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A CONTRIBUTION TO THE THEORY OF UPWELLING, PART II

Koji Hidaka

Research Conducted through the

Texas A. & M. Research Foundation

COLLEGE STATION, TEXAS

THE AGRICULTURAL AND MECHANICAL COLLEGE OF TEXAS
Department of Oceanography
College Station, Texas

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

Project 24

A CONTRIBUTION TO THE THEORY OF UPWELLING, PART II
The Most Favorable Condition for Upwelling and The Coastal Currents Induced by Winds

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by
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A CONTRIBUTION TO THE THEORY OF UPWELLING, PART II
The Most Favorable Condition for Upwelling and the Coastal Currents Induced by Winds

By

KOJI ITDAKA

ABSTRACT

In Technical Report No. 6 of this project the author showed the possibility of upwelling induced by a wind blowing parallel to the coast in such a manner that, in the northern (southern) hemisphere, the coast is on the left-hand (right-hand) side of an observer who looks in the direction of the wind. In this report he shows that an offshore wind blowing perpendicular to the coast can also give rise to a similar circulation in the same vertical plane, but in this case with much weaker intensity. From this result it can be concluded that the most intense upwelling will occur when the wind makes an angle of 21.5° with the coast line in an offshore direction. In addition to these circulations in the vertical plane there exist horizontal currents parallel to the coast. These will contribute to the formation of the coastal or longshore currents much debated recently. The horizontal components of the motion under consideration show a vertical variation very similar to the Ekman spiral at some distance from the coast.

1. <u>Upwelling Due to a Wind Not Parallel to the Coast</u>. In the first report on the subject of upwelling (Technical Report No. 6 of this project) the author has shown the possibility of upwelling induced by a wind blowing parallel to the coast. The result was that when a wind blows parallel to the coast in such a manner that, in the northern (southern) hemisphere an observer looking in the direction of the wind has the sea on his left-hand (right-hand) side, there exists an intense circulation in a vertical plane perpendicular to the coast ($x z$ -plane). This upwelling takes place close to the coast and has a velocity which is consistent with observations.

It may be anticipated, however, that a similar circulation can also be produced in the same vertical plane by an offshore wind, that is, a wind blowing offshore at right angles to the coast. To evaluate this circulation is, of course, possible by assuming a wind stress in the x- direction. Suppose we have a wind not perpendicular to, but making a certain angle with the coast, and divide the stress τ into its x- and y- components (Figure 1). Then the surface conditions (17) in Technical Report No. 6 should be replaced by

$$z = 0 . \quad -A_z \frac{\partial u}{\partial z} = \tau_x , \quad -A_z \frac{\partial v}{\partial z} = \tau_y , \qquad (.)$$

where u and v are the x- and y- components of current velocity, τ_x and τ_y are those of wind stress while the dynamical and continuity equations and boundary conditions remain the same.

In exactly the same manner as in the first report, we can integrate the equations of motion and obtain the stream function in the vertical plane perpendicular to the coast and the components u and v of the horizontal current. They are

$$\psi(x,z) = \text{real part of}$$

$$\frac{2\pi(\tau_x + i\tau_y)}{\rho\omega\sin\varphi}\int_0^\infty \frac{\sin\lambda\frac{x}{D_h}\left(1-\sin\lambda\frac{L}{D_h}\right)}{\lambda}\cdot\frac{e^{-\sqrt{\lambda^2+2\pi^2 i}\,\frac{z}{D_v}}-1}{\lambda^2+2\pi^2 i}\,d\lambda \qquad (2)$$

and

$$u+iv = \frac{2\pi(\tau_x+i\tau_y)}{\rho\omega\sin\varphi D_v}\int_0^\infty \frac{\sin\frac{x}{D_h}\left(1-\sin\lambda\frac{L}{D_h}\right)}{\lambda}\cdot\frac{e^{-\sqrt{\lambda^2+2\pi^2 i}\,\frac{z}{D_v}}}{\sqrt{\lambda^2+2\pi^2 i}}\,d\lambda \qquad (3)$$

respectively. These expressions can be put in real forms,

$$\psi(x,z) = \frac{2\pi\tau_x}{\rho\omega\sin\varphi}\Phi_x(x,z) + \frac{2\pi\tau_y}{\rho\omega\sin\varphi}\Phi_y(x,z), \qquad (4)$$

$$u = \frac{2\pi\tau_x}{\rho\omega\sin\varphi D_v}N(x,z) - \frac{2\pi\tau_y}{\rho\omega\sin\varphi D_v}M(x,z), \qquad (5)$$

$$v = \frac{2\pi\tau_x}{\rho\omega\sin\varphi D_v}M(x,z) + \frac{2\pi\tau_y}{\rho\omega\sin\varphi D_v}N(x,z), \qquad (6)$$

where

$$\Phi_x(x,z) = \int_0^\infty \frac{(P^2-Q^2)e^{-P\frac{z}{D_v}}\cos Q\frac{z}{D_v} - 1) - 2PQe^{-P\frac{z}{D_v}}\sin Q\frac{z}{D_v}}{(P^2+Q^2)^2}\,\frac{\sin\lambda\frac{L}{D_h}\left(1-\cos\lambda\frac{x}{D_h}\right)}{\lambda}\,d\lambda, \qquad (7)$$

