GMT to Onahama (st. g) at 1700 GMT as

shown in Fig. 1.

It may be worthwhile to emphasize from a practical point of view that the net change of pressure accompanied with the MD could amount to such a large value. The minimum pressure within the PD observed at, for example, Irozaki (st.b) was almost the same pressure (967.4 mb, at 1410 GMT) as that of

the typhoon center itself (965 mb) located about 100 km off the coast. Therefore miss judgments could be caused in the operational fix of the typhoon center. The greatest care must be taken for gusts, strong wind shears and sudden pressure changes accompanied with such a MD, and, especially, in the altimeter setting of aerial navigation. Also there would

Fig. 2 Synoptic charts at 1200 GMT Oct. 22, 1981, at the surface (a), at 850 mb (b), at 700 mb (c) and infrared cloud picture of GMS (d). Heavy solid and dotted line segments in (a) represent tracks of the typhoon and cyclones in every 12 hours. Solid lines in (a) show sea-level pressure. Solid and dashed lines in (b) and (c) represent geopotential height (in 10 m) and temperature (in °C).

be dangers of storm surge at each harbor in high tide if shock waves associated with a PD be superimposed on a rising tide by a typhoon depression.

The authors would like to discuss the structure of the MD which caused the PD phenomena, to show its IGW characteristics through comparisons between the results of analyses and a simple model presentation of the IGW in the atmosphere, and to propose that the nature of a PD phenomena within a typhoon is related to IGW motion.

2. Synoptic situation

A weak tropical depression generated on 14 October 1981 developed into the typhoon (T8124) two days later over the sea ($17^{\circ}10'N$, $144^{\circ}05'E$) to the west of Marianas. On 22 October this typhoon moved north-eastwards with a speed of 65 km/h to 60 km south of Miyake island (st. 21) at 1300 GMT. The central surface pressure was 965 mb at that time, and its size was fairly large (the 1,000 mb isobar was 1,200 km in diameter).

There was a rainy area ranging from north-west to the east of T8124, and the strong convective activity brought a large amount of precipitation all over Kanto District (most of stations in Fig. 1 are located). For instance the radar observation at Mt. Fuji (st. z) is missing for some time by a thunderbolt struck because of this strong convective activity in the typhoon rainbands.

Fig. 2 shows maps on the surface (a), on 850 mb (b) and on 700 mb (c) together with GMS IR imagery (d) at 1200 GMT immediately before the pressure anomaly appeared at Hamamatsu. Two developing cyclones located at Sikhote Alin Range (west of Sakhalin) and north part of Japan sea were lined up in almost N-S direction along with the typhoon, and brought a dry airmass from the Continent into the western quadrants of the typhoon. In 700 mb chart it is seen that the area of the MD development under investigation was already covered by the continental airmass in the upper layer. The clear cut cloud pattern in the western part of the typhoon may be regarded as a manifestation of airmass separation between the continental dry airmass and

the oceanic wet airmass. Meanwhile the GMS cloud imagery suggests the development of cloud system like a tapering cloud generating at 200 km northwest of the typhoon center, within which the MD under investigation took place. Developing cumulonimbi were generated at around the southernmost edge of this system.

Fig. 3 (a): Vertical profiles of the equivalent potential temperature at Tateno (1200 GMT, Oct. 22, 1981) (T_{12}), at Hamamatsu (1300 GMT, Oct. 22) (H_{13}) and at Tateno (1800 GMT, Oct. 22) (T_{18}). (b): The vertical cross section parallel to the MD trajectory for wind, θ_e and the Richardson number (along the broken line given in Fig. 12). Dashed lines show isotachs of east-west wind component in m/s, solid lines the isotherms of θ_e in $^{\circ}K$ and sparse and dense stipples the region of Richardson number less than 1.0 and 0.5. The wind in the shaded area has easterly component.

