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ONR ltr, 26 Oct 1977

A CONTRIBUTION TO THE THEORY OF
UPWELLING

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Research Conducted for the

Texas A. & M. Research Foundation

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Research conducted for the
Texas A. & M. Research Foundation

Project 24

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ABSTRACT

A theory of upwelling is presented taking into account the effects of the earth's rotation and vertical and horizontal mixing. It can be shown by this theory that upwelling is possible very close to a coast along which, in the Northern hemisphere, a wind blows in such a manner that the coast is on the left hand side of an observer who looks in the direction of the wind. A numerical computation indicates that the speed of upwelling is about 80 meters per month for a steady wind blowing at a speed of 5 to 6 meters per second. In addition to this vertical motion, there are horizontal currents parallel and perpendicular to the coast, the speeds of which are roughly 1000 times that of the speed of upwelling.

1. **Introduction.** There have been several descriptions of the upwelling noticed off California, Peru, West Africa and other coasts by Thorade (1909), McEwen (1912), Gunther (1936), Defant (1936), Sverdrup (1938) and Sverdrup and Fleming (1941). The explanation of this process given by Sverdrup in 1938 is worth attention. According to Sverdrup and Fleming, it is known from the analysis of the water masses that the water taking part in the process of upwelling off the coast of Southern California originates mostly in the layers from 200 to 300 meters below the surface. These are, however, all qualitative discussions and it has been as yet not possible to explain this phenomenon theoretically and predict the velocity and width of the coastal currents produced by the prevailing winds. Very recently Defant (1952) made a theoretical explanation assuming a sea consisting of two layers of water with different density. The present research is an attempt to solve this problem mathematically and to draw some quantitative conclusions concerning this process which is very important in all fields of oceanography.

The explanation of the upwelling seems to be satisfactory only by treating this problem thermodynamically as well as hydrodynamically. The following discussion will, however, be made only from a purely hydrodynamical standpoint, on the assumption that the seawater is of uniform density. This is a necessary consequence for simplifying the mathematics, but the author believes to have been able to obtain some results which are consistent with certain facts observed in this process.

It has been noticed that the upwelling takes place most favorably when, in the northern (southern) hemisphere, a wind blows in such a manner that the coast is on the left-hand (right-hand) side of an observer who looks in the direction of the wind. Thus the upwelling off the coast of California is most remarkable in early summer when northwesterly winds prevail for several weeks nearly parallel to the coast.

In this investigation the effect of the Earth's rotation and the frictional forces due to both vertical and horizontal mixing are taken into consideration. And it may be stressed that the horizontal mixing seems to play the most important role in the theoretical explanation of the process of upwelling.

2. **Theory.** Consider an infinitely long straight coast coincident with the axis of y , and take x -axis perpendicular to it in the offshore direction. (Fig. 1). Suppose a wind of constant force and direction is blowing steadily and uniformly in a belt of limited width L, parallel to the coast from negative to positive direction of y . This is a disposition favorable for the upwelling to actually occur. Take the z -axis vertically downwards.

In a steady state which is attained after a sufficiently long time since the wind began to blow constantly, the motion of water will be independent of y . This means that all the vertical and horizontal components of the currents can be determined as functions of x and z only. Moreover, the surface of the sea will not be a plane, but have a slope in

the x -direction. In such a case, the hydrodynamical equations of motion of sea water are, after several reasonable simplifications,

$$\frac{A_h}{\rho} \frac{\partial^2 u}{\partial x^2} + \frac{A_v}{\rho} \frac{\partial^2 u}{\partial z^2} + 2\omega \sin\phi\, v - g\frac{\partial \zeta}{\partial x} = 0$$

(1)

$$\frac{A_h}{\rho} \frac{\partial^2 v}{\partial x^2} + \frac{A_v}{\rho} \frac{\partial^2 v}{\partial z^2} - 2\omega \sin\phi\, u = 0$$

where u and v are the horizontal components of the current velocity in the x and y directions, ζ the surface elevation depending on x only, ρ the density, A_v and A_h the coefficients of vertical and horizontal mixing, of sea water, ω the angular velocity of the Earth and ϕ the geographic latitude. In addition to these, we have the equation of continuity in the form:

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0$$

(2)

where w is the vertical component of currents and represents the intensity of the upwelling, and because $\dfrac{\partial v}{\partial y} = 0$

Suppose the wind blows in the positive direction of y in a belt between the coast $x = 0$ and $x = L$. The wind velocity may vary in the offshore or x -direction. The conditions to be satisfied on the surface of the sea are therefore at

$$z = 0 \quad : \quad \begin{cases} -A_v \dfrac{\partial u}{\partial z} = 0 \\[2mm] -A_v \dfrac{\partial v}{\partial z} = \tau \quad \text{for } 0 \leqq x \leqq L \\[2mm] \qquad\quad = 0 \quad \text{for } L < x < \infty \end{cases}$$

(3)

where τ may be either a constant or a function of x. On the bottom we must have, no motion because of the vertical friction at

$$\bar{z} = h : \quad u = v = 0. \tag{4}$$

and along the coast which is considered to consist of vertical cliffs, we have because of the horizontal friction at

$$x = 0 \quad u = v = 0. \tag{5}$$

In the region very far from both the coast and wind region, we have at

$$x = \infty : \quad u = v = \frac{\partial \zeta}{\partial x} = 0. \tag{6}$$

Let us define D_v and D_h by

$$D_v = \pi \sqrt{\frac{A_v}{\rho \omega \sin \phi}} \quad ; \quad D_h = \pi \sqrt{\frac{A_h}{\rho \omega \sin \phi}}. \tag{7}$$

D_v is the depth of frictional influence defined by Ekman (1905) in his theory of ocean currents, and D_h is a quantity having a dimension of a length and may be called "frictional distance". This is a measure of the horizontal turbulence. Then the equations (1) now become

$$\frac{\partial^2 u}{\partial \xi^2} + D_v^2 \frac{\partial^2 u}{\partial z^2} + 2\pi^2 v - \frac{\pi^2 g}{\omega \sin \phi} \frac{\partial \zeta}{\partial \xi} = 0, \tag{8}$$

$$\frac{\partial^2 v}{\partial \xi^2} + D_v^2 \frac{\partial^2 v}{\partial z^2} - 2\pi^2 u = 0.$$

where

$$\xi = \frac{x}{D_h}. \tag{9}$$

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In order to solve the equations (5), suppose

$$u = \frac{2}{\pi} \int_0^\infty u_1(\lambda) \sin \lambda \xi \, d\lambda ,$$

$$u_1(\lambda) = \int_0^\infty u(\alpha, z) \sin \lambda \alpha \, d\alpha \qquad (10)$$

$$v = \frac{2}{\pi} \int_0^\infty v_1(\lambda) \sin \lambda \xi \, d\lambda ,$$

$$v_1(\lambda) = \int_0^\infty v(\alpha, z) \sin \lambda \alpha \, d\alpha \qquad (11)$$

$$\frac{\partial \zeta}{\partial \xi} = \frac{2}{\pi} \int_0^\infty \gamma(\lambda) \sin \lambda \xi \, d\lambda ,$$

$$\gamma(\lambda) = \int_0^\infty \frac{\partial \zeta}{\partial \alpha} \sin \lambda \alpha \, d\alpha . \qquad (12)$$

Next assume for the wind stress

$$-A_v \frac{\partial v}{\partial z} = \frac{2}{\pi} \int_0^\infty T(\lambda) \sin \lambda \xi \, d\lambda \qquad (13)$$

and

$$T(\lambda) = \int_0^\infty \left(-A_v \frac{\partial v}{\partial z} \right) \sin \lambda \alpha \, d\alpha$$

$$= \int_0^{L/D_h} \tau \sin \lambda \alpha \, d\alpha$$

$$= \frac{1 - \cos\left(\lambda \frac{L}{D_h} \right)}{\lambda} \tau \qquad (14)$$

if τ is independent of x .