$$\Phi_y(x,z) = \int_0^\infty \frac{(P^2-Q^2)e^{-P\frac{z}{D_v}}\sin Q\frac{z}{D_v}) + 2PQ(e^{-P\frac{z}{D_v}}\cos Q\frac{z}{D_v} - 1)}{(P^2+Q^2)^2}\,\frac{\sin\lambda\frac{x}{D_h}\left(1-\cos\lambda\frac{L}{D_h}\right)}{\lambda}\,d\lambda \qquad (8)$$

$$M(x,z) = \int_0^\infty \frac{Q\cos Q\frac{z}{D_v} + P\sin Q\frac{z}{D_v}}{P^2+Q^2}\,e^{-P\frac{z}{D_v}}\,\frac{\sin\lambda\frac{x}{D_h}\left(1-\cos\lambda\frac{L}{D_h}\right)}{\lambda}\,d\lambda \qquad (9)$$

$$N(x,z)=\int_0^\infty \frac{P\cos Q\frac{z}{D_v} - Q\sin Q\frac{z}{D_v}}{P^2+Q^2}\, e^{-P\frac{z}{D_v}} \; \frac{\sin\frac{x}{D_h}\left(1-\cos\frac{L}{D_h}\right)}{\lambda}\, d\lambda , \tag{10}$$

and we have

$$D_h = \pi\sqrt{A_h/\rho\omega\sin\phi} , \tag{11}$$

$$D_v = \pi\sqrt{A_v/\rho\omega\sin\phi} . \tag{12}$$

Here A_h and A_v are the coefficients of horizontal and vertical mixing, ρ the density of sea water, ω the angular velocity of the Earth's rotation and ϕ the geographic latitude.

If we put

$$\tau_x = 0 , \quad \tau_y = \tau ,$$

we have exactly the problem we discussed in the first report, in which the wind blows parallel to the coast.

The stream function $\Phi_x(x,z)$ and $\Phi_y(x,z)$, and the function $M(x,z)$ and $N(x,z)$ which give the horizontal components of velocity were evaluated by numerical computation against several values of x/D_h and z/D_v, and are compiled in Tables I to IV.

Since a wind of any direction can be divided into its two horizontal components perpendicular and parallel to the coast, it will be sufficient for us to discuss only the vertical circulations induced by the offshore and longshore winds and to combine the result.

2. <u>Pattern of the Vertical Circulations</u>. Figures 2 and 3 show the stream lines in the vertical circulations induced in the plane perpendicular to the coast (xz -- plane) by longshore and offshore winds respectively. There seems to be an essential difference between them. Both are clockwise circulations and the stream functions have negative values. In each of these two cases, the water is upwelled from levels deeper than $z = D_v$, but, as a whole, it appears that a longshore wind originates upwelled water from deeper layers than does an offshore wind. The circulation induced by a longshore wind has already been described in detail in the first report. The circulation due to an offshore wind of the same stress carries a much smaller amount of water than does that due to a longshore wind. Moreover, the former has a rather complicated structure, having two eddies (Figure 3) in upper layers, one situated close to the coast

and the other near the outer boundary of the wind belt. This means that upwelling due to a longshore wind is far more effective in lowering the surface temperature of the coastal regions than that induced by an offshore one, since the former carries much larger amounts of colder water to the surface from deeper layers than the latter.

3. <u>Most Favorable Condition for Upwelling</u>. The amount T of water upwelled across a long horizontal strip 1 cm wide extending from the coast to a distance x can be evaluated as

$$T = \int_0^x \rho w \, dx = \rho \int_0^x \frac{\partial \psi}{\partial x} \, dx$$

$$= \rho \psi(x, z) - \rho \psi(0, z)$$

$$= \rho \psi(x, z) , \tag{13}$$

since the stream function vanishes along the coastal wall ($x = 0$). Thus the vertical mass transport T will be largest when we take x at a distance at which the functions Φ_x and Φ_y have the largest negative value. By inspection of Tables I and II and Figures 2 and 3, we can estimate

$$T_y = -0.090 \times \frac{2\pi \tau_y}{\omega \sin \varphi} \quad \text{by longshore wind} \tag{14}$$

and

$$T_x = -0.0355 \times \frac{2\pi \tau_x}{\omega \sin \varphi} \quad \text{by offshore wind} \tag{15}$$

If the wind blows in a direction making an angle θ with the coast in an offshore direction (Figure 1), we have

$$\tau_x = \tau \sin \theta , \quad \tau_y = \tau \cos \theta \tag{16}$$

where τ is the absolute magnitude of the wind stress. Then we have from (14) and (15)

$$T = T_x + T_y$$

$$= \left(-0.0355 \sin \theta - 0.090 \cos \theta\right) \frac{2\pi \tau}{\omega \sin \varphi} . \tag{17}$$

This gives an approximate amount of water upwelled to the surface layers per unit length of the coast. The most intense upwelling will therefore take place when $dI/d\theta = 0$, or

$$\tan\theta = \frac{0.0355}{0.090} \quad \text{or} \quad \theta = 21.5^\circ \tag{18}$$

providing the magnitude of wind stress τ remains the same. This means that the upwelling will be <u>most intense when the wind makes an angle of 21.5° with the coast line in an offshore direction,</u> that is, when it is slightly deviated offshore from the coast line. Of course, we have more or less upwelling in a sector within ± 90° of this direction. According to Dr. Arnold Glaser, upwelling off the coast of Lima, Peru, appears to be more intense when the southerlies blow slightly deviated in an offshore direction rather than when they blow parallel to the coast.