Fig. 3(a) shows the stratifications at Tateno on 1200 GMT Oct. 22 (T_{12}), Hamamatsu on 1300 GMT Oct. 22 (H_{13}) and Tateno on 1800 GMT Oct. 22 (T_{18}). Since the PD has passed through Hamamatsu on 1210 GMT and Tateno on 1542 GMT, T_{12} , H_{13} and T_{18} are considered to represent the situation at 4 hours before, immediately after and 2 hours later the PD passage respectively. We find the typhoon airmass uniformly mixed in the lower troposphere (in T_{12}), the strongly stable layer above latently unstable one in the mid troposphere (in H_{13}) and the prevailing dry westerly in the mid and the upper troposphere (in T_{18}). In Fig. 3(b) is shown a vertical cross section of the MD. We observe easterly wind component of typhoon circulation in the lower troposphere to the right (in front) of the point indicated by the symbol H_{13} , and westerly winds prevailing elsewhere. We can find a stably stratified layer accompanied with vertical wind shear at about 700 mb above the area of PD development. The easterly region (shown by shade) corresponds to the cloudy region associated with the typhoon, and the clear cut cloud pattern to the west is linked to the dry westerly flow.

The Richardson number (Ri) is expressed in terms of the stability and the vertical wind shear

$$Ri = g(\ln \theta)_z / (U_z)^2 = (N/S)^2, \quad (1)$$

where N is the Brunt-Väisälä frequency, S the vertical wind shear. The distribution of Ri in the cross section of the MD is entered in Fig. 3(b). It is remarked that quite a favorable condition for large-amplitude IGW development was established above the PD location.

3. Observation of MD

3.1 Data and analysis

In Fig. 1 are given auto-barograph records at surface synoptic meteorological stations. Examples of meteorological records are given in Fig. 4(a) and (b). It is clearly seen that extremely sharp changes took place at many stations on surface pressure, wind and precipitation but neither on air temperature nor

Fig. 4 Records of surface meteorological elements (pressure: (a) and wind, precipitation, temperature and dew point temperature: (b)) as observed at Tateno (st. f) around the time of MD passage (1542 GMT).

on dew point temperature.

The typhoon passed through 45 km off the Boso peninsula. A remarkable MD was embedded in the northwestern quadrant of this typhoon. In order to extract the structure of the MD from the typhoon field, the authors define an anomaly (q) from the original data (Q),

$$q_i = Q_i - \frac{1}{13} \sum_{j=i-6}^{i+6} (Q_j) \quad (2)$$

which is the deviation from two hours running mean (as shown in Matsumoto & Akiyama, 1969). Here Q_j denotes every 10 min data such as surface pressure (P), surface wind (U, V), atmospheric temperature (T) and dew point temperature (Td). In Fig. 5 are given the pressure anomaly field and the typhoon general field. Let us discuss in the following sections the anomaly fields during the time interval under the typhoon influence.

Fig. 5 Detailed surface map at 1510 GMT Oct. 22, 1981. Solid lines are isobars either of the MD or of the typhoon, and dashed line shows the PD line. Shaded areas represent the area of precipitation equal to or greater than 0.5 mm, 5 mm and 10 mm respectively in 1500–1520 GMT. The enlargement of wind distribution in the square area is presented on the upper left.

3.2 Surface pressure disturbance field

In Fig. 6 is given the maximum value of negative anomaly of the pressure disturbance together with the PD line*. It is clearly seen that the MD traveled along the same direction as the typhoon movement with somewhat faster velocity. The anomalies amounted to more than 7 mb were recorded at Irozaki and Tateno (st. f). During the life cycle of this disturbance, we observe a fluctuation with about one hour period of the maximum pressure anomaly suggesting the existence of finer

* This is defined by the isochrone of the maximum value of negative pressure anomaly (PD)

structures, which will be discussed in the following sections.

During this event a very fine barograph array network (Table 1) distributed from north to south was operating in Mt. Tsukuba area indicated by a small shaded strip in Fig. 6. Analysing these records in detail, the authors found that the pressure disturbance has a longitudinally uniform structure extending nearly in N-S direction and propagating to the direction of 85° with the velocity of 66 km/h.

3.3 Wind Disturbance Field

The authors have carried out detailed analyses on the wind field by applying the

Fig. 6 The maximum pressure anomaly distribution and PD isochrones. A broken line A-A' is used for time section in Fig. 9. Shaded area around st. f indicates the special observation field implemented by five stations near Mt. Tsukuba. Heavy line shows the typhoon track.

time to space conversion method (as shown in Fujita, 1963) on the dense network data taking account of the steady displacement of the mesoscale system.

