

CURRENTS IN THE EASTERN MEDITERRANEAN (*)

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Ships have been ploughing through the Eastern Mediterranean since time immemorial.

The "sea-dogs" of olden times were certainly not unprovided with information about the currents in this region. However, literature on this subject has up till now been rather rare.

I. — SURFACE OBSERVATIONS

It is to NIELSEN (1912) that we owe the most commonly known picture of the currents in the Eastern Mediterranean. This consists of a West East circulation along the African coast, becoming approximately South-North off the Levantine states and taking finally an East-West direction starting from the Gulf of Iskenderun. This picture has been corroborated by the presence of waters of deltaic origin along the Israeli coast, as established by LIEBMAN (1930), as well as by the presence in Israel's coastal plain of alluvial deposits coming from the Nile (RIM, EMERY and BENTOR, EMERY and NEEV, NEEV). Furthermore, ROUCH (1945) attributes to the Nile the sudden decrease in salinity which he noted at Beirut after the floods of this river.

Nevertheless NIELSEN's point of view has had to be somewhat modified on account of later studies.

Indeed GRUVEL (1931) established the existence in the Eastern Mediterranean of counter-currents which he attributes above all to wind action.

Studying the influence of the Nile flood waters along the Israeli coast, OREN (1952) discovered two varieties of water developing side by side, each of different salinity and colouring. He found, moreover, that the least salt waters — i.e. those most affected by the Nile — are not necessarily to be found near the shores, and also that the line separating these two kinds of water moves away from the coast as it progresses northwards.

AKYÜZ (1957) in his study on hydrological conditions in the Gulf of Iskenderun shows that the surface circulation there forms two eddy units.

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In 1956 the *Calypso* Expedition directed by LACOMBE and TCHERNIA (1959) found currents with a southwesterly direction at its stations 173 and 174, as was also noted by EMERY and GEORGE (1963).

EMERY and NEEV (1960) in their study on the Israeli beaches draw attention to the North-South currents not far from shore — or rather, South-North currents that, following a semi-circular path, reverse their direction and become North-South. Simultaneously, the angle between the wave front and the coast progressively changes its orientation proceeding from the South to the North. Whereas this angle faces towards the North in the south of Israel it opens towards the South in the north of that country. These observations are moreover confirmed by coastal sand deposits. These accumulate from North to South above approximately latitude 32°, and in the opposite direction below this latitude.

EMERY and GEORGE (1963) when studying Lebanese beaches likewise notice that in the majority of their observational stations waves approach the shore from the Northwest in such a way that the angle made by the direction of the wave front and the shore generally faces South. South-North currents have only been observed by the authors in three border stations in the region of Tripoli.

Finally the Atlas Océanographique et Météorologique de la Méditerranée (1957), summarizing surface current directions for the whole extent of the grid coverage, does not indicate any particularly preferred direction in any grid square in the region.

The velocity of the currents seems seldom to rise above 6 - 12 miles per day (12 - 25 cm/s).

However, EMERY and NEEV (1960) in April 1959 found averages of about 50 cm/s in Israeli waters. This is also the magnitude assumed by GRUVEL for the Syrian countries.

On the contrary, HECHT (1964) adopts a value of 17 cm/s during the Nile's flood water period (September-October). At approximately the same time of year — in October 1962 — EMERY and GEORGE (1963) carried out velocity measurements off the Lebanon. They found predominant velocities of about 6 cm/s at the most, and a maximum of about 25 cm/s.

On the other hand OREN and KOMAROVSKY (1961) have summarized the recordings at the Ashdod station in Israel taken at a depth of 7 m using a Neyrpic currentmeter. The current velocities there varied from 0.00 cm/s to 21 cm/s, and the most frequent average was 10 cm/s.

II. — GEOSTROPHIC CIRCULATION

The general pattern of the geostrophic circulation in the region is shown in figures 2, 3 and 4.

These figures refer to the picture at the surface layer, and at depths of 20 m and of 500 m, respectively.

