Numerical Experiments on the Interaction of Ocean and Bay Breezes

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Abstract

On bay coasts, a characteristic local wind circulation is formed, because bay breezes and ocean breezes are caused by bay-land and ocean-land contrasts, respectively, and they interact with each other. For the purpose of examining basic characteristics of the circulation, several numerical experiments are performed. The three-dimensional model used in this study has a square bay of various sizes with the ocean on the south side, taking a model bay of 50 km in width and 100 km in length as our standard size.

Results for the standard bay model show noticeable differences between the east and west bay coasts. Near surfaces, hodographs rotate anti-clockwise on the east coast and clockwise on the west coast. Along the west coast, the depth of the ocean breeze is almost the same as that of ordinary sea breezes, and strong oceanward return currents blow aloft. On the other hand, along the east coast, the depth of the ocean breeze is much greater than that of ordinary sea breezes, and only a weak oceanward return flow can be seen.

The north-south circulation along the east or west coast is the composition of the real ocean breeze and the influence in the opposite direction of the northward bay breeze. Therefore, the scale of the total circulation is much larger than that of ordinary sea breezes, and can be extended to even more than 200 km. Through the influence of the eastward or westward bay breeze and its return current, the stratification surrounding the ocean breeze is stable. Since internal gravity waves are generated in the stably stratified layer, signals propagate at the group velocity of the waves. The propagation speed, although it somewhat slows down due to advection, is therefore faster than that of an ordinary sea breeze front.

The results from the standard width model show that circulations along the east and west coasts are independent of each other. On the other hand, the 20 km width causes the interaction of the circulation on both coasts, so that the ocean breeze is no longer deep even on the east coast. The model with the length of 40 km does not show a propagating phenomenon but a steady circulation.

1. Introduction

On bay coasts, a characteristic local wind circulation is formed, because bay and ocean breezes interact with each other, as found by McPherson (1970), Warner *et al.* (1978), Abbs (1986) and so on. Here "bay breeze" means a local wind caused by bay-land contrast, blowing from bay coasts toward lands, and "ocean breeze" is a local wind caused by ocean-land contrast, blowing from oceans to lands along bay coasts. In a similar manner, two land breezes, one of which is toward the ocean (hereafter referred to as Lo breeze) and the other is toward the bay (referred to as Lb breeze), may interact. This

interaction has not been documented so far, perhaps because the land breeze velocity is much weaker than that of ocean and bay breezes.

The purpose of the present study is to investigate the basic characteristics of this interacting circulation, by using three-dimensional numerical models. In order to get basic characteristics, we propose to use a simple configuration of coastlines, say a square bay together with flat land, rather than realistic topography. The standard size of the model bay is 50 km in width and 100 km in length (the depth is usually used instead of the length, but the word "depth" might cause confusion in this case, so we use the length instead in this paper), with an ocean on the south side. This size approximates the con-

Fig. 1. A map of the area near the Kii Channel and the model coastline of Experiment 1, where the shaded region shows land.

figuration around the Kii Channel in Japan (see Fig. 1), which is our main concern as later stated. However, the interaction between the two breezes largely depends on the bay size. We will therefore carry out several numerical experiments for bays of various sizes. Moreover, it is worthwhile to investigate differences in land and sea breezes between the two side coasts of the bay, *i.e.*, the east and west coasts. The two coasts have different relationships between the pressure gradient forces caused by bay-land and ocean-land contrasts and the Coriolis force, so that the circulation may also be different between the two coasts.

The main motivation of this study comes from observations around the Kii Channel. The Channel has the two side coasts, *i.e.*, the east coast of Wakayama side and the west coast of Tokushima side, with the Pacific Ocean on the south side. Therefore, this area is also where the interacting ocean-bay (channel) breeze can be seen. Three interesting phenomena discovered in this area should be pointed out.

The first one is a difference in the rotational direction indicated in hodographs of diurnal wind variation between the two coasts (*e.g.*, Itoh and Noda, 1989). That is, an anti-clockwise rotation is observed along the east coast, while a clear clockwise rotation is found along the west coast. There should be other differences between the two coasts, which

have not been observed. It is important to examine the differences and their causes.

Second, the ocean and Lo breezes are of considerable depths on the east coast (*e.g.*, Mizuma, 1985). Namely, the depths of the ocean and Lo breezes are about 1100 m and 800 m, respectively, which are much greater than those of sea and land breezes observed in general (Hsu, 1970; Lyons, 1972; and many others). On the other hand, the depth of the bay breeze is several hundred meters, which coincides with general observations. Why the depths of the ocean and Lo breezes are so great needs clarification. Further, it is interesting to investigate depths of these breezes on the west coast, which have not been observed so far.

Finally, the area with the anti-clockwise rotation of wind hodographs spreads over 100 km along the east coast (*e.g.*, Itoh and Noda, 1989). Since this rotation appears by the influence of the ocean breeze (Itoh and Kawazoe, 1983), we can infer that the scale of the ocean breeze is larger than 100 km. According to the linear theory of land and sea breezes (Niino, 1987), the scale is calculated as less than 100 km. Many observations also show similar scales (*e.g.*, Hsu, 1970; Lyons, 1972). Why does the ocean breeze here have such a large scale?