In the Gulf of Mexico, we have easterlies almost all the year round. This means that we can always expect upwelling along the northern coast of Cuba and Yucatan coast of Mexico. We do not have any distinct evidence of upwelling and the existence of colder water off these coasts. But we know that many American anglers go to fish in the waters off the Yucatan coast and special boats are sent from the United States coasts for these people. This fact may suggest the existence of upwelling in this region. According to the surface observations of the Department of Oceanography, Agricultural and Mechanical College of Texas made recently, there are also some indications of colder surface temperatures in the western part of the Gulf of Mexico very close to the coast. If they really exist, this temperature distribution may be ascribed to upwelling due to the easterly winds prevailing in these latitudes.

4. <u>Coastal Currents.</u> Recently several observations (Putnam, Munk and Traylor, 1949; Shepard and Inman, 1950, 1951; Inman and Quinn, 1953) on the flow of water close or adjacent to coasts have been reported. Many authors have discussed this subject, but they have not yet been able to arrive at a satisfactory explanation of possible causes of coastal or longshore currents. However, most of these currents are ascribed to the action of surf and waves. The longshore component which we have obtained may explain some part of these flows. The expression for this component is given by

$$v = \frac{2\pi \tau_x}{\rho\omega \sin\phi D_r} N(x,z) - \frac{2\pi \tau_y}{\rho\omega \sin\phi D_r} M(x,z) \tag{19}$$

where M and N are the functions compiled in Tables III and IV and illustrated in Figures 4 and 5. The two terms in the right-hand member of (19) show the longshore currents induced by offshore and longshore winds respectively. If the wind is parallel to the coast we have $\tau_x = 0$ and

TABLE I. VALUES OF THE STREAM FUNCTION

$\Phi_\psi(x, z) \times 10^4$

z/D_c \ x/D_h	0	.1396	.2793	.5585	.8378	1.1170	1.3963	1.6755	1.9548
0	0	0	0	0	0	0	0	0	0
.2	0	-175	-313	-449	-461	-465	-455	-425	-310
.4	0	-277	-499	-725	-748	-750	-734	-682	-496
.6	0	-321	-578	-843	-867	-866	-848	-788	-576
.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	-837	-823	-769	-558

TABLE II. VALUES OF STREAM FUNCTION

$\Phi_\psi(x, z) \times 10^4$

z/D_c \ x/D_h	0	.1396	.2793	.5585	.8378	1.1170	1.3963	1.6755	1.9548
0	0	0	0	0	0	0	0	0	0
.2	0	-168	-303	-301	-254	-270	-280	-305	-202
.4	0	-194	-342	-310	-230	-238	-257	-303	-199
.6	0	-179	-309	-234	-135	-135	-161	-226	-133
.8	0	-162	-275	-172	- 68	- 58	- 83	-162	- 85
1.0	0	-152	-256	-138	- 19	- 17	- 43	-129	- 61
1.2	0	-148	-248	-125	- 5	- 3	- 30	-107	- 54
1.4	0	-148	-274	-124	- 4	- 3	- 30	-107	- 54
1.6	0	-147	-248	-126	- 6	- 6	- 33	-110	- 57

TABLE III. VALUES OF THE FUNCTION

$N(x, z) \times 10^3$

z/D_v \ x/D_h	0	.1396	.2793	.5585	.8378	1.1170	1.3963	1.6755	1.9548
0	0	+240	+277	+284	+253	+252	+252	+266	+243
.2	0	+ 41	+ 64	+ 55	+ 35	+ 31	+ 38	+ 47	+ 22
.4	0	- 3	- 8	- 28	- 42	- 45	- 42	- 28	- 17
.6	0	- 10	- 20	- 38	- 47	- 50	- 48	- 39	- 29
.8	0	- 7	- 14	- 24	- 29	- 30	- 29	- 24	- 20
1.0	0	- 3	- 6	- 11	- 12	- 12	- 12	- 10	- 8
1.2	0	- 1	- 2	- 2	- 3	- 3	- 2	- 2	- 1
1.4	0	0	0	- 1	+ 1	+ 1	+ 1	+ 1	+ 1
1.6	0	0	+ 1	+ 1	+ 2	+ 2	+ 2	+ 1	+ 1

TABLE IV. VALUES OF THE FUNCTION

$M(x, z) \times 10^3$

z/D_v \ x/D_h	0	.1396	.2793	.5585	.8378	1.1170	1.3963	1.6755	1.9548
0	0	- 98	-174	-248	-255	-243	-263	-235	-174
.2	0	- 70	-127	-186	-191	-182	-197	-175	-129
.4	0	- 34	- 62	- 93	- 96	- 90	- 97	- 85	- 62
.6	0	- 12	- 21	- 31	- 30	- 26	- 30	- 26	- 18
.8	0	- 2	- 3	- 2	0	+ 3	+ 1	+ 1	+ 1
1.0	0	+ 1	+ 3	+ 6	+ 10	+ 10	+ 9	+ 8	+ 6
1.2	0	+ 1	+ 3	+ 6	+ 8	+ 8	+ 8	+ 6	+ 5
1.4	0	+ 1	+ 2	+ 3	+ 4	+ 4	+ 4	+ 3	+ 2
1.6	0	0	+ 1	+ 1	+ 1	+ 1	+ 1	+ 1	+ 1

$$v = -\frac{2\pi \tau_y}{\rho \omega \sin\phi D_v} M(\chi, z), \tag{20}$$

and if the wind blows in an offshore direction perpendicular to the coast we have

$$v = +\frac{2\pi \tau_x}{\rho \omega \sin\phi D_v} N(\chi, z). \tag{21}$$

From the Tables III and IV and Figures 4 and 5 it is possible for us to tell the difference between the two cases. The coastal current induced by a longshore wind is directed leeward as far down as the level $z = 0.9 D_v$, but the direction of flow is reversed at deeper levels. On the other hand, the longshore current produced by an offshore wind flows to the right of the wind direction only in layers shallower than $z = 0.3 D_v$. There is a secondary maximum of velocity in the opposite direction at the approximate level $z = 0.5 D_v$.

