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NUMERICAL CALCULATIONS OF THE WIND-DRIVEN CURRENTS IN LAKE ERIE AND COMPARISON WITH MEASUREMENTS

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ABSTRACT

The steady-state, wind driven velocities in Lake Erie have been calculated numerically using a shallow lake model. The three-dimensional velocities as a function of depth and horizontal position are displayed for the prevailing southwest winds. The results show that the velocities vary greatly from position to position and depend strongly on the bottom topography and boundary geometry. For the numerical calculations, a 0.805 kilometer grid size in an island region and a 3.22 kilometer grid size in the rest of the lake had to be incorporated to adequately represent the Lake Erie geometry.

The calculated velocities compare quantitatively very well with current meter measurements made at mid-depths in the Central and Eastern basins. The magnitudes of the average eddy viscosity used in the calculations agree with measurements made in the Great Lakes. Steady currents are shown to usually occur after two days of fairly uniform winds.

INTRODUCTION

Present day analyses of the effects of pollution in the Great Lakes require a detailed knowledge of the three-dimensional velocities at any location in the lakes. At last year's conference, the authors (Gedney and Lick, 1970) presented some preliminary calculated results for the local currents and velocities in Lake Erie. In that analysis, the presence of islands was considered, but the islands were approximated by underwater mounds. In the present paper, more extensive and improved results are presented. These results have been obtained by using a more realistic island geometry and a finer mesh in the numerical calculations.

In the present analysis, only the motions caused by steady winds and river through flows are considered. For deep lakes, in which the depth is much greater than the thickness of the friction or Ekman layers, the usual Ekman dynamics, in which the bottom stress is assumed proportional to the geostrophic velocity, can be used. For any of the Great Lakes, the use of Ekman dynamics is questionable since they all have shallow shore regions of considerable extent. In Lake Erie, for moderate winds, the thickness of the friction layer is comparable to the depth over much of the lake and therefore the use of Ekman dynamics is not valid. The necessary extension of the Ekman analysis to the case of a shallow lake has been given by Welander (1957) and that theory has been used here with slight modifications.

The shallow depth of Lake Erie does have an advantage in that the time to reach steady state for a specific wind condition should be less than for the other Great Lakes. The calculations performed are for steady state winds and comparison of the calculated results and measurements do show that the steady state limit is actually reached in Lake Erie. This is important since it makes the task of checking the current calculations with measurements much easier. It is only through this checking, which may indicate a need to modify the model, that a satisfactory model will be established. Once an accurate quantitative steady state model is established, the time-dependent case can be more readily analyzed. The comparison of our Lake Erie calculations with measurements is a major part of this article.

Shallow-Lake Model and Method of Solution

In the present analysis, the basic approximations are that the water density is constant, the vertical eddy viscosity is independent of depth but dependent on wind velocity, the pressure is hydrostatic, and the lateral friction and nonlinear acceleration terms can be neglected. The neglect of lateral friction means that the two transverse friction terms in the momentum equations are small compared to the vertical friction term. Lake Erie is stratified during the summer months and therefore the analysis presented here applies only to the fall, spring, and those periods in the winter when the lake is not iced over.

The above assumptions reduce the momentum equations to two equations containing the horizontal velocities and the surface slope as unknowns. The appropriate boundary conditions for these equations are a no-slip condition at the lake bottom and a specified shear stress (due to the wind) at the air-water interface. These equations and boundary conditions can be solved analytically to give the velocity as a function of depth with the surface wind stress and surface slope as parameters.

By vertically integrating the momentum equation, by using the vertically integrated continuity equation, and introducing a stream function ψ, defined by $M_x = \partial\psi/\partial y$ and $M_y = -\partial\psi/\partial x$ where M_x and M_y are the vertically integrated velocities in the x and y horizontal directions, one can obtain a single equation for the stream function, which is

$$\nabla^2\psi + g_1 \frac{\partial\psi}{\partial x} + g_2 \frac{\partial\psi}{\partial y} = g_3 \qquad (1)$$

Here g_1, g_2, and g_3 are functions of the depth and its derivatives and g_3 is also a function of the wind shear stress. Once ψ is known, the three components of the velocity can readily be found.

For details of the derivation of the above equation and a discussion of the approximations involved in the derivation, see Gedney (1971) and Gedney and Lick (1970). Gedney (1971) has shown that the approximations used, except for the approximation of a constant vertical eddy viscosity, induce at the most a small, $0\ (10^{-1})$, or very local error in the calculations. The error induced by assuming the vertical eddy viscosity is constant cannot easily be estimated a priori. A comparison of the calculated currents with measurements is necessary in order to determine the magnitude of the error in this assumption.

The region of Lake Erie for which the above stream function equation must be solved is multiply connected because of the islands in the Western Basin. The value of the stream function on the mainland shore can be specified within a single arbitrary constant as determined by the river inflows and outflows. The value of the stream function for the island boundaries is not known a priori. However Gedney (1971) has shown that if an incorrect value of an island boundary stream function is chosen, the condition that the surface elevation be continuous, i.e.,

$$\oint \frac{\partial \zeta}{\partial s}\, ds = 0 \tag{2}$$

will not be satisfied. In the present calculations for Lake Erie three islands were incorporated. The islands were Pelee, Kelley, and a single Bass island. The single Bass island was formed by neglecting the narrow channels between the South, Middle, and North Bass Islands. The value for the stream function on each island boundary was uniquely determined by summing four solutions, i.e.,

$$\psi = \psi^{o} + d_1\psi^{1} + d_2\psi^{2} + d_3\psi^{3}$$

The boundary and wind conditions for these four solutions are shown in table 1. In table 1, $\psi_r(s)$ is the external shore boundary which specifies the river inflows and outflows, and τ_x and τ_y are the wind shear stresses in the x and y horizontal directions.

The d_1, d_2, d_3 constants were then determined so that the line integral, equation (2), around each island was satisfied. The summed solution specifies the proper stream function value on each island boundary.

Equation (1) was solved by finite difference methods. At all grid points, a five point central difference equation was used. Lake Erie was divided into two regions. One was an island region as shown in Figure 1 in which a 0.805 kilometer (0.5 mile) square grid was used. The boundaries in this region were approximated by taking the closest

0.805 kilometer grid point as the boundary. The second region was composed of the remainder of Lake Erie and in this region a 3.22 kilometer (2 mile) square grid was used except for the points adjacent to the boundary. For these points, nonsymmetrical difference equations were written which used the actual distance from the grid point to the boundary. The island region contained some 2800 grid points and the rest of the lake contained some 2250 grid points. Figure 2 shows the island region in relation to the entire lake as well as the bottom topography used used in the calculations.