We must answer these questions, in addition to clarifying basic features of the interaction between the ocean and bay breezes. On the other hand, these observational facts should be used for the verification of our numerical models.

In Section 2, the three-dimensional model used in this study will be described, as well as the sizes of model bays. Our results will be given in Section 3. In the beginning of this section, results from the standard experiment will be shown, followed by discussion about differences between the two bay coasts. Basic characteristics of the ocean breeze will be summarized in the next subsection. In the final subsection in Section 3, results from various sizes of the model bay will be shown. In Section 4, we will give our conclusions and some remarks. In order to compare these results with those for an ordinary sea breeze under the same parameter values, we will also construct a two-dimensional model. The outline of the model and results for this experiment will be shown in Appendix.

2. Model description

The model used in this study is almost the same as the model of Ookouchi and Wakata (1984). As well as for the atmosphere, temperatures for the ground surface and the soil are calculated using the surface energy balance and soil temperature equations. On the other hand, sea surface temperatures are fixed. The main difference between our model and Ookouchi and Wakata (1984)'s is that, in our model, all land surfaces are flat, and therefore, the vertical

coordinate is the z -coordinate.

(a) *Basic equation in the atmosphere*

All time-dependent variables are deviations from basic unperturbed quantities, which are functions of z only. The basic equations are expressed as follows under independent variables, x (eastward), y (northward), z (upward) and t (time);

$$\begin{aligned} \frac{\partial u}{\partial t} = & -\frac{\partial}{\partial x}(u^2) - \frac{\partial}{\partial y}(uv) - \frac{\partial}{\partial z}(uw) \\ & +fv - \Theta \frac{\partial \pi}{\partial x} + \frac{\partial}{\partial x} \left(K_H \frac{\partial u}{\partial x} \right) \\ & + \frac{\partial}{\partial y} \left(K_H \frac{\partial u}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_V \frac{\partial u}{\partial z} \right), \end{aligned} \quad (1)$$

$$\begin{aligned} \frac{\partial v}{\partial t} = & -\frac{\partial}{\partial x}(vu) - \frac{\partial}{\partial y}(v^2) - \frac{\partial}{\partial z}(vw) \\ & -fu - \Theta \frac{\partial \pi}{\partial y} + \frac{\partial}{\partial x} \left(K_H \frac{\partial v}{\partial x} \right) \\ & + \frac{\partial}{\partial y} \left(K_H \frac{\partial v}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_V \frac{\partial v}{\partial z} \right), \end{aligned} \quad (2)$$

$$\begin{aligned} \frac{\partial \theta'}{\partial t} = & -\frac{\partial}{\partial x}(\theta'u) - \frac{\partial}{\partial y}(\theta'v) - \frac{\partial}{\partial z}(\theta'w) \\ & -Sw - \Gamma\theta' + \frac{\partial}{\partial x} \left(K_H \frac{\partial \theta'}{\partial x} \right) \\ & + \frac{\partial}{\partial y} \left(K_H \frac{\partial \theta'}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_V \frac{\partial \theta'}{\partial z} \right), \end{aligned} \quad (3)$$

$$\frac{\partial \pi'}{\partial z} = g \frac{\theta'}{\Theta^2}, \quad (4)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \quad (5)$$

where $\pi = C_a(p/p_o)^\kappa$. Symbols used in these equations are u , v and w (x , y and z components of winds); Θ and θ' (basic and perturbed potential temperatures); π and π' (total and perturbed Exner functions); p_o and p (reference pressure and pressure); T (temperature); K_H and K_V (horizontal and vertical eddy diffusion coefficients); C_a , f and g (specific heat of dry air at constant pressure, Coriolis parameter and gravitational acceleration); κ ($= R/C_a$; R is the gas constant for dry air); S and Γ (basic lapse rate of the potential temperature and Newtonian cooling rate). While the horizontal diffusion coefficient is constant, the vertical one is a time- and space-dependent variable. For the formulation of the vertical diffusion coefficient, refer to Ookouchi and Wakata (1984).

Although the reader may think that the term, $\partial/\partial z(K_V \partial \Theta/\partial z)$, should be added in Eq. (3), we checked the effect of this term and found that overall the features were similar, including the vertical profile of the potential temperature.

(b) *Surface energy balance and soil temperature equations*

The land surface temperature, T_G , is calculated from the surface energy balance equation, together with several deductive and empirical equations. For more detail, refer to Ookouchi and Wakata (1984) again. On the other hand, the sea surface temperature, T_S , is assumed to be constant. Further, the surface temperature on coastlines is calculated as $0.5(T_G + T_S)$. The soil temperature is computed from the following heat conduction equation:

$$\frac{\partial T_G}{\partial t} = -\frac{\partial}{\partial z} \left(K_g \frac{\partial T_G}{\partial z} \right), \quad (6)$$

where K_g is the thermal conductivity for the soil.

(c) *Initial and boundary conditions*

At the initial time, the atmosphere is assumed to be at rest, i.e.,

$$u, v, w, \pi', \theta' = 0. \quad (7)$$

The surface and underground temperatures are the same as the sea surface temperature, T_s . Since $\theta' = 0$, the potential temperature in the atmosphere is $\Theta(z) = \Theta(z=0) + Sz$.

