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Heavy Rainfall around the Suzuka Mountains

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Abstract

Analyzing the detailed data on rainfall during the warm season for three years (1967–1969) which include the data of our own observations around the Suzuka Mountains with other meteorological elements, a few characteristic features of the heavy rainfall were clarified and several problems concerning them were revealed.

In the case of heavy rainfall in which the maximum 24-hour precipitation was about 200 mm or more, strong easterly winds prevailed in the lower troposphere, the middle and lower troposphere was nearly saturated with water vapour, and the static stability was nearly neutral. The heavy rainfall area around the mountains shifted more or less westward according to the easterly component of the wind in the upper troposphere. Such contribution of the upper wind to the orographic rainfall was theoretically considered after the dynamical model proposed by Sarker (1966).

Comparing the results with the observation, it was suggested that our consideration was realistic.

1. Introduction

We have a share in the study of precipitation in the basin of Lake Biwa which is one of the research subjects of the project of the IHD (International Hydrological Decade) in Kyoto University. For that we began in 1967 the observation of rainfall amount during warm season around the Suzuka Mountains which lay along the south-east boundary of the basin of Lake Biwa. At our observation point initially set near to Mt. Gozaisho very large precipitations comparing to the other former stations around this mountain were often observed. So we tried to complete the network of rainfall observation for precise rainfall distribution in this area.

In this paper we shall investigate the relation between the detailed distribution of heavy rainfall around the Suzuka Mountains and the meteorological situation making use of data of three years (1967–1969) and point out some problems concerning them. Finally we consider theoretically the orographic rainfall in the case in which strong winds across mountains prevail in the lower troposphere, the air is saturated with water vapour, and the static stability is neutral.

2. Relation between Rainfall Distribution and Meteorological Situation

Lake Biwa and the Suzuka Mountains are located in the central part of Japan as shown in Fig. 1. The Suzuka Mountains extend nearly north and south with peaks of about 1000 m height as shown in Fig. 2. The observations of rainfall are carried out centering around Mt. Gozaisho of 1210 m height, which is the highest peak after Mt. Amagoi (1238 m) located west of Mt. Gozaisho. The observation points of rainfall were shown also in Fig. 2. However in this paper we do not make use of the data of observation points set in 1970.

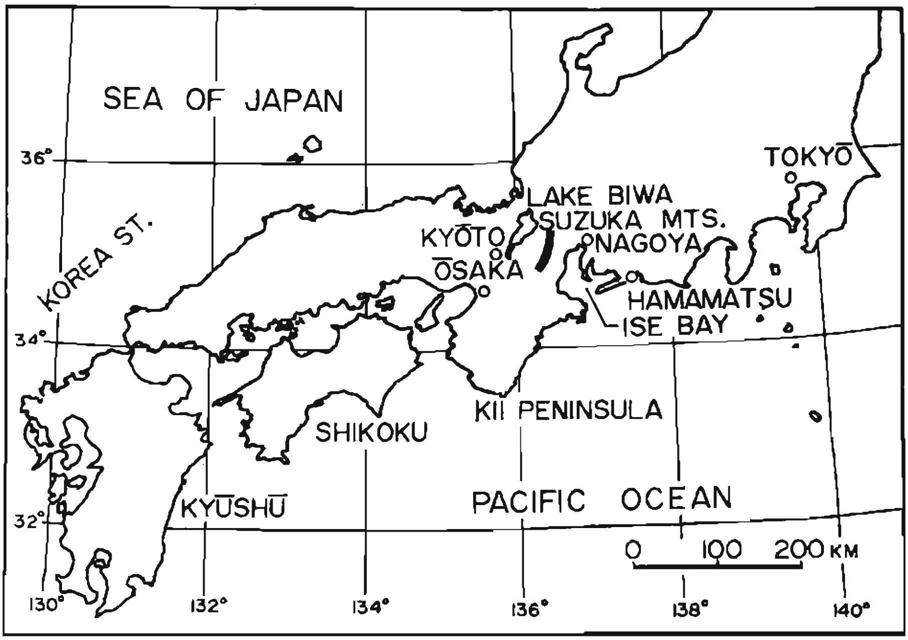


Fig. 1. Map of Central and Western Japan.

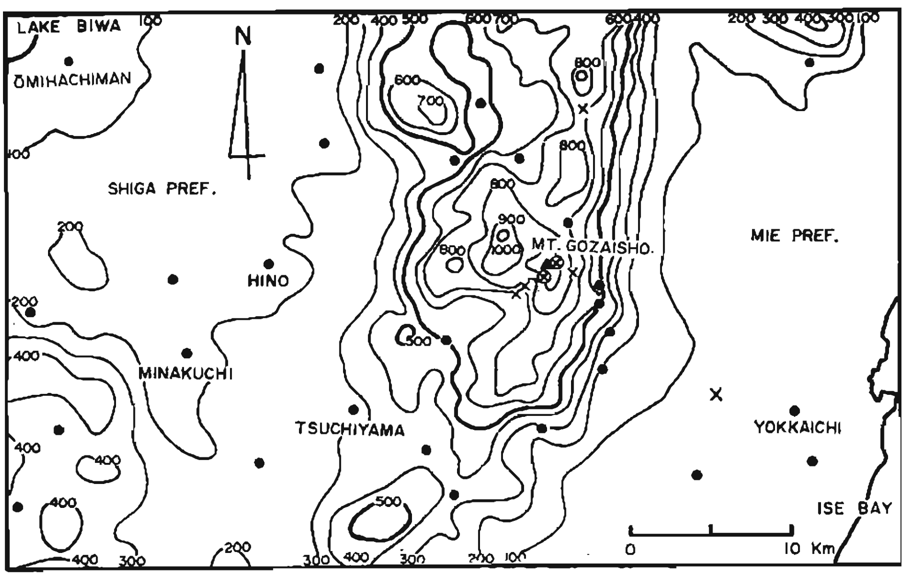
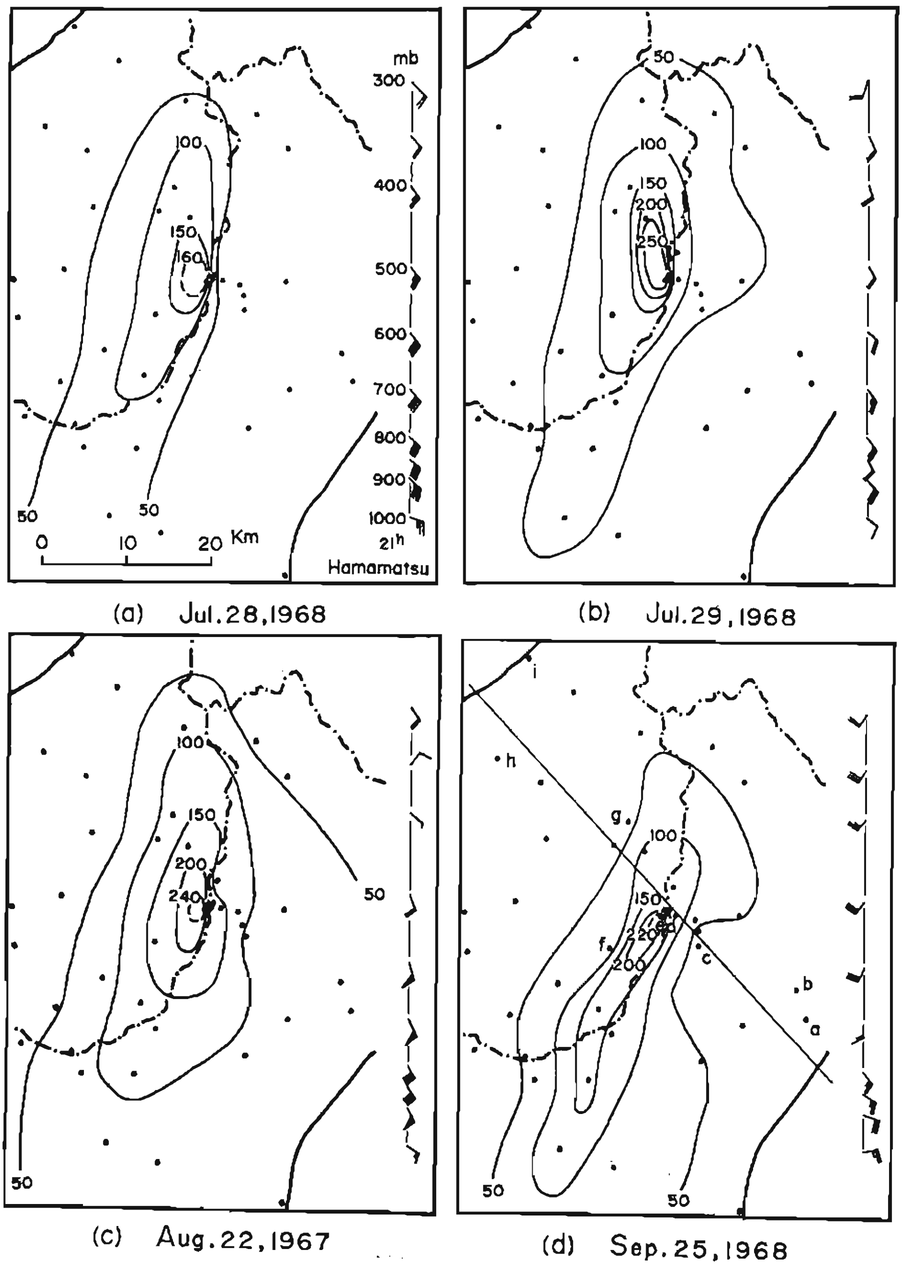