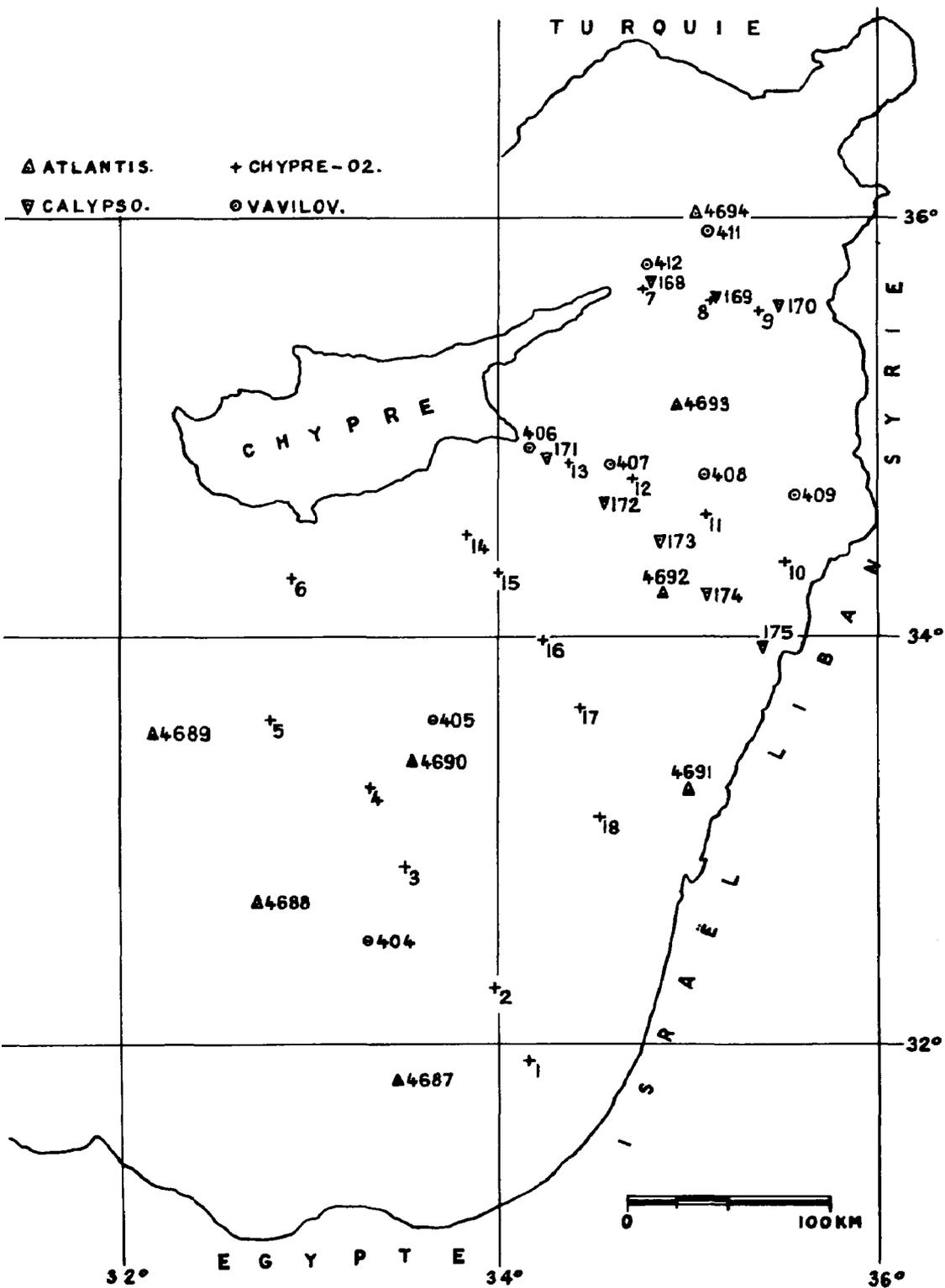


FIG. 1

Station de Recherches des Pêches maritimes à Caïffa (Israël)

Study of the Levantine Basin

Position Plan

Observations stations :

Atlantis. Chypre - 02.
 Calypso. Vavilov.