The upper and lower boundary conditions are

$$u, v \text{ and } w = 0 \quad \text{at } z = 0, \quad (8)$$

$$u, v, \theta' \text{ and } \pi' = 0 \quad \text{at } z = z_t, \quad (9)$$

$$\partial T_G/\partial z = 0 \quad \text{at } z = z_b, \quad (10)$$

where z_t and z_b are the height of the model top and the bottom depth of the model under the ground, respectively.

The lateral boundary conditions are

$$\frac{\partial}{\partial n}(v_n, \pi' \text{ and } \theta') = 0 \quad \text{and} \quad v_t = 0, \quad (11)$$

where v_n and v_t are the normal and tangential velocity at lateral boundaries, respectively.

(d) *Finite difference method*

The above basic equations are transformed into finite difference forms. A staggered grid is adopted, that is, θ' and π' are defined at same grid points, and u (v, w) is set at the center of two adjacent grid points for θ' and π' in the x (y, z) direction. Grid intervals in the x and y directions are 5 km. Vertical layers in the atmosphere are set at 50 m, 100m, and then every 200 m up to 2900 m ($=z_t$). Soil levels where the soil temperature is computed are 5, 15 and 35 cm below the ground surface. The flux is calculated between them, and z_b is -55 cm. Centered differencing is adopted as space differencing except for advective terms, for which the modified upstream method (Soong and Ogura, 1973) is used.

Time differencing for the atmospheric equations is done by the leap-frog method except for the vertical diffusion terms, for which the backward implicit scheme is used. In calculating the soil temperature, forward differencing is adopted, and the Newton method is used for computing the ground surface temperature. The time increment is 30 sec. Time integration starts at 6 LST and is carried out for 42 hours (*i.e.*, through 0 LST on the second day). After 22 LST on the second day, the model features are almost the same as on the first day (see Fig. 6), so that the 42 hour time integration is sufficient for our purpose. Data are stored at every 30 minutes, and results will be shown for the data from 0 LST on the first day through 0 LST on the second day.

The following values are input to the parameters: $T_s=298$ K, $S=0.0038$ K/m, $K_H=100$ m²/sec, $\Gamma=1/(2\text{days})$, $K_g=4\times 10^{-7}$ m²/sec, and the latitude is 34°N. Besides, the following parameters are necessary to calculate the surface energy balance; the solar declination = 20°, the solar constant = 1.353×10^3 J/(m²·sec), surface albedo = 0.3, the Boltzman constant = 5.67×10^{-8} J/(m²·sec·K⁴), water vapor pressure in the air = 27mb, air density = 1.2 kg/m³, the Bowen ratio = 0.0 when the stratification in the lowest layer is stable, and 2.0 when it is unstable, soil density = 2.5×10^3 kg/m³, the specific heat for the soil = 840J/(kg·K).

This solar declination corresponds to about May 21st or July 24th, and the sunrise and sunset times are 504 and 1857 LST, respectively. The amplitude of the ground surface temperature was about 9K for these parameters.

(e) Coastal configuration

The standard configuration of the model bay is shown in Fig. 1. This approximates the configuration of the land and sea around the Kii Channel, although Awaji Island and the Inland Sea are ignored. The bay size is 50 km in width and 100 km in length, and the other lengths of the configuration are shown in Fig. 2a. In this area, the lateral boundaries little affect motions around the model bay, because the motions there in this model are almost identical with those in a model with an extended area where 50 km is further added to each length shown in Fig. 2a. Three other bay sizes, 20 × 100 km, 50 × 200 km and 50 × 40 km, are taken to ascertain the dependency of the ocean-bay breeze interaction on the bay size. These numerical experiments are named Experiments 1, 2, 3 and 4, respectively.

Another series of experiments are performed for the configuration of Fig. 2b, which is the eastern part of Fig. 2a from the line along the west coast. The reduced area is to save the computation time. These experiments are planned to clarify the basic process of the ocean-bay breeze interaction. Experiment S1 is carried out for this configuration. The

Fig. 2. Configurations of the sea and land in this study. Units in length are km. (a) For Experiments 1, 2, 3 and 4. The bay size is changed with each experiment. The main wind directions of the B_N , B_E , B_W and ocean (abbreviated as O.) breezes are shown by arrows. (b) For Experiments S1, S2, S3 and SL.

north-west part labelled A in Fig. 2b is replaced with sea in Experiment S2, while the south-east part labelled B is replaced with land in Experiment S3.

(f) Definition of some words

A square bay causes bay breezes in three directions; northward breezes occurring on the north bay coast, eastward breezes on the east bay coast and westward breezes on the west bay coast. These will be called B_N , B_E and B_W breezes, respectively. The Lb breeze also occurs in three directions. Among them, only a southward breeze on the north coast will be referred to as L_N breeze.

Characteristics of the ocean and Lo breezes over areas far from the bay must be identical with those for ordinary sea and land breezes, which are not our concern. Our interest is in the interaction between the ocean and bay breezes or the Lo and Lb breezes. Therefore, attention is paid to the behavior of the ocean and Lo breezes along the bay coasts. The main wind directions of the B_N , B_E , B_W breezes and the ocean breeze along the bay coast are shown in Fig. 2a.

