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AN INSTABILITY THEORY OF AIR-SEA INTERACTION FOR COASTAL UPWELLING

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ABSTRACT

A surface wind (seabreeze), thermally generated by differential sea surface temperature, is introduced to *Gill-Clarke's model (1974) through* wind stress for investigating the effects of seabreeze on coastal upwelling. A coupled air-sea system is treated as an eigenvalue problem. The solutions show that the thermally forced local winds break down the coastal Kelvin wave into three parts: small-scale ($L<100$ km) growing and stationary modes, mesoscale (100 km$<L<$200 km) decaying and fast-moving modes, and 'large'-scale ($L>200$ km) coastal Kelvin modes. The consistency of the length scale between the most growing mode predicted by this model and the observed cold/warm alternation pattern of surface water near the Peruvian Coast (around 15 °S) implies that seabreeze may play some role in coastal upwelling.

I. INTRODUCTION

An upwelling study JOINT II (Moody et al., 1981; Stuart, 1981)along the Peruvian Coast (near 15°S)shows that sea surface temperature (SST) *reveals stationary* short wave pattern in the longshore direction (Fig. 1). The wave length, which is defined as the distance between two adjacent cold (or warm) plume centers, is around 55 km. What causes the formation of this small-scale wave pattern? It is an interesting problem, but hasn't been solved yet.

JOINT II data also show that the area and temperature of the plume are correlated with the wind stress (Stuart, 1981). Up to now, many theories treat coastal upwelling as a wind-forced problem, i.e., the surface wind is taken as a known function. In fact, surface temperature gradient generates surface winds, which in turn exert an extra force on the ocean surface. Therefore, coupling the air with ocean is important in investigating coastal upwelling. In this paper, a surface wind stress, brought on by thermally forced wind, is added to Gill-Clarke's model (1974). The model starts with rest ocean and atmosphere, and with the horizontal homogeneity. If there is a perturbation initially added to the air-ocean coupled system, will this perturbation be growing or decaying? be oscillatory or nonoscillatory?

The subsequent sections are tried to answer those questions, to prove the importance of the air-ocean coupling, and to predict the growth-rate and phase speed in the wavenumber spectrum. The length-scale corresponding to the maximum growth-rate and minimum phase speed of the coupled system is taken as the general appearing length scale in the ocean between the two adjacent cold (or warm) water plumes.

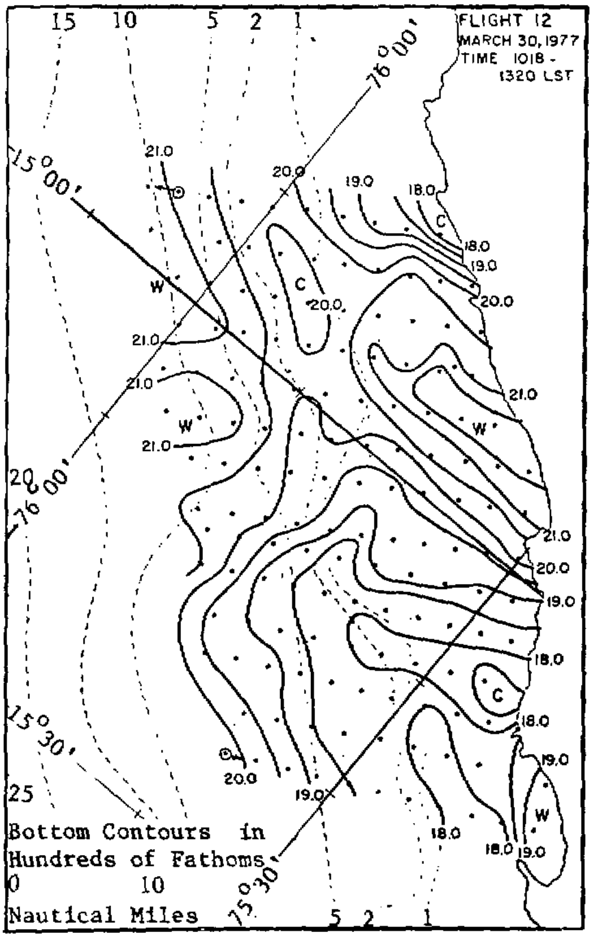

Fig. 1. Sea surface temperature near Peruvian Coast (after Stuart, 1981).

