A Numerical Study of the Local Downslope Wind “Yamaji-kaze” in Japan

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Abstract

In order to study the *Yamaji-kaze*—a typical downslope wind found in Japan, the two-dimensional flow over an asymmetric mountain is simulated by use of a non-hydrostatic model. The *Yamaji-kaze front* and the reversed wind behind the front—characteristic features of the *Yamaji-kaze*—are explained in terms of the internal hydraulic jump and its associated circulation.

Numerical experiments for a homogeneous atmosphere show that the behavior of the internal hydraulic jump is significantly affected by the inverse Froude number and the shape of the mountain. When the inverse Froude number is large, a quasi-steady state solution such that the hydraulic jump remains on the lee side of mountain is obtained, with the associated reversed flow being generated just behind the hydraulic jump. In the case of the *Yamaji-kaze*, the asymmetry of the Shikoku Mountains and the blocking effect of the Chugoku Mountains impede the propagation of the *Yamaji-kaze front* and allow the reversed wind to occur more readily.

For the case of the *Yamaji-kaze* observed on 21 April 1987, a notable inversion layer was found at a level near the mountaintop. It is confirmed, by numerical experiments of a heterogeneous atmosphere with the observed thermal stratification, that the surface wind strengthens in the presence of the inversion when compared to that without the inversion. Development and propagation of an internal hydraulic jump are qualitatively simulated under the observed thermal stratification and time-changing wind profile. On the basis of the experimental results a conceptual model of the *Yamaji-kaze* is proposed.

1. Introduction

The *Yamaji-kaze* is one of the most well-known local winds found in Japan. It is a strong downslope wind which occurs over the northern coastal plain of Shikoku Island when the lower-level synoptic wind is southerly. Observational investigations (for example, see Akiyama, 1956) have pointed out the following characteristic features of the *Yamaji-kaze*.

1) A northerly wind, opposite to the southerly synoptic wind is observed as one of the premonitory symptoms. 2) The surface wind direction exhibits a sudden variation along a line of discontinuity stretching from east to west. This line is called the *Yamaji-kaze front*, and a northerly wind opposite to the southerly downslope wind prevails on the north side of the front.

Although these phenomena are well known from previous observations, satisfactory explanations of their mechanisms have not been given. Arakawa and Kimura (1981) conducted a numerical experi-

ment of the *Yamaji-kaze* using a hydrostatic model with a horizontal grid distance of 10 km, but no opposite flow was simulated. As will be shown later, the *Yamaji-kaze* is a highly non-linear phenomenon having a hydraulic jump.

Similarities between the downslope winds in a stratified atmosphere and the flow in shallow water having a hydraulic jump have been known since old times. Long (1954) studied the behavior of the hydraulic jump in shallow water and the internal hydraulic jump on the interface of a two-fluid system, and discussed meteorological implications. Houghton and Kasahara (1968) theoretically explained the behavior of a non-linear shallow fluid flowing over a ridge. According to their hydraulic theory, the behavior of a shallow fluid can be classified by the velocity and depth of the fluid as well as by the mountain height into four regimes; a sub-critical flow, a quasi-steady flow with a stationary hydraulic jump, an unsteady flow with a propagating hydraulic jump, and a super-critical flow.

According to hydraulic theory, a downslope wind occurs when the fluid undergoes a transition from

an upstream sub-critical flow to downstream super-critical flow. The analogy between hydraulic fluids and the atmosphere is clearest in cases where there is a pre-existing layering of the environmental atmospheric stability. However, some questions have been raised concerning this simple analogy since downslope winds occur at times even when the atmosphere does not have such a pre-existing layered structure. Also, there is a difference between how the atmosphere and shallow water vertically transport the energy of internal gravity waves.

Another approach to study the mechanism of downslope winds is based on the linear theory of internal gravity waves in a continuously stratified infinite fluid. Klemp and Lilly (1975) calculated a linear analytic solution for a two-dimensional mountain flow in a stratified atmosphere. It was suggested that strong downslope winds occur when the atmosphere has a multilayer structure which allows an optimal superposition of the vertically propagating waves. However, strong downslope winds are usually highly non-linear phenomena, making it difficult to directly apply such a linear amplification mechanism to large-amplitude waves.

The importance of wave-breaking was discussed by Clark and Peltier (1977) (also see Clark and Peltier, 1984; Peltier and Clark, 1979), using numerical experiments. They suggested that when wave over-turning occurs in a highly non-linear mountain flow, a "wave-induced critical layer" (WICL) plays an important role in the occurrence of a strong downslope wind. They proposed that a resonant wave produced by reflection between the WICL and the ground surface creates a strong downslope wind.

Smith (1985) solved the Long's equation for a non-linear steady flow beneath a neutral, stagnant, well-mixed layer which was produced by wave breaking. The critical inverse Froude number was obtained where the mountain flow was transformed from sub-critical flow to super-critical flow. His theory suggested that a correspondence can be found between the flow in a stratified atmosphere and that in shallow water described by hydraulic theory.

Durran (1986) numerically examined the non-linear flow of a two-layer stratified fluid showing that there are similarities between the flow with strong downslope wind and the flow having a hydraulic jump in shallow water. Ikawa (1990a) also numerically examined the non-linear flow of a two-layer stratified fluid for the case of $l_2/l_1=0.4$ (l_1 , l_2 : the Scorer parameters of the lower and upper layers, respectively). It was shown that the response of the simulated flow to changes in the inverse Froude number and depth of the lower level does not significantly differ from Smith's theory, which is exactly valid for $l_2/l_1=0$. Durran and Klemp (1987) and Bacmeister and Pierrehumbert (1988) conducted numerical experiments with a pre-

existing critical level, showing that the character of high-drag states conforms well to the predictions of Smith's theory.

Although Smith's theory predicts the critical inverse Froude number for the transition to a high-drag state, at least in some cases, it does not yield any information concerning the transient features and the internal hydraulic jump. The internal hydraulic jump in a stratified atmosphere has not been examined sufficiently, either numerically or theoretically.

Recently, Ikawa and Nagasawa (1989) performed a numerical study of the dynamically induced foehn event observed in Hokkaido. It was reported that when the non-linearity of the mountain flow is strong and the internal hydraulic jump occurs, a reversed flow sometimes appears just behind the hydraulic jump. It appears that the hydraulic jump and the reversed flow have a number of similarities to the characteristic features of the Yamaji-kaze.

The main purposes of this study are 1) to examine the behavior of the internal hydraulic jump in a non-linear stratified atmosphere, and 2) to explain the *Yamaji-kaze front* and the northerly wind behind the front in terms of the internal hydraulic jump and its associated circulation. This will be accomplished by use of a non-hydrostatic model with a horizontal resolution of 2 km.

The organization of this paper is as follows. In Section 2, the geographical features of Shikoku Island, where the Yamaji-kaze occurs, are described and the climatological features of the Yamaji-kaze are examined. The conspicuous Yamaji-kaze, observed on 21 April 1987 is also discussed. In Section 3, a description of the numerical model and the design of the experiments are given. In Section 4, the effect of the shape of the terrain of Shikoku Island on the mountain flow is considered by use of an analytic solution, while the behavior of the internal hydraulic jump in a homogeneous atmosphere is examined by numerical experiments. In Section 5, numerical experiments with the observed thermal stratification are conducted to examine the influence of the inversion layer found at a level near the mountaintop on the Yamaji-kaze. The development and propagation of the internal hydraulic jump is simulated by incorporating the time-changing wind profile into the model. In Section 6, a conceptual model of the Yamaji-kaze is proposed, based on the experimental results. Summary and concluding remarks are given in Section 7.

2. The local downslope wind Yamaji-kaze

2.1 Geographical features of Shikoku island.

Figure 1 shows a geographical map of Shikoku Island, located in western Japan. The island is elongated in the east-to-west direction. In its northern central part, the Shikoku Mountains run from

Fig. 1. Geographical map of Shikoku Island, western Japan. The contour interval is 400 m, with the area above 800 m height shaded. Open circles denote the cities (Niihama and Mishima) where automated meteorological stations have been installed. Open squares denote the cities (Fukuyama and Kochi) where local meteorological observatories are located. The star symbol (☆) denotes the location of the town of Doi where the severest Yamaji-kaze occurs. The solid lines show the area over which the orography is averaged. The small rectangle indicated by broken line represents the domain shown in Fig. 4.

east to west. Mt. Ishizuchi (1981 m) and Mt. Tsurugi (1955 m) are located in the western and eastern parts of the Shikoku Mountains, respectively. North of Shikoku Island, the relatively flat Chugoku Mountains run from east to west over the central part of the Chugoku Peninsula. The two-dimensionality of the orography, framed by solid lines in Fig. 1, is generally good. The Yamaji-kaze occurs along the narrow northern coastal plain of Shikoku Island, from Mishima to Niihama, facing Hiuchi-nada (the south-

ern part of the Seto-uchi Sea). The most severe Yamaji-kaze occurs around Doi shown in Fig. 1 by the star symbol (☆).

