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ON THE INFLUENCES OF COASTAL UPWELLING ON THE DIURNAL  
VARIATION OF SEA BREEZE CIRCULATION

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## 1. INTRODUCTION

Observations indicate that the sea breeze circulations in many coastal regions are under the influence of coastal upwelling and this influence appears to be especially important along Oregon and Peru-Chile coasts. One most intriguing phenomenon along the Peru-Chile coast is the existence of a narrow cloud-free zone and the spreading of stratus clouds away from the coast. Clancy *et al.* (1979) developed a coupled air-ocean model and found that the air-sea feedback to the coastal upwelling process is exceedingly weak and could not explain this phenomenon. In this paper we shall adopt a consistent boundary-layer model which includes both the mechanical and the thermal influences associated with the longshore wind and the oceanic coastal upwelling. Assuming that the perturbation flow is independent of the longshore direction, we find that they are governed by the same set of equations derived and treated by Kuo (1973) in his planetary boundary layer analysis except that the representations of time and horizontal variations of the flow variables will not be restricted to a single Fourier component. We thus find that even though the coastal upwelling does not alter the one-cell type structure of the land-breeze circulation during the night, but it exerts a substantial influence on the day time sea-breeze circulation and changing it into a two-cell pattern, with rising motion over both the land and the open ocean regions and a strong descending motion over the coastal region. This circulation also produces a strong marine inversion over the coastal upwelling region, which fits the observational results reported by Elliott and O'Brien (1977) over the Oregon coast. However, because of the presence of the mean onshore wind in this middle latitude region, the Oregon coast is not dry.

## 2. PLANETARY BOUNDARY LAYER MODEL

## 2.1 Basic equations

We consider the same basic flow and use the same basic equations nondimensionalized by the same scheme as in our other paper (Kuo and Chu, 1984) in this Proceedings, but including the influence of the time variation. We then find that the stream function  $\psi$  for the perturbation velocity components  $u$  and  $w$  in the  $x$ - $z$  plane, the perturbation velocity  $v$  and potential temperature  $q$  are governed by the following dimensionless equations.

$$(M^2 + \eta^2) \psi_{\zeta\zeta} + R_1 \psi_{xx} = 0, \quad (1)$$

$$Mv = -\eta \psi_{\zeta} \quad (2)$$

$$Mq - R_1 \psi_x = 0 \quad (3)$$

where

$$M = \partial^2 / \partial \zeta^2 - \frac{\partial}{\partial t}, \quad \zeta = z/\delta, \quad \delta = \left(\frac{\nu}{Z_0}\right)^{1/2}, \quad (4)$$

$$X = x'/L, \quad R_1 = \frac{N^2 \delta^2}{\nu^2 L^2}. \quad \text{In addition, the}$$

basic vertical velocity  $w$  associated with the longshore wind also produce a change of the potential temperature and this change is also given by (3) but with  $\psi_x$  replaced by the dimensionless form of  $w_s$ . We expand the flow variables into double Fourier series in  $t$  and  $x$  as:

$$\psi(x, z, t) = R_e \sum_{k=-\infty}^{\infty} \sum_{n=1}^{\infty} \psi_n^{(k)}(z) \cdot \sin n\pi x \exp(i \frac{k\omega t}{2\Omega}) \quad (5a)$$

$$v(x, z, t) = R_e \sum_{k=-\infty}^{\infty} \sum_{n=1}^{\infty} v_n^{(k)}(z) \cdot \sin n\pi x \exp(i \frac{k\omega t}{2\Omega}) \quad (5b)$$

$$q(x, z, t) = R_e \sum_{k=-\infty}^{\infty} \sum_{n=0}^{\infty} q_n^{(k)}(z) \cdot \cos n\pi x \exp(i \frac{k\omega t}{2\Omega}) \quad (5c)$$

The basic equations for the Fourier components then become

$$D^6 \phi_n^{(k)} - 21K\sigma D^4 \phi_n^{(k)} + (F^2 - K^2 \sigma^2) D^2 \phi_n^{(k)} - (n\pi)^2 S^* \phi_n^{(k)} = 0 \quad (6a)$$

$$D^2 v_n^{(k)} - K\sigma i v_n^{(k)} = -FD \phi_n^{(k)} \quad (6b)$$

$$D^2 q_n^{(k)} - K\sigma i q_n^{(k)} = S^* (n\pi) \phi_n^{(k)} \quad (6c)$$

where

$$D = \frac{d}{d\zeta}, \quad F = \eta, \quad \sigma = \frac{\omega \delta^2}{\nu} \quad (7)$$