II. AIR-OCEAN COUPLED MODEL

In this study the air-sea coupling we consider is shown in Fig. 2. The atmospheric circulation, thermally generated by the SST gradient, drives the water current, which in turn changes the SST distribution. The salinity effect is neglected here. Consider the ocean of uniform depth H, with a straight coast line. The right-handed coordinate system is chosen that the x-axis is in the alongshore direction, and y-axis is in the cross-coastal direction. Furthermore, the y-axis is nearly coincident with 15°S latitude (Fig. 3). Therefore, the water thermal expansion equation is written by

$$\rho^{(w)}=\rho_0^{(w)}[1-\alpha(T_*^{(w)}-T_0^{(w)})], \tag{1}$$

where α is the water thermal expansion coefficient, $\rho_0^{(w)}$, $T_0^{(w)}$ are the basic states of water density and water temperature, $T^{(w)}=T_*^{(w)}-T_0^{(w)}$ is the temperature perturbation.

A state of rest with hydrostatic equilibrium and horizontally-uniform stratification is taken as a basic state, i.e.,

$$dp_0^{(w)}(z)/dz = -g\rho_0^{(w)}(z), \tag{2}$$

consider the theory of upwelling for the case where only small perturbations from this basic state are produced (Gill and Clarke, 1974). Let the perturbation pressure be $p^{(w)}$, the density perturbation be $\sigma^{(w)}$, and the velocity components be $(u^{(w)}, v^{(w)}, w^{(w)})$. With the Boussinesq approximation, these quantities satisfy the following equations:

$$u_t^{(w)} - fv^{(w)} = -p_x^{(w)}/\bar{\rho}_0^{(w)} + X_z/\bar{\rho}_0^{(w)} \tag{3}$$

$$fu^{(w)} = -p_y^{(w)}/\bar{\rho}_0^{(w)} + Y_z/\bar{\rho}_0^{(w)} \tag{4}$$

$$0 = -p_z^{(w)} - g\sigma^{(w)} \tag{5}$$

$$u_x^{(w)} + v_y^{(w)} + w_z^{(w)} = 0 \tag{6}$$

$$\sigma_t^{(w)} + w^{(w)}\rho_{0z}^{(w)} = 0, \tag{7}$$

where the subscripts (t, x, y, z) mean the partial derivatives with respect to those independent variables, and t is time, f the Coriolis parameter and g the gravitational acceleration. $\bar{\rho}_0^{(w)}$ is a constant which is a representative water density. The vector (X, Y) is tangential stress acting between horizontal planes. According to Gill and Clarke (1974) we have

$$(X_z, Y_z) = \begin{cases} (\tau_x^{(a)}, \tau_y^{(a)})/H_{mix}, & -H_{mix} < z < 0 \\ 0, & z < -H_{mix}, \end{cases} \tag{8}$$

where H_{mix} is the depth of ocean upper mixed layer. Water thermal expansion equation (1) and Boussinesq approximation give a relationship between water density perturbation $\sigma^{(w)}$ and water temperature perturbation $T^{(w)}$:

$$\sigma^{(w)} = -\alpha\bar{\rho}_0^{(w)}T^{(w)} \tag{9}$$

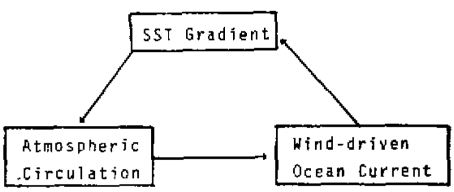

Fig. 2. Air-sea coupled system.

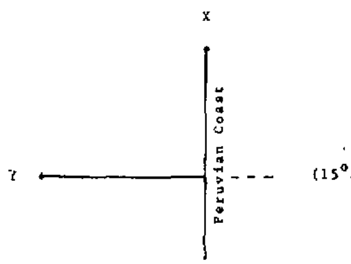

Fig. 3. The coordinate system.