Substituting (10), (11) and (12) into (5) and writing

$$u_1 + i v_1 = \overline{W} \qquad (15)$$

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the two equations (5) are combined into

$$D_h^2 \frac{d^2\overline{W}}{dz^2} - (\lambda^2 + 2\pi^2 i)\overline{W} - \frac{\pi^2 \mathcal{Y}}{\omega \sin\phi \, D_h} \, \gamma(\lambda) = 0 \qquad (16)$$

and the conditions to be satisfied along the boundaries now become

$$z = 0: \quad -A_v \frac{d\overline{W}}{dz} = \frac{1 - \cos(\lambda \, {}^L\!/D_h)}{\lambda} \cdot i\tau \qquad (17)$$

and

$$z = h: \quad \overline{W} = 0 \qquad (18)$$

The solution of the equation (16) subject to the conditions (17) and (18) is

$$\overline{W} = \frac{\pi^2 \mathcal{Y} \, \gamma(\lambda)}{(\lambda^2 + 2\pi^2 i)\sin\phi \, D_h}\left\{ 1 - \frac{\cosh(\sqrt{\lambda^2 + 2\pi^2 i} \; {}^z\!/D_v)}{\cosh(\sqrt{\lambda^2 + 2\pi^2 i} \; {}^h\!/D_v)} \right\}$$

$$+ \frac{i\tau \, D_v/A_v}{\sqrt{\lambda^2 + 2\pi^2 i}} \cdot \frac{1 - \cos(\lambda \, {}^L\!/D_h)}{\lambda} \cdot \frac{\sinh(\sqrt{\lambda^2 + 2\pi^2 i} \; \frac{h-z}{D_v})}{\cosh(\sqrt{\lambda^2 + 2\pi^2 i} \; {}^h\!/D_v)} \qquad (19)$$

If we separate the real part P of $\sqrt{\lambda^2 + 2\pi^2 i}$ from the imaginary part Q, we have

$$P = \sqrt{\frac{\sqrt{\lambda^4 + 4\pi^4} + \lambda^2}{2}} \qquad ; \qquad Q = \sqrt{\frac{\sqrt{\lambda^4 + 4\pi^4} - \lambda^2}{2}} \qquad (20)$$

Thus the real part of $\sqrt{\lambda^2 + 2\pi^2 i}$ is always greater than π. Hence, if the depth of the sea is sufficiently large ($h/D_v > 2$), the expression (19) can be given very accurately by

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$$\overline{W} = \frac{\pi^2 g \, \gamma(\lambda)}{(\lambda^2 + 2\pi^2 i)\sin\phi \, D_h}\left(1 - e^{-\sqrt{\lambda^2 + 2\pi^2 i}\,\frac{h-z}{D_v}}\right)$$

$$+ \frac{i\tau D_v}{A_v} \cdot \frac{1}{\sqrt{\lambda^2 + 2\pi^2 i}} \cdot \frac{1 - \cos(\lambda L/D_v)}{\lambda} \cdot e^{-\sqrt{\lambda^2 + 2\pi^2 i}\,z/D_v}.$$

(21)

Now we have, for the vertical component w ,

$$w = 0 \quad \text{at } z = 0 \text{ ; } z = h \tag{22}$$

since there can be no vertical motion on the surface and bottom of the sea. Integrating (2) with respect to z from the surface down to the bottom, we have

$$\frac{\partial}{\partial x}\int_0^h u\,dz = w\Big|_{z=0}^{z=h} = 0$$

This means that the integral $\int_0^h u\,dz$ is independent of x , or therefore a constant. But as this integral must vanish directly on the coast or at $x = 0$, we must have

$$\int_0^h u\,dz = 0 \tag{23}$$

always. Integrating (21) with respect to z from 0 to h and equating the real part of the resulting equation to zero, we have

$$\gamma(\lambda) = \frac{\dfrac{\pi^2 \tau/\rho}{\dfrac{\lambda^2}{2\pi^2}\left(h - \dfrac{P}{P^2 + Q^2}D_v\right) + \dfrac{Q}{P^2 + Q^2}D_v}} \cdot \frac{1 - \cos\lambda L/D_h}{\lambda} \tag{24}$$