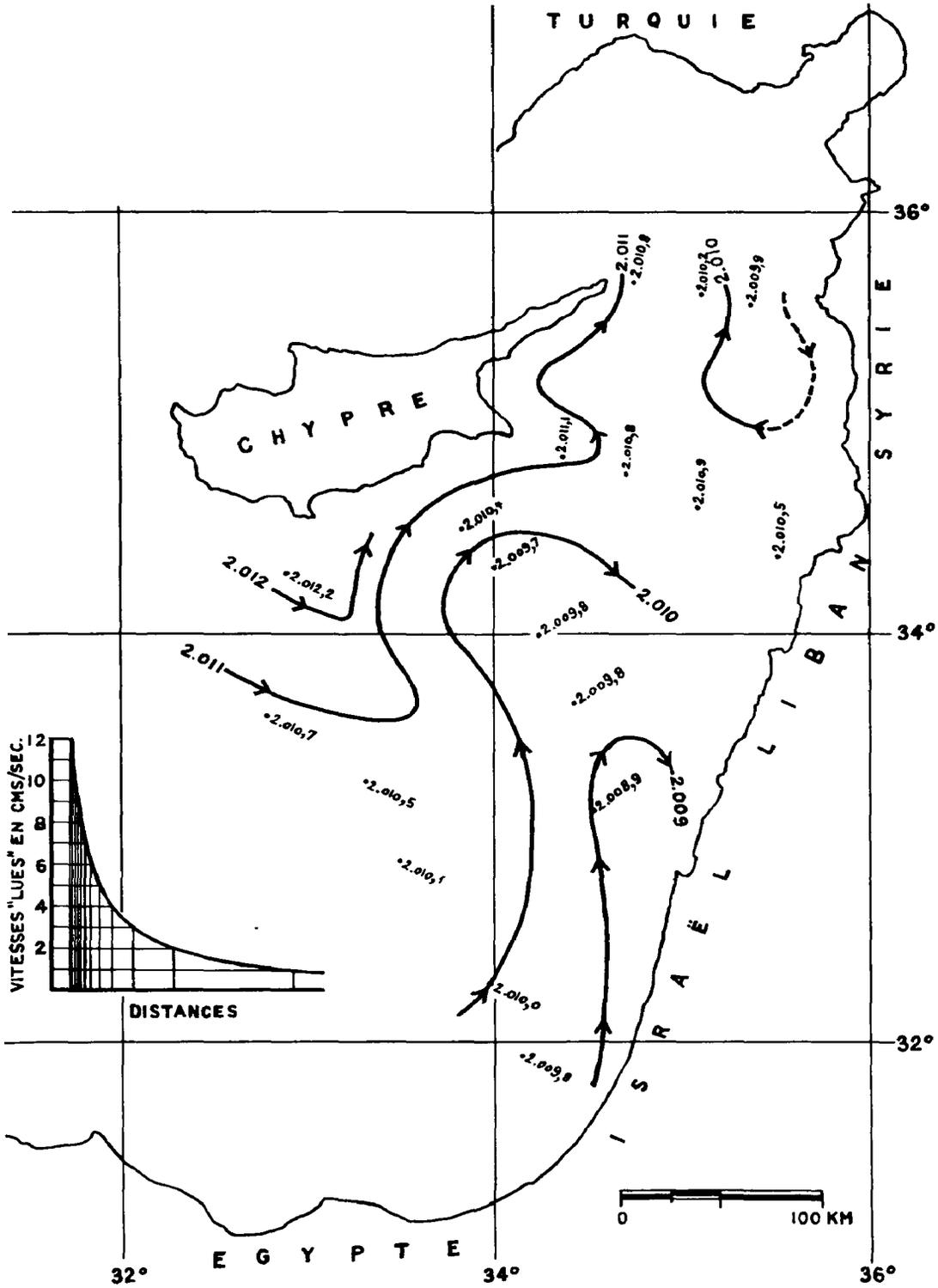


FIG. 3

Station de Recherches des Pêches maritimes à Caiffa (Israël)