Author 's earlier work (1985, 1986ab, 1987ab) shows that thermally forced surface winds blow across isotherms from cold to warm water with some deflection to the left (right) in the Southern (Northern) Hemisphere due to the earth rotation. The deflection angle δ (Fig. 4), which is positive (negative) in the Southern (Northern) Hemisphere, depends on the atmospheric stratification (Chu, 1986b, 1987a). $|\delta|$ is generally from 20° to 30°. The more steady the atmosphere is, the less deflection angle δ should be. The surface wind speed is proportional to the sea surface temperature gradient. Therefore, the thermally forced surface wind is computed by

$$V^{(a)} = K\begin{bmatrix} \cos\delta & -\sin\delta \\ \sin\delta & \cos\delta \end{bmatrix} \begin{bmatrix} \partial T^{(w)}/\partial x \\ \partial T^{(w)}/\partial y \end{bmatrix}_{z=0}, \tag{10}$$

where K is a proportionality constant which represents surface wind speed thermally driven by a unity surface temperature gradient. According to Defant (1950), a 1°C/100 km temperature gradient can generate a surface wind of about 2 m/s. Hence we take (Chu, 1987b)

$$K=(2\ \mathrm{m/s})/(1^{\circ}\mathrm{C}/100\ \mathrm{km}). \tag{11}$$

Using following bulk formula to calculate the surface wind stress

$$\tau^{(a)}=(\tau_x^{(a)},\ \tau_y^{(a)})=\rho_a C_D \mathbf{V}^{(a)}, \tag{12}$$

where ρ_a is air density at surface, and C_D the dimensional (m/s) air drag coefficient. Substituting (10) into (12) we have

$$\begin{cases}\tau_x^{(a)}=\rho_a C_D K(\cos\delta\partial T^{(w)}/\partial x-\sin\delta\partial T^{(w)}/\partial y)_{z=0}\\ \tau_y^{(a)}=\rho_a C_D K(\sin\delta\partial T^{(w)}/\partial x+\cos\delta\partial T^{(w)}/\partial y)_{z=0}.\end{cases} \tag{13}$$

Gill and Clarke (1974) shows that there exist eigenfunctions $\hat{p}_n(z)$, $\hat{w}_n(z)$ and eigenvalues c_n such that $u^{(w)}$, $v^{(w)}$, $w^{(w)}$, $p^{(w)}$, and $T^{(w)}$ can be expanded as follows:

$$u^{(w)}=\sum u_n^{(w)}(x,y,t)\hat{p}_n(z) \tag{14a}$$

$$v^{(w)}=\sum v_n^{(w)}(x,y,t)\hat{p}_n(z) \tag{14b}$$

$$p^{(w)}/\rho_0^{(w)}=\sum p_n^{(w)}(x,y,t)\hat{p}_n(z) \tag{14c}$$

$$w^{(w)}=\sum w_n^{(w)}(x,y,t)\hat{w}_n(z) \tag{14d}$$

$$\alpha g T^{(w)}/N^2(z)=\sum T_n^{(w)}(x,y,t)\hat{w}_n(z), \tag{14e}$$

where N is the Brunt-Vaisala frequency defined by

$$N^2(z)=-(g/\rho_0)d\rho_0(z)/dz. \tag{15}$$

The eigenfunctions and eigenvalue for barotropic mode are

$$\hat{p}_0=1,\quad \hat{w}_0=z+H,\quad c_0=\sqrt{gH}, \tag{16}$$

and the eigenfunctions for baroclinic modes are

$$\begin{aligned}&d^2\hat{w}_n/dz^2+(N/c_n)^2\hat{w}_n^2=0,\\ &\hat{p}_n=d\hat{w}_n/dz,\quad d\hat{p}_n/dz=-(N/c_n)^2\hat{w}_n,\quad n=1,2,3,\cdots\end{aligned} \tag{17a}$$

and satisfy the boundary conditions that (Gill and Clarke, 1974)