Figure 2 shows an orographical cross-section obtained by averaging the orography framed by the solid $x-y$ lines in Fig. 1 along the y -direction. As shown in this figure, the northern slope of the Shikoku Mountains facing Hiuchi-nada is steep. On the other hand, the southern slope of the Shikoku Mountains facing the Pacific Ocean is relatively gentle. In this study, this orography is regarded as the typical orography of the region where the Yamaji-kaze occurs. Since the most severe Yamaji-kaze occurs around Doi town, the observed Yamaji-kaze is likely influenced by the complicated three-dimensional effect of the actual orography.

2.2 Climatological features of the Yamaji-kaze.

The Yamaji-kaze develops when the lower-level synoptic wind is southerly, the result of a developed cyclone or typhoon being located in the Sea of Japan. Usually, it is accompanied by an increase of temperature, a drop of surface pressure, and no precipitation. In the past, many investigations have been conducted due to the severe damage that occurs. The Yamaji-kaze Countermeasure Headquarters was established by local municipalities and agricultural cooperative societies in 1951, and special observational experiments were conducted by local meteorological observatories. The results were summarized in an investigative report by the Osaka District Meteorological Observatory in 1958 (hereafter referred to as ODMO, 1958). The following characteristic premonitory symptoms and accompanying phenomena of the Yamaji-kaze were listed by ODMO (1958).

- 1) *Keta-kumo* (girder cloud); a stationary cloud appears on the mountain shoulder before the occurrence of the Yamaji-kaze.
- 2) *Sasoi-kaze* (invitation wind); a northerly wind opposite to the southerly synoptic wind is observed prior to the occurrence.
- 3) *Yama-nari* (mountain rumbling); a rumbling of the mountain occurs simultaneously with 1) and 2).
- 4) *Mayoi-kaze* (puzzled wind); the wind direction and wind speed widely vary during the early stage.
- 5) *Domai*; a northerly wind opposite to the southerly downslope wind prevails at Hiuchi-nada.
- 6) *Kawashi-kaze* (return wind); the wind direction

Fig. 2. Cross section obtained by averaging the orography framed by the solid lines in Fig. 1 along the y -direction (indicated by a vector).

Fig. 3. The analysis of surface winds of the Yamaji-kaze observed on 27 February 1954, adopted from Akiyama (1956). The number at the lower right-hand corner of each section shows the time in hours (Japan Standard Time, JST). The heavy solid lines show the streamlines on the surface and heavy broken lines show the streamlines separated from the ground. The broken line of each section shows the line where the streamlines detach from the ground surface.

changes from southerly to north-westerly during the decay stage.

Concerning the *Keta-kumo*, ODMO (1958) reported that the mountaintop could be seen *above* the *Keta-kumo*. It is conjectured that the *Keta-kumo* differs from the helm clouds often seen *over* the mountaintop.

The *Mayoi-kaze* denotes that both wind direction and speed vary continuously during the early stage of the Yamaji-kaze, while according to observational studies, the surface wind varies suddenly along a line

of discontinuity, stretching from east to west. Figure 3 shows a surface wind analysis of the 27 February 1954 Yamaji-kaze taken from Akiyama (1956). In this figure, the broken line of each section shows the region where the wind direction and speed suddenly change. Akiyama conjectured that the streamlines detach from the ground surface along this line. Such a line of discontinuity where the wind varies suddenly, is often observed during the early stage of the Yamaji-kaze, and is called the *Yamaji-kaze front*. Similar lines are sometimes observed in the downs-

lope winds of the Rocky Mountains, and are called the *Chinook front*. In Fig. 3, northerly winds corresponding to the *Sasoi-kaze* are seen from 09 JST to 13 JST.

The *Domai* is considered to be a *Sasoi-kaze* when *Yamaji-kaze front* has moved to Hiuchi-nada. Kobayashi (1954) concluded that *Domai* is not observed when the southerly synoptic wind is very strong, which often happens when a strong typhoon is located in the Sea of Japan.

The *Kawashi-kaze* is thought to be produced by the passage of a cold front, resulting from the migration of the cyclone in the Sea of Japan.

In addition to 1)–6), the following climatological features of the Yamaji-kaze have been pointed out.

7) A northerly wind is often observed along the southern coastal plain of the Chugoku Peninsula facing the Seto-uchi Sea (Furukawa, 1966).

8) The wind system conflicts with the pattern of the local surface pressure (Furukawa, 1966).

9) Strong winds are seldom observed on the windward side of Shikoku Island (Takami and Ohnishi, 1988).

10) Low surface pressures are observed from the northern coastal plain of Shikoku Island to Hiuchinada (Kobayashi, 1954). This area of low pressure is called the *Hiuchi-nada depression* by local meteorologists.

Although these characteristic features of the Yamaji-kaze are well-known from previous studies, satisfactory explanations for their mechanisms have not been found in observational investigations and numerical studies (Arakawa and Kimura, 1981) that have been conducted in the past.

2.3 The Yamaji-kaze of 21 April 1987

A strong Yamaji-kaze occurred on 21 April 1987, resulting in severe damage to houses and agricultural plants along the northern coastal plain of Shikoku Island.

Figure 4 shows the surface synoptic weather chart for 09 JST, 21 April 1987. A developing cyclone with a central pressure of 984 hPa was located west of the Korean Peninsula, creating a southerly synoptic wind in western Japan. This cyclone moved north-eastward along the track shown by the thick line.

Figure 5 displays the anemometer trace taken by the Doi Fire Department. The surface wind at Doi was weak and southerly as the synoptic wind of western Japan until 04 JST. However, the wind direction became variable after 04 JST, and changed to north-easterly at 05 JST. This northerly wind, which corresponds to the *Sasoi-kaze*, was observed until 09 JST. The wind direction then suddenly changed to southerly about 09 JST and a strong downslope wind began to blow. This sudden change of the surface wind corresponds to the onset of the Yamaji-kaze. The period of strongest winds was

Fig. 5. Anemometer trace recorded by the Doi Fire Department for the period from 21 JST 20 April to 03 JST 22 April 1987. Time runs from right to left. Top: wind speed (m/s). Bottom: wind direction.

Fig. 6. Trace of the surface temperature recorded by the Mishima Fire Department.

Fig. 7. Trace of the surface pressure recorded by the Mishima Fire Department.

spheric variables on 21 April 09 JST observed at Fukuoka, Yonago, Kagoshima, and Shionomisaki (for their location, see Fig. 4). From these figures, it can be seen that a notable temperature inversion existed in the lower levels over western Japan. The height of the inversion layer was relatively high in the windward region (1500–1700 m at Kagoshima and 1700–2200 m at Shionomisaki), while relatively low in the leeward area (600–900 m at Fukuoka and 600–1200 m at Yonago). Such an inversion layer is often observed during a strong Yamaji-kaze event. At all observation stations the synoptic wind was southerly near the surface, while south-westerly near the 850 hPa level. Above the 850 hPa level, westerly winds increased up to the tropopause. The lower southerly winds were strong at Fukuoka and Yonago, close to the cyclone, while they were relatively weak at Kagoshima and Shionomisaki farther from the cyclone.