Study of the Levantine Basin (Chypre - 02)

Contours of the pressures of the water column (in mb) at a depth of 20 m

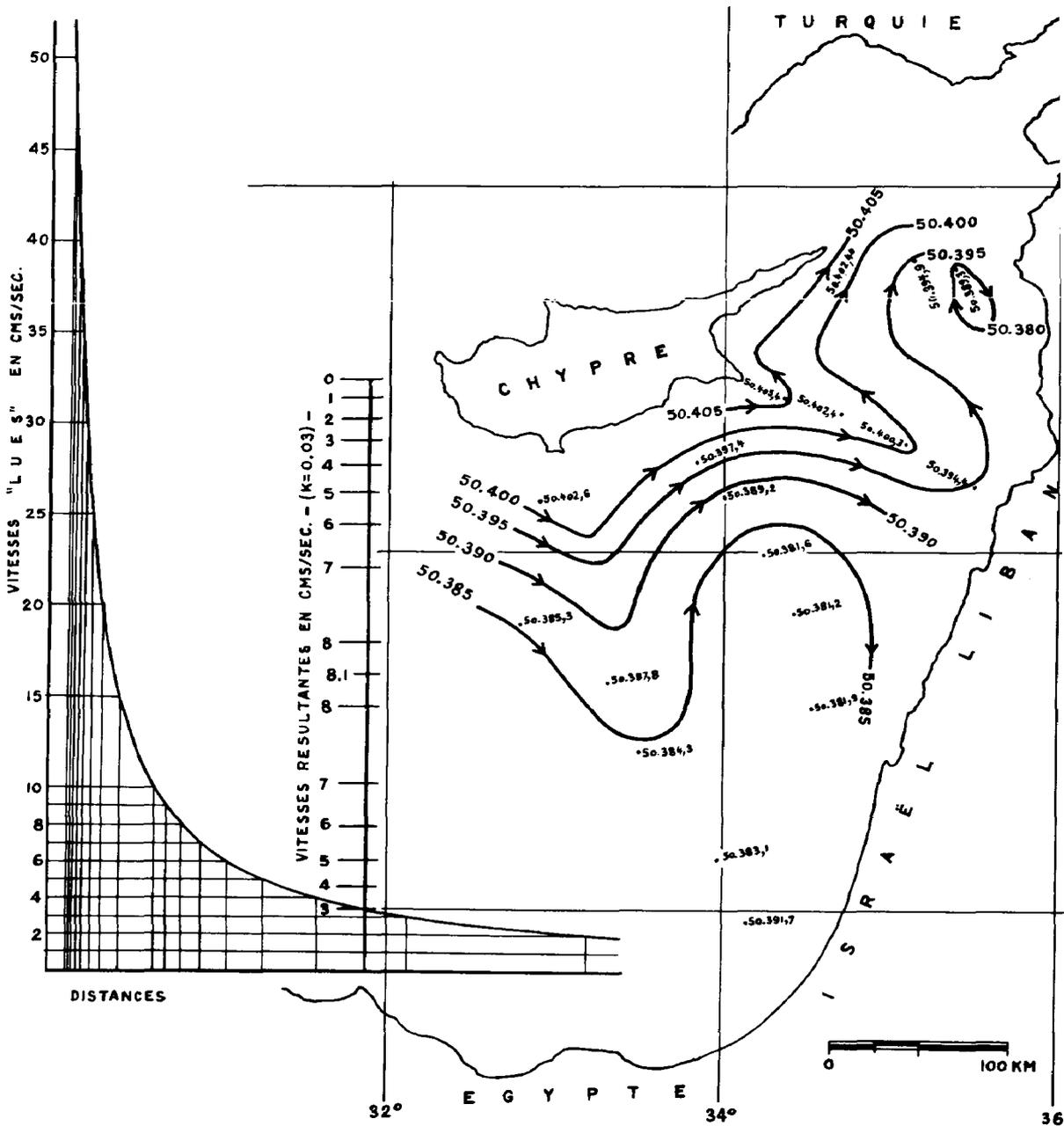


FIG. 4
 Station de Recherches des Pêches maritimes à Caïffa (Israël)
 Study of the Levantine Basin (Chypre - 02)
 Contours of the pressures of the water column (in mb) at a depth of 500 m