3. Results and discussion

(a) Results for the standard bay

In this subsection, results for Experiment 1 will be shown. Figure 3 illustrates horizontal wind vectors at 16 LST and 4 LST at the height of 100 m. The height of 100 m is where the strongest wind blows in general. The asymmetry between the east and west coasts can be seen at both times. Namely, at 16 LST, the northward component is dominant on

Fig. 3. Wind vectors at the height of 100 m at 16 LST (left) and 4 LST (right) in Experiment 1.

Fig. 4. Wind vectors and perturbed Exner functions [unit: 10^{-2} J/(kg·deg)] in Experiment 1 at 16 LST at 1300 m (left) and at 4 LST at 700 m (right).

the west coast, while the southward component can be seen near the center of the east coast. On the other hand, at 4 LST, the northwesterly can be seen on the west coast, while the east-north-easterly blows on the east coast.

The asymmetry between the east and west coasts is more conspicuous at upper layers. Figure 4 shows

wind vectors and pressure (perturbed Exner function) patterns at 16 LST at 1300 m and at 4 LST at 700 m. The heights, 1300 m and 700 m, are where return currents are the strongest at each time. At 16 LST, the northerly flow prevails on the west coast, but only very weak winds blow along the east coast. The flow pattern is strongly governed by an-

Site B on the West Coast

Site E on the East Coast

Fig. 5. Time-height cross sections of horizontal winds at site B on the west coast (upper) and site E on the east coast (lower) in Experiment 1. Sites B and E are indicated in Fig. 6. Directions of arrows show wind directions. Upside arrows mean a northerly flow. The number of barbs expresses wind velocity. $0.25 \text{ m/sec} < \text{one barb} < 0.75 \text{ m/sec}$, $0.75 \text{ m/sec} < \text{two barbs} < 1.25 \text{ m/sec}$, and so on. No arrow means that the velocity is less than 0.25 m/sec .

tycyclones formed over western and eastern land areas. At 4 LST, an anticyclone is formed over the north side of the bay. The wind is again affected by the anticyclone, and the northerly and easterly can be seen on the east and west coasts, respectively.

Figure 5 illustrates time-height cross sections of winds on the west and east coasts. This figure also shows the conspicuous asymmetry. The easterly and

southerly winds on the west coast in the day time, which are the bay breeze and the ocean breeze, respectively, have the same depth of about 500m. This depth coincides with that of the sea breezes observed in general (see also Appendix). On the other hand, on the east coast, the maximum depth of the southerly is as much as 1100m at 18 LST, while the depth of the westerly is about 500 m. Weak ocean-

Fig. 6. Wind hodographs at the 100 m level for 7 sites in Experiment 1. Sites A and D (B and E; C and F) are located 25 km (50 km; 75 km) north of the ocean coast. Site G is the center of the north coast. Numerals attached to the hodographs indicate LST. Units of wind velocity are m/sec.

ward return currents can be seen only above 1500 m. The asymmetry in the land breeze between the two coasts is also marked from 2 LST through 9 LST, although the feature that the oceanward flow exists above the thin bayward flow is the same. On the west coast, the depth of the oceanward flow is very thin and the return current with respect to the bayward flow is relatively deep. We cannot see any return current with respect to the oceanward flow here. On the east coast, however, the oceanward flow is of considerable depth. We can clearly see the return current with respect to the oceanward wind and there is no return current with respect to the bayward wind. These features on the east coast are similar to those observed by Mizuma (1985).

Next, we will show wind hodographs at several sites in Fig. 6. The three hodographs on the west coast show a clear clockwise rotation, compared with the hodograph at site G, which is similar to that for the ordinary sea breeze in Appendix (not shown). On the other hand, the rotational direction of the other three hodographs at the east coast is anti-

clockwise. These features are the same as observed. From more careful inspection, the hodographs have slight differences site by site on each coast. The south wind component increases at first at sites A, C, D and F, and later increases at points B and E. On the other hand, the time when the maximum southerly occurs is earlier at points A and D than at points B, C, E and F, where the time is almost the same.

The force that causes these differences is the pressure gradient in the northward direction, whose time changes along the east coast are shown in Fig. 7. The figure along the west coast is similar, although the force in the daytime is somewhat smaller. This clearly shows the propagation of the pressure gradient force from the ocean coast and the north bay coast to the middle of the east coast in the daytime. The former corresponds to the ocean breeze, and the latter is an influence of the B_N breeze in the opposite direction, which will be referred to as the backward influence of the B_N breeze. Except at places very close to the ocean and north bay coasts,

Fig. 7. Time changes of the pressure gradient force in the northward direction at 100 m level along the east coast (from the ocean coast to the north coast) in Experiment 1. Units are 10^{-4} N/kg.

the largest pressure gradient occurs earlier near the ocean coast (near site D) than near the north bay coast (near site F) and the middle of the east coast (near site E). These characteristics agree well with those of the wind hodographs given above. Thus, the circulation along the east coast (and therefore also along the west coast) is influenced by both the real ocean breeze and the backward influence of the B_N breeze. Since the former penetrates northward along the coast and the latter penetrates southward, it is natural that the apparent scale of the circulation along the coast is much greater than the scale of the ordinary sea breeze.