$$\hat{w}_n(0)=0,\quad \hat{w}_n(-H)=0. \tag{17b}$$

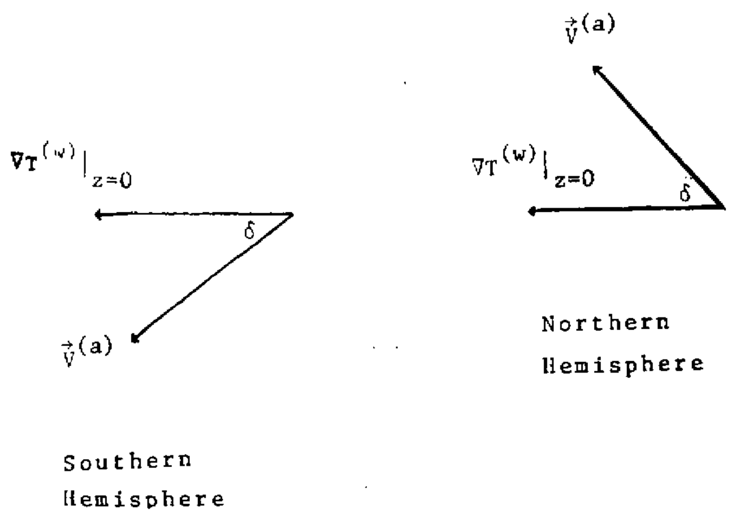

Fig. 4. Thermally forced wind and surface temperature gradient.

The forcing terms of the ocean, X_z and Y_z, are expanded in terms of the eigenfunctions $\hat{p}_n$, e.g.,

$$\begin{bmatrix} X_z \\ Y_z \end{bmatrix} = \sum (b_n / H_{mix}) \begin{bmatrix} \tau_x^{(a)} \\ \tau_y^{(a)} \end{bmatrix} \hat{p}_n(z), \tag{18}$$

where b_n ($n=0, 1, 2, \ldots$) are the coefficients of the expansion:

$$A(z) = \begin{cases} 1, & -H_{mix} < z < 0 \\ 0, & z < -H_{mix} \end{cases} = \sum b_n \hat{p}_n(z). \tag{19}$$

Integrating (19) with respect to z from $-H$ to 0 we find

$$b_0 = H_{mix}/H. \tag{20}$$

From the temperature expansion (14e) we have

$$T^{(w)}(x,y,0,t) = [N^2(0)/ag] \sum T_n^{(w)}(x,y,t) \hat{w}_n(0) \tag{21}$$

and from the top boundary conditions for eigenfunctions $\hat{w}_n(z)$, (16) and (17b), we obtain

$$T^{(w)}(x,y,0,t) = [N^2(0)H/ag] T_0^{(w)}(x,y,t). \tag{22}$$

Substituting water density perturbation (9) into hydrostatic equation (5) and using the expansions (14ce) and employing the relationship between eigenfunctions $\hat{p}_n$ and $\hat{w}_n$, (17a), we have

$$T_n^{(w)} = -p_n^{(w)}/c_n^2. \tag{23}$$

The surface water temperature perturbation is then written by

$$T^{(w)}(x,y,0,t) = -[N^2(0)/ag^2] p_0^{(w)}(x,y,t). \tag{24}$$

Substituting (24) into (13) the wind stress is

$$\begin{cases} \tau_x^{(a)} = -[\rho_a C_D K N^2(0)/ag^2](\cos\delta\ \partial p_0^{(w)}/\partial x - \sin\delta\ \partial p_0^{(w)}/\partial y) \\ \tau_y^{(a)} = -[\rho_a C_D K N^2(0)/ag^2](\sin\delta\ \partial p_0^{(w)}/\partial x + \cos\delta\ \partial p_0^{(w)}/\partial y). \end{cases} \tag{25}$$