Fig. 2. Topographic map around the Suzuka Mountains. Contour lines: 100 m intervals. Observation points: ● by Provincial Meteorological Observatories, Regional Construction Bureaus and others; ⊗ set by us in 1967 and 1968; × set by us in 1970.

As warm moist air can be directly carried to this area through Ise Bay from the Pacific Ocean, the Suzuka Mountains have high frequency of heavy rainfall. Heavy rainfalls above 200 mm/day occur at the rate of 0.5 times per year according to the observations by Japan Meteorological Agency.[1] But in the case in which the other observation points were added, heavy rainfalls above 200 mm/day and 100 mm/day occurred 3 days and 17 days respectively for the 3 years, 1967–1969. Typical examples of heavy rainfall among them are shown in the following:

2.1. Cases of 28 and 29 July 1968

The distributions of 24-hour (9^{h}–9^{h} (JST)) precipitation around the Suzuka



(a) Jul.28,1968 (b) Jul.29,1968
(c) Aug.22,1967 (d) Sep.25,1968

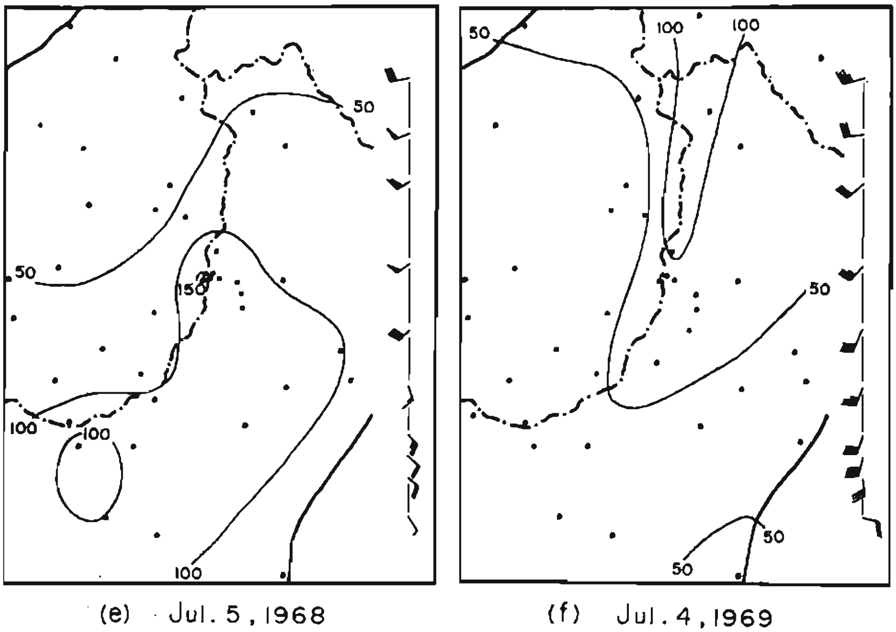
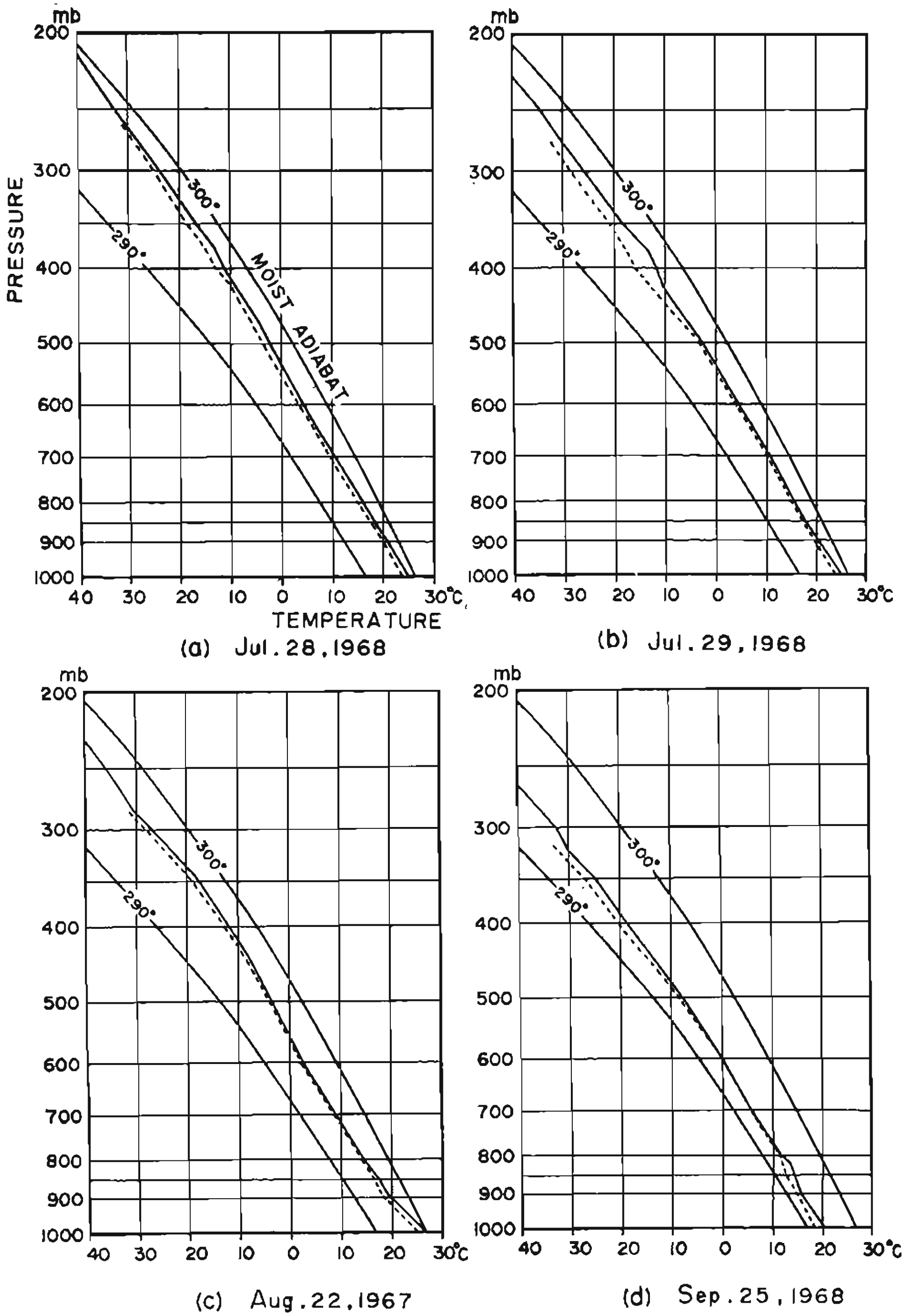


Fig. 3. The distributions of 24-hour precipitation (in mm) around the Suzuka Mountains and the vertical distributions of wind (a long barb 10 knots) for 21^h (JST) at Hamamatsu.