3. Numerical model and design of the experiments

3.1 Numerical model

The numerical model is a two-dimensional non-hydrostatic model using anelastic equations. The model is similar to that described by Ikawa and Nagasawa (1989) except for the turbulent closure model. The terrain-following coordinate, which is mainly based on Clark (1977), is given as

$$z^* = \frac{H(z - z_s)}{H - z_s}, \quad (1)$$

where z_s is the surface height, and H the height of the domain. The governing momentum and the thermodynamic equations are as follows;

$$\begin{aligned} \frac{\partial \bar{\rho} G^{\frac{1}{2}} u}{\partial t} + \frac{\partial G^{\frac{1}{2}} p'}{\partial x} + \frac{\partial G^{\frac{1}{2}} G^{13} p'}{\partial z^*} \\ = - \left(\frac{\partial \bar{\rho} G^{\frac{1}{2}} u u}{\partial x} + \frac{\partial \bar{\rho} G^{\frac{1}{2}} \omega u}{\partial z^*} \right) - DIF. u, \quad (2) \end{aligned}$$

Fig. 8. Vertical profiles of temperature (T), dew point temperature (T_d), potential temperature (θ), equivalent potential temperature (θ_e), and wind on 09 JST 21 April. The short barbs on an arrow indicate 5 m/s, long barbs indicate 10 m/s, while a pennant indicates 50 m/s. **a)** at Fukuoka, **b)** at Yonago, **c)** at Kagoshima, **d)** at Shionomisaki.

$$\begin{aligned} \frac{\partial \bar{\rho} G^{\frac{1}{2}} w}{\partial t} + \frac{\partial G^{\frac{1}{2}} p'}{G^{\frac{1}{2}} \partial z^*} + g \frac{G^{\frac{1}{2}} p'}{Cs^2} \\ = - \left(\frac{\partial \bar{\rho} G^{\frac{1}{2}} u w}{\partial x} + \frac{\partial \bar{\rho} G^{\frac{1}{2}} \omega w}{\partial z^*} \right) \\ + \bar{\rho} G^{\frac{1}{2}} g \frac{\theta'}{\bar{\theta}} - DIF. w, \end{aligned} \quad (3)$$

and,

$$\frac{\partial \bar{\rho} G^{\frac{1}{2}} \theta}{\partial t} = - \left(\frac{\partial \bar{\rho} G^{\frac{1}{2}} u \theta}{\partial x} + \frac{\partial \bar{\rho} G^{\frac{1}{2}} \omega \theta}{\partial z^*} \right) - DIF. \theta, \quad (4)$$

where Cs is the speed of the sound wave, $\bar{\rho}$ and $\bar{\theta}$ the density and potential temperature of the reference basic state, dependent only on z . The perturbation potential temperature and pressure for the basic state are given by θ' , p' . $G^{\frac{1}{2}}$, $G^{\frac{1}{2}} G^{13}$ are the tensors for the coordinate transformations and ω the vertical velocity in the z^* coordinate defined as

$$G^{\frac{1}{2}} = 1 - \frac{z_s}{H}, \quad G^{\frac{1}{2}} G^{13} = \left(\frac{z^*}{H} - 1 \right) \frac{\partial z_s}{\partial x},$$

and

$$\omega = \frac{dz^*}{dt} = \frac{1}{G^{\frac{1}{2}}} \left(w + G^{\frac{1}{2}} G^{13} u \right).$$

The continuity equation in an anelastic system is given as follows:

$$\frac{\partial \bar{\rho} G^{\frac{1}{2}} u}{\partial x} + \frac{\partial \bar{\rho} G^{\frac{1}{2}} \omega}{\partial z^*} = 0. \quad (5)$$

The pressure is calculated diagnostically from Eqs. (2), (3), and (5).

To determine the diffusion coefficients, the turbulent closure model is employed. The prognostic equation for sub-grid scale turbulent kinetic energy is given as follows:

$$\begin{aligned} \frac{\partial \bar{\rho} G^{\frac{1}{2}} E}{\partial t} = - \left(\frac{\partial \bar{\rho} G^{\frac{1}{2}} u E}{\partial x} + \frac{\partial \bar{\rho} G^{\frac{1}{2}} \omega E}{\partial z^*} \right) \\ - \frac{g K_h}{G^{\frac{1}{2}}} \frac{\partial \theta}{\partial z^*} - \overline{u'_i u'_j} \frac{\partial u_i}{\partial x_j} - \frac{C_e E^{\frac{3}{2}}}{l} - DIF. E, \end{aligned} \quad (6)$$

where

$$\overline{u'_i u'_j} = -K_m \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) + \frac{2}{3} \delta_{ij} E. \quad (7)$$

The diffusion terms in Eqs. (2)–(4) and (6) are given for u and w as

$$DIF. u_i = -\frac{\partial}{\partial x_j} \left(\overline{u'_i u'_j} \right), \quad (8)$$

and for θ and E as

$$DIF. \phi = -\frac{\partial}{\partial x_j} \left(K_\phi \frac{\partial \phi}{\partial x_j} \right). \quad (9)$$

The diffusion coefficients for momentum, heat, and turbulent energy are given respectively as follows:

$$K_m = 0.2lE^{\frac{1}{2}}, \quad (10)$$

$$K_h = P_r^{-1} K_m, \quad (11)$$

$$K_e = 2K_m. \quad (12)$$

The mixing length l is calculated according to the local stability N as follows:

$$l = \begin{cases} \Delta s & ; \text{for unstable case } (N \leq 0) \\ \min(\Delta s, 0.76E^{\frac{1}{2}}N^{-1}) & \\ & ; \text{for stable case } (N > 0) \end{cases} \quad (13)$$

where

$$N = \frac{g}{\theta} \sqrt{\frac{\partial \theta}{\partial z}}, \quad \Delta s = \sqrt{\Delta x \Delta z}.$$

Here, Δx and Δz are the horizontal and vertical grid distances. The coefficient for the viscosity dispersion term in Eq. (6) and the inverse Prandtl number in Eq. (11) are given respectively as follows:

$$C_e = 0.19 + 0.51 \frac{l}{\Delta s}, \quad (14)$$

$$P_r^{-1} = \left(1 + 2 \frac{l}{\Delta s} \right). \quad (15)$$

The above formulation is based on Klemp and Wilhelmson (1978), and Deardorff (1980).

The grid distances in the horizontal and vertical directions employed in this study are $(\Delta x, \Delta z) = (2000 \text{ m}, 200 \text{ m})$, and the time interval is $\Delta t = 8 \text{ s}$.

The lateral boundaries are open, where the radiation condition developed by Orlanski (1976) is employed. The upper and lower boundaries are free-slip walls, which are thermally insulated.

In order to prevent a false reflection from the upper boundary, an absorbing layer with Rayleigh friction is imposed at the upper layer of the model domain. In the upper layer ($z > z_d$), the following damping term is added to the time tendency of ϕ ;

$$Dub(\phi) = \left\{ 1 + \cos \left(\pi \frac{H - z}{H - z_d} \right) \right\} \frac{\phi_e - \phi}{150\Delta t}, \quad (16)$$

where z_d is the height above which the Rayleigh damping is imposed, ϕ denotes an arbitrary prognostic variable (u , w , and θ), and ϕ_e the environmental value of ϕ .

Near the lateral boundaries, weak Rayleigh friction is imposed to enforce the environmental condition on all prognostic variables except for the turbulent kinetic energy. Near the lateral boundaries ($|x_b - x| < 10\Delta x$), the following damping term is added to the time tendency term;

$$Dlb(\phi) = \left\{ 1 - \left| \frac{x_b - x}{10\Delta x} \right| \right\} \frac{\phi_e - \phi}{500\Delta t}, \quad (17)$$

where x_b is the location of the lateral boundary.

3.2 Design of the experiments

The numerical experiments are roughly classified into the three types listed in Table 1. In this table, a measure of the non-linearity of the mountain flow, the inverse Froude number F_r^{-1} is defined as

$$F_r^{-1} = lh_m = N \frac{h_m}{U}, \quad (18)$$

where N is the Brunt-Väisälä frequency, h_m the height of mountaintop, and U the environmental wind speed. For Ex-2 and Ex-3, F_r^{-1} is roughly estimated by using the averaged tropospheric variables of N and U .

In Ex-1, the behavior of the internal hydraulic jump in a homogeneous atmosphere with constant N and constant U is examined for the simplest case. A fixed potential temperature lapse rate of 3K/km is used for the stability of the environmental atmosphere.

Use is made of three different mountain shapes, i.e., a symmetric bell-shaped mountain given as

$$h(x) = \frac{h_m}{1 + \left(\frac{x}{a} \right)^2}, \quad (19)$$

($h_m = 1050 \text{ m}$; the height of the mountaintop,
 $a = 12 \text{ km}$; the half width),

the averaged orography of the Shikoku and Chugoku Mountains shown in Fig. 2, and the averaged orography of the Shikoku Mountains alone. In order to simplify the interpretation of the experimental results, the Boussinesq approximation is employed, and the density of the basic reference state $\bar{\rho}$ is set to 1 kg/m^3 , independent of height. The height z_d in Eq. (16) is set to 8.5 km . The mountain height is initially set to zero for a smooth spin-up, and is raised linearly to the height of the mountaintop, h_m , during the first 600 time steps, equivalent to a model time of $t = 80 \text{ min}$. Each simulation is continued until $t = 8 \text{ hrs}$.

In Ex-2, the role of the inversion layer at the level near the mountaintop in the Yamaji-kaze is examined. The environmental thermal stratification is

Table 1. Classification of the numerical experiments.