As pointed out in Section 1, the characteristic wind change accompanied with the PD phenomena was veering winds followed by backing winds with strong gusts. Easterly wind anomaly is growing stronger in front of the PD line, and weaker in the rear of it (see Fig. 4(b) and enlarged figure in Fig. 5). This situation suggests the divergence and the convergence region to the east and to the

west of the PD line respectively. It is also found that the wind anomaly field as well as the pressure anomaly field, shows a uniform structure in N-S direction. Meanwhile, we cannot find any characteristic wind field outside of the MD region, except that only a weak and nonsystematic wind anomaly field existed.

3.4 Precipitation, air temperature and dew point temperature

Precipitation areas where rainfall amount over 0.5 mm were observed in 1500–1520 GMT, are entered by shade in Fig. 5. There appears a large precipitation area extending to the east of the PD line, but no precipitation is observed on and just west of it. This situation is very much similar to that analyzed by Bosart & Cussen (1973). A weak precipitation area was recognized to the north of 35.5°N in the western region of the PD line.

The time series of (t) and (td) assumed almost constant values before and after the PD line passage and there were not any characteristic changes, indicating that this PD phenomenon was quite different from so-called frontal phenomena.

3.5 Doppler radar analysis

The vertical structure of the MD is studied by using the data obtained by a doppler type meteorological radar installed in MRI which is located in Tsukuba area. Detailed investigations with the radar data reveal fine structures within the large cloud system, and the precipitation area will be classified into strong rainbands (or enhanced convective activity) and weak rainbands (or suppressed convective activity) in the following discussion.

Since the radar backscattering intensity is

Table 1 Station list of the arrayed network implemented around Mt. Tsukuba area. The time of the PD passage is entered in the second column.

Station	GMT	N	E	height
a) Makabe Machi	1541	36°16.56'	140°05.70'	40 m
b) Tsukuba Y.H.	1541	13.87	06.32	550
c) Mt. Tsukuba Obs.	1541	12.59	06.12	250
d) Tsukuba Machi	1539	10.23	05.50	20
e) Tatenno (MRI)	1542	03.12	07.59	25

Fig. 7 Fine structure of convective rainband as observed by Doppler-Radar at 1504 GMT Oct. 22, 1981. The broken line labelled with PD indicates the PD line on the surface, and D and E region represent the long systematic suppressed and enhanced cloud system respectively. The deviation of radar intensity for the cross section perpendicular to the PD line is presented on the upper right.

subject to the loss effect by raindrops, the authors tried to eliminate its effect from the net intensity by a running mean method. The deviation from the running mean value of radar intensity on each azimuthal scanning line is considered to represent relatively enhanced or suppressed cloud activity. The averaging width perpendicular to the MD is 15 km. Shaded areas in Fig. 7 show enhanced clouds region, where the radar backscattering intensity were relatively strong. There is a suppressed rainband D to the east of the PD line. Region E, appearing around the PD line, is a well organized enhanced rainband. On the other hand, there are some less organized, smaller scale rainbands at some distance from the PD line.

The movement of D and E is tracked by using sequential radar data. It is found that the velocity of D and E was equal to the

steering velocity of the PD line.

4. Structure of MD

4.1 Horizontal structure of MD

Based on the surface wind data, we derived higher order fields such as the divergence, the vorticity and the two deformations denoted by $D=U_x+V_y$, $R=U_y-V_x$, $A=U_x-V_y$ and $B=U_y+V_x$ respectively. We also introduced the anomaly fields by taking deviations from running means as $d=u_x+v_y$, $r=u_y-v_x$, $a=u_x-v_y$ and $b=u_y+v_x$, where x and y are the horizontal coordinate towards east and north respectively, u the horizontal velocity component in x -direction and v in y -direction.

In Fig. 8 is shown the time series of d , r , a and b at Tokyo (st. d). We find remarkable changes with the passage of the PD line (1510 GMT). A very similar change found in d and a suggests that the change of u_x term was dominant in the time series. This is undoubtedly related to the change of wind direction before and after the PD line passage, and also to the uniformity along y -direction, as we have already seen in Fig. 5.