Mountains and the vertical distributions of wind for 21^h(JST) on the days over Hamamatsu at a distance of about 120 km from the mountains to south-east are shown in Fig. 3. (a) and (b) in Fig. 3 are in the case of 28 and 29 July, 1968. The maximum total precipitation for three days (7/28–30) was 566 mm at the point *e* in Fig. 3 (d). In this period typhoon 6804 moved slowly in the vicinity of Shikoku and Kyushu, and south-easterly winds prevailed almost throughout the troposphere and the maximum wind speed reached 40–45 knots in the lower troposphere as seen in Fig. 3. Such a synoptic situation continued for the three days. The vertical distributions of temperature and pseudo-wet-bulb temperature corresponding to the case indicated in Fig. 3 are shown in Fig. 4. As seen from (a) and (b) in Fig. 4 at least middle and lower troposphere was nearly saturated with water vapour and the static stability was nearly neutral.

Comparing the distributions of precipitations in these two days, for instance the shapes of isohyets of 100 mm are not so different from each other. But the maximum precipitation of the case of (a) is 163 mm and that of (b) is 259 mm, the latter is larger than the former by about 100 mm. On the other hand, comparing the upper winds of these, the wind speed in the case of (a) is larger than that of (b) by about 10 knots at any level. The data of surface winds indicate also the same tendency. Although such strong winds across any mountains are expected to be favourable for orographic rainfall, the maximum precipitation of the case (b) is larger than the case (a). This is an interesting problem. But, if we compare the vertical distributions of both temperature and pseudo-wet-bulb temperature in Fig. 4 (a) and (b) the static stability above about 550 mb level is more stable in the case of (a) than in the case of (b).


mb
PRESSURE
TEMPERATURE
°C
300°
290°
MOIST ADIABAT
(a) Jul. 28, 1968
(b) Jul. 29, 1968
(c) Aug. 22, 1967
(d) Sep. 25, 1968

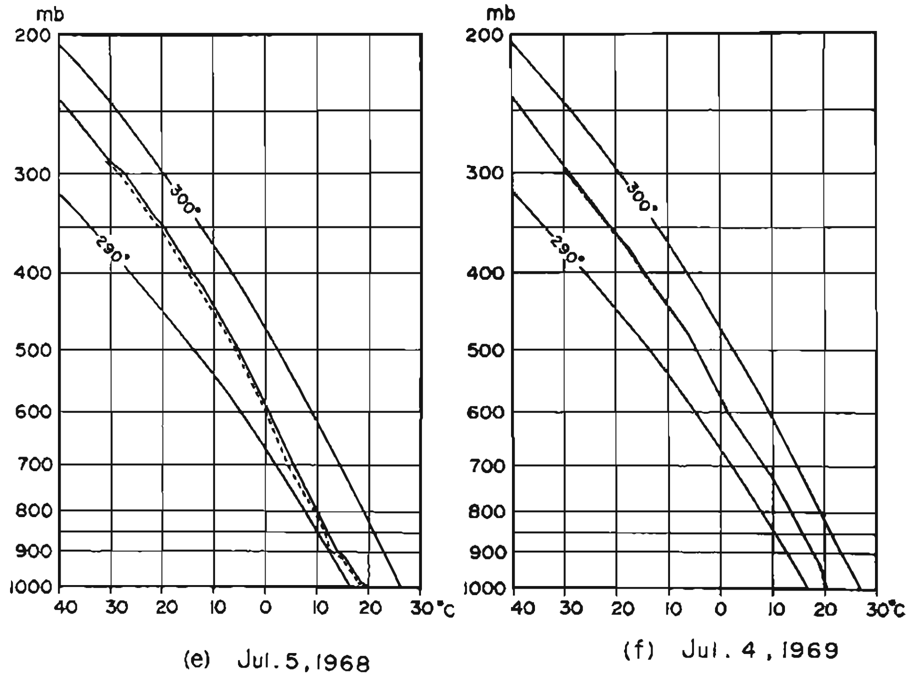


Fig. 4. Emagrams for 21^h (JST) at Hamamatsu. Solid lines: temperature. Dashed lines: pseudo wet-bulb temperature.

But, because of the insufficiency of observation points of rainfall at this stage, the isohyets in Fig. 3 (a) and (b) are not so precise, so we cannot continue the detailed analysis. It is, therefore, clear that new observation points set in 1970 as shown in Fig. 2 are very important.

2.2. Case of 22 August 1967

The distribution of 24-hour precipitation for 22 August 1967 is shown in Fig. 3 (c). Comparing this case with the case (b), the area of this heavy rainfall is broader considerably than the case (b) and the area enclosed by the isohyet of 100 mm, for instance, spreads out to the eastward though with less maximum precipitation, that is 240 mm. As synoptic situation typhoon 6718 was moving north-eastward along the south-east coast of Kii Peninsula. A very strong south-easterly wind prevailed under about 600 mb level with maximum speed of about 50 knots, but the wind above this level was very weak. Though there was a thin unstable layer under 900 mb level, as seen from Fig. 4 (c), the stability was on the whole nearly neutral for moist air, as in the case of Fig. 4 (b).

We will mention the comparison of all these cases in 2.6.

2.3. Case of 25 September 1968

The distribution of 24-hour precipitation for 25 September 1968 is shown in Fig. 3 (d). The maximum total precipitation at the point *e* for three days (9/24–26) was 380 mm. The maximum precipitation was observed at the same point as in those cases above mentioned and the area of heavy rainfall extended also along the mountains. But the area enclosed by the isohyet of 100 mm spreads out more to the eastward than the others. As synoptic situation during the three days typhoon 6816 almost always stayed west of Kyushu. In the layer, from the surface to about 800 mb level, east and south-east winds prevailed with a maximum speed of about 30 knots, but the wind changed to south and south-west above the 800 mb level. Though temperature is lower than in the case of (c) by about 5°C throughout troposphere, as seen from Fig. 4 (d), the air is nearly saturated and the stability is almost neutral also.