As pointed out by Kuo (1973) the complimentary solutions  $v_n^{(k)}$  and  $q_n^{(k)}$  of (5c) and (6c) must

satisfy the thermal wind relation:

$$F D V_{nc}^{(k)} = Q_{nc}^{(k)} \quad (6d)$$

## 2.2 Diurnal variation of net radiation

According to Kondratyev (1969, Fig. 10.1) the diurnal variation of the net radiation can be represented by the following formula:

$$R_{o,L}(t) = \begin{cases} \bar{R}_{o,L} (d_2 \cos \omega t + d_1) & 0 \leq |t| \leq t_o \\ \bar{R}_{o,L} (d_2 \cos \omega t_o + d_1) \frac{\pi}{\omega} > |t| > t_o \end{cases} \quad (8)$$

where  $\bar{R}$  is the annual mean of the latitude dependent net radiation and subscripts o and L are for ocean and land, respectively, and  $t = 0$  corresponds to noon time,  $t_o$  the sunset time, and  $-t_o$  the sunrise time. In the computation we take

$$t_o = 2\pi/3\omega, \quad d_1 = 0.1224408, \quad d_2 = 0.336746$$

$$d_3 \equiv d_2 \cos \omega t_o + d_1 = -0.04592,$$

$$\bar{R}_o = 133.31 \text{ Watts m}^{-2}$$

$$\bar{R}_L = 86.88 \text{ Watts m}^{-2}.$$

These values of  $\bar{R}$  are for  $\phi = 20^\circ S$  and are adopted from Table 16 in Sellers' (1965) book. Expanding  $R_o(t)$  and  $R_L(t)$  into Fourier series in "t" as  $R_{o,L}(t) = \sum_{k=-\infty}^{\infty} R_{o,L}^{(k)} \exp(i \frac{k\omega t}{2\Omega})$ . (9)

where the components are given by

$$R_{o,L}^{(k)} = \frac{\bar{R}_{o,L}}{k\pi} (d_1 - d_3) \sin\left(\frac{2}{3} k\pi\right) + \frac{d_2 \bar{R}_{o,L}}{\pi(1-k^2)} \cdot \left[ \cos\left(\frac{2}{3} k\pi\right) \sin\left(\frac{2}{3} \pi\right) + k \sin\left(\frac{2}{3} k\pi\right) \cos\left(\frac{2}{3} \pi\right) \right] \quad (k \neq -1, 0, 1)$$

$$R_{o,L}^{(0)} = \frac{\bar{R}_{o,L}}{3} (2d_1 + d_3) + \frac{\bar{R}_{o,L}}{\pi} d_2 \sin\left(\frac{2}{3} \pi\right)$$

$$R_{o,L}^{(1)} = \frac{\bar{R}_{o,L}}{\pi} (d_1 - d_3) \sin\left(\frac{2}{3} \pi\right) + \frac{\bar{R}_{o,L}}{4\pi} d_2 \cdot \left(\frac{4}{3} \pi + \sin\left(\frac{4}{3} \pi\right)\right)$$

$$R_{o,L}^{(-1)} = R_{o,L}^{(1)} \quad (10)$$

## 2.3 The solutions

Using the same surface heating, no-slip and normal boundary conditions at the surface ( $\zeta = 0$ ) as in our other paper (Kuo and Chu, 1984) in the proceedings, we find that the solutions of the equations (6a) (6b) and (6c) can be written as

$$\phi_n^{(k)}(\zeta) = C_n^{(k)} \left( e^{\lambda_1 \zeta} + \frac{\lambda_1 - \lambda_3}{\lambda_3 - \lambda_2} e^{\lambda_2 \zeta} + \frac{\lambda_2 - \lambda_1}{\lambda_3 - \lambda_2} e^{\lambda_3 \zeta} \right) \quad (11a)$$

$$V_n^{(k)}(\zeta) = D_n^{(k)} e^{Pk\zeta} - C_n^{(k)} \left( \frac{\lambda_1}{\lambda_1^2 - P_k^2} e^{\lambda_1 \zeta} + \frac{(\lambda_1 - \lambda_3)\lambda_2}{(\lambda_2^2 - P_k^2)(\lambda_3 - \lambda_2)} e^{\lambda_2 \zeta} + \frac{(\lambda_2 - \lambda_1)}{(\lambda_3^2 - P_k^2)(\lambda_3 - \lambda_2)} e^{\lambda_3 \zeta} \right) \quad (11b)$$