The Lake Erie bottom depth at the regular grid points was determined by curve fitting irregularly spaced data taken from the U.S. Lake Survey Charts which are published by the U.S. Army Corps of Engineers. The bottom topography determined from these curve fits is very representative of the actual lake and contains considerable amount of detail. When a rapid change occurred in the bottom topography over one or two grid spaces, the topography was locally smoothed by a least squares smoothing program. This was done because the numerical solution could not accurately predict the effect of the local rapid change for the grid spacing used.

A combination of successive over-relaxation by points and lines was used to solve the system of finite difference equations. A complete iteration was performed in one region. Then from these new values, interim interface values between the two regions were formed for the second region. A complete iteration was then performed in the second region. From these new values in the second region, interface values were in turn formed for the first region. The process was repeated until the maximum relative error between successive iterations at any grid point was less than 10^{-5}. The grid points where the interface values between the two regions were formed are shown in Figure 1.

To evaluate equation (2), 3.22, 1.61, and 0.805 kilometer grid sizes were tried in the island region. It was found that the 0.805 kilometer grid size was necessary in order to obtain consistent and accurate line integrals around each island. As can be seen in Figure 1, the 0.805 kilometer grid size is also necessary to accurately represent the island boundaries.

Complete details on the method of solution can be found in Gedney (1971).

Results of Numerical Calculations

The numerical solution for the stream function and velocities in Lake Erie for the two region configuration was obtained for friction depths of 18.2 meters (60.0 ft) and 27.4 meters (90.0 ft). These depths correspond respectively to winds of 5.2 meters/second (11.8 mph) and 10.1 meters/second (22.7 mph). Here the friction depth is $\pi\sqrt{2\nu/f}$ where ν is the kinematic viscosity and f is the Coriolis parameter. The winds quoted are those at 6 meters above the water. The shear stress used in the calculations was determined by

$$\tau^W = 0.00273\ \rho_a W_a^2$$

where ρ_a is the air density and W_a is the wind velocity. The 0.00273 drag coefficient was determined by Wilson (1960) by analyzing wind shear stress measurements made by 47 different experimenters. In the calculations, the wind was assumed to be uniform over the entire Lake Erie surface. The uniform wind condition does not occur all the time on Lake Erie but it was found to be a good approximation for the periods for which the calculations and measurements were compared. The 18.2 and 27.4 meter friction depths correspond, respectively, to eddy viscosities of 16.8 and 38.0 cm^2/sec. These friction depths were determined to be the proper ones for the 5.2 and 10.1 meter per second winds because they provided the best agreement with current meter measurements.

All the results shown herein include a Detroit River inflow of 5380 m^3/sec and an equal outflow via the Niagara River. Other calculations have shown that the remaining rivers only modify the flow locally near the mouth of the rivers for moderate winds such as used here.

Figure 3 shows the stream function plot for a West 50 degrees South wind at 10.1 meters per second. This plot shows an upper clockwise and lower counterclockwise integrated mass flux gyre in both the Central and Eastern basins. In the Eastern basin, the upper cell has been deformed into two connected cells due to the peculiar geometry effect. The two cell configuration in each basin is essentially due to the dish (bowl) shape of the bottom topography.

The complete lake velocity plots for the West 50 degrees South wind are shown in Figure 4. In this figure, plots a through c give the horizontal velocities, respectively, at a constant 0.4, 9.9, and 14.9 meters from the surface. Plot d gives the horizontal velocities at a constant 1.2 meters from the bottom. Plot e gives the vertical velocities at mid-depth. The beginning of the arrow represents the actual location of the current represented by the arrow. The magnitude of the velocity can be determined from the velocity scale indicated on the figure. Note that the velocity scale is different from plot to plot. In plot e, arrows pointing toward the top of the plot represent the vertical velocities toward the lake surface.

Figure 4(a) shows that a top-surface mass flux is being transported toward the eastern and southern boundaries. A subsurface current returns the surface mass flux in the opposite direction as shown in Figures 4(b) through (d). In the Central and Eastern basins, surface currents at 0.4 meter are in general smaller in the center of the lake than near the shore. This effect is essentially due to the relatively large subsurface return current down the center of the lake which is opposite in direction to the surface current and subtracts from it.

The mass flux gyres shown in Figure 3 are confirmed by the Figure 4 velocity plots. The flow at all levels at many locations at the south shore is essentially parallel to the shore in an easterly direction. The bottom currents shown in Figure 4(d) indicate where these parallel shore currents occur. The width of this band of parallel flow near the bottom is only 3 to 4 kilometers wide. In the Eastern basin, the depths are great enough near the shore so that at some locations the bottom currents are more perpendicular than parallel to the shore. The near parallel shore currents on the south shore of Long Point and then the abrupt northern flow at the tip of Long Point are in agreement with the mass flux gyre shown in Figure 3. The location of Long Point is shown on Figure 2.

More detailed plots of the horizontal currents in the Western basin for a West 50 degrees South wind at 10.1 meters per second are shown in Figures 5(a) and 5(b). These plots are for depths of 4.5 and

7.6 meters, respectively. The effect of the inflow at the mouth of the Detroit River is particularly evident in Figure 5(a). As can be seen, most of the Detroit River through flow is going north of Pelee Island across the tip of Pelee Point. The subsurface return flow from the lower half of the Central basin prevents the Detroit River flow from taking a southward path through the islands. This flow blockage produces the clockwise gyre below Pelee Point which is evident in Figure 5(a). At the 6 meter level and below, a large portion of the flow in Pelee Passage goes directly south to a point where it is turned either into the Central basin or toward Pelee Island. This southerly flow is caused by an underwater ridge which extends south from Pelee Point. The underwater ridge also prevents the subsurface return flow in the northern part of the Central basin from entering the Western basin. As a result, the clockwise gyre is formed east of the ridge at 6 meters and below. The effect of the flow blockages at the Pelee Point underwater ridge and at the islands are also reflected in the vertical velocities shown in Figure 4(g). This figure shows that the blockage creates significant areas of upwelling. It should be noted that the path that the Detroit River through flow takes is highly dependent on the wind direction. For a northwest wind most of the through flow takes a southern path between the islands.