Therefore the basic equations for the n-th mode are

$$\partial u_n^{(w)}/\partial t - f v_n^{(w)} = -\partial p_n^{(w)}/\partial x - \gamma_n(\cos\delta\ \partial p_0^{(w)}/\partial x - \sin\delta\ \partial p_0^{(w)}/\partial y) \tag{26}$$

$$f u_n^{(w)} = -\partial p_n^{(w)}/\partial y - \gamma_n(\sin\delta\ \partial p_0^{(w)}/\partial x + \cos\delta\ \partial p_0^{(w)}/\partial y) \tag{27}$$

$$\partial p_n^{(w)}/\partial t + c_n^2(\partial u_n^{(w)}/\partial x + \partial v_n^{(w)}/\partial y) = 0 \tag{28}$$

where $n=0, 1, 2, \ldots$, and

$$\gamma_n = b_n \gamma, \quad \gamma \equiv \rho_0^{(a)} C_D K N^2(0)/(\bar{\rho}_0^{(w)} H_{mix} a g^2).$$

Here the nondimensional parameter γ is a coupling factor. If there is no air-sea coupling, $\gamma=0$, the system is reduced to the Gill-Clarke model.

III. INSTABILITY CRITERION

Eqs. (26)–(28) show that the barotropic mode ($n=0$) is a closed system, and the baroclinic modes ($n=1,2,3,\ldots$) are forced by the barotropic mode. If under some circumstances unstable barotropic mode is generated, then all the baroclinic modes are correspondingly unstable. In this section we discuss the instability criterion.

Setting $n=0$ in (26)–(28), substituting the differentiation of (25) with respect to y from the differentiation of (26) with respect to x, and using the continuity equation (28), the result is

$$\partial/\partial t[(1/f)\partial u_0^{(w)}/\partial y + p_0^{(w)}/c_0^2] = (\gamma_0 \sin\delta/f)\nabla^2 p_0^{(w)}. \tag{29}$$

Eliminating $u_0^{(w)}$ from (27) and (29) then gives

$$\partial/\partial t[1/f^2\ \partial^2 p_0^{(w)}/\partial y^2 - p_0^{(w)}/c_0^2 + \gamma_0/f^2(\sin\delta\ \partial^2 p_0^{(w)}/\partial x\partial y + \cos\delta\ \partial^2 p_0^{(w)}/\partial y^2)] = -(\gamma_0 \sin\delta/f)\nabla^2 p_0^{(w)}. \tag{30}$$

Let the solution have the following form

$$p_0^{(w)}(x,y,t)=\varphi^{(w)}(y)\exp(\sigma t+ikx). \tag{31}$$

Substituting (31) into (30) we have a second order ordinary differential equation for $\varphi^{(w)}$

$$d^2\varphi^{(w)}/dy^2+[ik\gamma_0\sin\delta/(1+\gamma_0\cos\delta)]d\varphi^{(w)}/dy$$
$$-[f^2/(1+\gamma_0\cos\delta)](\gamma_0k^2\sin\delta/f\sigma+1/c_0^2)\varphi^{(w)}=0. \tag{32}$$

Far away from coast perturbations should vanish, i.e.,

$$\varphi^{(w)}\to 0 \text{ as } y\to\infty. \tag{33}$$

Right at the coast $v_x^{(w)}$ should also vanish, i. e.,

$$fv_0^{(w)}=\partial u_0^{(w)}/\partial t+\partial p_0^{(w)}/\partial x+\gamma_0(\cos\delta\ \partial p_0^{(w)}/\partial x-\sin\delta\ \partial p_0^{(w)}/\partial y)=0,$$
$$\text{at } y=0. \tag{34}$$

Eliminating $u_0^{(w)}$ from (27) and (34), and using (31) we have the boundary condition for $\varphi^{(w)}$

$$d\varphi^{(w)}/dy=ik\varphi^{(w)}(1+\gamma_0\cos\delta-\sigma\gamma_0\sin\delta/f)/[\sigma/f+\gamma_0(\sin\delta+\cos\delta)],$$
$$\text{at } y=0. \tag{35}$$