The broken line A-A' in Fig. 6 is obtained by connecting stations near the center of the PD line passage within a dense wind data network. The time sections of D , d , r and a along this line are shown in Fig. 9(a), (b), (c) and (d) respectively. It is clearly seen in the time section of D that the divergent field is discontinuous in the region of the MD just as

Fig. 8 The time variation of d , r , a and b at Tokyo (st. d). The passage of the PD line took place at 1510 GMT Oct. 22, 1981, and remarkable changes appeared at the time.

Fig. 9 Time sections along the A-A' line given in Fig. 6, (a); *D* (b); *d* (c); *r* and (d); *a*. Solid lines in (a) represent the isobars of the pressure anomaly (mb), and shaded area the negative value region. The unit of each section is 10^{-4} s^{-1} .

suggested by the characteristic wind variation accompanied with the passage of the PD line. A wave-like structure around the PD line appears clearly in the anomaly time section *d* but not so in *D*, and the wave length is found to be about 90 km. The time section of *r* shows that the vorticity is rather weak just on the PD line.

The time section of *a* shows a more remarkable response to the passage of the MD than the case of *d*. A positive region in front of the PD line and a negative region in the rear of it appeared clearly throughout 1300–1700 GMT. The positive *a* is a deformation rate made up of expansion in the east-west and contraction in the north-south. Therefore the systematic negative *a* in the rear of the PD line indicates that the western region of the MD was characterized by pre-

dominant u_x (contraction in the east-west as we have already seen in Fig. 5) reinforced by expansion in the north-south direction.

4.2 Fine Structure of MD Observed at MRI

Since the wind near the surface is considerably influenced by the ground roughness, the signal-to-noise ratio on the surface wind recorder is inferior to that on the upper wind. Estimating of the ratio for the meteorological tower in MRI, we found the value at the higher level (213 m height) showed five times as much as one at the lower level (10 m height).

By using every one minute digitized wind data of the higher level, the authors examined the detailed anomaly wind field. The direction of u is chosen along the steering direction of the PD line determined by the baro-

(a)

(b)

Fig. 10 (a) : The power spectra of the anomalies (p , u and v). (b) : The time series of the anomalies (dotted lines) and the bandpass-filtered anomalies. Solid and broken lines represent the 80 min and 30 min period bandpass-filtered anomalies respectively.

graph array observation near Mt. Tsukuba (see Fig. 6). The authors also calculated the pressure anomaly field in more detail by using the pressure data at Tateno (located in the same place as MRI), and the sampling interval of pressure and wind data are 2.5 and 1.0 min

Table 2 Response of the bandpass filter used in this analysis.

No.	T_l	T_c	T_s
1	160 min	80	53
2	40	32	27

respectively.

In Fig. 10(a) is shown the power spectra of p , u and v . It is seen that the spectral peak around 80 min is dominant. In order to examine fine structure of the MD, a recursion type digital *bandpass filter** was applied to the original time series (as shown in Murakami, 1979). This filter is designed to retain the oscillation around the central period (T_c), as shown in Table 2.

In Fig. 10(b) are presented the anomalies (p , u and v) and the bandpass-filtered anomalies (p_{80} , u_{80} and v_{80} ; p_{30} , u_{30} and v_{30}). The MD is recognized as a plane wave propagating eastward with a period of 80 min (or about 90 km for wave length) to the first approximation. The pressure anomaly (p_{80}) shows an almost *one to one* correspondence to the wind anomaly (u_{80}). This linear type connection between pressure and wind strongly supports the idea that the PD phenomenon was characterized by IGW.

The second mode fluctuation with a period of about 30 min is noticed in Fig. 10(a), and is clearly depicted in the bandpass-filtered anomalies, p_{30} , u_{30} and v_{30} . The corresponding wave length of about 30 km well coincides with the interval of rainbands appearing in Fig. 7. Therefore we can infer that the fine structure of the MD derived from doppler radar analyses represents a characteristic field of the second mode fluctuation. This second mode fluctuation within the immediate vicinity of the PD line assumes the effect to strengthen the through of the IGW. Meanwhile, the first mode fluctuation with a period of about 80 min corresponds to the wave length of 90 km which we have already seen in the horizontal structure of MD as shown in the former sec-

* The response function is designed to retain more than 50% of the amplitudes of the oscillation between the shorter period (T_s) and the longer period (T_l).

tion.