The detailed velocities in the island region for the West 50 degrees South wind case are plotted in Figures 6(a) and 6(b). The results here are the same as already seen in the Western basin plots except in more detail. We see that all flow south of Kelleys Island is in a westerly direction. Very little flow goes southward between the Bass Islands and

Pelee Island. Some of the southerly flow through the Pelee Passage which is below 6.0 meters rises and goes over the Pelee Point underwater ridge. The fact that this does happen is indicated by the large easterly bottom currents on top of the ridge as shown in Figure 6(b).

The interested reader can find complete velocity plots at many more levels for the entire lake, Western Basin, and island region for a variety of wind directions in Gedney (1971).

Comparison of Lake Erie Measurements with Numerical Calculations

In an effort to determine the prevailing circulation patterns in Lake Erie, the U.S. Federal Water Pollution Control Administration (now part of the Environmental Protection Agency (EPA)) established a system of automatic current metering stations in Lake Erie in May 1964. The metering program was maintained until September 1965. The reduction of data was made by the EPA for part of this period but was never published in detail. Fortunately, the EPA loaned the reduced data to the authors and it is this data that we now compare with our calculations.

The EPA data consisted of both atmospheric wind and water current velocity readings taken every 20 or 30 minutes. In order to compare the water current measurements with our calculations, the 20 and 30 minute readings were vectorially summed over a 24-hour period to form a resultant current or wind. The resultant magnitude is equal to the vector sum magnitude divided by the numbers of readings. The time periods used for comparison purposes were those for which the wind was fairly steady for 2 or more days so that any seiche currents were small in magnitude. The vector summing of the data over a 24-hour period should remove part, if not all, of any small seiche current since the largest seiche period is 14 hours.

The wind data available from the EPA lake stations was for only certain limited periods. Therefore, it became necessary to use the 24-hour resultant winds recorded at Cleveland, Erie, and Buffalo by the U.S. Weather Bureau. The U.S. Weather Bureau 24-hour resultant winds at these land stations agreed fairly well in direction with the EPA resultant wind determined at the various lake stations. However, the "over-the-lake" wind magnitudes were on the average 1.48 times the corresponding land values.

The procedure incorporated for determining what wind to use in the numerical calculations for a particular day was to take the average of both the direction and magnitude of the 24-hour resultant wind at Cleveland, Erie, and Buffalo as published by the U.S. Weather Bureau. This average wind with its magnitude increased by 1.48 was then used to determine the shear stress at the water surface. On May 24, 1964, the resultant wind was determined to be 10.1 meters per second with direction West 50 degrees South. The resultant winds for the prior 2 days had been within 20° of this and at somewhat less magnitude. However, on May 21, 1964, the resultant winds were approximately East 50 degrees South at roughly 1 meter per second which is significantly different from the wind on May 24th. The current meter data for May 24, 1964, as measured at 10 and 15 meters below the surface is shown in Figures 4(b) and (c). Note that the position of the measurements are different from those of the calculated current vectors. The agreement is markedly good in both magnitude and direction. The discrepancy between the magnitude of the measurements and calculations at point A at 10 meters is believed to be a measurement error since this meter became erratic at a later date. The magnitude of the measurements in the Long Point region may at first appear to be considerably different from the calculated values. However, the agreement between measurements and calculations is believed to be satisfactory when one considers that the currents are changing rapidly with distance in the Long Point area. It should be noted that data from all the stations are not shown. This is because either the data were not available from the EPA or the data showed the meters did not respond to wind changes.

Also, in Figures 4(b) and (c), the meter measurements are plotted for a West 43 degrees South wind at a velocity of 8.6 meters per second. These measurements were taken on October 25, 1964. The agreement is again quite good except for the measurement at point B in Figure 4(c). The small value of this reading may indicate a measurement error. The winds for these measurements differ from the May 24, 1964 case by 1.5 meters per second and 7^{o} in direction. However, both sets of measurements agree quite well with the calculations and each other. The May 24, 1964, data was taken during a period of heating which tends to suppress turbulence while the October 25, 1964, data was taken during a period of cooling which increases turbulence. It is possible that the difference in wind properties may be offset by a difference in eddy viscosity. Other reasons may also be responsible for the seemingly same results for the two slightly different cases. In any event this serves to illustrate the difficulty in establishing precise values for the eddy viscosity (friction depth) from measurements.

In Figure 7(a) and (b), the meter measurements are plotted for a

West 67 degrees South wind at 4.8 meters per second (dark arrow) and a West 67 degrees South wind at 5.2 meters per second (dashed arrow).

The current calculations were made using a wind velocity of 5.2 meters per second. The first set of measurements were taken on July 16, 1964, when the thermocline should have been well established in the Central and Eastern basins. The second set of measurements were taken on September 10, 1964, when the thermocline was very near the bottom of the central basin and was quite weak. In the Central basin, the thermocline is usually gone by October 1st while, in the Eastern basin, the thermocline disappears sometime later. The measurements in the Central basin agree fairly well with the calculations which may indicate that the steady state currents in the Central basin upper epilimnion are not appreciably affected by the lake stratification during the middle of July and September. However, some of the fairly large measured currents in the Eastern basin which differ from the calculations may be indicating a thermocline effect. More recent measurements made by the EPA in the Central basin in later July and early August do indicate an appreciable effect of stratification on epilimnion velocities.

The automatic current meters used for making the measurements quoted above were set in late May and were left unattended after that time. Because of this, it is felt that the confidence level of the measurements made in late summer and fall is less than the earlier measurements. This factor should also be taken into account when assessing the agreement between the measurements and calculations.

A point which experimenters making new measurements in Lake Erie should consider is well illustrated by the currents shown in Figure 4. This is that the location of the instrument both horizontally and vertically must be well known. As we see, there are areas in Lake Erie where the currents change rapidly with horizontal position. In addition, in the top ten meters of the lake the currents are also changing rapidly in the vertical direction.

CONCLUDING REMARKS

The results of the numerical steady-state calculations of the wind driven current in Lake Erie compare favorably with current meter measurements made at mid-depth in the Central and Eastern basins. This agreement shows that quasi-steady currents do exist in Lake Erie. The quasi-steady currents usually occur after 2 days of fairly uniform winds.

The agreement of calculations and measurements also shows that the shallow lake model which uses a constant eddy viscosity is capable of predicting accurately the local three-dimensional velocities at mid-depths. Current meter measurements near the bottom and surface of the lake in conjunction with measurements at mid-depths are needed to really determine the error that may result from the constant eddy viscosity assumption. The calculated velocities near the bottom of the lake, where the eddy viscosity is known theoretically to reduce to some small value, must especially be checked by measurements.