Until now, we have used the term "ocean breeze" ambiguously. We will hereafter use it to mean the real ocean breeze, and the circulation along the bay coast influenced by both the ocean breeze and the backward influence of the B_N breeze will be called the *total ocean breeze* circulation.

We can also estimate the moving speed of the ocean breeze front from Fig. 7. The time of the front passage can be defined as abrupt increases of the pressure gradient force, and may be objectively determined as the time when the value, $\tan^{-1}[(P^{k+1} - P^k)/\Delta P_s] - \tan^{-1}[(P^k - P^{k-1})/\Delta P_s]$,

Fig. 8. The vertical structure of the perturbed potential temperature and the (v, w) components along the east coast (0 km corresponds to the ocean coast) in Experiment 1. (a) at 18 LST, and (b) at 6 LST. The contour interval is 1K. Shading: $>0K$, and dark shading: $>3K$.

is the largest, where P^k is the pressure gradient at time k and ΔP_s is a standard change of the pressure gradient per 30 minutes. Although ΔP_s is taken to 10^{-4} N/(kg·30min), we can obtain similar results, even by taking other values around this ΔP_s . In this and later cases, the front passage thus defined almost coincides with the contour of 10^{-4} N/kg from the ocean coast northward to about 45 km. The propagation speed can be estimated at about 1.9 m/sec, which is larger than the speed of the south wind itself (see Fig. 6).

Figure 8a illustrates the vertical section of the perturbed potential temperature and the (v, w) components at 18 LST along the east coast. This figure clearly shows that the ocean breeze is suppressed by downward motions. These downward motions are caused by the return flow of the B_E breeze as shown in Fig. 9a. Further, the stratification surrounding the ocean breeze is almost stable (see also Fig. 11).

At night, the story is almost reversed. The circu-

Fig. 9. The horizontal structure of vertical motions in Experiment 1. (a) at 18 LST at 600 m, and (b) at 6 LST at 400 m. 600 m and 400 m are the heights where the upward motions show their maxima at each time. Contour intervals are written at the bottom of each figure. Shading indicates upward motion.

lation along the east or west coast is the sum of the L_N breeze and the backward influence of the L_o breeze, although there are several differences from the daytime circulation. As a matter of course, the winds and pressure gradient force at night are smaller than in the daytime (see Figs. 3, 4, 5, 6 and 7). Although the L_N breeze front cannot be determined definitely, it proceeds without accompanying vertical motions, as shown in Fig. 8b. No vertical motion along the bay coast can be seen from Fig. 9b showing the vertical motions at 400 m at 6 LST. The penetration of the L_b breeze into the bay is negligible, so that the bay coast is the border for upward and downward motions.

(b) *Discussion of the differences between east and west coasts*

The main differences in the daytime circulation between the east and west coasts may be summarized as follows; anti-clockwise versus clockwise rotation of wind hodographs near surfaces, thick as opposed to ordinary depths of the southerly, and weak versus strong oceanward return currents. These differences can easily be understood from the pressure gradient and Coriolis force, as shown in the following.

The pressure gradient force in an east-west direction has its maximum at about noon on the east or west coast at the 100 m level (not shown). On the other hand, the maximum pressure gradient in

a north-south direction occurs in the late afternoon, except near the ocean coast or north bay coast, as a function of distances from the two coasts (see Fig. 7). According to Kusuda and Alpert (1983), this time lag between the eastward and northward pressure gradient, as well as the fact that the amplitude of the northward pressure gradient is not so small, is of great assistance to the anti-clockwise rotation of wind hodographs on the east coast. In other words, the lag between the B_E breeze and the total ocean breeze results in anti-clockwise rotation. On the west coast, on the other hand, this relation causes clockwise rotation.

Next, let us explain the reason why anticyclones are formed over the eastern and western land areas in the daytime (see Fig. 4) before discussing upper layer motions. We do this because the upper layer winds are strongly governed by anticyclones. In the lower layers, the maximum convergence occurs at areas where the B_E or B_W breeze and the ocean breeze are confluent (see Fig. 3). These areas should be located near $N45^\circ E$ ($N45^\circ W$) going from the junction of the ocean coast and the east (west) coast. Upward motions show their maxima there, and the anticyclones are formed at return current layers. Since the convergence and upward motions are more intense than those over the northern land areas, the anticyclones are stronger than the high pressure formed in upper layers over the northern land.

Table 1. Differences between the total ocean breeze along the bay coast and the ordinary sea breeze.

		total ocean breeze	ordinary sea breeze
1	horizontal scale	large	small
2	wind velocity	weak	strong
3	accompanying vertical motion	downward	upward
4	stratification	stable	unstable
5	speed of the fronts	fast	slow

Considering the Coriolis force, the anticyclone and high pressure act in concert on the west coast so that a strong northerly blows. On the other hand, since they act in opposition of the east coast and the anticyclone dominates the high pressure, the southerly can rather be seen at ordinary return current levels. The depth of the southerly, which resembles the depth of the ocean breeze, is therefore great.