This determines the relation between the wind stress τ and slope of the sea surface induced by the former. Substitution of (24) in (21) gives

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u_1 , and v_1 . The substitutions of u_1 , v_1 and $\gamma(\lambda)$ thus obtained into (10), (11) and (12) give u , v and the surface slope $\frac{\partial \zeta}{\partial \eta}$. The vertical velocity w can be derived from the equation of continuity (2) as

$$w = - \frac{\partial}{\partial x} \int_0^z u \, dz \tag{25}$$

3. **Upwelling in a Deep Sea.** When the sea is sufficiently deep and the ratio h/D_v increased indefinitely, we have from (24)

$$\gamma(\lambda) = 0 \tag{26}$$

and (21) becomes

$$\overline{W} = \frac{i\tau \, D_v/A_v}{\sqrt{\lambda^2 + 2\pi^2 i}} \cdot \frac{1 - \cos(\lambda^L/D_h)}{\lambda} \cdot e^{-\sqrt{\lambda^2 + 2\pi^2 i} \, z/D_v} \tag{27}$$

Then we have

$$u(x,z) = \frac{2\pi \tau}{\rho \omega \sin \phi \, D_v} \int_0^\infty M(\lambda z) R(\lambda,x) \, d\lambda \tag{28}$$

$$v(x,z) = \frac{2\pi \tau}{\rho \omega \sin \phi \, D_v} \int_0^\infty N(\lambda,z) R(\lambda,x) \, d\lambda \tag{29}$$

$$w(x,z) = \frac{2\pi \tau}{\rho \omega \sin \phi \, D_v} \left(\frac{D_v}{D_h} \right) \int_0^\infty L(\lambda,z) S(\lambda x) \, d\lambda \tag{30}$$

where

$$M(\lambda,z) = \frac{Q \cos(Q^z/D_v) + P \sin(Q^z/D_v)}{P^2 + Q^2} \, e^{-P^z/D_v} \tag{31}$$

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$$N(\lambda,z) = \frac{P\cos(Q\,z/D_v) - Q\sin(Q\,z/D_v)}{P^2 + Q^2}\, e^{-P\,z/D_v} \tag{32}$$

$$L(\lambda,z) = \frac{2PQ\left\{\cos(Q\,z/D_v)\,e^{-P\,z/D_v} - 1\right\} + (P^2 - Q^2)\sin(Q\,z/D_v)\,e^{-P\,z/D_v}}{(P^2 + Q^2)^2} \tag{33}$$

$$R(\lambda,x) = \sin(\lambda\,x/D_h)\cdot\frac{1-\cos(\lambda\,z/D_h)}{\lambda} \tag{34}$$

$$S(\lambda,x) = \cos(\lambda\,x/D_h)\cdot\left\{1-\cos(\lambda\,l/D_h)\right\} \tag{35}$$

and P and Q are the real and imaginary part of $\sqrt{\lambda^2 + 2\pi^2\lambda}$ whose expressions are given by (20).

These results show that, by the effect of the wind blowing parallel to the coast, we can expect a vertical circulation in the plane perpendicular to the coast in addition to a coastal current parallel to the direction of the wind. The vertical component of this circulation evidently represents the upwelling.

From the expressions (28), (29) and (30) for u, v and w, it can be expected that the horizontal velocity of the water in this process is approximately D_h/D_v or $\sqrt{A_h/A_v}$ times as large as the vertical velocity. This result will be very useful in estimating the approximate speed of upwelling. But this kind of vertical circulation can be noticed best in the case of a very deep sea where there is very little current produced by the slope of the surface of the sea.

If we define a function $\psi(x, z)$ as

$$\psi(x,z) = \frac{2\pi\tau}{\rho\omega\sin\phi} \, \Re \int_0^\infty \left(\frac{\sin\lambda \, x/D_h (1-\cos\lambda \, L/D_h)}{\lambda \sqrt{\lambda^2 + 2\pi^2\lambda}} \right) \left\{ e^{-\sqrt{\lambda^2 + 2\pi^2\lambda} \, z/D_v} - 1 \right\} d\lambda \quad (36)$$

we can show that this is the stream function in the plane perpendicular to the coast and u and ω are given by

$$u = -\frac{\partial \psi}{\partial x} \quad ; \quad \omega = \frac{\partial \psi}{\partial x} \quad (37)$$

so that any curve

$$\psi(x,z) = \text{constant}$$

represents a stream line.