For 24–26 September we tried to analyze the meteorological surface data somewhat in detail. The time change of hourly precipitation for this period at each point shown by the alphabet in Fig. 3 (d) is indicated in Fig. 5. Rainfall begins during the night of the 24th at the earliest point. This time, as seen from the anemographs at Yokkaichi (*b* in Fig. 3 (d)) of Fig. 6, corresponds to the time when the east wind in the lowest

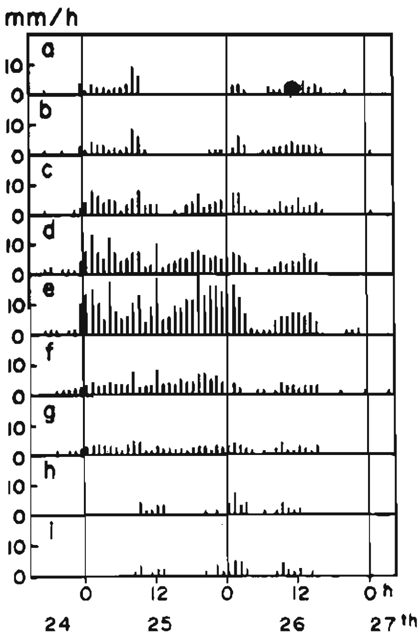
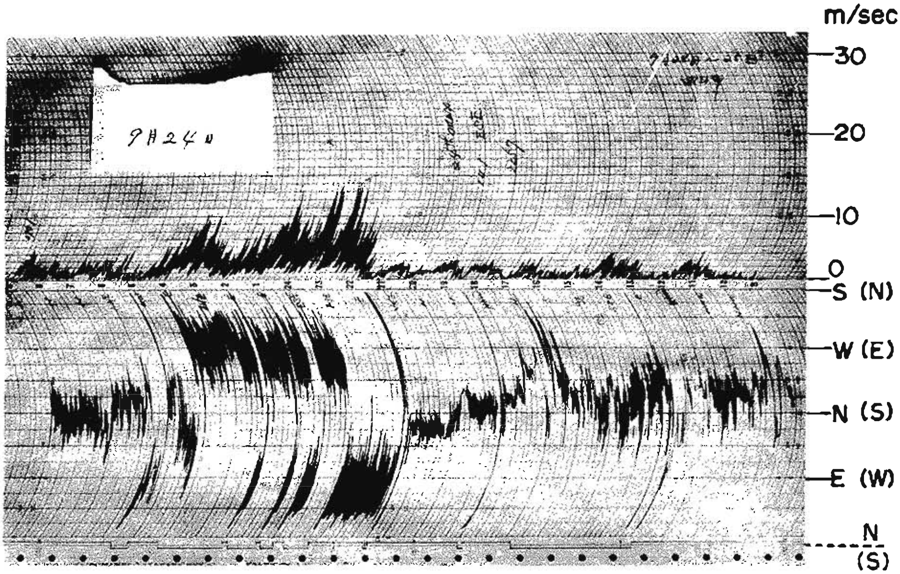
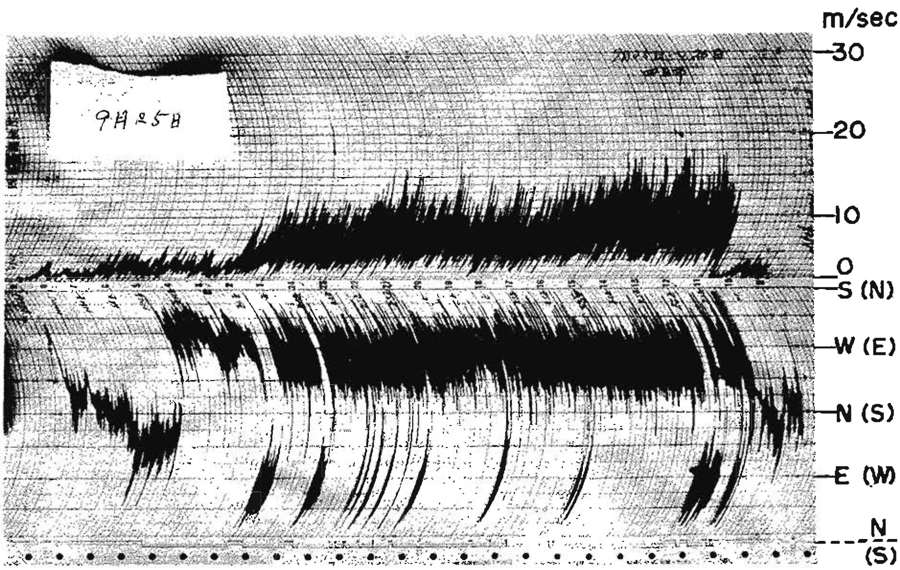


Fig. 5. The time change of hourly precipitation at each point shown in Fig. 2(d).

troposphere became strong. From about the noon of the 25th rainfall stopped for a while over the lowland (points *a*, *b*, *h*, and *i*), but over the mountains (points *c*, *d*, *e*, *f*, and *g*) rainfall continued, especially at the points *d*, *e*, and *f* rainfall became rather heavier. In Yokkaichi rainfall stopped at about 11^h on the 25th, after which east-south-east winds became strong, having been weak for a while, as seen from Fig. 6 (a) and (b). At the same time the humidity at Yokkaichi also fell from about 95 to about 80% (not shown as a figure). However the meteorological state of upper air over Hamamatsu at 9^h on the 25th was almost same as the state at 21^h (Fig. 3 (d) and Fig. 4 (d)), therefore it is difficult to determine whether this fall of humidity in



(a) Sep. 24～25, 1968



(b) Sep. 25～26, 1968

Fig. 6. Anemographs at Yokkaichi.

Yokkaichi is the cause or the result of no rain.

The time changes of wind speed at the top of Mt. Gozaisho and 10 minutes precipitation at point *e* where the maximum precipitation was observed are shown in Fig. 7. Wind direction was not observed, but presumed to be south-east. It is seen that such

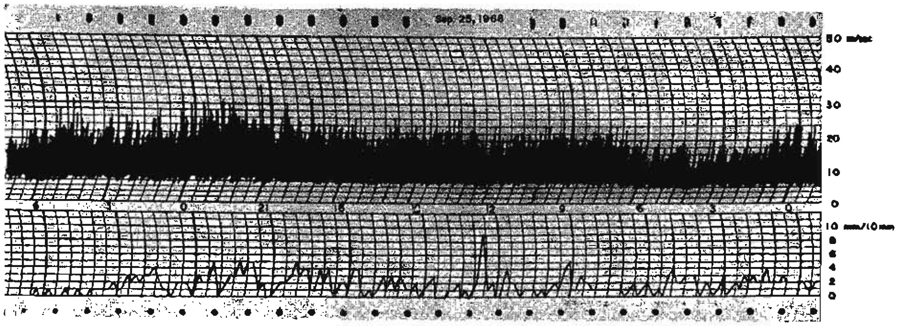


Fig. 7. The time changes of wind speed at the top of Mt. Gozaisho and 10 minutes precipitation at *e* point shown in Fig. 2(d).

rainfall over the mountains has considerably intense time change. On the whole, the rain intensity tends to increase according to wind speed. The same relation is seen also in detail, though the correspondence at 10^h–13^h on the 25th is wrong. Rainfall became light on the 26th, but the wind almost did not change. However the stability becomes considerably stable, though the air remains almost as saturated as before.

2.4. Case of 5 July, 1968

The distribution of 24-hour precipitation for 5 July, 1968 is shown in Fig. 3 (e). In all of the above mentioned cases rainfalls occurred in connection with typhoons, but this rainfall occurred when a developed cyclone moved eastward along the south coast of Western Japan. The maximum precipitation was also observed at the same point as in the previous cases, that is located west of Mt. Gozaisho, but it is not so large comparing the surrounding. It is a characteristic that the precipitation is large over the lowland of windward (east) side of the mountains though not so large. The upper wind was south-east with the maximum speed of about 25 knots under 700 mb level, but above this level the relatively strong south-west wind prevailed with a maximum speed of about 60 knots. As seen from Fig. 4 (e) which shows the vertical distributions of temperature and pseudo-wet-bulb temperature, the air is almost saturated and the stability is, on the whole, considerably stable, except for the thin layer in the lowest troposphere.

2.5. Case of 4 July, 1969

The distribution of 24-hour precipitation for 4 July 1969 is shown in Fig. 3 (f). In all of the above mentioned cases strong easterly winds (mainly south-east wind) prevailed in the lower troposphere under, at least, the 800–700 mb level, but in this case westerly winds prevailed on the whole, except in the lowest layer, which had east to south winds. As the synoptic situation there was a warm front along the south coast of Japan with a cyclone moving eastward in the vicinity of the Strait of Korea. The number of cases in which heavy rainfalls occur with such a westerly upper wind is three times in these three years, which are the cases of 25 August, 1968 and 25 June, 1969, and this case. In these cases, precipitations over the usually less rainy lowland are several tens of mm and not too small. Though the rainfall is certainly heavy along the mountains, the maximum precipitation is about 150 mm in these years. Therefore the rainfall distributions are relatively uniform. The shape of heavy rainfall area is nearly symmetrical for the ridge, though the details are not clear on account of the insufficiency of the number of observation points. The air is almost saturated and the stability is nearly neutral as seen from Fig. 4 (f).