The same can be said about the asymmetry in the land breeze. The anticyclone is formed over the north side of the bay in layers where return currents can be usually seen, because strong convergence and upward motions occur there (see Figs. 3 and 9). The land breeze is so weak that wind motions in these layers are governed only by the anticyclone. Therefore, the northerly (easterly) can be seen at the 500~900 m level on the east (west) coast. The northerly and the easterly look like the Lo breeze and the return current with respect to the Lb breeze, respectively, but the mechanism is different as explained above.

(c) *Basic characteristics of the total ocean breeze and their causes*

Comparing results in Subsection (a) with those in the Appendix, many characteristics of the total ocean breeze are very different from those of the ordinary sea breeze. The differences are summarized in Table 1. Observational support of items 1, 4 and 5 in the ordinary sea breeze can be found in, *e.g.*, Hsu (1970) and Lyons (1972).

These differences come from two factors. One factor is that the total ocean breeze interacts with the bay breeze (B_E or B_W breeze), while the ordinary sea breeze has no interaction. Another factor is a difference in differential heating. The ordinary sea breeze results from the direct sea-land contrast. On the other hand, the total ocean breeze is caused by the ocean-northern land contrast. We still do not know which factor is responsible for which difference in Table 1. It is especially interesting to examine what determines the fast propagation of the ocean breeze front, because this makes the scale of the total ocean breeze additionally large. Moreover, we do not understand the relation between the total

ocean breeze and the ocean-northern land contrast: It is clear from Fig. 7 that the total ocean breeze is not caused by the direct ocean-northern land contrast, and the ocean breeze (the backward influence of the B_N breeze) seems independent of the northern land (the ocean). We must clarify this relation. These problems will be pursued in the remaining of this subsection.

A way to clarify the relation between the total ocean breeze and the ocean-northern land contrast is to perform and compare Experiments S1, S2 and S3. Namely, if results for Experiment S1, in which both the northern land and the ocean are retained, are identical with a sum of results for Experiments S2 and S3, in which the northern land and the ocean are eliminated, respectively, the northern land (the ocean) exerts no influence on the ocean breeze (the backward influence of the B_N breeze).

Results for Experiment S1 are similar to those for Experiment 1 (not shown). Results for Experiments S2 and S3 are shown in Figs. 10a and b, respectively, in terms of the time change of the northward pressure gradient along the coast oriented in a north-south direction. As a matter of course, the position of the maxima moves northward (southward) with time in Fig. 10a (Fig. 10b), and the maximum values become smaller. Figure 10c is the sum of Fig. 10a and b, and we can say that it is almost the same as Fig. 7. It is quite like the figure for Experiment S1. Thus, the ocean breeze (the backward influence of the B_N breeze) along the coast is independent of the northern land (the ocean). It might be said that the ocean breeze (the backward influence of the B_N breeze) is caused by the ocean-coast (the coast-northern land) contrast. Here "coast" has a wide meaning; it means a zone along the east or west coast, which has an average property of the land and bay.

Next, it is interesting to see the role of advective nonlinearity in the ocean breeze, particularly in relation to the behavior of the front, because sea breeze fronts behave like gravity currents, which are highly nonlinear, as observed by Simpson *et al.* (1977) and Mitsumoto *et al.* (1983). This also gives a clue to interpret the fast propagation of the ocean breeze front, as will be later shown. Thus, we perform Ex-

Fig. 10. As in Fig. 7 except (a) in Experiment S2, (b) in Experiment S3, (c) the sum of (a) and (b), and (d) in Experiment SL.

periment SL, in which all advective terms are eliminated and the coastline is the same as in Experiment S1. The results are shown in Fig. 10d. This figure is similar to Fig. 7, except for two minor differences: There is no north-south asymmetry and there is no abrupt increase in the pressure gradient. These characteristics can easily be interpreted from the well known property of advective nonlinearity. Therefore, the overall circulation of the total ocean breeze can be understood without the advective effect.

Further, it can be calculated from this figure that the propagation speed of the ocean breeze front is about 2.9 m/sec. This speed is identical with the group velocity of internal gravity waves under the following values and formulation; $S=0.0038$ K/m, $K_V = 5$ m²/sec, and $c = \sqrt{\alpha g S} \cdot K_V^{1/2} \omega^{-1/2}$ (Niino, 1987), where α is the expansion coefficient of the air and $\omega = 2\pi/(24\text{hours})$. In other words, without advective terms, the propagation of signals is done by internal gravity waves, as suggested by Kimura and Eguchi (1978) and Niino (1987). On the other hand, when advective terms are taken into account, the propagation speed of the front (1.9 m/sec in Experiment 1) cannot be completely identified as the speed of the internal gravity waves, but slows down due to the influence of advection.

Now we can proceed to the problem of cause and effect in the total ocean breeze circulation. For this purpose, we again use the two-dimensional model described in Appendix. A difference from the model in Appendix is in the model surface, which is composed of an ocean, a "coast" and a land. Their lengths are 50 km, 100 km and 100 km, respectively. The "coast" means a place where the surface temperature is calculated as $0.5(T_G + T_S)$. In this model, we will see a "modified total ocean breeze", which means that it is caused by the ocean-coast and coast-land contrasts but never interacts with bay breezes. This experiment is called Experiment T.