$$Q_n^{(k)}(\zeta) = FP_k D_n^{(k)} e^{Pk\zeta} + (n\pi) S^* C_n^{(k)} \cdot \left( \frac{e^{\lambda_1 \zeta}}{\lambda^2 - P_k^2} + \frac{(\lambda_1 - \lambda_3) e^{\lambda_2 \zeta}}{(\lambda_2^2 - P_k^2)(\lambda_3 - \lambda_2)} + \frac{(\lambda_2 - \lambda_1) e^{\lambda_3 \zeta}}{(\lambda_3^2 - P_k^2)(\lambda_3 - \lambda_2)} \right) \quad (11c)$$

Where  $C_n^{(k)}$  and  $D_n^{(k)}$  are the integration constants determined by the lower boundary conditions and  $\lambda_1, \lambda_2, \lambda_3$  are the roots with negative real parts from the following sextic equations:

$$\lambda^6 - 2ik\sigma\lambda^4 + (F^2 - k^2\sigma^2)\lambda^2 - (n\pi)^2 S^* = 0 \quad (12)$$

$$\text{and } Pk = \sqrt{\frac{k\sigma}{2}} (-1 - i).$$

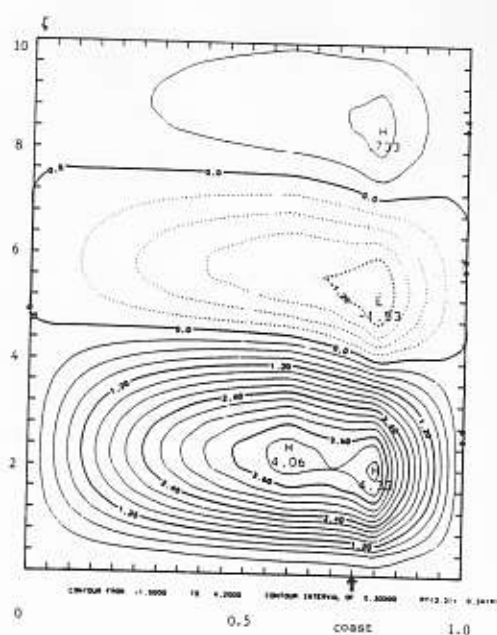
It is noticed that in (12) we neglect the subscripts "n" and superscripts "k" for  $\lambda_1, \lambda_2, \lambda_3$ .

In this paper we set the Bowen ratios over land and water as  $B_s = 50, B_w = 0.1$ , respectively. The equations are solved under two different conditions, namely, 1) the conditions depicted in the previous subsections (control run), and 2)  $\Delta T_w^* = 0$ , that is, no coastal upwelling case.

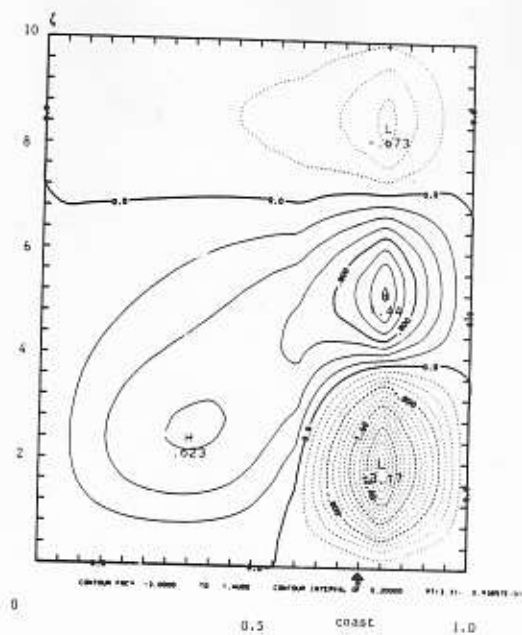
## 3. Model results and discussion

In all the figures in this paper, the units of  $x$  and  $\zeta$  are 200 km and the Ekman depth respectively.

Figs. 1, 2 show the stream function and flow field in the cross-coastal section during daytime (2:00 p.m.) and night (2:00 a.m.), respectively. Notice that the stream function is only for disturbances, whereas the vertical velocity in Fig. 2 includes both  $\psi_x$  and the basic flow vertical velocity  $W_s$  from the longshore wind.