5. <u>Vertical Variation of Horizontal Currents: Ekman Spiral.</u> The expressions

$$\left.\begin{aligned} u &= \frac{2\pi \tau_x}{\rho \omega \sin\phi D_v} N(\chi, z) - \frac{2\pi \tau_y}{\rho \omega \sin\phi D_v} M(\chi, z), \\ v &= \frac{2\pi \tau_x}{\rho \omega \sin\phi D_v} M(\chi, z) + \frac{2\pi \tau_y}{\rho \omega \sin\phi D_v} N(\chi, z), \end{aligned}\right\} \tag{22}$$

for the horizontal currents show that the angle between the wind and current induced at a given depth <u>depends only upon the distance from the coast and not on the direction of wind</u>. The angle between the wind and the surface current induced by it is computed for several points at different distances from the coast and compiled in Table V.

Table V

The Angle Between the Wind and Surface Current

χ/D_v	0.0000	.1396	.2793	.5585	.8378	1.1170	1.3963	1.6755	1.9548
Angle	0°	22.2°	32.2°	41.2°	45.2°	44.3°	46.3°	41.5°	35.7°

Thus the angle between the wind and surface current is nearly 45° at the middle part of the wind belt, but decreases both toward the coast and outer margin of the wind belt. Along the coast there is a very weak current in approximately the same direction as the wind. The vertical variation of currents is practically equal to the Ekman spiral along the

median line of the wind belt, but is more or less flattened close to the coast and near the outer margin of the wind belt. Some of these spirals are illustrated in Figure 6.

6. **Acknowledgements.** 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 Department. He is also much obliged to Dr. Leipper, Robert O. Reid and Dr. Arnold Glaser for their valuable suggestions and information.