The representation of the geostrophic current in these figures is made by using the pressures in the hydrostatic column, with the sea surface as reference level.

The basic data for drawing up these figures were obtained from the samples taken during the "Chypre-02" project of the Haifa Fisheries Research Station in Summer 1963. Under the supervision of O.H. OREN 18 stations, as indicated in figure 1, for conventional measurement of temperature and salinity were occupied.

Computational technique

It is well known that the geostrophic current can be defined by considering the pressure differences between two points.

Charts can be drawn up, using the known pressure distribution within a sea, which will show the current lines, their directions, and data on current velocity.

a) Computation of hydrostatic pressure

Contrary to the usual practice of using a deep reference level, here the sea surface was chosen for this purpose.

This procedure was at first adopted not only because the Mediterranean Sea in the summer months has a calm surface but also because we do not know the depth of a plane surface in full equilibrium. Later on it will be seen that our assumptions were well founded.

Numerical results were obtained for layers equal in depth to the distances between two standard depths, applying the formula :

$$p_z = p_0 + g \rho_m h \quad (1)$$

where :

p_z = the total pressure of the hydrostatic column at z depth and at the base of the layer being considered;

p_0 = the pressure at the top of this layer;

ρ_m = the mean value for densities at the top and the bottom of this layer;

$$\rho_m = \left(\frac{\rho_0 + \rho_z}{2} \right)$$

h = the height of the layer;

g = the gravity value at the station's latitude φ and at the z depth at the bottom of the layer. The values of g have been computed by first of all applying the Helmert formula (SVERDRUP-JOHNSON-FLEMING, p. 404) :

$$g_0 = 9.80616 (1 - 0.002644 \cos 2\varphi + 0.000007 \cos^2 2\varphi),$$

then the formula :

$$g_z = g_0 + 2.202 \cdot 10^{-6} z.$$

With the value of hydrostatic pressure thus obtained at each station, isobaric contours were then drawn. Figures 3 and 4 give examples for depths of 20 m and 500 m respectively.

For the isobars in figure 2 the formula expressing pressures (1) was reduced to the terms :

$$p = g_0 \cdot \rho \cdot \text{unit of height} \quad (2)$$

with

- p = the pressure of the hydrostatic column;
- g_0 = the gravity value at the sea surface;
- ρ = the density of sea water, sampled at the surface.

In the results, calculated by means of the above formula, by taking account of decimals for the height instead of the unity figures we arrive at a picture of the geostrophic circulation on the surface. This is represented in figure 2 which in addition offers the possibility, in identical conditions, of a comparison with the theoretical contours, such as those of B. SAINT GUILY (1961).

It is obvious that current lines resulting from the above computations can differ from the actual current lines as a result of atmospheric conditions. We may note in this respect that the average variation in atmospheric pressure between the Island of Cyprus and the Levantine coast rises in summer to about 1 - 2 millibars. This variation is comparable in size to the variations in hydrostatic pressure obtained at the same places at a depth of 20 m. However the size of these differences in atmospheric pressure seems insufficient to affect appreciably the distribution of pressures at a depth of 500 m.

Moreover, a comparison of figures 2, 3 and 4 shows that the aspect of the isobars remains approximately the same from the surface down to considerable depths. It follows first of all that the distribution of hydrostatic pressure depends at all depths on the same causes, and then that this distribution is independent of temporary atmospheric conditions, the densities being more widely distributed in time.