The solution of (32) satisfying the boundary condition (33) is written by

$$\varphi^{(w)}(y)=c\exp(\beta y), \tag{36}$$

where β is a root with negative real part of the following quadratic equation

$$\beta^2+[ik\gamma_0\sin\delta/(1+\gamma_0\cos\delta)]\beta-f^2(\gamma_0k^2\sin\delta/f\sigma+1/c^2)/(1+\gamma_0\cos\delta)=0. \tag{37}$$

Substituting (36) into the boundary condition (35) we obtain

$$\beta=ik(1+\gamma_0\cos\delta-\sigma\gamma_0\sin\delta/f)/[\sigma/f+\gamma_0(\sin\delta+\cos\delta)]. \tag{38}$$

Eliminating β from (37) and (38) reaches the dispersion relation

$$[A_1^2+A_5+\mu A_6](\sigma/f)^3+[A_4+2A_3(A_5+\mu A_6)+2A_1A_2](\sigma/f)^2$$
$$+[A_2^2+2A_3A_4+A_3^2(A_5+\mu A_6)](\sigma/f)+A_4A_3^2=0, \tag{39}$$

where

$$\begin{aligned}
\mu &\equiv f^2/c_0^2k^2\\
A_1 &\equiv 0.5\gamma_0\sin\delta/(1+\gamma_0\cos\delta)-\gamma_0\\
A_2 &\equiv 1+\gamma_0\cos\delta+0.5\gamma_0^2\sin\delta(\sin\delta+\cos\delta)/(1+\gamma_0\cos\delta)\\
A_3 &\equiv \gamma_0(\sin\delta+\cos\delta)\\
A_4 &\equiv A_3\sin\delta/(1+\gamma_0\cos\delta)^2\\
A_5 &\equiv -0.25\gamma_0\sin^2\delta/(1+\gamma_0\cos\delta)^2\\
A_6 &\equiv 1/(1+\gamma_0\cos\delta).
\end{aligned} \tag{40}$$

The standard values of parameters representing the Peruvian coastal region near 15°S are given in Table 1. We solve the characteristic equation (39) with different values of k and δ. k varies from 10^{-3} m^{-1} to 10^{-6} m^{-1}, and δ from 20° to 30°. The instability criterion for coastal upwelling in the air-sea coupled system is written as

$$\mathrm{Re}\,(\sigma)\begin{cases}<0 & \text{decaying}\\ =0 & \text{neutral}\\ >0 & \text{growing}\end{cases}\qquad \sigma=\sigma_1,\sigma_2,\sigma_3, \tag{41}$$

where σ is the root of cubic equation (39). The oscillation criterion for coastal upwelling is given by

$$\mathrm{Im}\,(\sigma)\begin{cases}=0, & \text{non-oscillatory}\\ \neq 0, & \text{oscillatory}\end{cases}\qquad \sigma=\sigma_1,\sigma_2,\sigma_3. \tag{42}$$

The cubic equation (39) has three sets of roots: first one contains only negative real roots for all k, which represents damping non-oscillatory modes. The other two sets contain complex conjugates. In this paper we are interested in growing modes, so that we pay less attention to the first set of roots. Furthermore, without the air-sea coupling

($\gamma=0$), the solutions of basic system (3)–(7) should be coastal trapped Kelvin wave propagating in the alongshore direction with the coast on the left (right) in the Southern (Northern) Hemisphere (Gill and Clarke, 1974). The solutions of the coupled air-sea system should reduce to the coastal trapped Kelvin wave, therefore, we should choose σ with

$$\text{Im }(\sigma)>0 \tag{43}$$

as our solutions.

Table 1. The Standard Values of Model Parameters

$f=-0.366\times10^{-4}\ s^{-1}$,	$g=9.81\ m\ s^{-2}$,	$C_D=1.3\times10^{-2}\ ms^{-1}$,
$N^2(0)=3.27\times10^{-1}\ s^{-2}$,	$H=182.88$ m (100 Fathoms),	
$\rho_a=1.29\ kg\ m^{-3}$	$\rho_w=10^3\ kg\ m^{-3}$,	
$\alpha=2\times10^{-4}\ K^{-1}$,	$k=2\times10^3\ m^2s^{-1}K^{-2}$	