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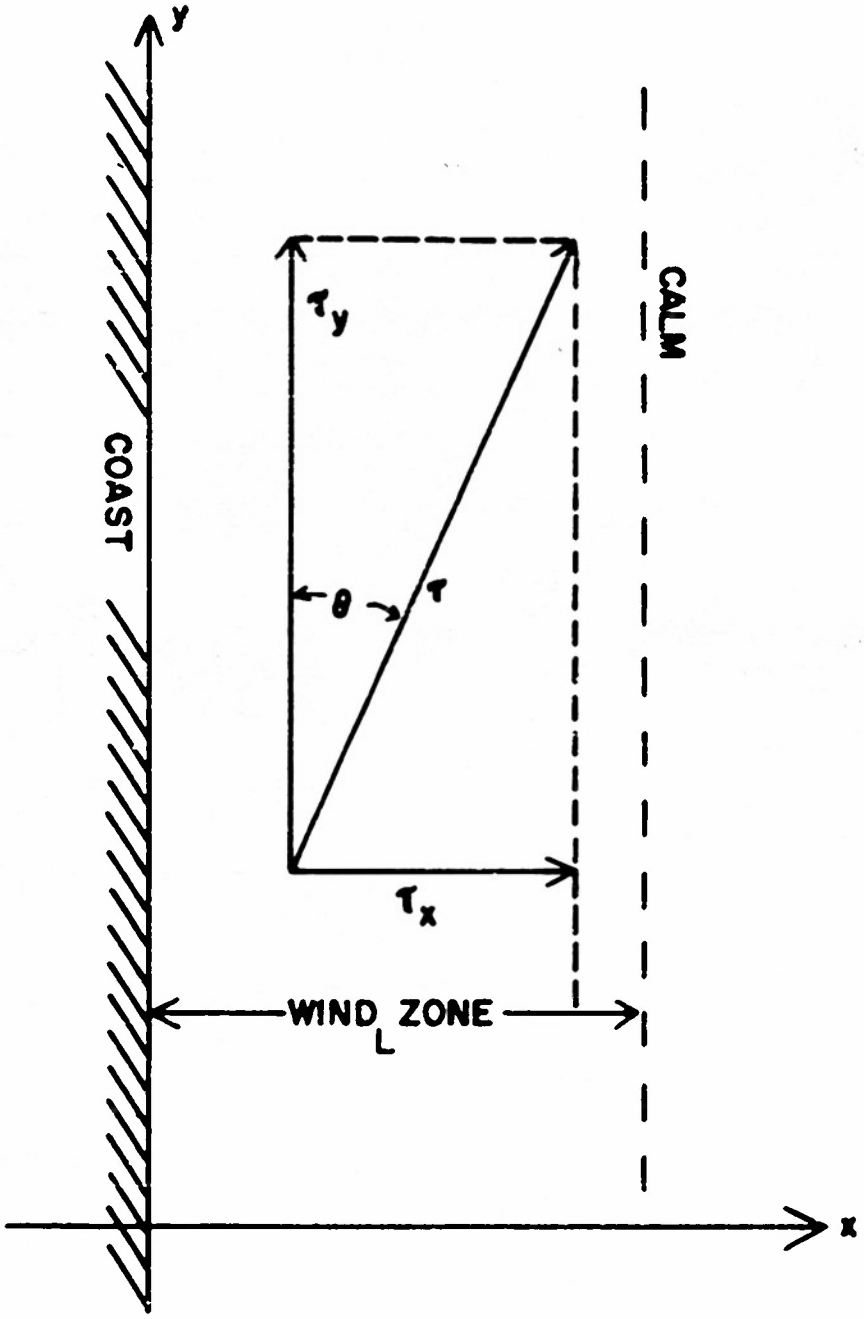
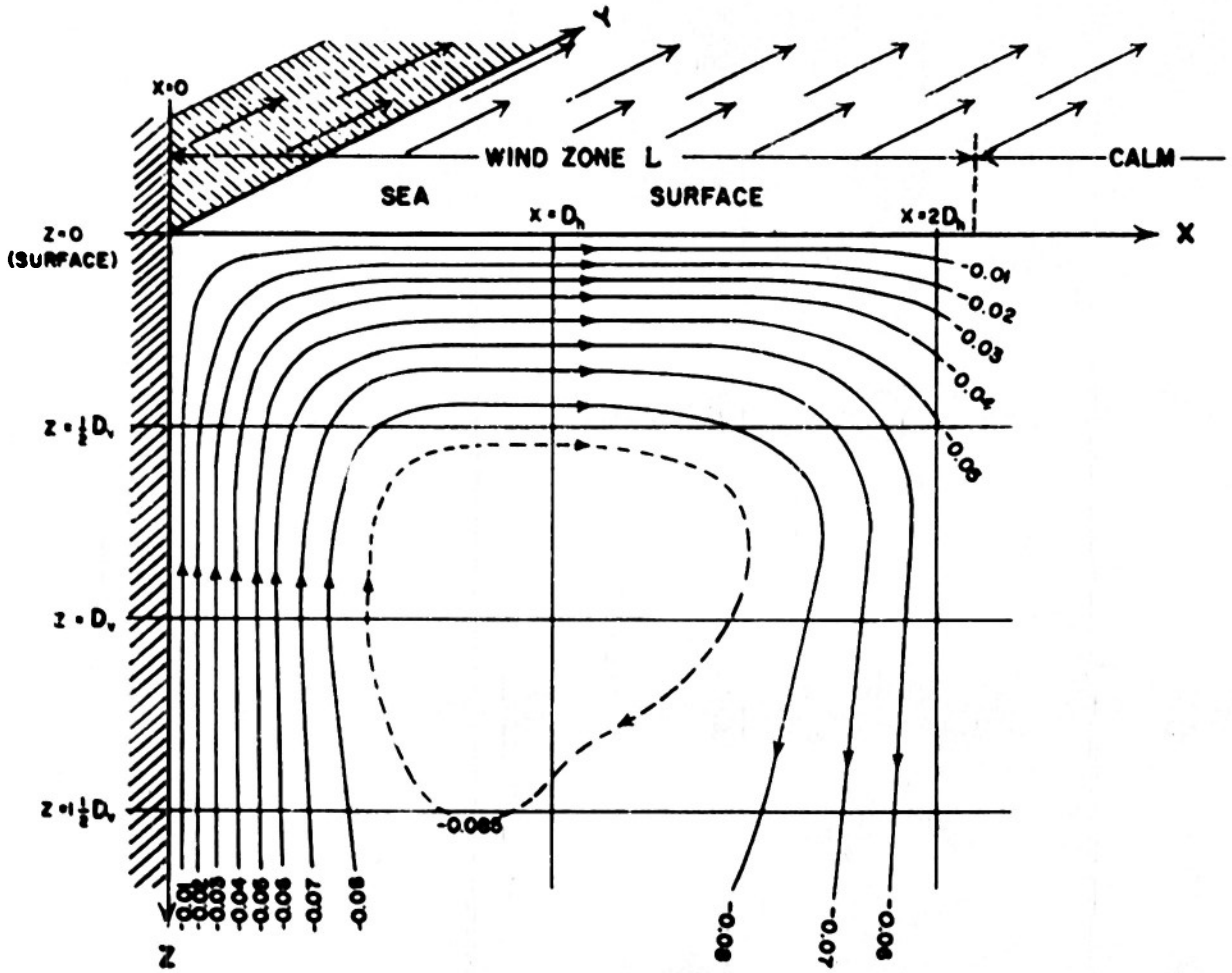


FIGURE 1

WIND MAKING AN ANGLE
θ WITH THE COAST



UPWELLING AS INDUCED BY A WIND PARALLEL TO THE COAST
(ILLUSTRATED BY THE STREAMLINES IN THE VERTICAL PLANE PERPENDICULAR TO THE COAST)