On the other hand by again comparing these figures we see that they show clearly that the distribution of hydrostatic pressure is liable to alteration on account of bottom conditions. The bathymetric chart of GIERMANN (1960) shows, in fact, that there exists a sill in the neighbourhood of the line between Cape Andreas in Cyprus and Latakia in Syria. This sill limits any possibility of circulation towards the North beyond a depth of 550 m. The isobars then take the opposite direction, i.e. towards the South. This fact is already apparent in figure 4 for a depth of 500 m. Thus with increasing depths the pressures would become more and more distributed according to rings corresponding to the bottom topography.

b) *The study of velocities*

SVERDRUP-JOHNSON-FLEMING (p. 391) state :

$$v = g \frac{\Delta p}{1.458 \cdot 10^{-4}} \cdot \sin \varphi \quad (3)$$

where :

- v = the current velocity;
- g = the mean gravity value for the place under consideration;
- Δ_p = the difference in hydrostatic pressure between the two given points;
- φ = the mean latitude for the place.

The nomograms given in the margins of figures 3 and 4 were established using this formula.

It is well known that velocities thus computed have only a relative value, and moreover that they are affected by the forces of friction.

Thus the *virtual viscosity effect* in its most simple expression is reduced to the relation (DEFANT, p. 314) :

$$R = -k\rho v^2 \quad (4)$$

where :

- R = the resultant force of friction expressing the loss of velocity;
- ρ = the density of water;
- v = the velocity of flow;
- k = the coefficient of virtual viscosity.

However, the numerical application of this relation is restricted beforehand to the problem of the determination of k , the coefficient of virtual viscosity.

In fact the sole relevant value available is that given by DEFANT (p. 317), from the work of TAYLOR, namely

$$k = 2.6 \cdot 10^{-3}.$$

In the above formula this value of k restricts the current velocities to about 95 cm/s.

However, TAYLOR's value is valid for particular conditions, and its author warns us that the value could change for other cases and could increase up to 100 times this value.

This means that this coefficient's value must be sought by means of computation.

To this end we note that the relation used by TAYLOR is similar to the equation of Eckmann (DEFANT, p. 420). This last, expressing the interaction of wind and sea circulation, is written :

$$T = 2.6 \cdot 10^{-3} \cdot \rho' \cdot w^2 \quad (5)$$

where :

- T = the force resulting from wind friction on the sea surface;
- ρ' = the density of air;
- w = the velocity of the wind expressed in cm/s.

This formula has been checked and found valid for wind velocities of 20 m/s and more.

On the other hand the average for relative velocities, deduced from the distance between isobars, for a depth of 500 m is of the order of 20 cm/s.

Thus taking the scale ratio into account, we obtain for the ratio of the above equations :

$$\frac{T}{R} = \frac{2.6 \cdot 10^{-3} \cdot 1.29 \cdot 10^{-3} \cdot 4.10^{+6}}{2.6 \cdot 10^{-3} \cdot 1.02922 \cdot 400} \approx 10$$

We know also that the effect produced on an environment depends solely on the acting force and on the particular structure of this environment. In other words, the action of one kind of water on another kind, or of the wind, is the same provided that the forces involved are identical.

It follows that we should find equality in the results of these two relations, i.e. $T = R$.

In order to obtain this it is necessary and sufficient to adjust the R value so as to arrive at this equality, or else — and this comes to the same — to multiply the sole unknown (i.e. the Taylor coefficient) by the result of the above mentioned ratio.

Thus we finally arrive at a value of the order of $k = 3 \cdot 10^{-2}$ for the coefficient of the virtual viscosity of sea water.

Furthermore, in a recent publication, written by OREN and ENGEL (1965), it was stated that the waters of the Eastern Mediterranean could be divided into three superimposed layers according to their σ_t gradients, the depths of the theoretical limits of these layers being defined by the points of intersection of their best gradient straight lines. The third and deepest of these layers is characterized by nil gradients, a fact relating to conditions that tend to a final equilibrium in deep water.