Experiment	Thermal stratification	Environmental wind		inverse Froude number	Orography	Boussinesq approximation	Rayleigh friction	Integration time
		shear	magnitude (m/s)					
Ex-1	homogeneous ($N=0.01/s$)	no	2, 3, 3.5, 4, 5	5.2~2.1	Bell	employed	$z_d=8.5$ km	8 hours
			2, 3, 4	5.2~2.6	Shikoku			
			2, 3, 4, 6	5.2~1.7	Shikoku-Chugoku			
Ex-2	observation	no	10~18	1.1~0.6	Shikoku-Chugoku	employed	$z_d=13.5$ km	4 hours
	smoothed							
Ex-3	observation	observation	dependent of height and time	$\infty \sim 0.6$	Shikoku-Chugoku	not employed	$z_d=13.5$ km	26 hours

obtained from the composited observations of 21 April 1987 (Fig. 16a). The environmental wind U is constant, independent of height. The height z_d is set to 13.5 km, which corresponds to the height of the tropopause. The application of the Boussinesq approximation and the spin-up procedure are the same as in Ex-1. Each simulation is continued until $t=4$ hrs.

In Ex-3, the development and propagation of the internal hydraulic jump is simulated on the basis of the observed data of the Yamaji-kaze on 21 April 1987. In this experiment, the vertical profiles based on the observations are used not only for the thermal stratification but also for the environmental wind (Fig. 16a, b). Unlike Ex-1 and Ex-2, the Boussinesq approximation is not employed to express the amplification of the mountain wave due to the decrease of the mean density. For the initial spin-up procedure, the environmental wind is gradually increased from zero to the observed vertical profile with a fixed mountain height during the first 24 hours of model time.

In these experiments, the use of a non-hydrostatic model is more desirable than the use of a hydrostatic model, since wave breaking is expected to occur. Although the effect of the Coriolis force is not necessarily negligible, all experiments are conducted without the Coriolis force since its inclusion results in no large difference in the features of the hydraulic jump (Houghton, 1969; Ikawa and Nagasawa, 1989) and its use would only complicate the results of experiments.

In the experiments, the size of the model domain is set to $(X_L, H)=(440 \text{ km}, 16 \text{ km})$. For the case of a large inverse Froude number, the decelerated layer on the windward side of the mountain propagates far upstream with time (Pierrehumbert and Wyman, 1985). To check if the horizontal domain is sufficient, several additional experiments—in which horizontal domain of 440 km is enlarged on the inflow side to 640 km—are performed for Ex-1 and Ex-2. As will be shown later, the differences in the experimental results due to the size of the horizontal domain are small during the simulation. For Ex-3, although the influence of the lateral bound-

ary is inevitable due to its long term simulation, the Rayleigh damping imposed near the lateral boundary acts as an external forcing and represents the time change of the environmental state.

4. Flow in a homogeneous atmosphere

4.1 The effect of the asymmetric orography of Shikoku Island on the mountain flow: theoretical consideration

In this sub-section, the effect of the shape of the terrain of Shikoku Island on the mountain flow is studied, using an analytic solution in advance of the numerical experiments. Concerning this problem, Lilly and Klemp (1979) obtained an analytic solution for Long (1953)'s equation taking into account the non-linear lower boundary condition (see Eq. (8) of Lilly and Klemp, 1979, hereafter called the 'non-linear solution'). According to Lilly and Klemp, strong downslope winds tend to occur on the steep slope of an asymmetric mountain. As shown in Fig. 2, the orography of Shikoku Island has a gentle slope on the southern side and a steep slope on the northern side, and appears to be suitable for the occurrence of a downslope wind when a southerly synoptic wind is present.

Figures 9a, 9b show the vertical cross-sections of the horizontal wind obtained by a linear analytic solution of Long's equation. In these figures, the analytic solutions are computed by use of a Hilbert transform. $N=0.0097/s$, $U=12 \text{ m/s}$ is used for the typical stability and environmental wind. In this case, the inverse Froude number is 0.85. As shown in these figures, the surface wind over the Shikoku Mountains, due to the asymmetry, is stronger than that of a bell-shaped mountain. No wave overturning is seen in these figures.

Figures 9c, 9d show the vertical cross-sections of the horizontal wind obtained by a non-linear analytic solution. The surface wind on lee side of the Shikoku Mountains is very large, and wave overturning is seen at heights of 5 km and 13 km above the Shikoku Mountains.

Figure 10 shows the magnitude of the maximum surface winds determined by analytic solutions as functions of U . In this figure, values of U less

Fig. 9. Vertical cross-sections of the horizontal wind for $U=12$ m/s and $N=0.0097$ /s.

a) Linear analytic solution for a bell-shaped mountain ($h_m=1050$ m, $a=12$ km).

b) As in a) except for the Shikoku Mountains.

c) Non-linear analytic solution for a bell-shaped mountain.

d) As in c) except for the Shikoku Mountains.

e) Numerical solution for a bell-shaped mountain at $t=240$ min.

f) As in e) except for the Shikoku Mountains.

Fig. 10. The magnitude of the maximum surface winds obtained from analytic solutions as functions of U for the case of $N=0.0097/s$ and $h_m=1050$ m.

than zero denote northerly environmental winds. As shown in this figure, the surface wind over the northern slope of the Shikoku Mountains with a southerly environmental wind tends to become larger than that on the southern slope with a northerly environmental wind. Actually, north-westerly monsoon winds often prevail over western Japan during winter, but strong downslope winds are seldom observed over the southern coastal plain of Shikoku Island.

In practice, the analytic solution becomes meaningless when wave over-turning occurs. According to the non-linear analytic solution, the critical value of the inverse Froude number above which wave over-turning occurs is 0.85 for a symmetric bell-shaped mountain, while it is 0.76 for the orography of Shikoku Island. Therefore, the mountain wave over the Shikoku Mountains tends to be enhanced when the environmental winds are southerly, and wave breaking easily occurs for small inverse Froude numbers. In Fig. 10, the values of the maximum

surface wind are shown by broken lines in the region where wave over-turning occurs. Generally, once wave breaking occurs, the pattern of mountain flow becomes quite different from that resulting from the analytic solution, and a hydraulic jump is seen, which will be described in the next sub-section.

The vertical cross-sections of the horizontal wind obtained by the numerical model (at $t=240$ min) are shown in Figs. 9e, 9f for reference. The amplitude of the mountain wave decreases as height increases, owing to the absorption by Rayleigh damping imposed in the upper layer of the model domain. For the orography of Shikoku Island, wave over-turning is seen at the height of 4 km above the northern slope of the Shikoku Mountains.

4.2 The internal hydraulic jump in a homogeneous atmosphere

In this sub-section, the flow with a hydraulic jump after wave breaking occurs in a homogeneous atmosphere is numerically examined, making use of various values of U and a constant $N=0.01/s$. The results of the experiments are summarized in Table 2. In this table, C denotes the propagation speed of the hydraulic jump and h the height of the mountain slope where the jump remains stationary. In the column headed hydraulic jump, open circles denote the cases in which a hydraulic jump appears, and double circles signify that the jump remains stationary. The triangle means that the appearance of the hydraulic jump is somewhat unclear. In the column headed *rev.W*, open circles denote the cases in which the reversed flow behind the hydraulic jump is found. The triangle denotes the case in which the appearance of reversed wind is intermittent, and crosses signify that the reversed flow does not appear.

a) for a bell-shaped mountain

Figures 11a–11c show the vertical cross-sections of the horizontal wind u at various model times for the case of a bell-shaped mountain and $U=4$ m/s ($F_r^{-1}=2.6$). At $t=80$ min, at the time when the

Table 2. Results of experiments for a homogeneous atmosphere (Ex-1).