No current meter measurements were available in the Western basin. Measurements are needed here to determine the importance, if any, of time dependent and nonlinear convection effects in producing quasi-steady currents. Measurements are also needed to determine if a smaller eddy viscosity should be used in the Western basin because of its extreme shallow depth.

The calculations predict that the flow from the Western basin to the Central basin takes place mainly via the Pelee Passage when the winds have a southwest direction. The blockages of flow that occur between the Western and Central basins are reflected in an extensive gyre east of the islands. These and other flow patterns such as the strong currents at the tip of Long Point are the results of the bottom topography and shape of the shore boundaries. The bottom topography and boundary shape influence the currents in Lake Erie so strongly that they must be accurately represented in any model predicting the currents. For the numerical calculations, the relatively small mesh sizes of 0.805 and 3.22 kilometers had to be incorporated to adequately include the Lake Erie geometry. Improvements to the model could be made by using even smaller mesh sizes for certain local areas.

Eddy viscosities of 16.8 and 38.0 centimeters squared per second were used in the calculations, respectively, for wind velocities of 5.2 and 10.1 meters per second. These values are the same order of magnitude as measurements made by Nobel, et al. (1960) in Lake Michigan. Platzman (1963) used 40 centimeters squared per second for his time dependent calculations of wind tides on Lake Erie. This value cannot really be compared with the ones used here since Platzman used the single value for a range of wind speeds. A wind shear coefficient of 0.00273 seems to give good results.

The calculated results show that for a large percentage of Lake Erie the bottom velocities are not in the direction of the horizontal mass flux. This indicates that for shallow Lake Erie, there may be large differences between the solution presented herein and one where the bottom stress is made proportional to the horizontal mass flux.

Based on measurements, Hartley (1968) believes that the steady state velocities in the hypolimnion of Lake Erie's Central basin are very small and the thermocline which is almost horizontal acts as the summer lake bottom. Since the hypolimnion in the Central basin has a maximum thickness of 6 meters, one might expect the upper epilimnion velocities not to differ appreciably from the nonstratified case. Comparison of calculations with mid-depth measurements made during the summer are inconclusive as to whether or not this is true. Eastern basin measurements however, seem definitely to indicate that the thermocline has significant effect on upper epilimnion velocities.

ACKNOWLEDGMENTS

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TABLE 1. - STREAM FUNCTION BOUNDARY CONDITIONS

ψ^j solution	ψ^j boundary condition				Wind condition
	External shore	Island 1	Island 2	Island 3	
ψ^o wind only	0	1	1	1	τ_x, τ_y
ψ^o river only	$\psi_r(s)$	1	1	1	0
ψ^o wind and river	$\psi_r(s)$	1	1	1	τ_x, τ_y
ψ^1	0	1	0	0	0
ψ^2	0	0	1	0	0
ψ^3	0	0	0	1	0

FIGURE LEGENDS

NUMERICAL CALCULATIONS OF THE WIND-DRIVEN CURRENTS IN LAKE ERIE AND COMPARISON WITH MEASUREMENTS BY R. GEDNEY AND W. LICK

Figure	
1	0.805 kilometer (0.5 mile) grid for island region.
2	3.22 and 0.805 kilometer grid region for Lake Erie.
3	Lake Erie stream function for a W50S wind at 10.1 meters/sec.
4(a)	Horizontal velocities at a constant 0.4 m (1.5 ft) from surface.
4(b)	Horizontal velocities at a constant 9.9 m (32.8 ft) from surface.
4(c)	Horizontal velocities at a constant 14.9 m (49.2 ft) from surface.
4(d)	Horizontal velocities at a constant 1.2 m (4.0 ft) from bottom.
4(e)	Vertical velocities at mid-depth.
5(a)	Horizontal velocities at a constant 4.5 m (15.0 ft) from surface.
5(b)	Horizontal velocities at a constant 7.6 m (25.0 ft) from surface.
6(a)	Horizontal velocities at a constant 4.5 m (15.0 ft) from surface.
6(b)	Horizontal velocities at a constant 1.2 m (4.0 ft) from bottom.
7(a)	Horizontal velocities at a constant 9.9 m (32.8 ft) from surface.
7(b)	Horizontal velocities at a constant 14.9 m (49.2 ft) from surface.

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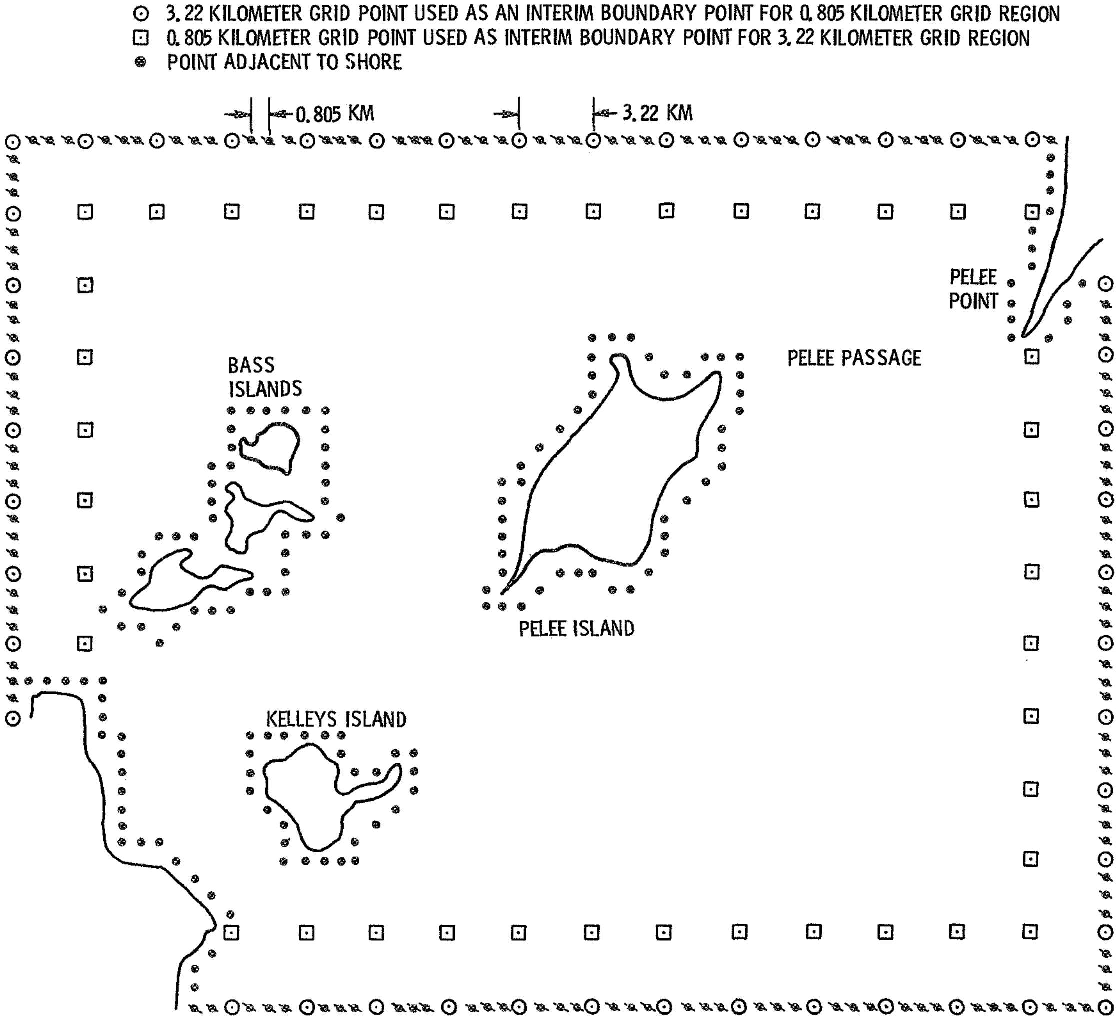