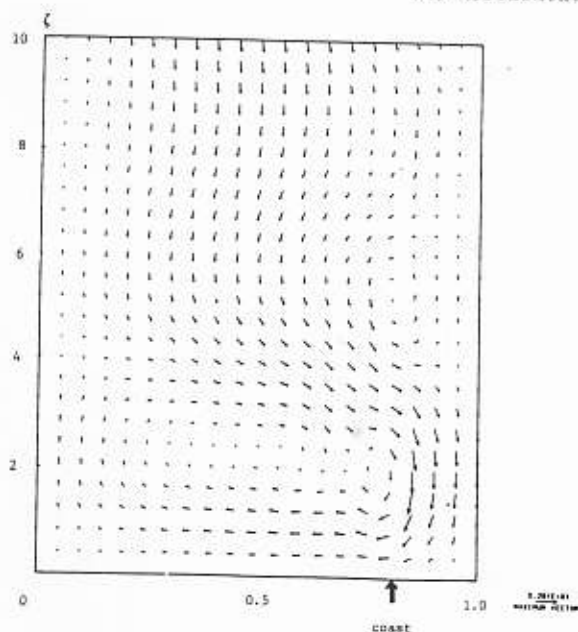


a. 2:00 a.m.

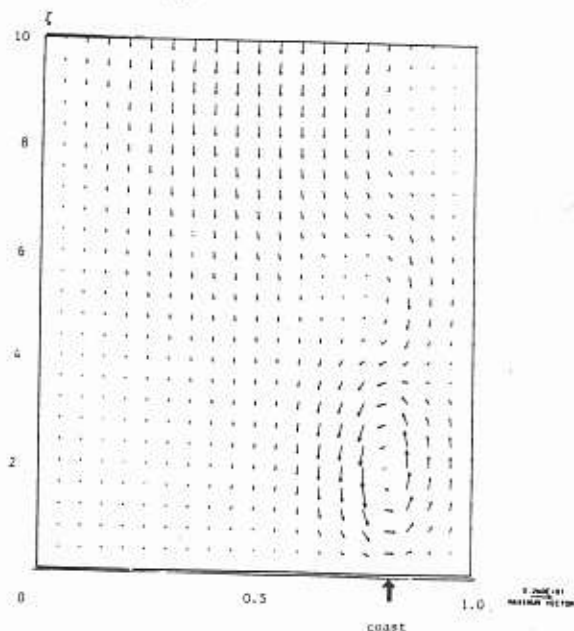


b. 2:00 p.m.

Fig. 1. Stream function under diurnal variation of radiation. Unit =  $132.5 \text{ m}^2 \text{ s}^{-1}$ , the horizontal length-scale is 200 km.



a. 2:00 a.m.



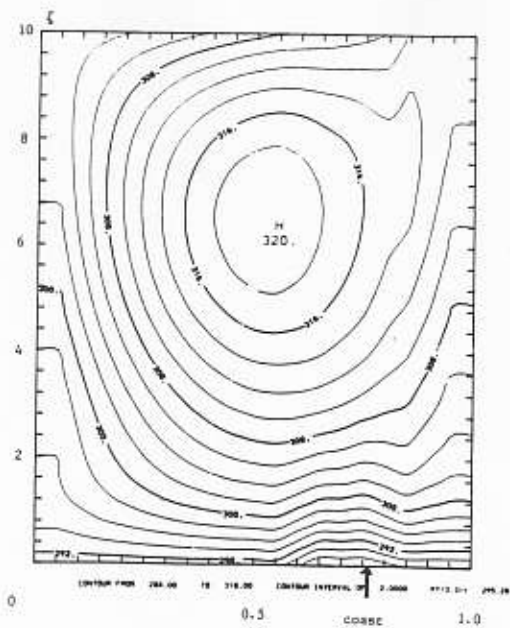
b. 2:00 p.m.

Fig. 2. Velocity vector in XZ-plane under diurnal variation with  $w$  enlarged 100 times.