4. A Numerical Example. So far the author has elucidated the process of upwelling in a quantitative manner and obtained the expressions representing the motion of water produced by a wind blowing parallel to the coast in a belt of finite width. The stream-function $\psi(x, z)$ can be computed from the formula (36) for any distance x/D_h and for any depth z/D_v below the sea surface where D_h and D_v are the distance and depth specifying the intensity of the horizontal and vertical mixing respectively. The result of computation of the stream function is given in the Table I and illustrated by the diagram in Figure 2. The unit is given by

$$\frac{2\pi\tau}{\rho\omega\sin\phi} \quad (38)$$

From the table and diagram it can be easily shown that the vertical circulation is most strongly developed close to the coast and in the upper layers of the sea directly below the surface swept by the wind. An intense upwelling can be seen in the belt within $0.5 \, D_h$ from the coast-line and the stream-lines go down gradually outside the wind zone. This means that beyond the wind zone there occurs the process of sinking.

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TABLE I

NUMERICAL VALUES OF THE STREAM FUNCTION

IN THE VERTICAL PLANE PERPENDICULAR TO THE COAST

$$\left(\chi \frac{2\pi \tau}{\omega^2 h} \cdot 10^{-4} \right)$$

$z/D =$ \ $x/h =$	0	0.1396	0.279	0.559	0.839	1.117	1.396	1.676	1.953
0	0	0	0	0	0	0	0	0	0
0.2	0	-175	-313	-449	-461	-465	-455	-425	-310
0.4	0	-277	-499	-725	-748	-750	-734	-682	-496
0.6	0	-321	-578	-843	-867	-866	-848	-788	-576
0.8	0	-334	-600	-871	-891	-887	-869	-808	-589
1.0	0	-332	-598	-865	-890	-873	-856	-797	-579
1.2	0	-329	-592	-852	-863	-853	-837	-782	-568
1.4	0	-327	-587	-844	-853	-843	-827	-773	-561
1.6	0	-326	-585	-840	-848	-857	-823	-769	-558

.The fact that the expressions for the velocity components all include $\sin\phi$ in the denominator shows that the lower the latitude the more intense will be the process of upwelling. Perhaps the strong upwelling off the Peruvian coast may be ascribed to this theoretical result.

It will be interesting and useful to compute the magnitude of the off-shore currents and the velocity of upwelling from the stream-function given by (36) and to compare them with the values fromerly estimated from various sources.

It is of course not easy to estimate the magnitude of the coefficients of mixing. The vertical mixing coefficient may be estimated at something like 1000 c.g.s. If we adopt this value, D_v is about 162 meters for a latitude of 30°N. To estimate A_h is even more difficult. But actual oceanographic observations show that $A_h/A_v = 10^6$ approximately. This means that D_h is just about 1000 times as large as D_v, or 162 kilometers. Furthermore, we do not know much about the width of the coastal wind belt. In this computation, the author tentatively assumed $L = 2.094 \times D_h$ that is, about twice as large as D_h; or at 339 kilometers.

It will be still more difficult to estimate the wind velocity of the northwesterlies prevailing off the coast of Southern California in the earlier summer months. The author took $\tau = 1$ c.g.s. This corresponds to a wind velocity between 5 and 6 m/s. If we consider the upwelling off the coast of Southern California and take $\phi = 30°N$, we have

$$\frac{2\pi\tau}{\rho\omega\sin\phi} = 2.37 \times 10^5 \text{ sec}^{-1}$$

From the table we can compute the average velocity between the surface and the layer $0.2\,D_v$ deep by

$$\frac{\Delta\Psi}{\Delta z} = 2.37 \times 10^5 \times \frac{0.045}{0.2\,D_v} = 3.35 \text{ cm/sec}$$

This is the maximum velocity of the offshore current in the layer between the surface and the 32.4 meters level.