FIGURE 1. - 0.805 KILOMETER (0.5 MILE) GRID FOR ISLAND REGION.

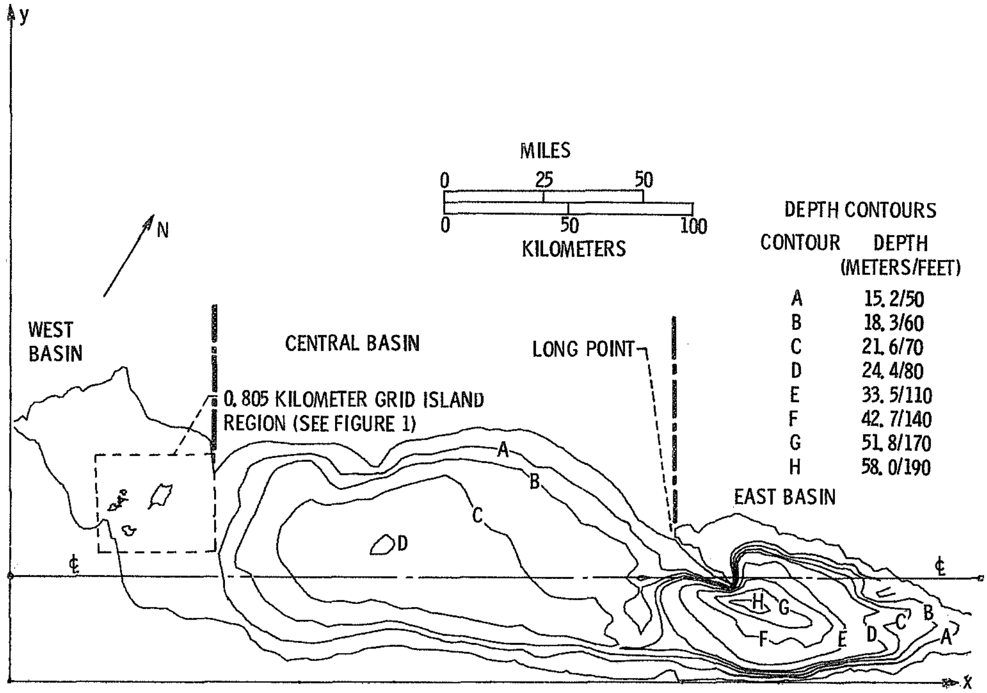

FIGURE 2. - 3.22 AND 0.805 KILOMETER GRID REGIONS FOR LAKE ERIE.

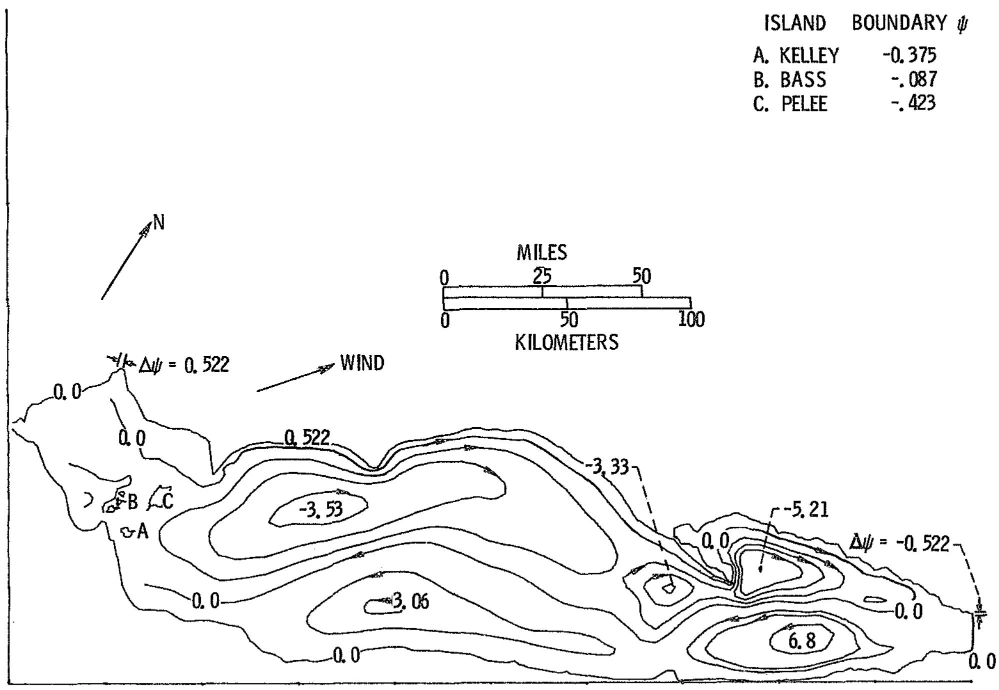

FIGURE 3. - LAKE ERIE STREAM FUNCTION FOR A W50S WIND AT 10.1 METERS/SEC.