FIGURE 2

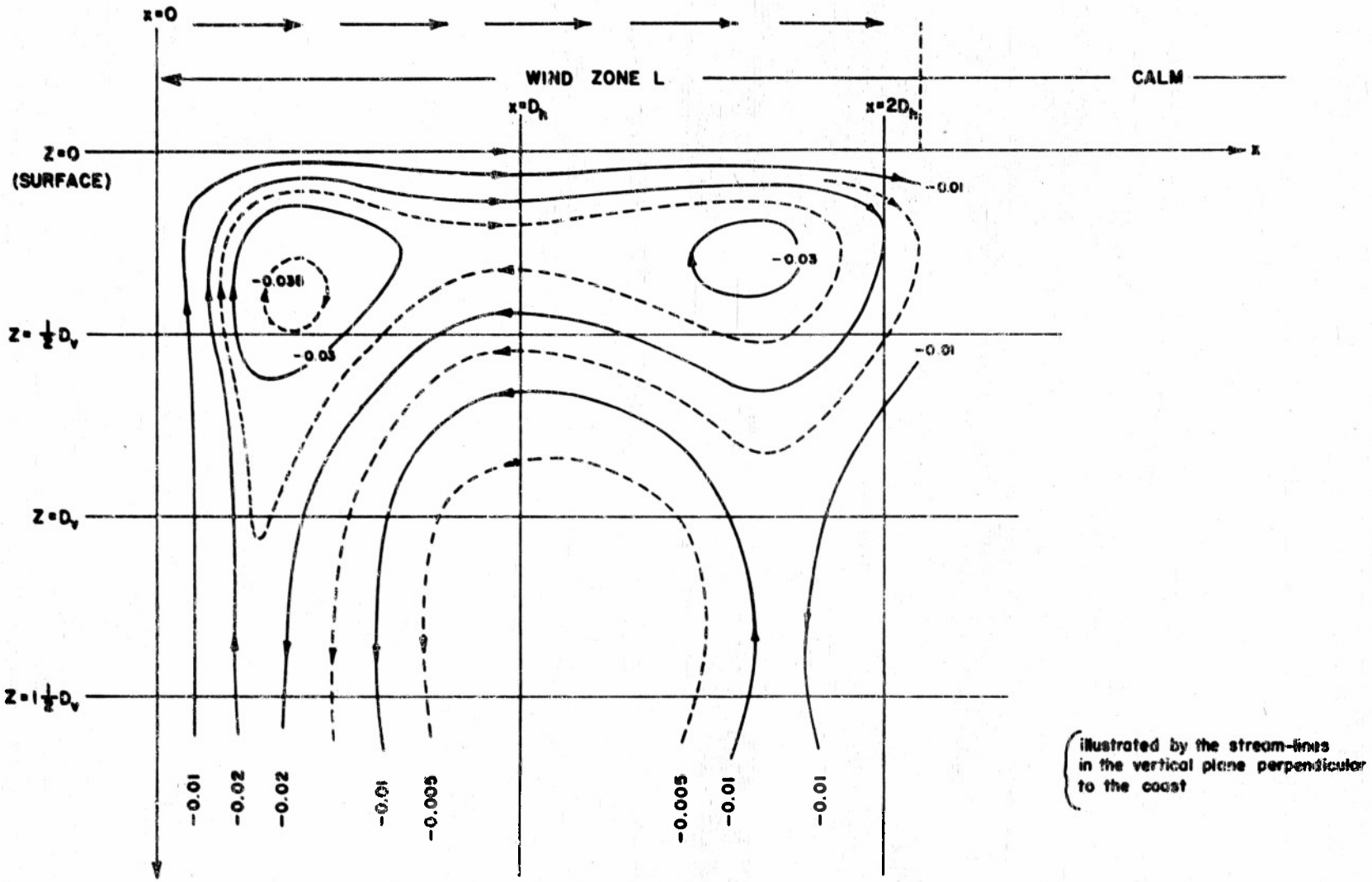
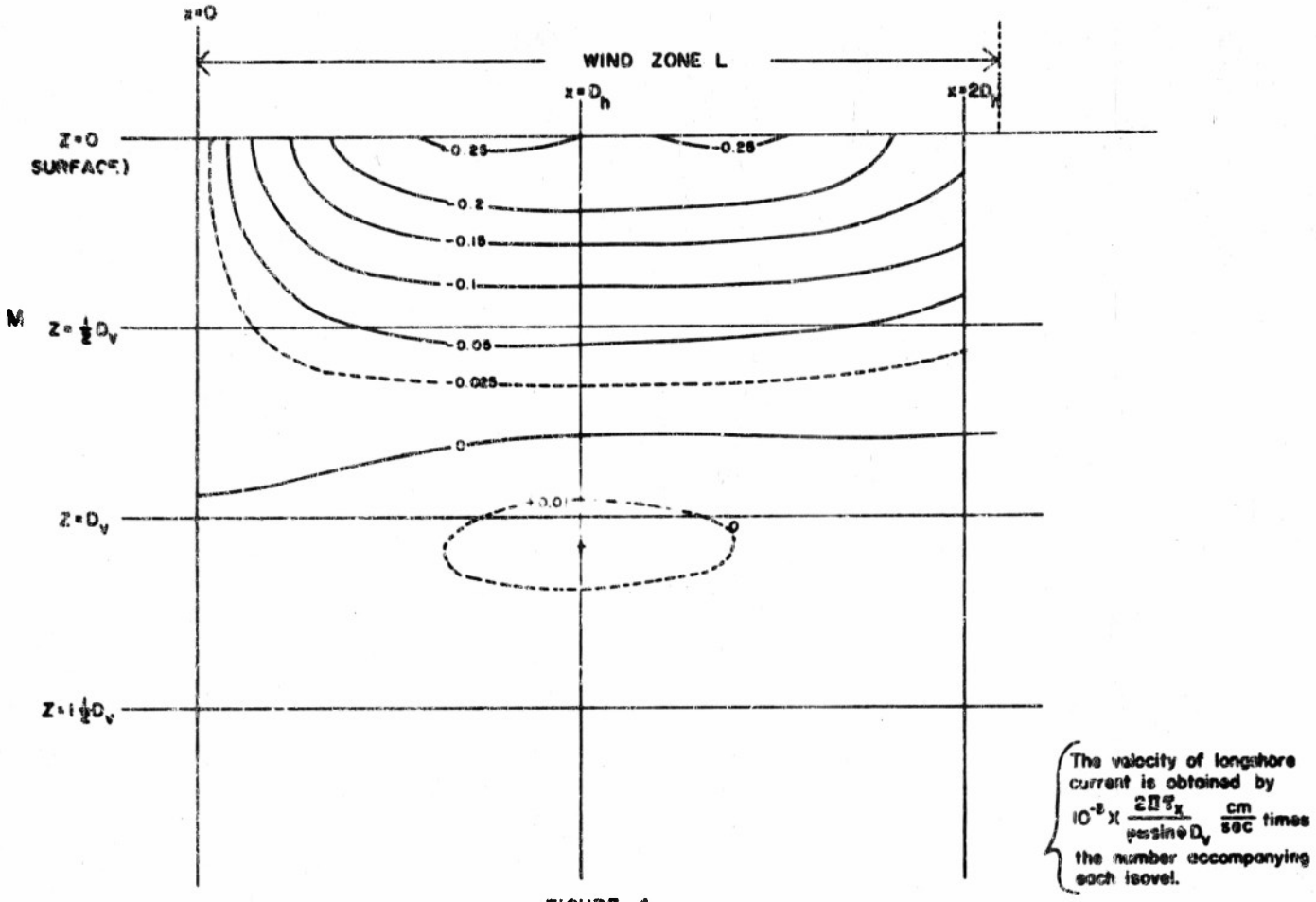