2.6 Comparison of these cases

The typical heavy rainfalls around the Suzuka Mountains are mentioned and several problems concerning them are pointed out above. Here, these cases are compared each other.

Comparing (a), (b), (c) and (d) in Fig. 3 with each other, the maximum precipitations are observed at the same point *e* indicated in Fig. 3 (d) that is located west of Mt. Gozaisho. The maximum precipitations are much larger than the precipitations in the surrounding and we can say that the heaviest rainfall areas lie just west (the leeward for easterly winds) of the mountains. The fact that the heavy rainfall area is along the mountains seems to indicate the rainfall to be orographic. As the values and locations of the maximum precipitations cannot be known precisely, as mentioned in 2.1, we have given attention to the relative position of the isohyets of 100 mm. Though the positions of them in (a) and (b) are almost same, in (c) it shifts eastward somewhat, while in (d) the eastward shift is greatest among them. Comparing the vertical distributions of upper winds: though in every case strong easterly winds prevail in the lower troposphere, in (a) and (b) relatively strong easterly winds prevail also until the upper troposphere, but in (c) it becomes weak above about 500 mb level and in (d) it changes to the south-west wind above about 600 mb level. As seen from (a), (b), (c) and (d) in Fig. 4, in every case the air is almost saturated and the stability is nearly neutral. In every case the meteorological state is considerably steady during a rainfall. It seems that these facts indicate that the easterly component of the wind in the upper troposphere shifts the heavy rainfall area to the westward, when strong easterly wind prevails in the lower troposphere, the upper air is almost saturated and the stability is nearly neutral. Moreover, as long as we compare (b), (c) and (d) in Fig. 3, the maximum precipitations increase according to the westward shift of the heavy rainfall area. However, more precise facts will make a clearer picture in future. Those mentioned above seem to be explained easily by the strength of wind, but the correspondence between precipitation and wind velocity is not always good. When strong easterly wind prevails in the lower troposphere, the heavy rainfall area over the south-eastern part of the Kii Peninsula tends also to shift more or less westward according to the easterly component of the wind in the upper troposphere[2]. Therefore this seems to have generality, so we try to consider such effect of upper wind theoretically in the next section.

Comparing (e) with (d) in Fig. 3, the isohyets of 100 mm in the case (e) shifts strickingly eastward and the south-west wind in upper troposphere becomes stronger. This is apparently similar to those mentioned above. However, as seen from Fig. 4 (e), the stability is clearly stable differently from others, though the air is almost saturated. Moreover we have two cases (on 13 September, 1968 and 8 July, 1969) similar to this, in every case vertical distribution of temperature indicates stable stability. Consequently we think that this case must be distinguished from the cases of 2.1, 2.2 and 2.3, though the effect of such stability is not yet clear.

Giving attention to the isohyets of 200 mm in (b) and (d) of Fig. 3, though both are west of Mt. Gozaisho, one in the case (b) extends northward and in the case (d) southward. Examining the upper winds corresponding them, it is found that the southerly component of easterly strong wind in the case of (b) is larger than (d). In this way the heavy rainfall area in the case of strong easterly wind tends to shift more or less northward according to the southerly component of the wind. It seems to be caused by the effect of Ise Bay. But those will become more evident with the future observations.

On the case of 2.5, we can only say that upper winds are, on the whole, westerly, and distribution of precipitation has somewhat different features from others.

3. Some Considerations in Orographic Rainfall

As described in the previous section, in the case of heavy rainfall in which the maximum 24-hour precipitation near the ridge of the Suzuka Mountains was above about 200 mm, strong easterly wind prevailed in the lower troposphere, under at least 700–800 mb level, the middle and lower troposphere was almost saturated with water vapour, and the static stability was nearly neutral.

Sarker (1966)[3] presented the dynamical model for orographic rainfall with particular reference to the Western Ghats Mountains in India, which assumes a saturated atmosphere with pseudo-adiabatic lapse rate and is based on two-dimensional and linearized equations. He computed the rainfall as case studies and obtained considerably good results. The width of the Western Ghats Mountains is about 100 km, on the other hand that of the Suzuka Mountains is only about 20 km. Applying simplified Sarker's model to the mountains with the scale such as the Suzuka Mountains, we consider the orographic rainfall:

We assume also two-dimensional flow in the vertical plane with the z axis vertical and the x axis in the direction of the undisturbed wind. The undisturbed wind speed, U is a function of z and its direction is normal to the ridge line of the mountains at any level. The origin of the x axis is located on the ridge and the lee side of the mountains is positive. All quantities are invariable in the direction of the ridge line because of the two-dimensional flow. Though Sarker considered the asymmetric mountains for ridge, we considered the symmetric mountains for simplicity. The height of the surface of the mountains ζ_s are expressed by

$$\zeta_s(x)=\frac{a^2b}{a^2+x^2}, \qquad (1)$$

where the height of the ridge b is 0.8 km and the half width a is 4 km. These values are determined with reference to Fig. 2 which indicates the topography averaged with 1 km mesh. The shape of the mountains is shown in Fig. 8 and those that follow.

In order to compute the precipitation or rainfall rate, we must know vertical speed of air flow. Then we try to obtain the vertical speed of the flow over the mountains. If we put the vertical speed

$$w(x, z)=Re\left(\frac{\rho_0}{\rho_z}\right)^{\frac{1}{2}}W(z)\cdot e^{ikx}, \qquad (2)$$

w can be obtained approximately from the solution of the following differential equation.

$$\frac{d^2 W}{dz^2}+(l^2-k^2)W=0, \tag{3}$$

where l^2 is called Scorer parameter,

$$l^2(z)=\frac{g(\gamma^*-\gamma)}{U^2 T}-\frac{1}{U}\frac{d^2 U}{dz^2}. \tag{4}$$

While g is acceleration due to gravity, T and ρ, are undisturbed temperature and density of the air respectively (when $z=0$, $\rho=\rho_0$), γ^* is adiabatic lapse rate, γ is actual lapse rate in the undisturbed atmosphere. k is wave number and equation (2) is later integrated with k by the method of Fourier Integral. As mentioned above we assume saturated atmosphere with neutral stability, so $\gamma^*-\gamma=0$, and

$$l^2(z)=-\frac{1}{U}\frac{d^2 U}{dz^2}. \tag{5}$$

Therefore l^2 (z) is determined only by undisturbed wind, $U(z)$. To solve the equation (3) we divide the atmosphere into several layers with constant l^2. Sarker considered the 3 layers in which values of l^2 are approximately constant respectively. But we considered the 2 layers have positive and exactly constant l^2. For that we represent $U(z)$ by the form of trigonometrical function that satisfies the equation (5).