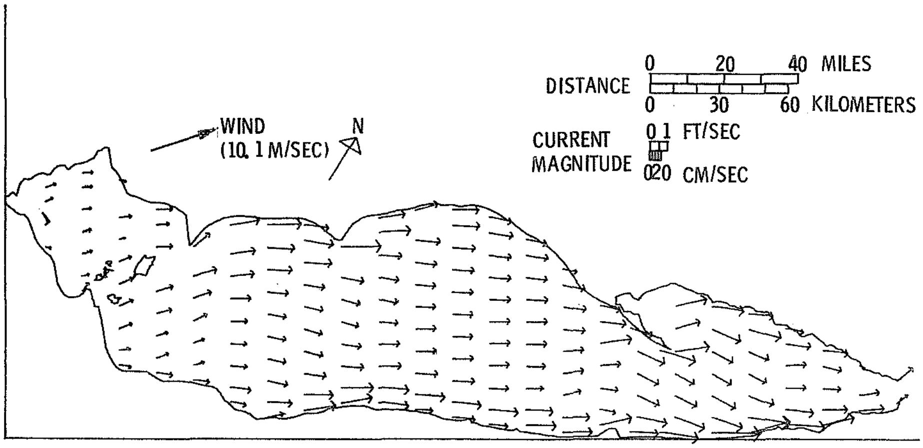

FIGURE 4(a). - HORIZONTAL VELOCITIES AT A CONSTANT 0.4 M (1.5 FT) FROM SURFACE.

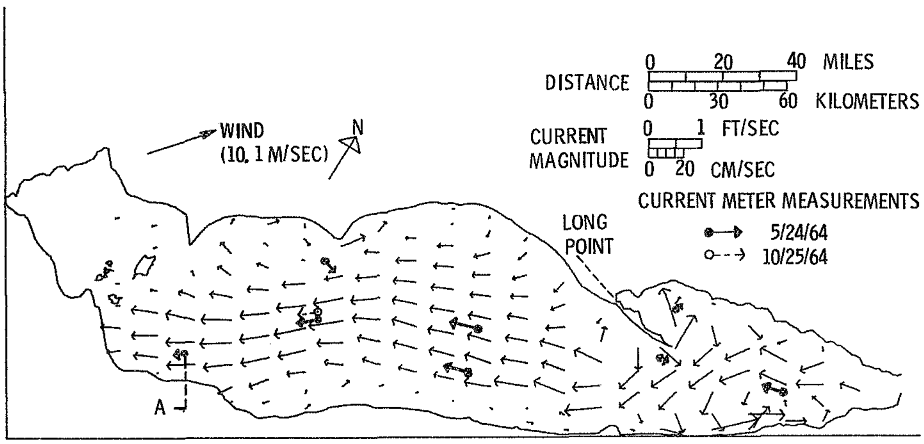

FIGURE 4(b). - HORIZONTAL VELOCITIES AT A CONSTANT 9.9 M (32.8 FT) FROM SURFACE.

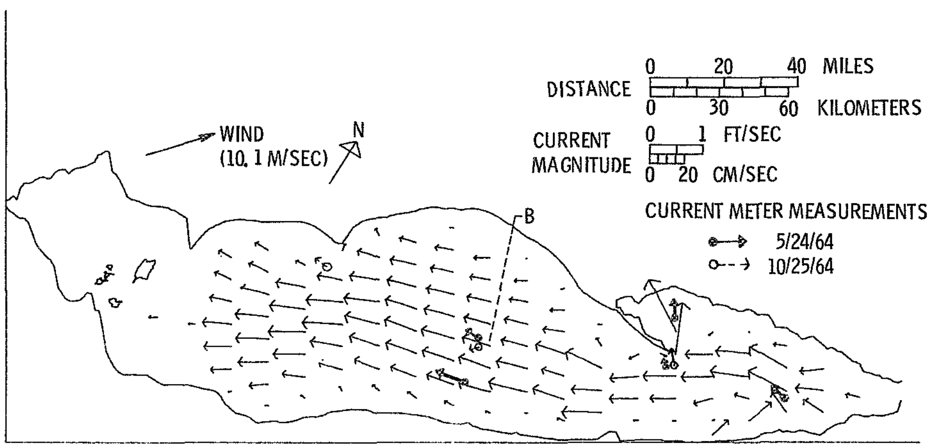

FIGURE 4(c). - HORIZONTAL VELOCITIES AT A CONSTANT 14.9 M FROM SURFACE.

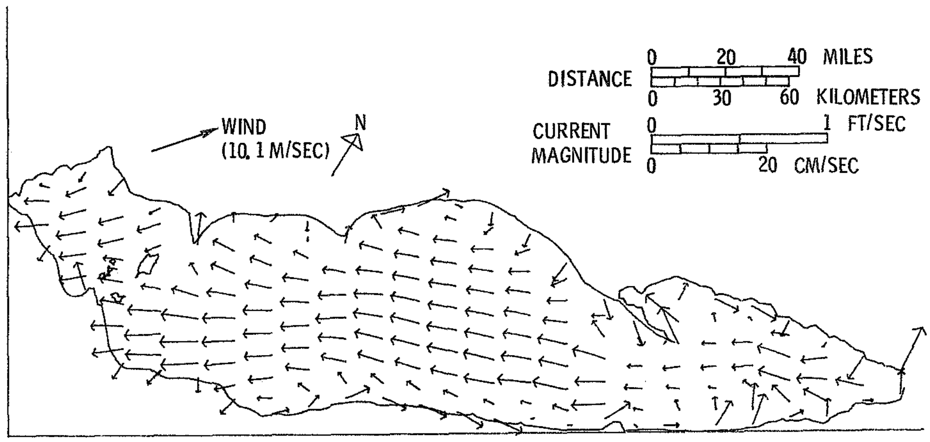

FIGURE 4(d). - HORIZONTAL VELOCITIES AT A CONSTANT 1.2 M (4.0 FT) FROM BOTTOM.