This tendency towards complete stability should also be reflected in the velocity of currents by a progress towards a total slackening. Furthermore, the graphic representation of velocities should confirm the fact by showing a break in the curve at the depth of the intersection of the best straight lines relative to the σ_t gradients.

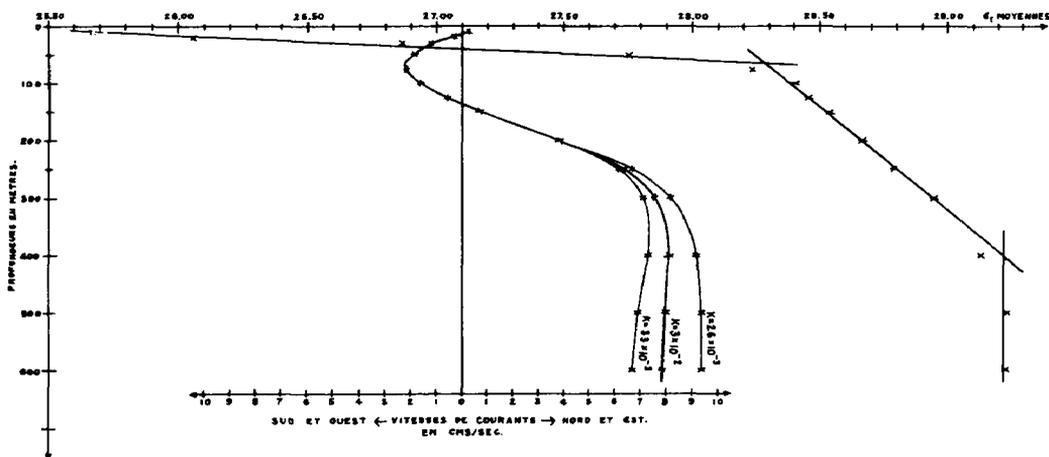


FIG. 5

Station de Recherches des Pêches maritimes à Caïffa (Israël)
 Computation of currents — Stations Nos. 15 and 16
 (Chypre - 02)

In figure 5 the relevant mean values for σ_t obtained at the border stations 15 and 16 are shown, and also the velocities resulting from differences in hydrostatic pressure at these same points. These velocities therefore arise from the application of the general formula (3) given at the beginning of this section and whose results have been corrected for friction effects by means of relation (4).

As to the virtual viscosity term, the following k values have been taken into account :

$k = 26 \cdot 10^{-3}$; a value deduced from the magnitude of the ratio $T/R \approx 10$.

$k = 33 \cdot 10^{-3}$; a value deduced from the result of the ratio $T/R = 12.6$.

$k = 3 \cdot 10^{-2}$; a value obtained by trial and error.

As we see on the graph, for $k = 3 \cdot 10^{-2}$ the corresponding velocity curves reach their maximum at the "critical" depth of the intersection of the slopes relative to the σ_t gradients. Beyond this depth the remaining points of the curve show up a regular decrease in velocity.

It should be noted, however, that this same k value is noted also at the border of the other two layers.

Another example of breaks refers to stations 14 and 15 and is illustrated in figure 6. Nevertheless the maximum in resultant velocities is here reached at the theoretical border between the two upper layers. There it can moreover be seen that the break indicating the beginning