Figs. 5 and 6 show $\text{Re}(\sigma)$ and $\text{Im}(\sigma)$ versus the negative logarithmic wave number with different diffraction angle δ (20°, 25°, and 30°), respectively. They indicate following results:

(a) Thermally forced local winds produce instability in the coastal upwelling system. The growth rate has a single peak in the wave-number spectrum. For $\delta=20°$ (very stable atmospheric stratification), the peak is located nearly at $k=10^{-3.875}\ m^{-1}$, the corresponding wave-length is $L=2\pi/k=47$ km. For $\delta=25°$ (stable atmospheric stratification), the peak is located at $k=10^{-4.078}\ m^{-1}$, the corresponding wavelength is $L=2\pi/k=75$ km. For $\delta=30°$ (less stable atmospheric stratification), the peak is located at $k=10^{-4.156}\ m^{-1}$, the corresponding wave-length is $L=2\pi/k=90$ km. It shows that the wave-length of the most unstable mode enlarges with the decrease of the atmospheric stratification.

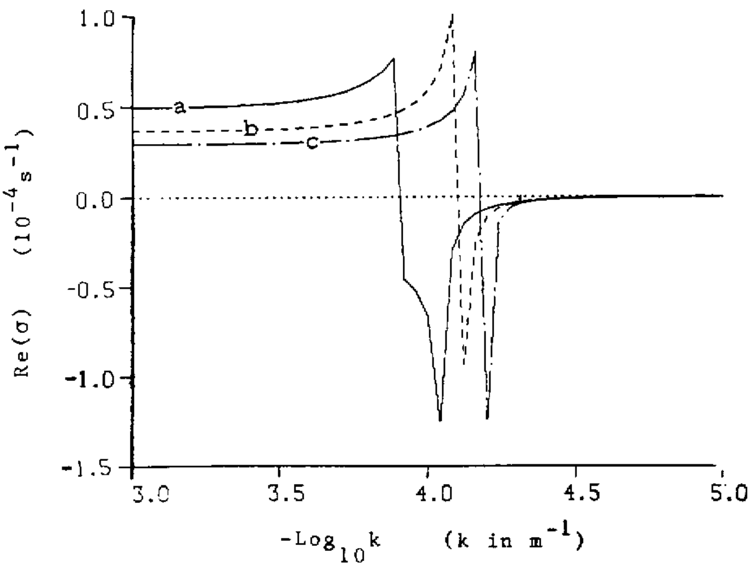

Fig. 5. Growth rate Re (σ) versus wavenumber K for three different deflection angles; (a) $\delta=20°$, (b) $\delta=25°$, and (3) $\delta=30°$.

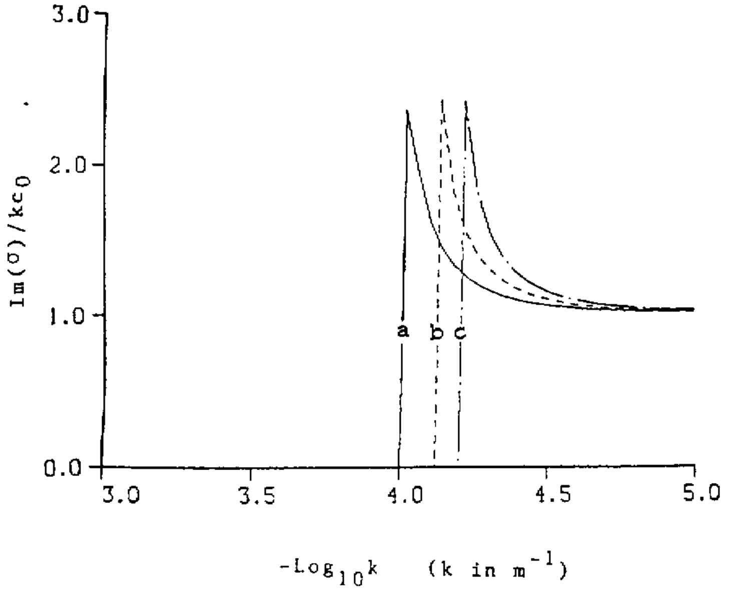

Fig. 6. Phase speed versus wavenumber K for three different deflection angles: (a) $\delta=20°$, (b) $\delta=25°$, (c) $\delta=30°$.