4.3 Vertical structure of MD observed by doppler radar

In Fig. 11 are shown the cloudiness and horizontal velocity distribution in a cross section perpendicular to the MD. This is obtained by composing every two minutes interval PPI observations from 1502 to 1516 GMT Oct. 22, 1981 and applying the same running mean treatment as mentioned in Section 3.5. F and G show the region of weak radar backscattering intensity lower than 40 dB. The surface precipitation area of more than 0.5 mm during the period from 1500 to 1520 GMT is also presented. No precipitation is found on and to the west of the PD line. We have already seen in Fig. 7 the horizontally stripe-like structure of enhanced and suppressed elongated rainband cloud systems, and the corresponding structure appears in the vertical cross section presented in Fig. 11.

The horizontal wind distribution is entered by broken lines. A remarkable vertical wind shear is observed in the level around 2.5 km

Fig. 11 The composite vertical cross section of the MD revealed by Doppler-Radar in MRI. The shaded regions correspond to relatively strong rainband cloud system. The isotachs of horizontal wind estimated by the doppler velocities are given by broken lines. Dotted line represents the position of the maximum wind in each level. The amount of precipitation on the surface for the period from 1500 to 1520 GMT is shown on the bottom.

height. The actual height of this wind shift from easterly to westerly would be somewhat higher if we take a falling velocity of raindrop into consideration for the apparent doppler velocity, and this level would be found at about 700 mb height. The existence of this remarkable wind shear verifies the assumption mentioned in Section 2, that a stably stratified layer would be built up with the upper airmass flowing from the northern cyclones (dry and westerly) and the lower airmass flowing from the typhoon (wet and easterly).

The doppler velocity V_d is defined by

$$V_d = V_h \cos \chi + (V_w - V_r) \sin \chi, \quad (3)$$

where V_h is horizontal air flow in the vertical plane under consideration, V_w updraft, V_r raindrop falling velocity and χ radar elevation angle, respectively. In the case of $\chi \ll 1$, we have

$$V_d \cong V_h = \bar{U} \sin \phi + u, \quad (3)'$$

where $\bar{U}$ is wind speed of the typhoon circulation estimated by running mean, ϕ wind direction relative to this section and u anomaly wind speed respectively. The variation of V_d is mainly dependent on the variation of u since $\bar{U} \sin \phi$ is of synoptic scale and uniformly distributed in this section. The ridge of the isotach in the lower layer is found to coincide with the PD line on the surface. It should also be pointed out that not only in the lower layer but also in the upper layer u had extremum values above the PD line. The dotted line is the axis of this extreme horizontal wind. Fig. 11 also suggests that there would be a divergence and a convergence region in front and in the rear of the dotted line respectively in the lower layer, and the situation is reversed in the upper layer.

The cloud system D and E show well organized structures in accordance with their horizontal structure presented in Fig. 7. On the other hand, the cloud system A, B and C are poorly organized. The cloud system D and E obviously play an important part in the second-order anomalies (p_{30} , u_{30} and v_{30}), and the PD line is found to be located on the boundary between D and E. Therefore, it is inferred that the MD is characterized by

downdrafts in front and updrafts in the rear of the PD line. In this respect it may be said that the second-order anomalies show up more clearly the gravity wave characteristics.

4.4 Summary of the observation

A model structure of the MD observed in T8124 is shown in Fig. 12. Above the MD, we observe a stratified wind system in which a dry westerly (W_U) originated from the Continent prevails in the middle troposphere, and a wet northeasterly (W_L) typhoon circulation is found underneath. A stably stratified layer with a strong vertical shear of horizontal wind was formed at about 700 mb level. This

situation is likely to be maintained in the northwest quadrant of T8124.

The minimum surface pressure of the MD was almost comparable to the central pressure of the typhoon itself, and its pressure gradient was one order larger than that of the typhoon. The disturbance had a lifetime of more than 5 hours, and this PD phenomena is not considered as just a temporal deformation of the typhoon, but as a substantial phenomenon associated with the typhoon structure.