FIGURE 3

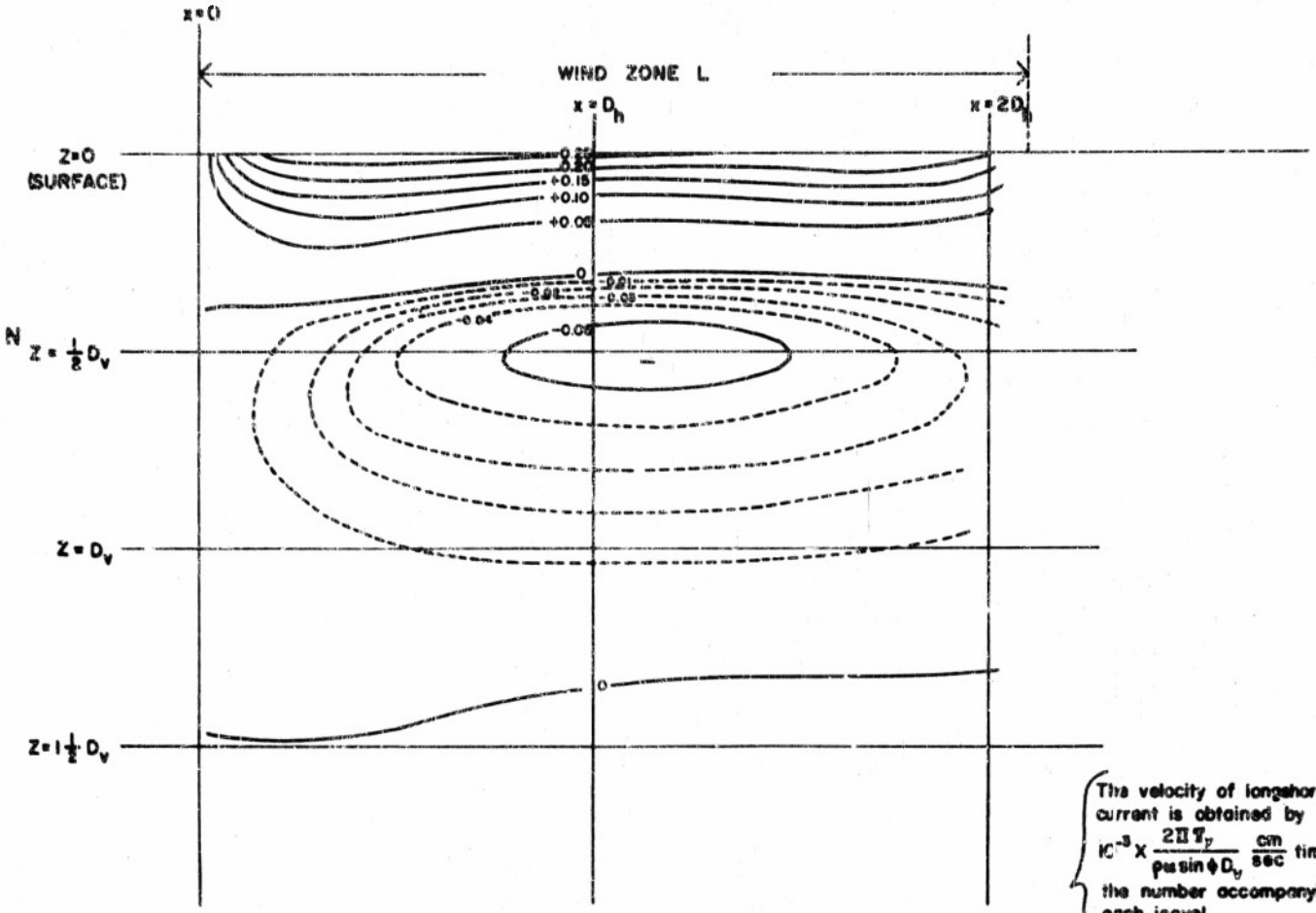
UPWELLING AS INDUCED BY AN OFFSHORE WIND



The velocity of longshore current is obtained by $10^{-2} \times \frac{2\pi\tau_x}{\rho_s \sin\theta D_v} \frac{cm}{sec}$ times the number accompanying each isovel.