Orography	U (m/s)	F_r^{-1}	hy.jump	C (m/s)	C^* ($=C/U$)	h (Km)	rev.W
Bell	2	5.2	⊙	—	—	0.23	○
	3	3.5	○	0.21	0.07	—	○
	3.5	3.0	○	0.67	0.19	—	○
	4	2.6	○	1.47	0.37	—	×
	5	2.1	△	2.80	0.57	—	×
Shikoku	2	5.2	⊙	—	—	0.24	○
	3	3.5	○	0.07	0.02	—	○
	4	2.6	○	0.97	0.24	—	○
Shikoku-Chugoku	2	5.2	⊙	—	—	0.38	○
	3	3.5	⊙	—	—	0.06	○
	4	2.6	○	0.54	0.13	—	○
	6	1.8	○	1.43	0.23	—	△

mountain has just been raised to its ordinary height, the pattern of u is still similar to that given by the linear analytic solution. After the occurrence of wave breaking, the surface wind on the lee side of the mountain increases. At $t=240$ min, wave over-turning occurs 2 km above the mountain, and a wave-induced stagnant layer spreads leeward as shown in Fig. 11b. Below the stagnant layer, an area of strong surface wind also spreads leeward. The surface wind abruptly changes near the forefront of the area of strong surface wind at $x=235$ km. This sudden change in the surface wind corresponds to the location of the internal hydraulic jump. As time elapses, the hydraulic jump propagates leeward with the area of spreading strong surface winds. As shown in Fig. 11c ($t=400$ min), the hydraulic jump is located at $x=250$ km, with weak winds less than 2 m/s seen at the surface from $x=262$ km to $x=272$ km. The time evolution of the surface wind at an interval of 80 min is shown in Fig. 12a. In this case, the mean propagating speed of the hydraulic jump during $t=240$ min to $t=480$ min is 1.46 m/s, and the ratio to U is 0.37.

For the case of $U=3$ m/s ($F_r^{-1}=3.5$), the hydraulic jump develops more in the windward direction compared with the case of $U=4$ m/s, and its propagation speed decreases to 0.21 m/s. The deceleration of the surface wind behind the jump becomes clear and a reversed flow opposite to the environmental wind appears. For the case of yet a smaller environmental wind as $U=2$ m/s ($F_r^{-1}=5.2$), the area of strong surface wind no longer spreads leeward, and the jump remains on the lee slope of the mountain. By contrast, for the case of a stronger environmental wind of $U=5$ m/s ($F_r^{-1}=1.8$), the propagation speed of hydraulic jump increases, while the jump itself becomes somewhat unclear. These experimental results are summarized in Table 2 and Fig. 15.

To examine the differences in the experimental results due to the size of horizontal domain, additional experiments—in which horizontal domain is set to 640 km by the addition of 200 km on the inflow side—are also performed for $U=2$ m/s and $U=4$ m/s. For the case of $U=4$ m/s, the difference in the location of the hydraulic jump between the different domains is less than 2 km (one grid distance), and there is no difference in the propagation speeds of the hydraulic jump. When $U=2$ m/s, the jump tends to develop slightly to the windward side in the experiment of the enlarged domain, while the difference in the location of hydraulic jumps at $t=480$ min is only one grid distance.

The reversed flow behind the hydraulic jump develops when $U \leq 3.5$ m/s ($F_r^{-1} \geq 3$). Ikawa and Nagasawa (1989) reported the occurrence of a similar reversed flow in experiments of $F_r^{-1}=1.875$ ($N=0.01$ /s, $U=4$ m/s, a bell-shaped mountain with $a=6$ km, $h_m=750$ m). One of the causes for this dif-

ference can be explained by the deceleration of the environmental wind due to the cyclic boundary condition employed by Ikawa and Nagasawa.

b) for the orography of Shikoku Island

Figure 12b shows the time evolution of the surface wind when the orography of the Shikoku Mountains but not with Chugoku Mountains is used ($U=4$ m/s). The change in the surface wind associated with the hydraulic jump becomes more distinct while its propagation speed is smaller by 0.97 m/s, than that for the bell-shaped mountain. The reversed flow behind the hydraulic jump appears at $t=240$ min and spreads leeward as time elapses. These differences are caused by the asymmetry of the Shikoku Mountains.

As given in Table 2, the propagation speeds of the hydraulic jump are smaller than those for the bell-shaped mountain, and the jump remains on the lee side of the Shikoku Mountains for the case of $U=2$ m/s. The reversed flow behind the hydraulic jump develops for all cases of $U \leq 4$ m/s ($F_r^{-1} \geq 2.6$).

c) for the orography of Shikoku Island and the Chugoku Mountains

Figure 13a shows the time evolution of the surface wind when the orography of the Shikoku Mountains and the Chugoku Mountains are included ($U=4$ m/s). When compared with the case of Shikoku Island only, the maximum surface wind slightly decreases, the hydraulic jump almost remains on the lee side of the Shikoku Mountains, and the reversed flow behind the jump is more distinct. These differences are ascribed to the deceleration of the environmental wind due to the blocking effect of the Chugoku Mountains. Figures 13b, 13c show the vertical cross-sections of the horizontal wind (u) and wind vectors at $t=240$ min. On the northern slope of Shikoku Island, a downslope wind of greater than 10 m/s occurs. On the other hand, at Hiuchi-nada, from $x=222$ km to $x=250$ km, an opposite wind behind the hydraulic jump is found. As shown in Fig. 13a, this opposite wind spreads as time passes, eventually covering the surface from $x=230$ km to $x=280$ km at $t=400$ min. This opposite wind is quite analogous to the characteristic features of the Yamajikaze such as the *Domai* and the northerly wind over the southern coastal plain of the Chugoku Peninsula, described in Section 2.

Figures 14a, 14b show vertical cross-sections of u and wind vectors for $U=2$ m/s ($t=400$ min). The surface wind over the northern slope is about 7 m/s, which is greater by a factor of three than the environmental wind, while a weak opposite wind is seen over Hiuchi-nada and the northern coastal plain of Shikoku Island. This opposite wind is quite analogous to the *Sasoi-kaze*, which is the characteristic premonitory symptom of the Yamajikaze.

When the Chugoku Mountains are added to the

Fig. 11. a) Vertical cross-section of the horizontal wind from $x=80$ km to $x=300$ km by EX-1 with $U=4$ m/s and a bell-shaped mountain at $t=80$ min. b) $t=240$ min. c) $t=400$ min.

Fig. 12. a) Time evolution of the surface wind from $x=80$ km to $x=300$ km for every 80 min by Ex-1 with $U=4$ m/s and a bell-shaped mountain. The broken line shows the linear analytic solution for the bell-shaped mountain.

b) As in a) except for the Shikoku Mountains without the Chugoku Mountains. The broken line shows the linear analytic solution for the orography of the Shikoku Mountains.

Fig. 13. a) As in Fig. 12b) except for the Shikoku Mountains and the Chugoku Mountains.

b) As in Fig. 11b) except for the Shikoku Mountains and the Chugoku Mountains.

c) As in b) except for the wind vectors from $x=150$ km to $x=260$ km. The lower-right arrow shows the scale of 8 m/s. Wind vector components in the vertical are enlarged by a factor of ten.

orography of the model island, the propagation speed of the hydraulic jump decreases as shown in Table 2. The jump remains on the lee side of the Shikoku Mountains whenever $U \leq 3$ m/s ($F_r^{-1} \geq 3.5$), while it is found more to the windward side than that in the experiments without the Chugoku Mountains.

The reversed flow develops for all cases of $U \leq 6$ m/s ($F_r^{-1} \geq 1.8$); however it is intermittent for the case of $U=6$ m/s ($F_r^{-1}=1.8$).

4.3 Analogy to the hydraulic jump in shallow water

The behavior of shallow water fluid over a ridge, including the formation of a hydraulic jump, has been previously studied (Long, 1954; Houghton and Kasahara, 1968; among others). Following hydraulic theory, the behavior is classified into four regimes: a sub-critical flow, a quasi-steady flow with a stationary hydraulic jump, an unsteady flow with a propagating hydraulic jump and a super-critical flow. According to Houghton and Kasahara (1968; Figs. 3 and 9), the behavior of the hydraulic jump is dictated by the fluid velocity when the mountain height

and the fluid depth are fixed. The hydraulic jump remains on the lee slope of the mountain if the fluid velocity is weak, while it propagates leeward if the fluid velocity is strong. In a quasi-steady regime, in which the jump remains stationary, the height of the mountain slope where the jump is located increases with the decrease of the fluid velocity. In an unsteady regime in which the jump propagates, the propagation speed of the hydraulic jump increases with the increase of the fluid velocity.

Figure 15 shows the dependence of the propagation speed or the location of the hydraulic jump on U , obtained by the numerical experiments described in the former sub-section. For the case of a bell-shaped mountain with $U=2$ m/s, a quasi-steady state is obtained in which the internal hydraulic jump is fixed on the lee side of the mountain. When $U \geq 3$ m/s, the hydraulic jump propagates leeward and its propagation speed increases as U increases. For the case of the orography of the Shikoku Mountains, the propagation speeds of the jump are somewhat smaller than those for the bell-shaped mountain, while the same tendency is found. When the

Fig. 14. a) As in Fig. 13b) except for $U=2$ m/s. The contour interval is 1 m/s.

b) As in Fig. 13c) except for $U=2$ m/s, with the lower right arrow representing 4 m/s.