The MD had the spatial scale; a length extending from north to south of 200 km; a width ranging from east to west of 90 km; a depth of 5 km. It moved almost eastward keeping a wave front structure. The isochrones of the PD line indicate apparent northeastward movement, inferring the continuous generation of new cells at the extreme north and dissipation of old ones at the south. This situation is very similar to the displacement of rainbands within a typhoon, in which new rain cells appear at the windward end of the rainbands and old ones dissipate at the leeward end.

According to rain gauge observations, the principal rain area is found to the east of PD line just as shown by Bosart & Cussen (1973). They pointed out that there appears to be a *shutting off* of precipitation associated with rapid decrease in convergence after the wave passage. Meanwhile the doppler radar observation revealed smaller scale structure of the rain area around PD line. A divergence region accompanying downdraft is found immediately to the east of PD line, where convective activities are suppressed but still precipitations are observed. To the west of PD line, on the other hand, a convergence region accompanying updraft is observed, where convective activities are enhanced although no raindrop reaches to the ground. In the case studied by Uccellini (1975), the peak in magnitude of rainfall rates was observed after the passage of the wave trough just as shown by Matsumoto & Ninomiya (1965) or by Matsumoto & Akiyama (1969). Concerning the divergence distribution around the PD line, it is pointed out that a maximum

Fig. 12 Model structure of the MD within the typhoon. A shaded region represents the PD zone, and the arrow labelled W_U and W_L represent the trajectories of the upper and lower air mass respectively. The pressure profile on the bottom is for the solid line B-B'. The minimum value of 967.4 mb was observed at Irozaki (1410 GMT). Along the broken line $T_{18}-H_{13}-T_{12}$, the vertical cross section of MD presented in Fig. 3(b) was composed.

convergence is observed after the pressure minimum passage in each of the cases.

It should be emphasized that the small Ri region over the PD line is influenced considerably by the strong vertical wind shear in stratified layer about 700 mb level (see Fig. 3(b)) and that the MD was generated somewhere over the sea, and therefore, the orographic interaction is considered to be of secondary effect.

5. IGW in two-dimensional two-layer model

There have been many studies about IGW occurred in a stably stratified layer with a marked vertical shear (Goldie, 1925; Ewing, 1950). And some theoretical modellings for IGW have been developed (Thompson, 1961; Gedzelman & Rilling, 1978).

A simple model of two-dimensional flow with two density layers, was applied to investigate the characteristics of the IGW. The flow is assumed to be incompressible, hydrostatic and inviscid. In this model, the atmosphere is regarded as two homogeneous layers of fluid with uniform densities ρ_1 and ρ_2 without a vertical velocity shear, separated by a surface of density discontinuity, the height of which is H . It will also be supposed that there exists some level in the upper layer (with density ρ_1) where the horizontal pressure gradient vanishes. It is assumed that the wave motion is in x -direction. Governing equations in the lower layer are given (Thompson, 1961) by,

$$P_x = g(\rho_2 - \rho_1)H_x \quad (4)$$

$$U_t + UU_x + \rho_2^{-1}P_x = 0 \quad (5)$$

$$H_t + UH_x + HU_x = 0 \quad (6)$$

where g is the gravitational acceleration, P the pressure, H the height, and U the velocity of the lower layer.

We substitute the expressions $(\bar{u} + u)$ and $(\bar{h} + h)$ for U and H in the above equations, and linearize them. If we introduce perturbations and assume that all the perturbations can be expressed by $\exp(i(\kappa x - \omega t))$, we obtain

$$c = \omega/\kappa = \bar{u} + (g\bar{h}(1 - \rho_1\rho_2^{-1}))^{1/2}, \quad (7)$$

where c is the phase velocity of the disturbance.

Considering a stably stratified layer between 600 and 700 mb level height, we assume

$$\begin{aligned} 1 - \rho_1\rho_2^{-1} &\cong 1 - \theta e(\text{lower})/\theta e(\text{upper}) \\ &= 1 - 342.5/347.5 = 1.4 \times 10^{-2}, \end{aligned}$$

and

$$\bar{h} = H(700 \text{ mb}) \cong 3.0 \times 10^3 (\text{m}).$$

If we substitute these values into equation (7), we obtain

$$c - \bar{u} = 21 \text{ m/s} = 74 \text{ km/h}.$$

This gives a good agreement with the propagation velocity of the MD as shown in Section 3.2, since $\bar{u}$ has obviously a negative value.