3.1. Cases in which the vertical wind shears are different above the level of the maximum wind speed

As mentioned in 2.6 we are interested in the relation between the vertical distribution of upper wind and rainfall distribution in the atmosphere with neutral stability for the saturated air, that the smaller the rate of decreasing of wind speed above the low level strong wind across the mountains, the more heavy rainfall area shifts leeward and rainfall becomes heavy. In this section we intend to consider this relation theoretically:

We consider the case in which wind direction does not change upward, because the case in which it inverts is difficult to treat theoretically. We consider first two cases with the same maximum speed of 20 m/sec at the height of $2\pi/3$ km (about 2 km) and the same surface wind speed of 10 m/sec as shown left in Fig. 8. But in the case (a) the rate of decreasing of wind speed above 2 km height is smaller than in the case (b). Of course these values of wind are given, taking the above examples into consideration. At the heights of about 14 km in the case (a) and about 7 km in the case (b), the wind direction inverts. As it is known that such an upper boundary condition has not so great an influence on the stream line in the lower layer, the results obtained seem to be significant at least to the height of about 5 km even in the case (b). As indicated in Fig. 8 the forms of undisturbed wind $U(z)$ are different in the layers, upper and lower from the level of maximum wind, and the values of l are expressed by the coefficients of z in the trigonometrical function. Therefore in the case (a) values of l^2 are 1/4 km^{-2} and 1/64 km^{-2} for the lower and upper layers

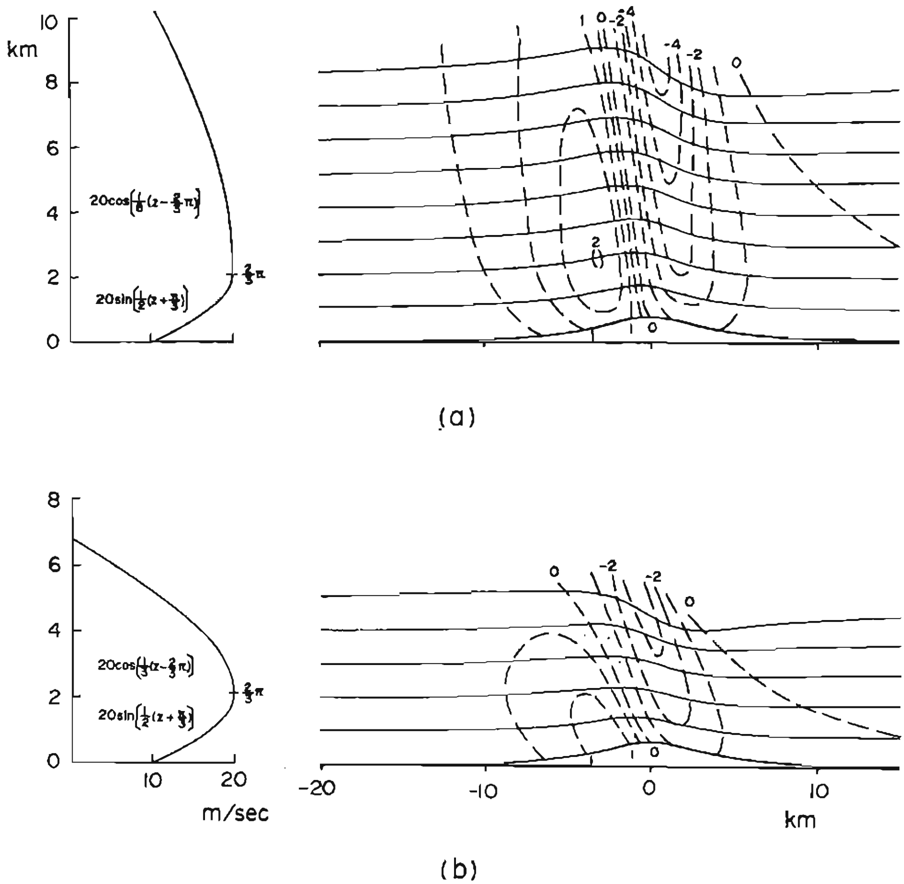


Fig. 8. Vertical profiles of wind (left), streamlines over the mountains and distributions of vertical velocity (in m/sec) (right).

respectively, and in the case (b) $1/4\ \mathrm{km}^{-2}$ as same as in the case (a) and $1/9\ \mathrm{km}^{-2}$ for the lower and upper layers respectively. The vertical distribution of temperature is assumed to be the moist adiabat of 297°K (24°C at 1000 mb), which is used to obtain the density and mixing ratio of the air and the condensation rate of the vapour.

The computed stream lines and distribution of vertical speed are shown in the right of Fig. 8, the upward speed is expressed for every 0.5 m/sec and the downward speed for every 1 m/sec. Later we try to compute the condensation rate of vapour for such vertical speed under the condition of neutral stability for moist air and rainfall rate under the assumption that the all condensed water reaches the ground surface. When we assume the atmosphere to be saturated with vapour everywhere, the condensed water must evaporate in the region of down-draught. This is inconsistent with the above assumption. Therefore the value of downward speed may be not so precise, but it is shown for the sake of comparison. Then, comparing the distribution of upward speed for the two cases, in the case (a) the maximum speed is about 2 m/sec and about two times the speed in the case (b), and the location of the maximum speed is higher in the case (a) than (b). The windward tilt of region of up-draught is greater in the case (b) than (a).

The distributions of condensation rate of vapour computed for each vertical speed are shown in Fig. 9. The condensation rate is indicated by the depth (in mm) of

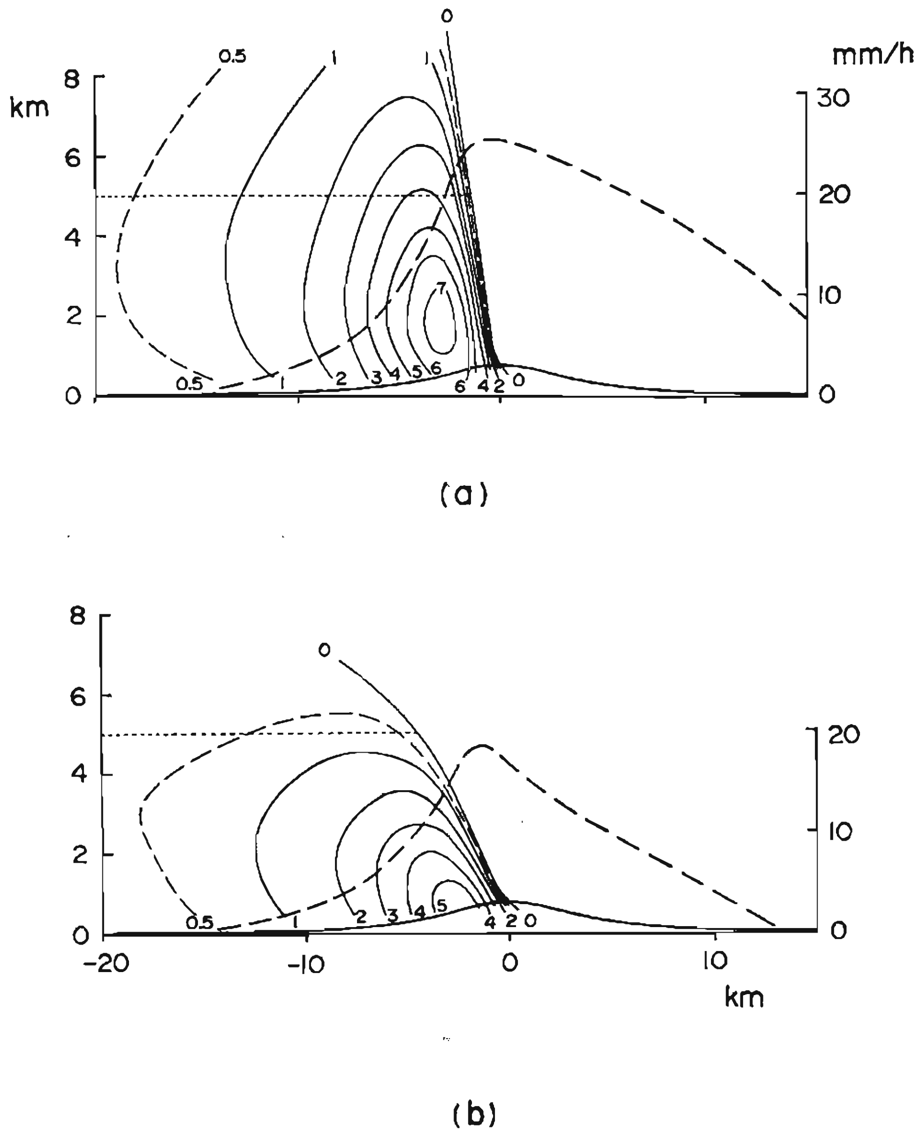


Fig. 9. Distributions of amount of condensed water (as depth in mm) for an hour in the layer whose thickness is 500m and hourly precipitation (thick dashed line) corrseponding to the case shown in Fig. 8.