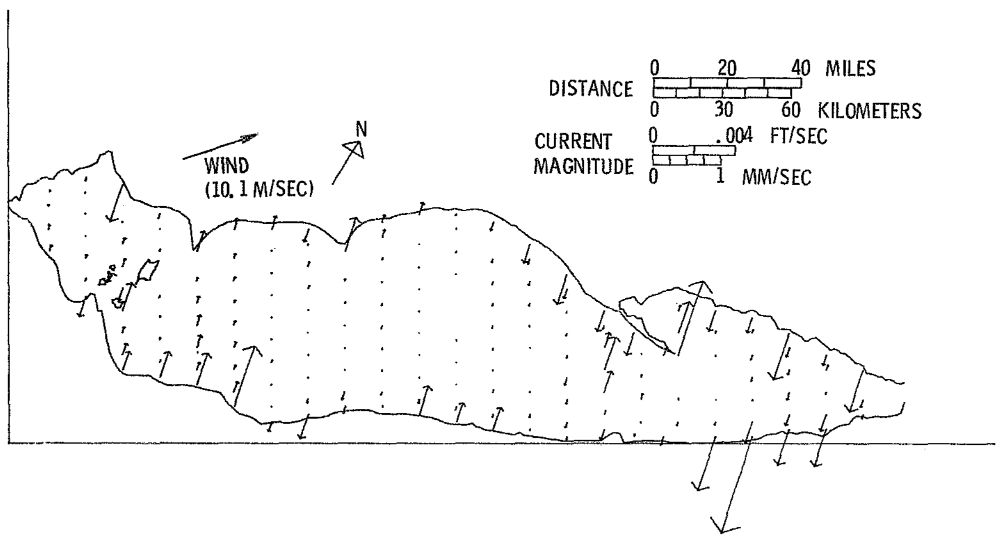

FIGURE 4(e). - VERTICAL VELOCITIES AT MID-DEPTH.

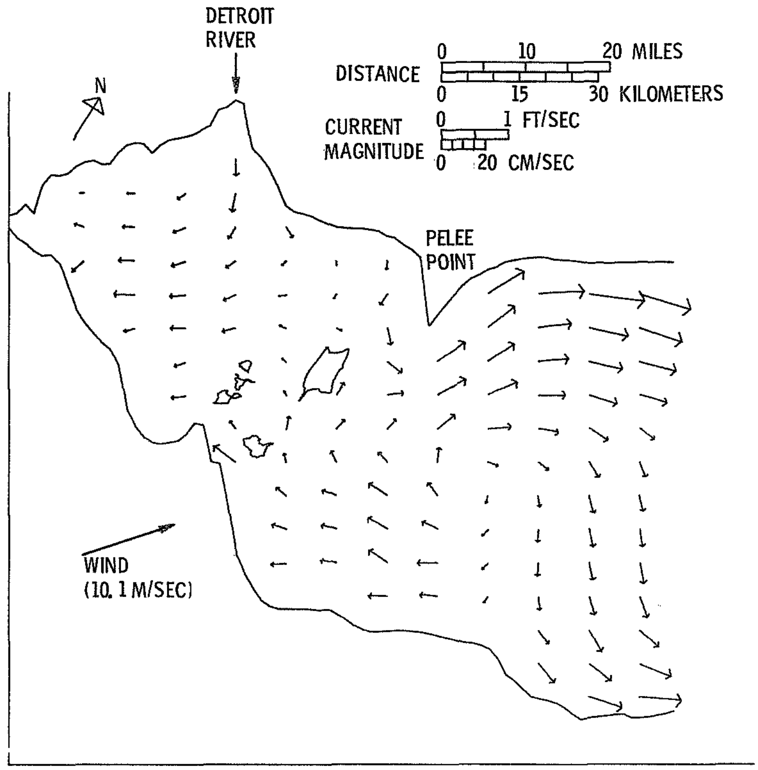

FIGURE 5(a). - HORIZONTAL VELOCITIES AT A CONSTANT 4.5 M FROM SURFACE.

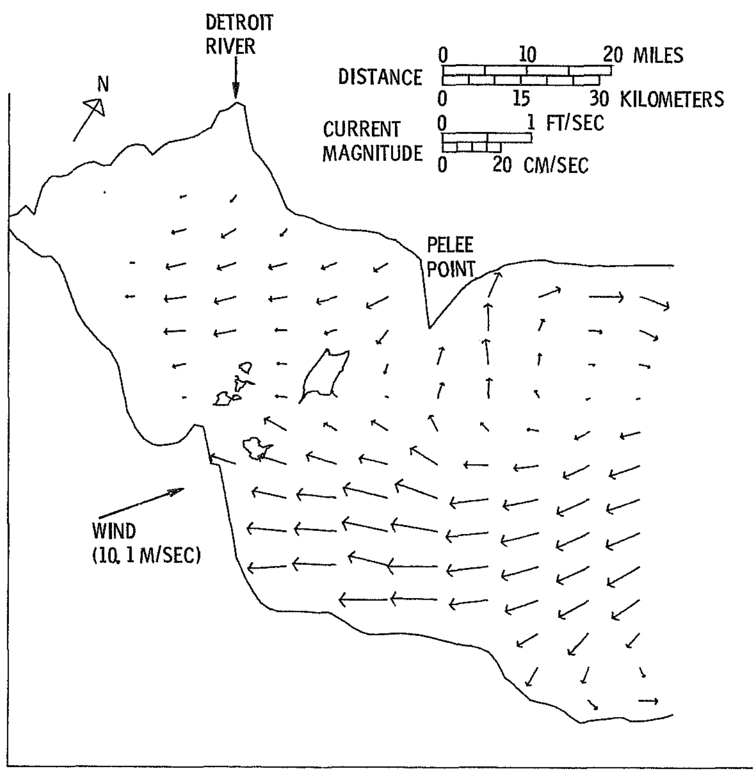

FIGURE 5(b). - HORIZONTAL VELOCITIES AT A CONSTANT 7.6 M FROM SURFACE.