FIGURE 4

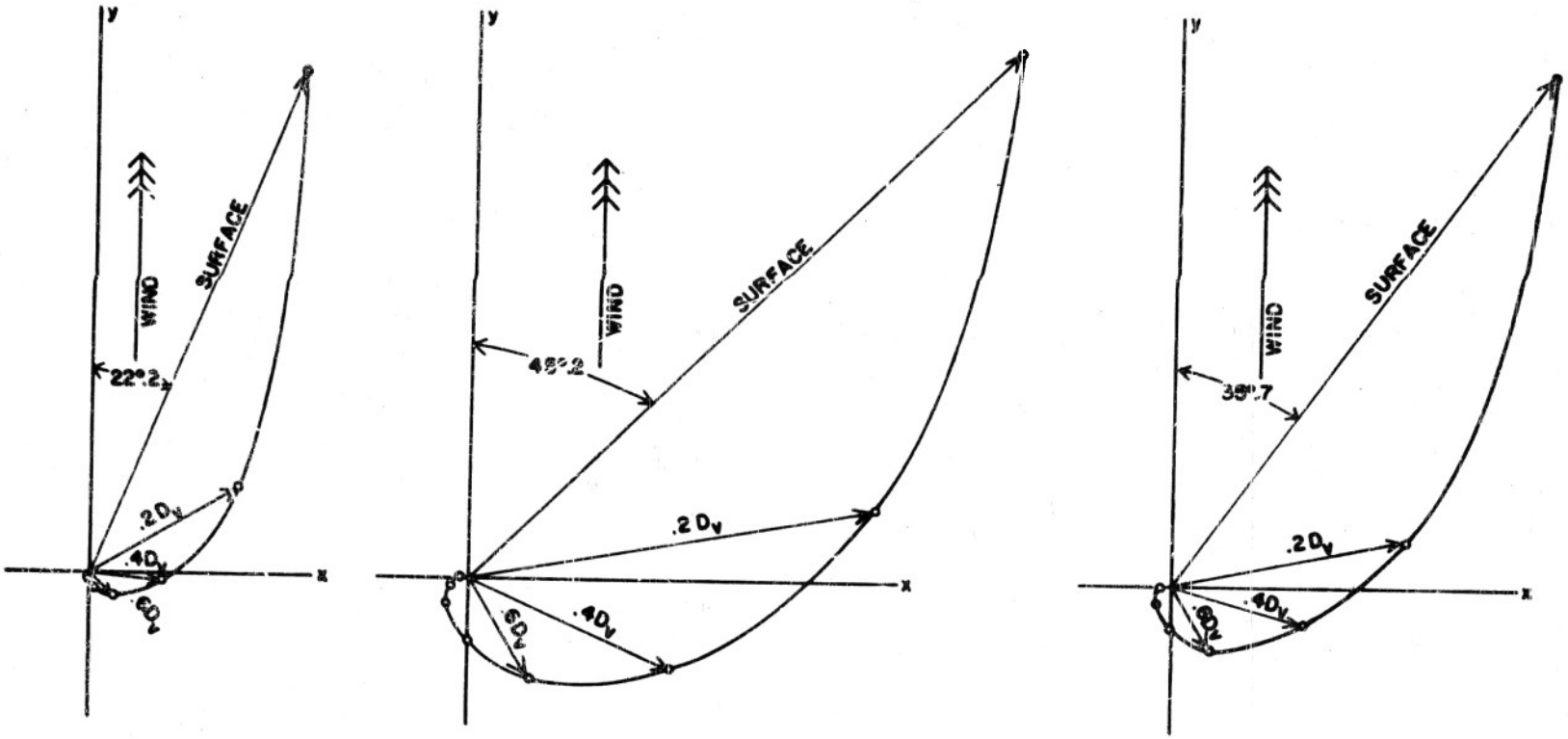
DISTRIBUTION OF LONGSHORE CURRENTS M(x, z) INDUCED BY A WIND τ_x PARALLEL TO THE COAST



The velocity of longshore current is obtained by $10^{-3} \times \frac{2\Pi\tau_y}{\rho u \sin\phi D_v} \frac{cm}{sec}$ times the number accompanying each isovel.

FIGURE 5

DISTRIBUTION OF LONGSHORE CURRENT N(x,z) INDUCED BY AN OFFSHORE WIND τ_y



$\frac{x}{D_h}$ = .1396
(CLOSE TO THE COAST)

$\frac{x}{D_h}$ = .8378
(MIDDLE PART OF WIND BELT)

$\frac{x}{D_h}$ = 1.9548
(CLOSE TO OUTER PART OF WIND BELT)

FIGURE 9
VERTICAL VARIATION OF DRIFT CURRENTS

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