Fig. 15. The U -dependence of the normalized propagation speed ($C^*=C/U$) of the internal hydraulic jump or the normalized height where jump remains over the lee slope ($M^*=h/h_m$) obtained from Ex-1. C means the propagation speed of the jump. h and h_m denote the height where the jump remains, and the height of the mountaintop, respectively. The open circles, triangles, and crosses show the results of the experiments with a bell-shaped mountain, the Shikoku Mountains only, and both the Shikoku and Chugoku Mountains, respectively.

orography of the Chugoku Mountains is added to the mountain shape, the propagation speed of the hydraulic jump becomes smaller, and the jump remains stationary when $U \leq 3$ m/s. It is considered that the blocking effect of the Chugoku Mountains reduces the effective magnitude of environmental wind. When the jump is stationary, the height of mountain slope where the jump is fixed increases with the decrease of U .

The response of the internal hydraulic jump to the magnitude of the environmental wind speed shown

in Fig. 15 is analogous to the behavior of hydraulic jump in shallow water. However, the reverse flow behind the hydraulic jump never occurs in the shallow fluid. More studies should be conducted in order to further the understanding of the relationship between the two.

5. Flow in the atmosphere with observed thermal stratification

5.1 The influence of an inversion layer on the Yamaji-kaze

As described in Section 2.3, a notable inversion layer was found at a level near the mountaintop for the Yamaji-kaze observed on 21 April 1987. Klemp and Lilly (1975) pointed out that similar inversion layers are usually found when strong downslope winds are observed over the Rocky Mountains. They calculated a linear analytic solution for a multi-layered atmosphere, and indicated that the amplitude of a linear mountain wave is significantly influenced by the atmospheric stability and wind profiles. They suggested that strong downslope winds occur when a certain optimal condition is satisfied. However, such a linear amplification mechanism for mountain waves is not applicable to non-linear mountain flow. Recently, Ikawa (1990b) made a theoretical investigation of the weakly non-linear effect on the flow of a two-layered stratified fluid past a two-dimensional mountain, and indicated that the applicability of linear theory can be unexpectedly narrow in some cases. According to his calculation, a large difference in the mountain drag sometimes occurs between the value obtained by linear theory and that by weakly non-linear theory, even when the inverse Froude number of the lower layer is only 0.2. In this sub-section, the nonlinear influence of the inversion layer at the level near the mountaintop on the Yamaji-kaze is numerically examined by comparative experiments using the observed thermal stratification.

Fig. 16. a) Composite of the vertical profile of temperature (T) and potential temperature (θ) for the soundings at Yonago, Kagoshima, and Shionomisaki at 09 JST 21 April 1987. The broken lines show the vertical profiles of T and θ of the reference smoothed atmosphere (*i.e.*, no low-level inversion). b) Time evolution of the vertical profile of the composited SSE-ly component of the wind for every 6 hours from 21 JST, 20 April to 15 JST, 21 April 1987.

Figure 16a shows the vertical profile of thermal stratification of the atmosphere used in the numerical experiments. This profile is a composite of the vertical profiles of temperature observed at Kagoshima, Shionomisaki and Yonago on 21 April 1987 at 09 JST (Fig. 8b–d). The composite was accomplished by averaging these three profiles using weighting coefficients which varied according to the distances between Shikoku Island and the observation sites. Near the inversion, the soundings were vertically adjusted before averaging to avoid the smoothing out of the low level inversion. A reference profile in which the inversion was smoothed out is shown by the broken lines. This smoothed profile is used in the comparative experiments.

Figures 17a–17d show the vertical cross-sections of the horizontal wind (u) and the deviation of potential temperature (θ) for the numerical experiments with and without the inversion ($U=12$ m/s, $t=240$ min). Strong downslope winds are found on the lee side of the Shikoku Mountains in both cases, and there is no significant difference in the magnitude of the maximum surface wind between the two.

Figures 18a–18d show the vertical cross-sections of u and θ for the case of $U=16$ m/s ($t=240$ min). Strong downslope wind is only seen in the case with the inversion, and no strong mountain wave is found in the case without the inversion. In these experiments, the inverse Froude numbers estimated by use of the mean tropospheric stability ($N_t=1.07 \times 0.01$ /s) are 0.93 for $U=12$ m/s and 0.70 for $U=16$ m/s. In Fig. 18c, the inversion layer ap-

pears to rise around $x=90$ km. This is the atmospheric bore propagating upstream, excited by the mountain.

For the case of $U=16$ m/s, additional experiments were performed in which horizontal domain was expanded to 640 km. No remarkable differences in the flow patterns were seen, and the differences in the magnitude of the maximum surface wind and the pressure drag were also small (the results are indicated by Δ^* and o^* in Fig. 19a, b).

Figure 19a shows the simulated maximum surface wind as a function of U . The magnitude of the maximum surface wind is always larger in the experiments with the inversion than in those without the inversion except for $U=10$ m/s, with the differences increasing as U increases. Wave breaking (areas in which u becomes negative or almost zero) was seen in all cases having the inversion, while it was not found in the cases of $U \geq 16$ m/s without the inversion ($F_r^{-1} \leq 0.7$).

Figure 19b shows the values of the surface pressure drag on Shikoku Island, defined as

$$DRAG = \int_{x=139 \text{ km}}^{x=225 \text{ km}} \left(p' \frac{dz_s}{dx} \right) dx. \quad (20)$$

For the cases without an inversion, the values of $DRAG$ are very small for $U=16$ m/s and $U=18$ m/s. These small values are thought to be associated with the non-linear “low drag state” reported by Bacmeister and Pierrehumbert (1988) and Ikawa (1990a).

Fig. 17. a) Vertical cross-section of the horizontal wind (u) at $t=240$ min for Ex-2 with an inversion layer and $U=12$ m/s. b) As in a) except without the inversion layer. c) As in a) except for potential temperature deviation (θ). d) As in b) except for θ .

Fig. 18. a)-d) As in Figs. 17a)-17d) except for $U=16$ m/s.

Fig. 19. a) The maximum surface wind as a function of the environmental wind speed U obtained from Ex-2. F_r denotes the value of the inverse Froude number estimated from the mean tropospheric stability. Δ indicates the use of an inversion layer, $\circ$ signifies no inversion layer. The experimental results for a horizontal domain of 640 km are indicated by Δ^* and $\circ^*$.

b) As in a) except for the surface pressure drag on Shikoku Island.

It is suggested by these experiments that when the environmental wind is not very strong, the Yamaji-kaze occurs regardless of the presence of the inversion layer. On the other hand, when the environmental wind is strong, the inversion layer near the mountaintop plays an important role in the occurrence of the strong Yamaji-kaze. There are still many uncertainties concerning the role of the inversion layer in non-linear mountain flow, and systematic investigations of these uncertainties should be the subject of a future study.

5.2 Time evolution of the Yamaji-kaze under a change in the environmental state

In the experiments for a homogeneous atmosphere with the Shikoku and Chugoku Mountains, the internal hydraulic jump remained on the lee slope of Shikoku Mountains when $U \leq 3$ m/s, and remained fixed over Hiuchi-nada when $U = 4$ m/s. The progression of a typical Yamaji-kaze follows steps from the occurrence of *Sasoi-kaze* to the onset of the Yamaji-kaze as described in Section 2. It is assumed that

this progression of the Yamaji-kaze is closely related to the propagation of the hydraulic jump under a time-changing environmental state. To check this hypothesis, a numerical experiment of the Yamaji-kaze observed on 21 April 1987 is made with the time-changing environmental wind taken into account.

Figure 16b shows the time evolution of the SSE-ly component of the composited wind for every 6 hours from 21 JST, 20 April to 15 JST, 21 April 1987. The procedure of the composition is same as that used in the composition of the temperature mentioned in the former sub-section. As shown in this figure, the synoptic southerly wind gradually increased over a day in response to the approach of the cyclone shown in Fig. 4. In this sub-section, the vertical profile of the composited wind on 15 JST, 21 April 1987 is taken as the representative wind profile during the mature stage of the Yamaji-kaze on 21 April 1987. The wind profile is characterized by the southerly wind having its maximum at $z = 1600$ m, decreasing with the height above $z = 1600$ m, and becoming negative above $z = 13$ km, with a critical level occurring in the environmental wind profile. Such a critical level at times plays an important role, depending on its height, in the occurrence of the downslope wind (Tomine, 1984; Durran and Klemp, 1987; Bacmeister and Pierrehumbert, 1988). However, it was considered that the pre-existing critical level did not play an important role in the occurrence of the Yamaji-kaze on 21 April 1987, since its height was sufficiently high and another critical level was induced below it by the action of wave breaking (WICL).