Figure 11 illustrates the vertical profiles of the potential temperatures at 10 LST and 16 LST over the center of the coast. For comparison, the profiles over the center of the east coast in Experiment 1 are also shown. At 10 LST when the B_E breeze just begins to blow in Experiment 1 (see Fig. 6), both profiles are almost the same. On the other hand, at 16 LST, the potential temperature in Experiment 1 is cooler in the lower layer and warmer in the upper layer than in Experiment T. It is clear that the cool air in the lower layer is brought by the B_E breeze and the warm air in the upper layer is due to the downward motions. Therefore, the stratification at 16 LST in Experiment 1 is stable except below the 50 m level. On the other hand, Experiment T shows that the stratification is unstable in the deep layer below 500 m at 16 LST, even though the surface heating over the coast is weaker than over the land.

Figure 12 illustrates the time change of the pressure gradient force along the coast. The maximum values are comparable to those in Fig. 7 and smaller than in Fig. A1. This similarity and difference can also be seen in the wind velocity (not shown), and clearly come from those in the differential heating, *i.e.*, the strong sea-land contrast vs. the weak ocean-coast and coast-land contrasts.

The propagation speed of the front is about 1.0 m/sec in this figure. This speed, as well as that of the ordinary sea breeze, is much slower than that of the ocean breeze. The penetration of the front is therefore much weaker than in Experiment 1 or S1.

The slow propagation speeds of the fronts in this experiment and the experiment in Appendix can be interpreted as follows: Since the stratification is unstable in the deep layer in these experiments, internal gravity waves are not generated there. Therefore, signals of the ocean-coast (sea-land) contrast cannot propagate at the fast group velocity of the waves, although we still do not know what determines the slow speeds in these cases. This interpretation can be verified from the behavior of a "modified Lo breeze" front in Fig. 12. The contour line of -10^{-4} N/kg associated with this front is quite similar to that in Experiment 1 or S3. This is probably because the "modified Lo breeze" occurs in stably stratified layers, and internal gravity waves are generated even in this experiment as well as the other experiments. This interpretation is further consistent with the results of Estoque (1962), in which the propagation speed of sea breeze fronts is fast, when the stratification is stable. Physick (1980) also obtained the fast propagation of sea breeze fronts in his model. This is probably because the unstable stratification immediately disappears by the convective adjustment incorporated in his model.

Thus, most of the characteristics for the total ocean breeze listed in Table 1, *i.e.*, items 3, 4 and 5, are mainly caused by the interaction with the bay breeze (B_E or B_W breeze). Item 1 is also greatly affected by the bay breeze, via item 5. Figure 13 shows the relationship between the causes and results in the total ocean breeze circulation.

(d) Results for the model bays with the different widths and lengths

In this subsection, we will show results from the experiments with the different bay sizes. Only main differences from the results for Experiment 1 will be mentioned.

First, the width of the model bay is changed. We can however find from Figs. 3 and 4 that the winds on the east coast are almost independent of those on the west coast, *i.e.*, closed circulation occurs on each coast. A model with a greater width than 50 km would give similar results to Experiment 1. Then, the width is reduced to 20 km. Figure 14 shows

Fig. 11. The vertical structure of the potential temperatures at 10 LST (left) and 16 LST (right) over the center of the "coast" in Experiment T (solid line) and over the center of the east coast in Experiment 1 (dashed line).

wind vectors at 100 m and 1300 m at 18 LST in this experiment. Southerlies are dominant all over the bay at 100 m. Also at the 1300 m level, northerlies are prominent, even along the east coast. We can say that the winds on one coast interact with the winds on the other coast at this width. The behavior of the pressure gradient force is also different from that in Experiment 1, as shown in Fig. 15. The force is larger at the southern part of the bay than at the northern part. In accordance with this property, the penetration of the ocean breeze front is deeper than that for Experiment 1, so that the asymmetry with respect to the north-south direction is more conspicuous. In other words, the influence of the ocean breeze is relatively large, while the backward influence of the B_N breeze is small.

Next, returning to the standard width, we will see effects of the length. Experiment 3 is performed, with a length of 200 km. Figure 16 illustrates a wind hodograph at the 100 m level on the site located 100 km north of the ocean coast along the east coast, together with a hodograph at the same point in Experiment S2. In Experiment 3, the hodograph still shows anti-clockwise rotation. We can say from this fact that the total ocean breeze has the scale of more than 200 km under the parameters adopted in the present study. However, the anti-clockwise rotation in this experiment is produced by the joint influence of the ocean breeze and the backward influence of the B_N breeze, because the hodograph in Experiment S2, in which there is only the ocean breeze, does not show any anti-clockwise rotation. Therefore, the scale of 200 km seems almost the maximum.

The length in Experiment 4 is 40 km. The time changes of the pressure gradient and the v compo-

Fig. 12. As in Fig. 7 except along the "coast" in Experiment T.

nent on the east coast are shown in Fig. 17. The results for the west coast are similar. Although the pressure gradient and the southerly show larger values on the nearer site to the north coast than on the central and south parts of the east coast, the time

Fig. 13. The relationship between the causes and results in the total ocean breeze circulation.

Fig. 14. As in Fig. 3 except at 100 m (left) and 1300 m (right) at 18 LST in Experiment 2.

when they have their maxima is almost the same at every site. In other words, there is no propagating phenomenon but a steady circulation. The structure of the circulation may be identified as steady heat island convection, which Kimura and Eguchi (1978) regarded as the basic process of land and sea breezes in the case of a short length.