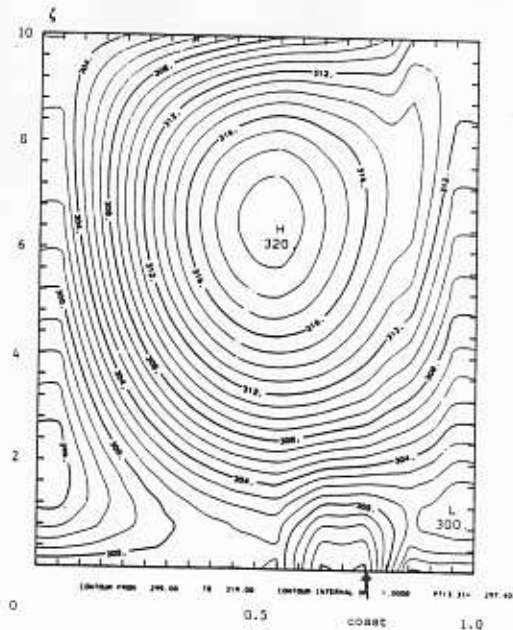
We can see that the coastal upwelling does not affect the land breeze during the night too much, but it substantially affects the sea breeze during the day, changing the one-cell type circulation into a two-cell structure. This result can explain the observed phenomena along the Chilean coast, namely, along the northern two-thirds of the Chilean coastline, persistent stratus and

fog extend hundreds of kilometers out over the ocean (Miller, 1973), and there is a narrow cloud-free zone along the Peru-Chile coast and a spreading of stratus and fog away from the coast (Miller, 1971).

Fig. 3 (a, b) shows the distributions of the potential temperature during daytime and

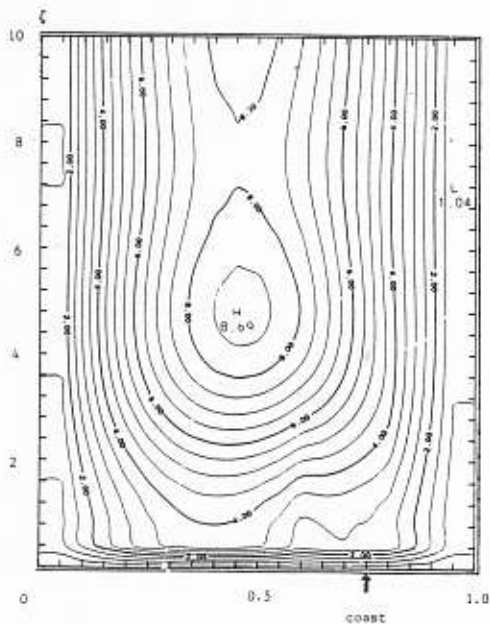


a. 2:00 a.m.

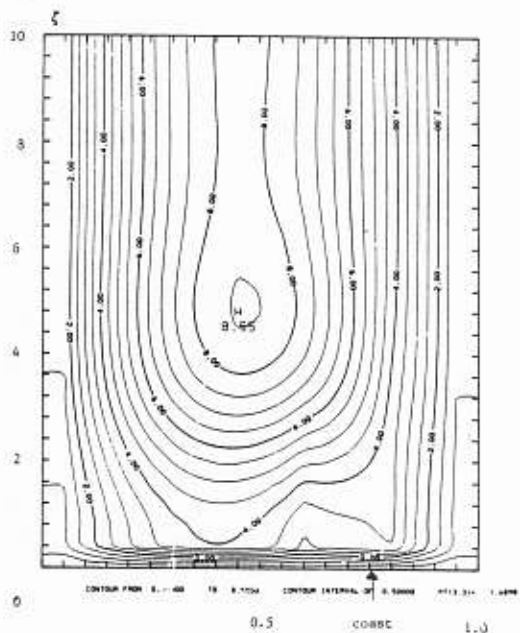


b. 2:00 p.m.

Fig. 3. Potential temperature in XZ-plane (in  $^{\circ}\text{K}$ ).



a. 2:00 a.m.



b. 2:00 p.m.

Fig. 4. Longshore wind in  $\text{m s}^{-1}$ .

during night. We can see the stable stratification over the coast.

Fig. 4a,b show the diurnal variation of the longshore wind. We see that a maximum zone over the ocean about 40-50 km away from the coast. The longshore wind decreases both seaward and

landward from the maximum zone. This distribution fits the observational results reported by Elliott and O'Brien (1977) over the Oregon coast (Fig. 3 in their paper). The difference between the control run and no upwelling run gives the effects of coastal upwelling.