The maximum vertical velocity can be estimated in a similar way, viz., at

$$\frac{\Delta\Psi}{\Delta x} = 2.37 \times 10^5 \times \frac{0.033}{0.14\,D_h} = -3.14 \text{ cm/sec (upward)}$$

$$= -2.7 \text{ m/day.}$$

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This speed of upwelling is just about <u>80 meters per month</u>.

G. F. McEwen (1929) estimated the speed of upwelling off the coast of Southern California at about 10-20 meters per month. The present result appears to show a speed a little too high, but may be suggestive of the order of magnitude of ascending motion in the process of upwelling.

From the diagram in Figure 2, we can see that the water mass partici- pating in this process comes up from the layers from $Z = D_v$ to $1.5\, D_v$ or more. If we take $D_v = 162\,m$, the layers from which the upwelled waters originate are located somewhere around the layers 200 meters or deeper. This also agrees with Sverdrup-Fleming's estimation derived from practical observations off the coast of Southern California.

5. <u>Coastal Currents</u>. In addition to the circulation in a vertical plane perpendicular to the coast, we have a current parallel to the coast. The author thinks that this will be another subject of major interest.

The model treated in this research is very simple, the winds being assumed always to blow parallel to the coast. But it may be considered that a certain pattern of water motion will always correspond to the wind of any direction. The investigation into this problem seems suggestive of the explanation of several facts observed close to the shore in relation to the motion of the water.