(b) Fig. 5 indicates that according to the growth rate distribution, the air-sea coupled system is separated into three parts: small-scale ($L<100$ km) growing modes, meso-scale (100 km$<L<$200 km) decaying modes, and large-scale ($L>200$ km) neutral modes. However, Fig. 6 shows that according to the wave-speed distribution, the coupled system is also separated into three parts: small scale ($L<100$ km) stationary modes, mesoscale (100 km$<L<$200 km) fast-moving modes (phase velocity is greater than c_0), and large-scale ($L>200$ km) coastal trapped Kelvin wave (phase velocity is c_0). Therefore, the system has three parts: small-scale stationary and growing modes, meso-scale faster-moving and decaying modes, and large-scale ($L>200$ km) coastal trapped Kelvin modes. It means that the coupling effects only have an influence on the small and meso-scale modes, not on the largescale modes.

(c) Comparing Figs. 5 and 6 with Fig. 1, which shows the observed stationary alternation pattern of plumes of cold and warm surface seawater along the Peruvian coast, we find that length-scale corresponding to the maximum growth-rate (which is stationary) fits the length-scale between two adjacent cold (or warm) plumes of the surface seawater. It implies that thermally forced local winds may play some role in coastal upwelling.

IV. RESULTS

This coupled air-sea model shows that the thermally forced local winds break down the coastal trapped Kelvin wave into three parts: small-scale growing and stationary modes, mesoscale decaying and fast-moving modes, and largescale ($L>200$ km) coastal trapped Kelvin modes. The alternation pattern of cold and warm plumes of surface seawater observed by Moody et al. (1981) and Stuart (1981) can be predicted by this model as a small-scale growing and stationary mode. The wave-length corresponding to the maximum growth-rate also fits the length-scale of the two adjacent cold (or warm) plumes very well.

In this model we neglect the nonlinearity and the salinity effect on the water density. We should include these effects in the future research.

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REFERENCES

Chu, P.C., (1986a), An instability theory of ice-air interaction for the migration of the marginal ice zone. *Geophys. J. Roy Astron. Soc.*, **86**: 863–883.

Chu, P.C. (1986b), An air-ice-ocean coupled model for the formation of leads or polynyas. *MIZEX Bull., VII*, 79–88.

Chu, P.C., (1987a), An instability theory of ice-air interaction for the formation of ice edge bands. *J. Geophys. Res.*, **92**: *C7*, 6966–6970.

Chu, P.C., (1987b), Generation of unstable modes of the iceward attenuating swell by icebreeze. *J. Phys. Oceanogr.*, **17**: 828–832.

Chu, P.C., (1987c), An icebreeze mechanism for an ice divergence-convergence criterion in the marginal ice zone. *J. Phys. Oceanogr.*, **17**, 1627—1632

Defant, F., (1950). Theorie der Land- und Seawinde. *Arch. Meteor. Geophys. Biokl.*, **2(A)**: 404–425.

Gill, A.E. and A.J. Clarke, (1974), Wind-induced upwelling, coastal currents, and sea-level changes. *Deep-Sea Res.*, **21**: 325–345.

Kuo, H.L., (1973), Planetary boundary layer flow of a stable atmosphere over the globe. *J. Atmos. Sci.*, **30**: 53–65.

Moody G.L., D.W. Stuart, A.I. Watson, and M.M. Nanney, (1981), Sea-surface temperatures and winds during JOINT II: part I mean conditions. In: *Coastal Upwelling*, (Richard, F.A. Ed.), Amer. Geophys. Union, Washington, D.C., 21–31.

Stuart, D.W. (1981), Sea-surface temperatures and winds during JOINT II: part II, temporal fluctuations. In: *Coastal Upwelling*, (Richard, F.A. Ed.), Amer. Geophys. Union, Washington, D.C., 32–38.