liquid water condensed per an hour in every layer of thickness of 500 m and it is computed above the height of 500 m. As seen from Fig. 8 and 9, the distribution of condensation rate reflects the distribution of vertical speed and may be regarded as shape of cloud. The thick dashed line in Fig. 9 indicates the precipitation (in mm) for an hour, which is computed by the assumption that the all condensed water, in the layer from 500 m to 5 km (temperature of about 2°C) in height and with the condensation rate of above 0.5, composed of the raindrops with the fall speed of 4.5 m/sec, falls on the ground surface while drifting with the undisturbed flow. Sarker added snow, which was made in the layer above the height of 0°C and fell there at the speed of 1 m/sec, to the precipitation, but in our case it was negligible over the extent under consideration. Such a method of computation of rainfall rate seems to have rough propriety, though some problems still remain in the case because of the small scale of the Suzuka Mountains compared to the Western Ghats Mountains. In the case of the Western Ghats Mountains the maximum rainfall rate appears always over the windward slope of the mountains, but in our case the rainfall rate over the lee side slope of the mountains is the larger. It seems to be caused by the difference in the scales of the mountains. Then we compare (a) with (b) in Fig. 9. It is found that over the windward slope of the mountains the distributions of rainfall rate are nearly same on the whole, though the rainfall rate is a little larger in the case (a) than in the case (b), and that over the lee side slope the rainfall rate is much larger in the case (a) than in the case (b) and the heavy rainfall area in the case (a) shifts considerably lee ward. Such results of theoretical consideration agree qualitatively with the above mentioned results of observation. However, since the maximum rainfall rates are 20–30 mm per an hour in these results of theoretical consideration, these computed values may be too large when compared with the observed maximum 24-hour precipitation of 200–300 mm. Therefore it seems that it is necessary to take the evaporation of rain drops in the region of down-draught into account. On that account the computed rainfall rate becomes small, but it cannot yet be estimated.

3.2. Cases with stronger wind than in the cases of 3.1

Next we consider the cases in which the wind speeds are 1.5 times larger than in the cases of 3.1 though with the similar shapes of vertical distribution of wind as the cases of 3.1. The distributions of condensation rate and rainfall rate in these cases are shown in Fig. 10 corresponding to the shapes of vertical distribution of wind in (a) and (b) of Fig. 8. Comparing the distributions of rainfall rate in Fig. 10 with Fig. 9, it is found that those in both cases (a) and (b) over the windward side from the maximum rainfall rate differ little from those in the cases of 3.1 respectively, but that the both rainfall rates over the lee side from the maximum become large respectively the farther from the ridge.

3.3. Cases in which the levels of the maximum wind speed are different

It is expected that the upward speed of wind over the mountains becomes large according to the strength of low level wind. Here we consider the cases in which the heights of maximum wind speed are π km (about 3 km) and $\pi/3$ km (about 1 km) taking the case of Fig. 8 (a) into account. The former and the latter cases are shown in Fig. 11 (a) and (b) respectively. The undisturbed wind speed at $z=0$ and the maximum are 10 m/sec and 20 m/sec in both cases respectively, as in the cases of 3.1. The values of l^2 are $1/64\ \mathrm{km}^{-2}$ in both cases above the level of the maximum wind, but $1/9\ \mathrm{km}^{-2}$ in the case (a) and $1\ \mathrm{km}^{-2}$ in the case (b) under the level. As expected the upward speed is larger and the location of the maximum speed is lower in the

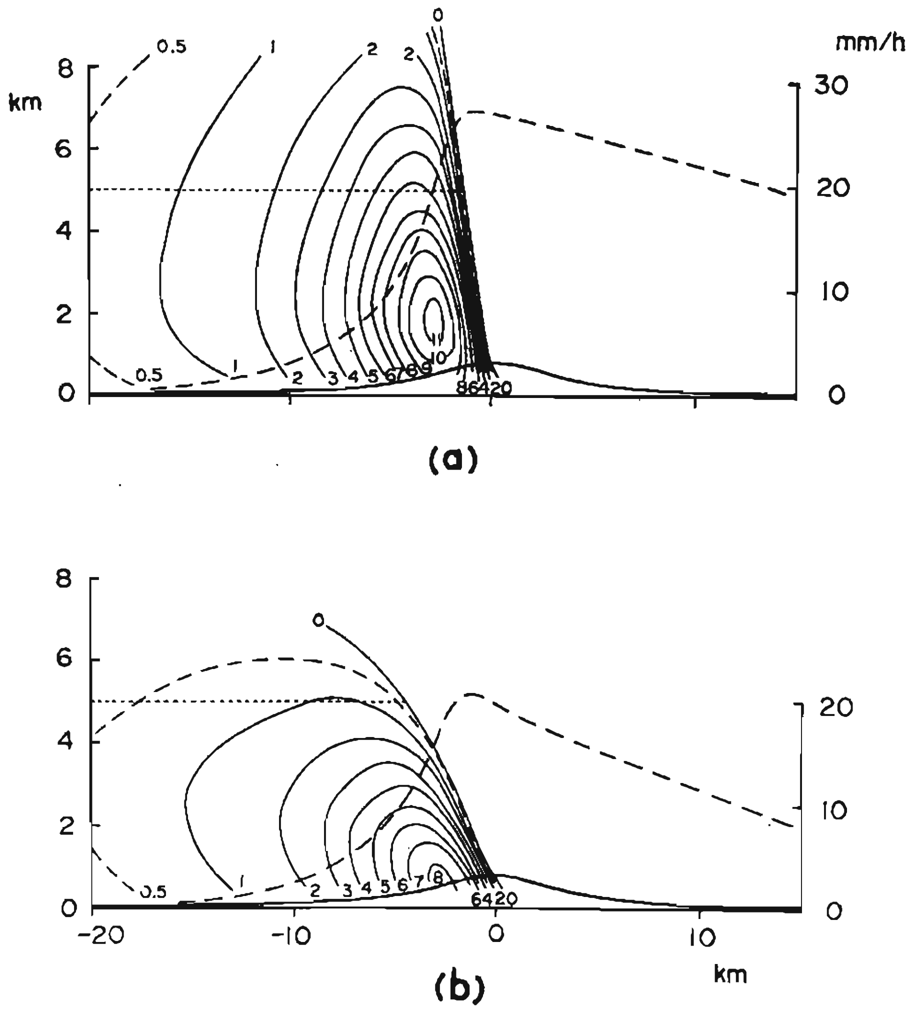


Fig. 10. Distributions of amount of condensed water (as depth in mm) for an hour in the layer whose thickness is 500m and hourly precipitation (thick dashed line) corresponding to the case in which wind speed is larger by 1.5 times than in the case shown in Fig. 8.

case (b) than the case (a). The distributions of condensation rate and rainfall rate corresponding to Fig. 11 are shown in Fig. 12. Both the condensation and rainfall rates are larger in the case of (b) than in the case (a), and the difference of rainfall rate is larger over the lee side than the windward side.

3.4. Discussions

It is found that the strong wind across mountains in the lower troposphere is certainly a favourable condition for orographic rainfall, and that the strong wind in the higher level than that of the maximum speed is also favourable especially for the rainfall over the lee side.