In order to incorporate the time-change of the synoptic wind, the experiment in this sub-section is initialized with fixed mountains. During the first 24 hours of model time, the environmental wind is increased linearly from zero to the composited wind profile at 15 JST, 21 April (Fig. 16b). After this period, the experiment is continued for 2 hours with the fixed environmental wind. The vertical profile of temperature shown in Fig. 16a is adopted for the environmental thermal stratification. The Boussinesq approximation is not employed to express the amplification of mountain waves due to the decrease of the mean density.

Figures 20a, 20b show the vertical cross-sections of the horizontal wind (u) and deviation of potential temperature (θ) at $t = 4$ hrs. At this time, the magnitude of the environmental wind is 1/6 of the final wind profile. The internal hydraulic jump is located on the northern slope of the Shikoku Mountains and an opposite wind is seen just behind it. This pattern is quite similar to the result of the experiments for a homogeneous atmosphere of $U = 2$ m/s, shown in Figs. 14a, 14b. The surface wind increases on the lee side near the top of the Shikoku Mountains,

Fig. 20. a) Vertical cross-section of the horizontal wind (u) from $x=140$ km to $x=400$ km at $t=4$ hrs by Ex-3. The contour interval is 2 m/s. b) As in a) except for the deviation of potential temperature (θ). The contour interval is 3K.

while it remains weak over the Chugoku Mountains. The isentropes dip on the lee slope of the Shikoku Mountains and are broken by the mountain surface, while the increase of surface temperature is obscure over the northern coastal plain of Shikoku Island, around $x=220$ km. As time elapses, this hydraulic jump propagates slowly leeward in response to the increase of the environmental wind, and crosses the northern coastal plain at $t=6$ hrs.

Figures 21a, 21b show the vertical cross-sections of u and θ at $t=9$ hrs 20 min. At this time, the hydraulic jump is located over Hiuchi-nada at $x=230$ km. A strong downslope wind greater than 18 m/s is blowing over the northern coastal plain of Shikoku Island. The opposite wind is seen over Hiuchi-nada from $x=235$ km to $x=256$ km. This wind pattern is quite similar to that shown in Figs. 13b, 13c, while the magnitude of the downslope wind is much stronger than that shown in Fig. 13b. The surface wind at the top of the Chugoku Mountains is strengthening, while a downslope wind is not found over the northern coastal plain of the Chugoku Peninsula. The surface potential temperature in the region from the northern coastal plain of Shikoku Island to the southern part of Hiuchi-nada has increased by about 2K. This simulated foehn is obviously a *dynamically* induced phenomenon (Ikawa and Nagasawa, 1989). Since the difference of inflow potential temperature between the surface and the level of the mountaintop (1050 m) is about 2K,

this increase in surface temperature means that the warm air mass near the mountaintop is brought down to the surface. It is considered that the vertical mixing due to wave breaking also plays an important role in this foehn event. As the integration continues, the hydraulic jump rapidly propagates leeward in response to the increase of the environmental wind. At $t=14$ hrs, it reaches the Chugoku Mountains and then dissipates.

Figures 22a, 22b show the vertical cross-sections of u and θ at $t=24$ hrs. At this time, the entire surface on the lee side of the Shikoku Mountains is covered by strong southerly winds. The stable layer is located between heights of 1100 m and 1600 m at the inflow boundary at $x=0$ km, while it is displaced to heights lower than 1000 m over the lee side of the Shikoku Mountains. On the other hand, the break of the isentrope by the ground surface is no longer seen, and the obvious foehn is no longer found. Such a flow pattern of which the streamlines are displaced downward is quite similar to the transitional flow in hydraulic theory (Houghton and Kasahara, 1968) or the theory given by Smith (1985). The experiment was continued with the fixed environmental wind until $t=26$ hrs, when the flow pattern became almost invariant.

5.3 Comparison of the experimental results with observations

The experimental results are compared with the

Fig. 21. a), b) As in Figs. 20a), 20b) except at $t=9$ hrs 20 min.

Fig. 22. a), b) As in Figs. 21a), 21b) except at $t=24$ hrs. The contour interval of u is 4 m/s.

Yamaji-kaze observed on 21 April 1987. It is not appropriate to compare directly the simulated surface wind (the horizontal wind at the lowest level) with the observed wind because surface friction was not included in the simulation, and the lowest model

level of the horizontal wind is set at a height of 100 m above the surface. Also, wind perturbations due to sub-grid-scale turbulent eddies are included in the observed wind. Therefore, the lowest-level ($z^*=100$ m) simulated wind at $x=220$ km is compared with

Fig. 23. a) Record of the surface wind at Doi (gust=solid line) and the simulated surface wind at $x=220$ km (broken line) in Ex-3. Solid straight line shows the time-change of the environmental wind U at $z=1600$ m supplied to the model. The symbol Δ shows the composited wind for the soundings at $z=1600$ m.

b) Time evolution of the simulated surface potential temperature variation at $x=220$ km and the observed surface temperature change due to the foehn event.

c) Time evolution of the simulated surface pressure deviation at $x=220$ km and the observed pressure change.

the gusts at Doi shown in Fig. 5. Figure 23a shows the time evolution of the two winds. In this figure, the observed gusts are expressed by negative values when the main wind direction is northerly, and by positive values when the main wind direction is southerly. Except for a quantitative difference, the two winds are similar in the pattern of the evolution. The quantitative difference is mainly caused by a time lag between their variations. In the ob-

served results, the northerly wind corresponding to the *Sasoi-kaze* was recorded from 05 JST to 09 JST, 21 April, while in the model results, the opposite wind was seen from $t=3$ hrs to $t=6$ hrs. If the time axis in Fig. 23a is enlarged by a factor of three for only the simulated wind, the evolutions of the two closely agree. Since the environmental wind used in the simulation was increased linearly until $t=24$ hrs as shown by the straight line, it is thought that the magnitude of the environmental wind required for the onset of the simulated Yamaji-kaze is one third that of the actual composited wind. The evolution of the simulated surface wind shown in this figure is also similar to the record of the surface wind in the analysis of the Yugoslavian Bora by Smith (1987) as can be seen in his Fig. 2.

Figure 23b shows the time evolution of the simulated surface potential temperature variation at $x=220$ km and the observed surface temperature change due to the foehn event. In this figure, the temperature change due to a foehn is calculated by subtracting the daily variation (about 4 C for a cloudy day) from the temperature variation recorded at Mishima, shown in Fig. 6. The model potential temperature exhibited sudden increase just after the passage of the hydraulic jump, and then attained a maximum value of 2.4K. This maximum increase of temperature is somewhat smaller than that observed (about 4 C). As in the case of the surface wind, the onset of the foehn is too early in the simulation, but the patterns of the time evolution are similar. The model surface potential temperature decreased after $t=12$ hrs. This evolution could be analogous to the decrease of the observed temperature at the zenith of the Yamaji-kaze, about 18 JST.

Figure 23c shows the time evolution of the simulated surface pressure deviation at $x=220$ km and the observed pressure change. In this figure, the observed pressure change is calculated by subtracting the pressure variation due to the migration of the cyclone from the observed pressure variation at Mishima, shown in Fig. 7. As in the evolutions of wind and temperature, the onset of the variation occurs too early in the simulation, but the patterns are similar. The maximum decrease in the surface pressure of the model is 8.5 hPa and is somewhat larger than that observed (about 6 hPa). The low pressure often observed over the northern coastal plain of Shikoku Island is a dynamically induced event.

6. Conceptual model of the Yamaji-kaze

Through the experiments given in Section 5.2, the development and propagation of an internal hydraulic jump was qualitatively simulated. The hydraulic jump propagated leeward in response to the increase of the environmental wind and a reversed flow developed just behind it. The characteristic

Fig. 24. Schematic representation of the conceptual model of the *Yamaji-kaze*. a) Premonitory stage, b) Mature stage, c) Strong wind case.

features of the *Yamaji-kaze* described in Section 2.2 can be explained by the evolution and migration of the hydraulic jump and its associated flow.

Figure 24a shows a schematic representation of a conceptual model of the *Yamaji-kaze* in its premonitory stage. When the southerly synoptic wind is still weak, the hydraulic jump occurs on the lee slope of the Shikoku Mountains. A reversed flow develops just behind the hydraulic jump and a northerly wind is observed over the northern coastal plain of Shikoku Island. This opposite wind corresponds to the *Sasoi-kaze* which is one of the premonitory symptoms of the *Yamaji-kaze*. At this time, the surface wind near the top of the Shikoku Mountains has already strengthened, and the *Yama-nari*, rumbling of the mountain can be detected in the cities on the northern coastal plain of Shikoku Island where the wind is still calm. The *Keta-kumo* may be explained by a stationary cloud which appears in the updraft at the hydraulic jump. On the other hand, on the windward side of the Shikoku Mountains, the southern coastal plain of Shikoku Island, is calm due to the blocking effect of the Shikoku Mountains.