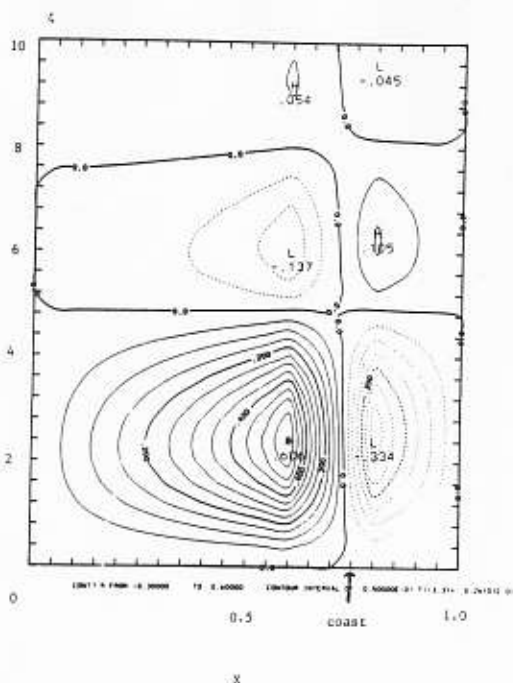


Fig. 5. Stream function from daily mean ocean surface temperature difference  $\Delta T_w$ .

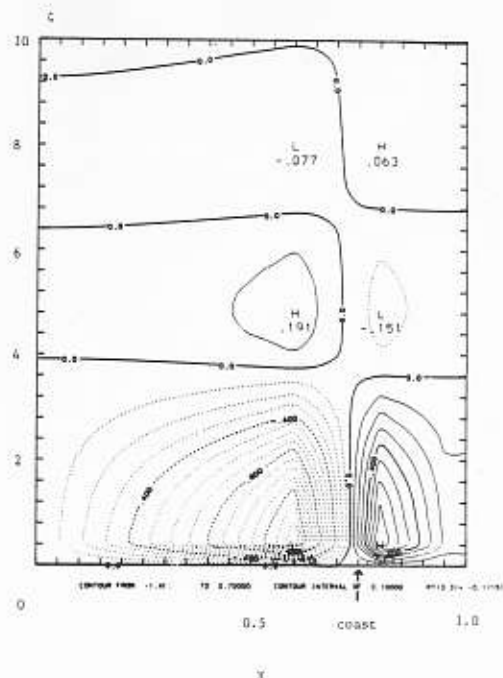


Fig. 7. Longshore wind from daily mean  $\Delta T_w$ , in  $m s^{-1}$ .

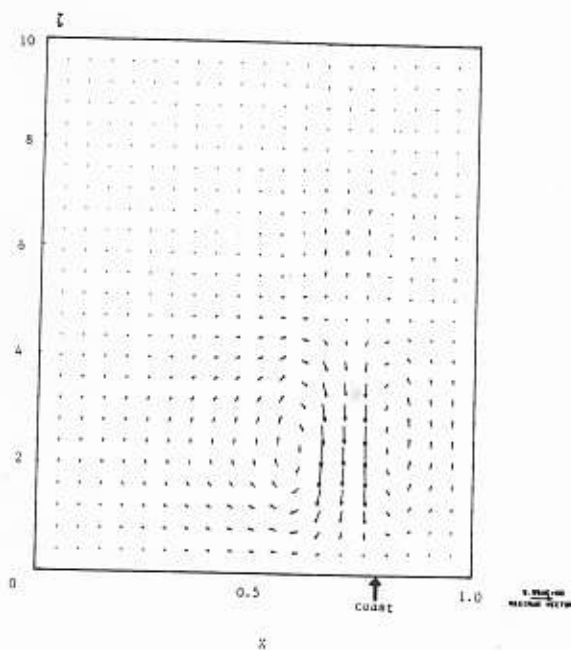


Fig. 6. Velocity vector in XZ-plane from daily mean  $\Delta T_w$ , with  $w$  enlarged 100 times.

Fig. 5 shows the influences of coastal upwelling on the diurnal variation of the stream function. The effect is the same during daytime and at night. The coastal upwelling tends to break the sea breeze into two cells.

Fig. 6 shows the influences of coastal upwelling on the flow field. We can see the same results as in Fig. 5.

Fig. 7 shows the influences of coastal upwelling on the longshore wind. This effect is rather small compared to the longshore wind itself. This conclusion is in agreement with the results of Clancy et al. (1979).

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