4. Conclusions and remarks

Several numerical experiments have been performed to understand the basic features of the interaction between the ocean and bay breezes. Special

attention is paid to the behavior of the ocean breeze along the bay coasts where the bay breeze is dominant. In the three-dimensional model used in this study, the ground surface temperature shows diurnal variation, being calculated from the surface energy balance equation, while the sea surface temperature is fixed. The model has a square bay with an ocean at the south side. The standard size of the model bay is 50 km in width and 100 km in length, which approximates the coastline around the Kii Channel in Japan. The other sizes of 20 × 100 km, 50 × 200 km and 50 × 40 km are also used to see how the

Fig. 17. Time changes of the pressure gradient force (left; unit: N/kg) and the v component (right) at 10 km (solid line), 20 km (dashed line) and 30 km (dash-dotted line) north of the ocean coast along the east coast in Experiment 4.

scale of more than, but not much more than 200 km. The model with the length of 40 km does not show a propagating phenomenon but a steady circulation.

At least two phenomena among these conclusions have not been observed yet around the Kii Channel. They are the backward influence of the northward bay breeze and the strong southward return current on the west coast. It should be checked whether they are really observable or not. However, we might have difficulty observing the former, because pressure gradient data cannot be obtained and it is difficult to find out the backward influence from wind data only. On the other hand, it will be easy to observe the latter phenomenon, if it exists.

Our model topography is too simple. Therefore, as a matter of course, there are several differences from observations. One of the biggest differences is that the southerly is weak in our model, especially over the bay area. There are some possible causes: One is the sea surface temperature. There might be a difference in the sea surface temperature between the Kii Channel and the Pacific Ocean. Further, the sea surface temperature used in this experiment might be higher than the real one. The second is the realistic topography, especially the orographic effect and the fact that the width of the Channel decreases as it goes northward. When the ocean breeze was strong, the characteristics of the total ocean breeze could be different from those obtained in this study. It will require further work to pursue these problems.

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Appendix

Two-dimensional model and results—Ordinary sea breeze

The equations, the boundary and initial conditions and the finite differencing in the two-dimensional model are identical with those in the three-dimensional model, except that the v component is computed at the points where θ' and π' are defined. The model length is 250 km, which is composed of a sea region of 100 km and a land region of 150 km. The sea breeze obtained from this model will be referred to as the ordinary sea breeze.

The time change of the pressure gradient force in the landward direction is shown in Fig. A1. The maximum values are much larger than those in Fig. 7. In accordance with this result, wind speeds of the sea breeze (not shown) are also larger than those associated with the ocean breeze. The propagation speed of the sea breeze front which is estimated to be 1.4 m/sec is slower than that for the ocean breeze front. As a result, the scale of the sea breeze is less than 100 km, which is estimated from the inland penetration of the sea breeze front in the daytime. Figure A2 illustrates the vertical section of the perturbed potential temperature and the (v, w) components at 18 LST. The depth of the sea breeze is

Fig. A1. As in Fig. 7 except for the ordinary sea breeze.

Fig. A2. As in Fig. 8a except for the ordinary sea breeze.

about 500 m. Further, it can be seen that the sea breeze is accompanied by a strong updraft, and the stratification is unstable in the deep layer. Of course, this strong instability is artificially brought on by the hydrostatic approximation, therefore this unstable layer should be regarded as a mixed layer.

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大洋風と湾風との相互作用に関する数値実験

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湾岸域では、大洋–陸間および湾–陸間の熱的コントラストによって生じる大洋風および湾風の相互作用のため特有な循環が形成されている。この基本的な過程を明らかにする目的で、3次元数値モデルを用いていくつかの数値実験を行った。モデルは長方形の湾を持ち、その南に大洋がある。基準の実験における湾の大きさは、幅 50 km・奥行き 100 km である。

結果は、湾の東岸と西岸の循環に顕著な差があることを示す。地表近くのホドグラフは、観測されるように、東岸で反時計回り回転、西岸で時計回り回転を示す。さらに、西岸においては大洋風の厚さが一般的な海風の厚さと同程度で、その上で大洋に向けた強い反流が吹くのに対し、東岸では大洋風はかなりの厚さを持ち、反流はかなり上空で見られるのみである。

東岸または西岸の南北循環は、真の大洋風と、北向きの湾風の逆方向への影響が合成されたものになっている。従って、その循環のスケールは一般的な海風のそれよりずっと大きく、200 km にも及ぶ。また、大洋風循環はいくつかの点で海風の特徴と異なっている。東向きまたは西向きの湾風とその反流の影響で、大洋風は下降流を伴い、安定成層中で生じる。このなかで内部重力波が形成されるので、シグナルの伝播にはその群速度が重要な役割を果たす。従ってその伝播速度は一般的な海風のそれよりかなり速い。

幅 50 km の湾では、東岸と西岸の循環はほぼ独立であったが、幅 20 km になると相互に関係し、東岸の大洋風ももはやそんなに厚くない。奥行き 40 km ではほとんど伝播する現象は見えず、ほぼ定常な循環となる。