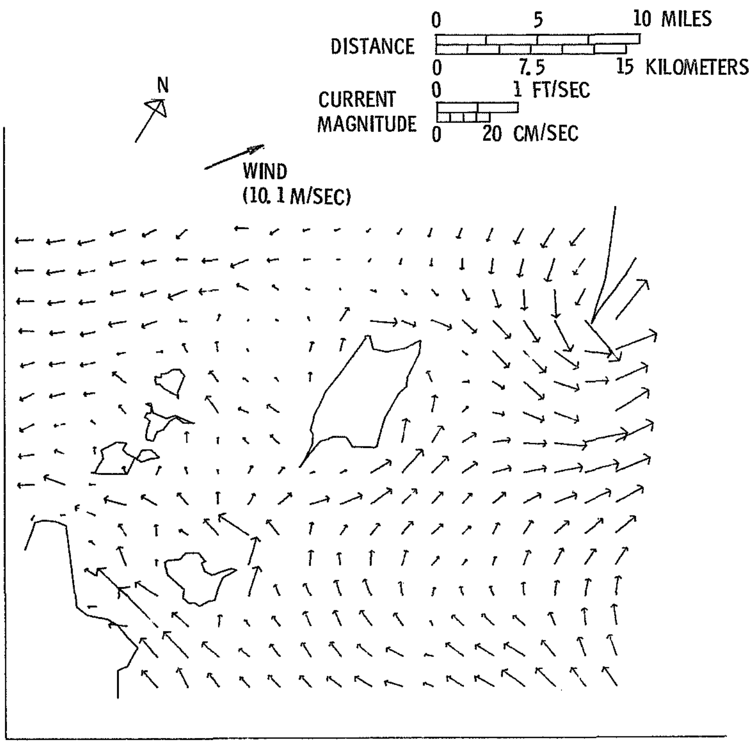

FIGURE 6(a). - HORIZONTAL VELOCITIES AT A CONSTANT 4.5 M FROM SURFACE. (SEE FIGURE 1 FOR IDENTIFICATION OF ISLANDS AND LANDMARKS.)

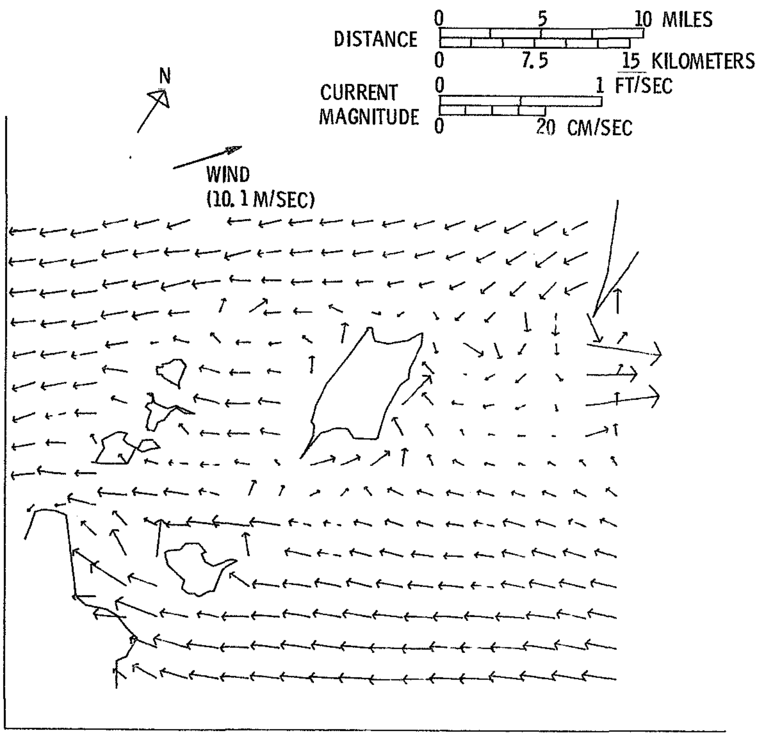

FIGURE 6(b). - HORIZONTAL VELOCITIES AT A CONSTANT 1.2 M FROM BOTTOM. (SEE FIGURE 1 FOR IDENTIFICATION OF ISLANDS AND LANDMARKS.)

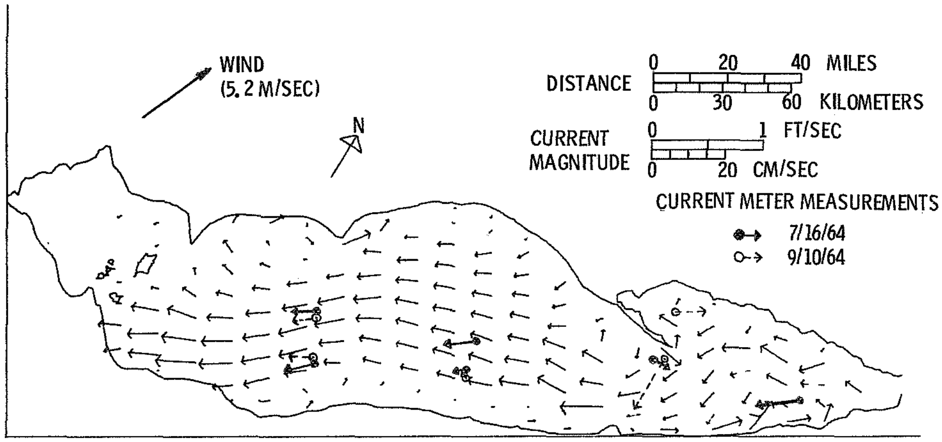

FIGURE 7(a). - HORIZONTAL VELOCITIES AT A CONSTANT 9.9 M FROM SURFACE.

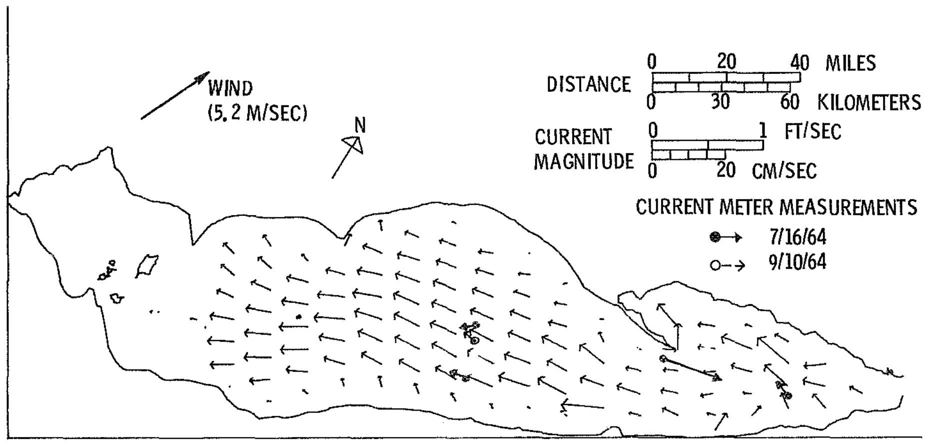

FIGURE 7(b). - HORIZONTAL VELOCITIES AT A CONSTANT 14.9 M FROM SURFACE.