6. <u>Acknowledgements</u>. The author hereby expresses his sincerest appreciation to Dr. Dale F. Leipper, Head of the Department of Oceanography, Agricultural and Mechanical College of Texas, who encouraged the author in carrying out the present research and publishing the result during his stay in the depart- ment. He is also much obliged to Mr. Robert O. Reid and Dr. John T. Hurt who kindly discussed on the result.
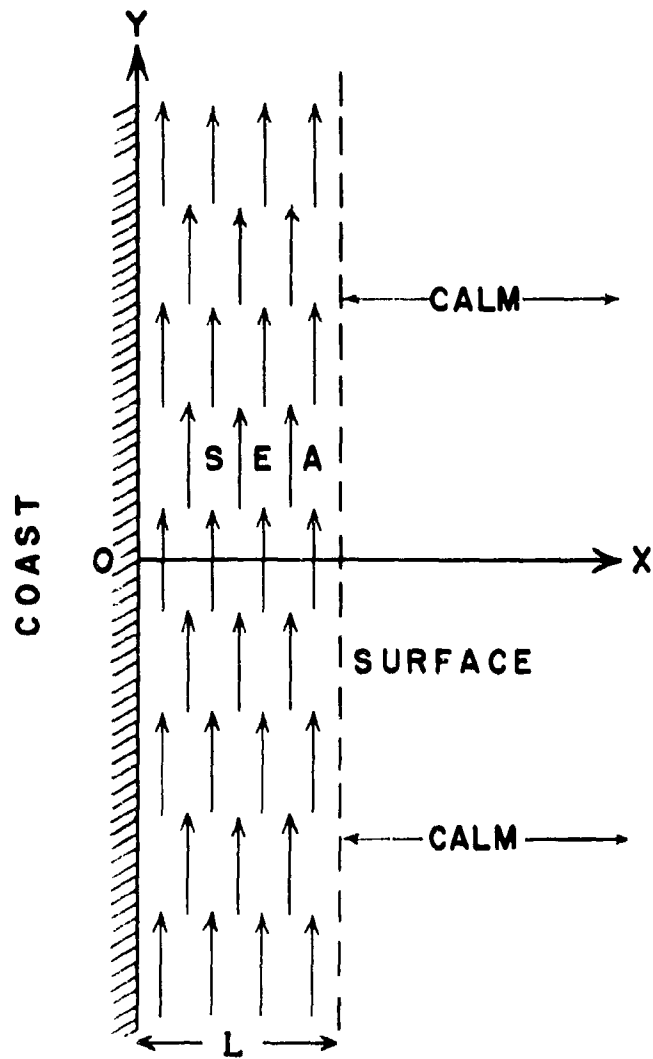
WIND BELT
SCHEMATIC DIAGRAM OF WIND TO COAST RELATIONSHIP
FOR THE DEVELOPMENT OF UPWELLING IN THE NORTHERN HEMISPHERE
FIGURE I
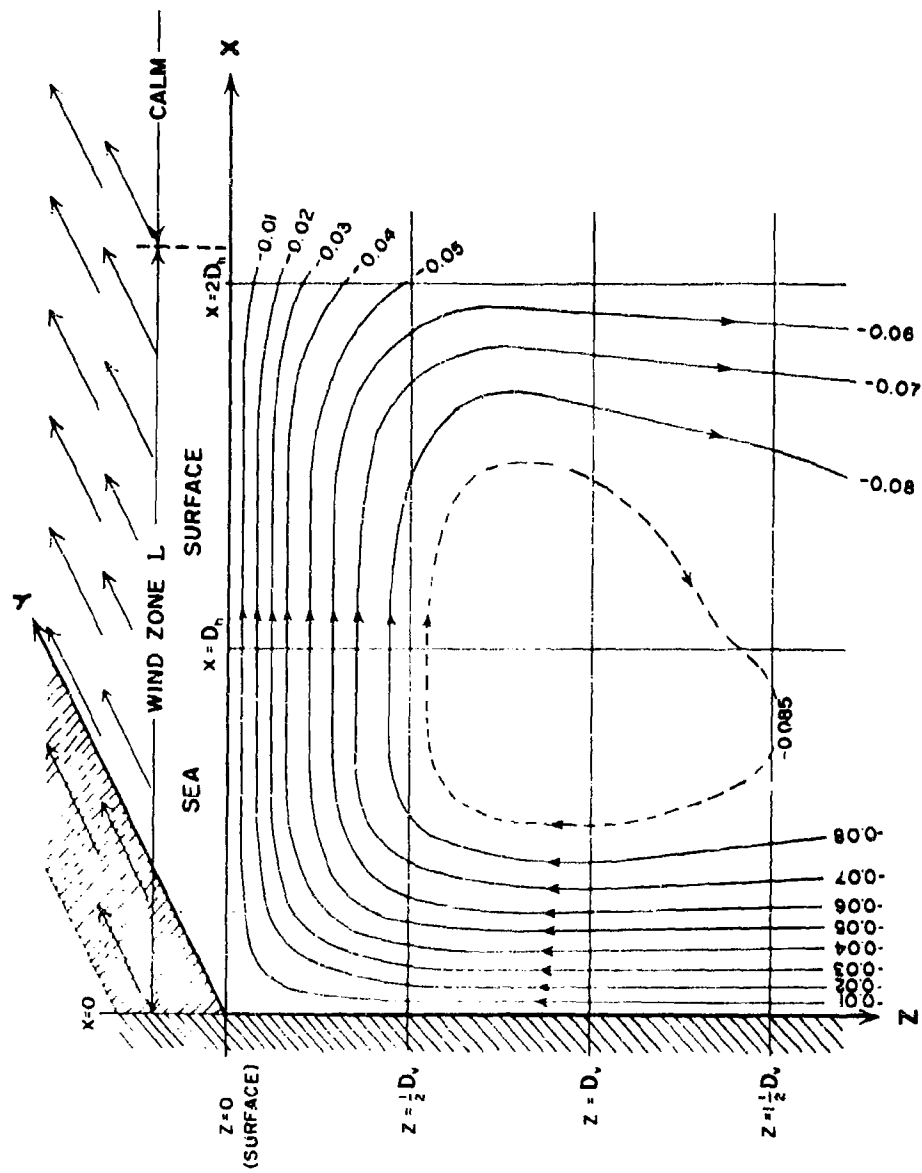
UPWELLING AS INDUCED BY A WIND PARALLEL TO THE COAST
(ILLUSTRATED BY THE STREAMLINES IN THE VERTICAL PLANE PERPENDICULAR TO THE COAST)

FIGURE 2

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