As time passes, the hydraulic jump moves leeward according to the increase of the synoptic wind and the *Yamaji-kaze* begins to blow on the northern coastal plain of Shikoku Island. The surface wind varies suddenly at the location of the hydraulic jump, and the *Yamaji-kaze front* is observed. Before and after the onset of the *Yamaji-kaze*, the surface wind easily varies, reflecting the slight movement of *Yamaji-kaze front*. This variation corresponds to the *Mayoi-kaze*.

Figure 24b displays a schematic diagram of the

Yamaji-kaze in its mature stage. The hydraulic jump moves to Hiuchi-nada and strong a downslope wind blows over the northern coastal plain of Shikoku Island. In this area, surface temperatures are increased by the foehn event. The surface pressure decreases in the region from the lee slope of the Shikoku Mountains to Hiuchi-nada, and the *Hiuchi-nada depression* is formed. A northerly wind, opposite to the southerly downslope wind, prevails from the southern coastal plain of the Chugoku Peninsula to the hydraulic jump. This opposite wind corresponds to the *Domai*. On the upstream side of the Shikoku Mountains, the southerly wind is usually weak due to the blocking effect of the Shikoku Mountains.

The schematic representation of the conceptual model of the *Yamaji-kaze* when exceptionally strong synoptic winds occur is shown in Fig. 24c. Here, the hydraulic jump no longer appears, and strong southerly winds prevail over the entire area from the northern coastal plain of Shikoku Island to the southern coastal plain of the Chugoku Peninsula. This state corresponds to the case in which the *Domai* is not observed as described in Section 2.2. The *Yamaji-kaze* is observed over the northern coastal plain of Shikoku Island, while it is not always very strong, dependent on the thermal stratification. The increase in the surface temperature due to the foehn event is usually considered to be small compared to the case of Fig. 24b.

7. Summary and concluding remarks

The *Yamaji-kaze*—a local downslope wind which occurs over the northern coastal plain of Shikoku

Island—was numerically studied. The averaged orography of Shikoku Island in the east-west direction was regarded as the typical orography of Shikoku Island, exhibiting a steep slope on the northern side and a gentle slope on the southern side.

Steady linear and non-linear analytic solutions (Lilly and Klemp, 1979) of the two-dimensional mountain flow in a homogeneous atmosphere were calculated, and the effect of the asymmetric orography of Shikoku Island on the mountain flow was studied. It was found that the mountain wave over the Shikoku Mountains tends to be enhanced for the case of a southerly environmental wind and wave breaking easily occurs even for small inverse Froude numbers.

Numerical experiments were performed with a homogeneous fluid to examine the relationship between the behavior of the internal hydraulic jump conformed qualitatively to the hydraulic theory by Houghton and Kasahara (1968). For the case of a weak wind, a quasi-steady state was attained in which the hydraulic jump remained on the lee slope of the mountain. For the orography of Shikoku Island, the reversed flow just behind the hydraulic jump, as reported by Ikawa and Nagasawa (1989), was generated at smaller inverse Froude numbers than for the isolated bell-shaped mountain. When the orography of the Chugoku Mountains was added to the Shikoku Mountains, the propagation speed of the hydraulic jump decreased, and the opposite wind was generated for smaller inverse Froude numbers. It was concluded that the blocking effect of the Chugoku Mountains resulted in a weak effective magnitude of the environmental wind, which allowed the reversed flow to occur more readily.

The influence of an inversion layer at the level near the mountaintop on the Yamaji-kaze was examined by comparative numerical experiments using the observed thermal stratification and constant wind. It was indicated that when the environmental wind is not very strong, the Yamaji-kaze occurs regardless of the presence of the inversion layer. When the environmental wind is strong, the inversion layer near the mountaintop plays an important role in the occurrence of the strong Yamaji-kaze.

The development and propagation of an internal hydraulic jump were simulated by a numerical experiment using the observed thermal stratification and time-changing wind profile. The hydraulic jump propagated leeward in response to the increase of the environmental wind while a reversed flow developed just behind the jump. The evolution of the simulated surface wind, temperature, and pressure agreed qualitatively with the observed results.

In the experiments for a bell-shaped mountain with a homogeneous fluid described in Section 4, the reversed flow behind the hydraulic jump devel-

oped when $U \leq 3.5$ m/s ($F_r^{-1} \geq 3.0$), while it did not occur with $U \geq 4$ m/s ($F_r^{-1} \leq 2.6$). Namely, the critical value of inverse Froude number above which the reversed flow occurs was found to be between 2.6 and 3.0. This critical value may be somewhat dependent on the formulation of the turbulent closure model. In the experiments where the simpler turbulent closure model by Klemp and Wilhelmson (1978) was employed, the critical value was found to be between 2.1 and 2.6.

The reversed flow simulated in the present study differs from the reversed flow which occurs in the lee of an obstacle in a viscous fluid as a result of the separation of the viscous boundary layer. The former—which develops even behind a hydraulic jump propagating leeward far from the mountain—is affected by the magnitude of inverse Froude number, while the latter depends on the Reynolds number. It is known that a stagnant layer occurs in periodic valleys. Kimura and Manins (1988) simulated the occurrence of lee rotors in periodic valleys with the use of a hydrostatic model. However, the reversed flow simulated in the present study occurred even for an isolated mountain. Baines and Hoinka (1985) reported, from laboratory experiments of a two-dimensional stratified flow, that a blocked flow appeared on the lee of a ridge when the inverse Froude number was larger than 2. However, in their experiments, the blocking had a wave-like structure.

The behavior of the hydraulic jump conformed qualitatively to the hydraulic theory by Houghton and Kasahara (1968). However, the hydraulic jump in hydraulic theory is not accompanied by a reversed flow behind it. Recently, Durran (1986), Durran and Klemp (1987), Smith (1987), and Smith and Sun (1987) have attempted to make a correspondence between the behavior of a continuously stratified flow over a mountain in a nonlinear regime and the hydraulic theory by introducing the concept of “internal hydraulic theory”. The theoretical description of the behavior of the internal hydraulic jump, including the reversed flow behind it, is the subject of a future study.

Although the experimental results in Section 5.2 agreed qualitatively with observed results, some quantitative difference in the time of occurrence between the two remained. It appears that these differences were caused by the two-dimensionality of the model and the neglect of physical processes in the simulation. Arakawa (1969) indicated that the downslope wind becomes the strongest in the lee of a col of the mountain range. The actual Yamaji-kaze is probably influenced by the complicated three-dimensional effect of the real orography. Smolarkiewicz and Rotunno (1989) reported that a reversed flow was simulated on the lee side of a three-dimensional isolated mountain in a highly nonlinear flow. Such a reversed flow may keep the hy-

draulic jump stationary. The neglect of the surface friction appears to create large quantitative differences. Recently Richard *et al.* (1989) and Saito and Ikawa (1990) have reported that surface friction delays the onset of strong surface winds and also prevents the downstream propagation of the hydraulic jump. The results of the three-dimensional simulation of the Yamaji-kaze, including physical processes, will be reported at a later date.

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やまじ風の数値的研究

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日本における代表的なおろし風「やまじ風」を対象に、非対称な山に対する2次元山越え気流が非静水圧モデルを用いてシミュレートされる。やまじ風の特徴である「やまじ風前線」とその後面での逆向きの風が、内部ハイドロリックジャンプとそれに伴う循環によって説明される。

一様流体に対する数値実験によれば、内部ハイドロリックジャンプの振舞は一般流の大きさや山の形に強く影響される。大きな逆フルード数の場においては山の後面でジャンプが殆ど停滞する準定常なレジームが得られ、その後面に一般場とは逆向きの風が発生する。やまじ風のケースでは四国山地の非対称性と中国山地の存在がジャンプの停滞とその後面での逆風をより生じやすくしている。

1987年4月21日のやまじ風では山頂付近の高度に明瞭な逆転層が存在していた。逆転層がある場合には無い場合と比較して地表風はより大きくなる事が、実況データに基づく大気プロファイルを用いた数値実験により示される。一般流の大きさを時間とともに変化させる実験により、内部ハイドロリックジャンプの発生と移動が定性的にシミュレートされ、やまじ風の概念モデルが示される。