The problem on location of the maximum rainfall rate is very interesting, which is

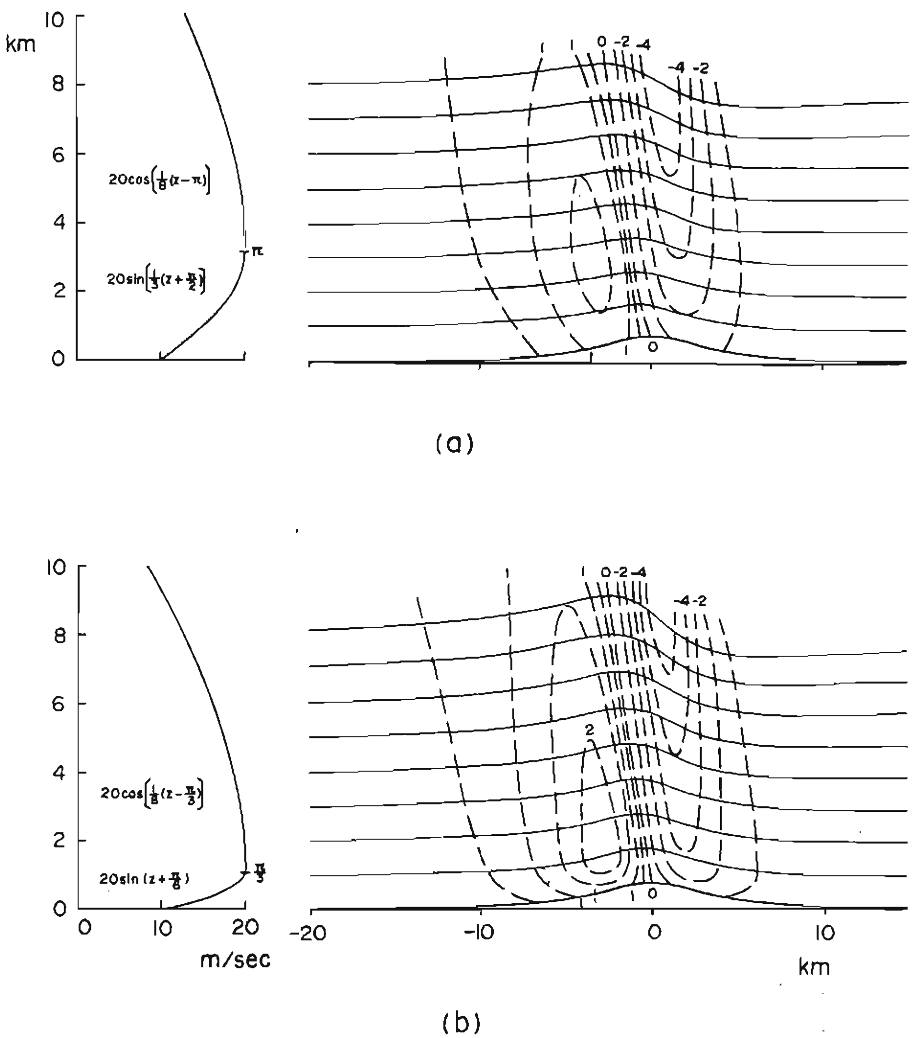
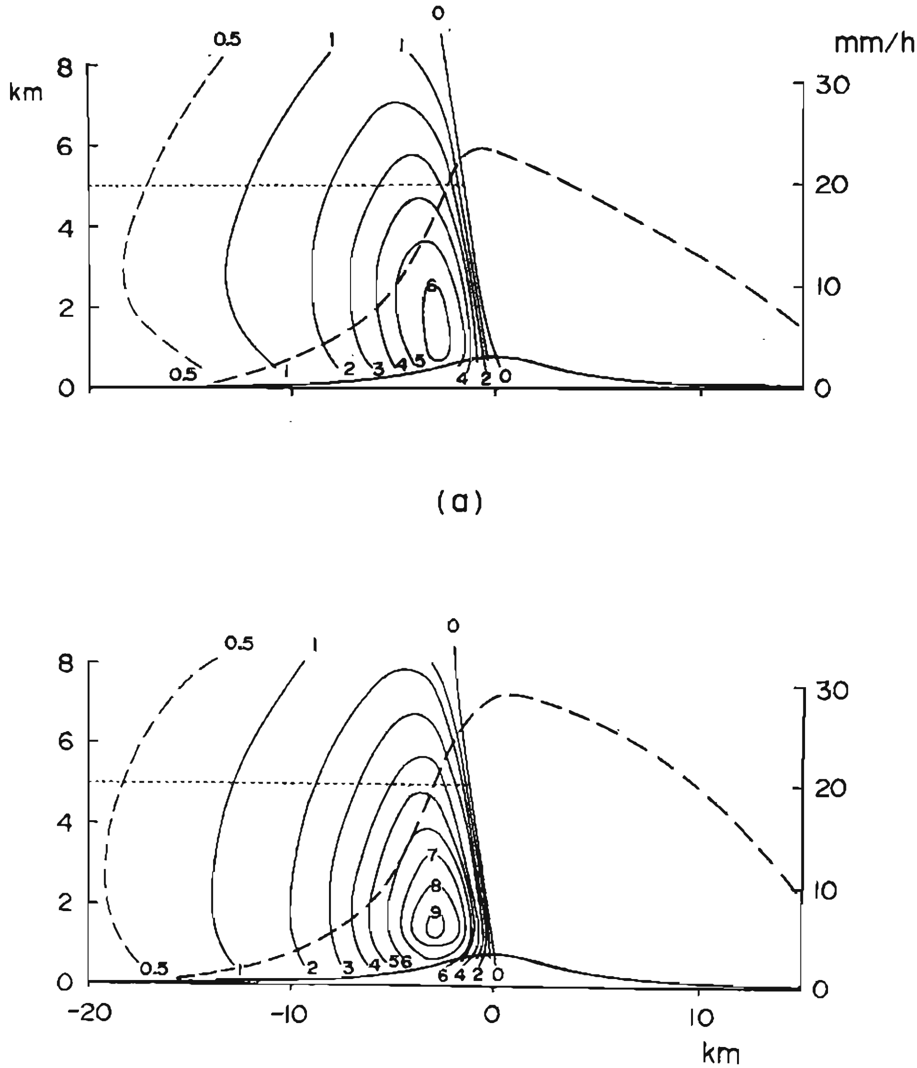


Fig. 11. Vertical profiles of wind (left), streamlines over the mountains and distributions of vertical velocity (in m/sec)(right).

in connection with the effect of spillover. However, assuming that in this model falling raindrops drift with the undisturbed flow, the effect is not taken into account. Because this approximate solution may have not enough accuracy especially in the neighbourhood $x=0$, as mentioned above, the stability in the region of down-draught may not be able to be regarded as neutral, and the trajectories of raindrops cannot be computed on account of insufficiency of the knowledge on the actual raindrop sizes.

As mentioned above over this region strong winds in the lower troposphere often appear, when a typhoon is south to west of this region. The above considerations can be applied to the case of low level jet-stream.

Takeda (1971)[4] investigated the role of the vertical wind shear in the formation



(b)

Fig. 12. Distributions of amount of condensed water (as depth in mm) for an hour in the layer whose thickness is 500 m and hourly precipitation (thick dashed line) corresponding to the case shown in Fig. 11.

of a cloud by numerical simulation of a precipitating convective cloud. Though in our case convection is not considered, it is necessary to take into account the more precise cloud physical processes. On our consideration the neutral stability is thought to be severe restriction. We intend to clarify such problems by further observation and analyses.

4. Concluding Remarks

Analyzing the detailed data on rainfall around the Suzuka Mountains with other meteorological elements during the warm season for three years, a few characteristic features of the heavy rainfall were clarified and several problems concerning them were revealed. In the case of heavy rainfall such as the maximum 24-hour precipitation of above 200 mm, strong easterly wind prevailed in the lower troposphere and the stability was nearly neutral for the saturated air. In such a case the heavy rainfall area lies just west of the ridge and along the mountains. The heavy rainfall area tends to shift more or less westward and rainfall tends to become heavy according to the easterly component of the wind in the upper troposphere.

Such contribution of the upper wind to the orographic rainfall was theoretically considered after the dynamic model presented by Sarker. With the above considerations the results of our observations were qualitatively explained. However the details concerning them will be clarified by further investigation.

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