Now, for the pressure perturbation, we obtain

$$p_0 = \rho_2(c - \bar{u})u_0, \quad (8)$$

where p_0 and u_0 is the amplitude of the pressure and velocity perturbation respectively. This indicates a linear connection between the pressure anomaly and the wind anomaly, as shown in Section 4.2. If we take $p_0 = 4.5 \times 10^2 (\text{Pa})$ from the full amplitude of p_{80} , we then obtain $u_0 = 22 (\text{m/s})$. This value is nearly equal to the full amplitude of u_{80} . The same statement is true for the relation between p_{30} and u_{30} .

Since p is in proportional to u , the front region of the IGW (where $p_x > 0$) is to be divergent (because $u_x > 0$), the rear region (where $p_x < 0$) be convergent (because $u_x < 0$). These situations have good correspondence with the PD phenomena.

Applying similar governing equations to the upper layer, the pressure perturbation would be positive while the pressure perturbation in the lower layer is negative. This suggests that there would be a reversed *pressure jump-drop* phenomenon in the upper layer higher than 700 mb level. In fact, a *pressure jump-drop* of +1.2 mb was observed in the barograph trace at Mt. Fuji (3,773 m height) at 1410 GMT. The divergence distribution in the upper layer should be reversed in the lower layer, which is consistent with the observation (see Fig. 11).

6. Remarks

Generally speaking a pressure drop phenomena is caused by a decrease in the weight of air-column. Fujita (1952) considered removal of water vapor as raindrops within rain cloud. For a pressure jump phenomena, Fujita (1959) suggested a role of evaporation process on the generation of cold air resulting in *meso-high*. However, the expected pressure difference from these theories are at most 5 mb respectively.

We have already observed in Fig. 2(d) a tapering cloud pattern, which was also detected by the Mt. Fuji radar observation. It is found to be made up of cumulonimbi with 8 km height. Generation of new cumulonimbi is often found just to the north of southern edge of such a system, and a strong updraft accompanies with its growth (Anderson, 1974). It is quite feasible that such a strong updraft may be one of the triggers for the development of an IGW motion in the atmosphere.

Although IGW is originally a stable oscillation, the convective activity is considered to have an effect of increasing the amplitude of IGW (Matsumoto & Ninomiya, 1969), and it is obviously related with enhancement of precipitation rate in the MD. The GMS cloud imageries and the Mt. Fuji radar observations showed the existence of cloud system above the region of the PD line passage, indicating the existence of strong convective activities which would provide a necessary condition for the IGW development.

While there is still uncertainty about the origin of the typhoon rainband, a variety of physical mechanisms were proposed to explain the formation of it. Since there are many reports in which the typhoon rainbands had IGW characteristics, Willoughby's study (1978) in which the rainband was some form of IGW is a leading investigation for the origin of it. It is quite natural that an appropriate stratified condition in the vicinity of a rainband which is a systematic disturbance organized in a line would maintain an IGW and enable us to observe a strong pressure anomaly.

The convective activity corresponding to E (in Fig. 11) was located just after the PD line.

Here it is perhaps appropriate to assume a possible effect on decreasing precipitation due to the strong updraft accompanied with the convergence of -10^{-3} s^{-1} (as suggested in Mitsuta *et al.*, 1979), and, also due to the insufficient supply of water vapor in the westerly dry airmass.

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台風 8124 号で観測された重力波について

松 本 誠 一*, 岡 村 博 文

気象研究所

台風 8124 が房総沖を通過した頃、その北西象限に中規模擾乱に伴う気圧の dip 現象が現れた。その最低気圧は、台風本体の中心示度 (965 mb) にほぼ近い 967.4 mb を示すものであった。擾乱は台風よりやや速く進み、5 時間以上にわたって確認されている。地上観測と気象研究所のドップラーレーダー観測などにより、この擾乱について調べた結果、風の変化や雨雲の分布等が、よく知られている内部重力波の様相を示していた。また、この擾乱は台風のレインバンドの微細構造と密接な関係があり、dip 現象は東西波長 90 km で南北に連なる積雲活動列と結びついている。

* 現在：日本気象協会・中央研究所