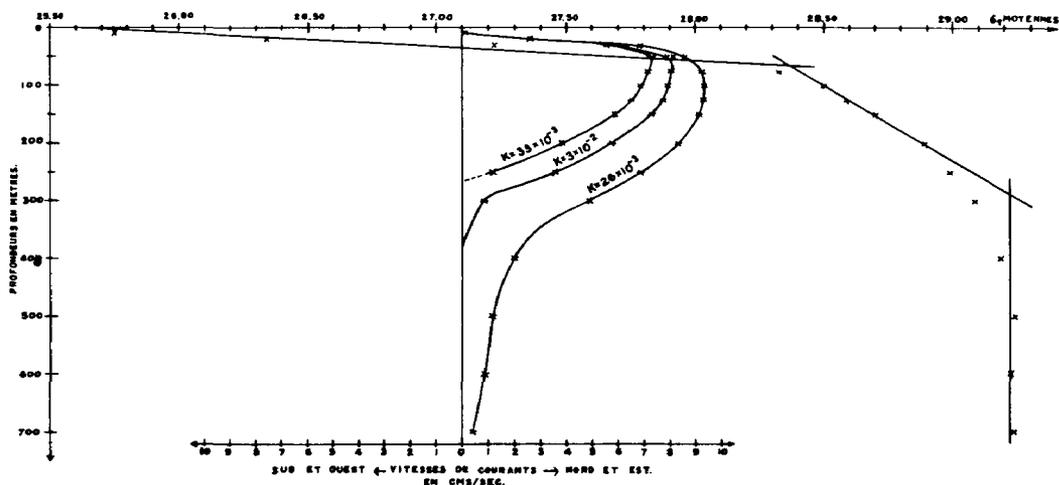


FIG. 6

Station de Recherches des Pêches maritimes à Caïffa (Israël)
Computation of currents — Stations Nos. 14 and 15
(Chypre - 02)

of total slackening stands out clearly even among the decreasing resultant velocities, as a result of the continuous increase in relative velocities.

For other values of k , on the contrary, the breaks occur either below or above the "critical" depths of the intersections.

This fixes the virtual viscosity value at $k = 3 \cdot 10^{-2}$ (*).

The primary consequence of this value of $k = 3 \cdot 10^{-2}$ is that in deep sea it restricts the current velocities to about 8 cm/s — or more exactly 8.15 cm/s. This maximum for resultant velocities corresponds to relative speeds of the order of 14-18 cm/s, and all motion halts for relative velocities of about 32 cm/s.

The scale for converting relative velocities into resultant velocities is given beside the nomogram shown in figure 4.

Computations similar to those set out above, but carried out with other results from the "Chypre-02" Project confirm everywhere the validity of the $k = 3 \cdot 10^{-2}$ value, as far as relative velocities of more than 14 cm/s are concerned. For cases of smaller velocities this computation procedure proves somewhat uncertain. There are then places where satisfactory results are obtained, but in other places — in spite of the existence of a break (possibly shown up by tangents) at the critical depth of the intersection — the maximum in velocities only appears below this depth.

Nevertheless this anomaly seems to be due less to the uncertainty of the value already found for the virtual viscosity coefficient than to conditions peculiar to the flow. There may be cases where either the flow may follow eddy paths or else places where there are significant alterations in the circulation between the two points being considered.

We could, on the other hand, envisage cases where the positions of these velocity "breaks" near the sea surface would allow not only the limits of the field of application of the virtual viscosity coefficient determined above to be fixed, but also a relation tending towards the value given in the Eckmann and Taylor equations.

Having thus arrived at calibrating the current velocity, it will not be without interest to determine its velocities on the median line of two adjacent isobars. Then, by joining up the points of the same value, to plot lines of equal velocity of current. Figure 7 is a diagram of this at a depth of 500 m.

Comparison of this figure with the bathymetric chart of the region (GIERMAN, 1960) shows that the places where the maximum velocities are found correspond to the deepest parts of the bottom, whereas the areas of nil velocity are generally confined to the smallest depths.

It follows that the development of geostrophic currents and consequently the resulting density distribution take place in sea water according to layers moulded on the bottom topography. This result is, moreover, in agreement with NEUMANN's deductions (SVERDRUP-JOHNSON-FLEMING, p. 469), which state that the bottom conditions will be reflected in the temperature and salinity distribution up to levels near the sea surface.

(*) This is, moreover, a value very close to that for the constant in the virtual friction expression given by SVERDRUP-JOHNSON-FLEMING (p. 480) where :

$$\tau = \frac{0.0302}{\left(\log \frac{z + z_0}{z_0} \right)^2} \cdot v^2$$

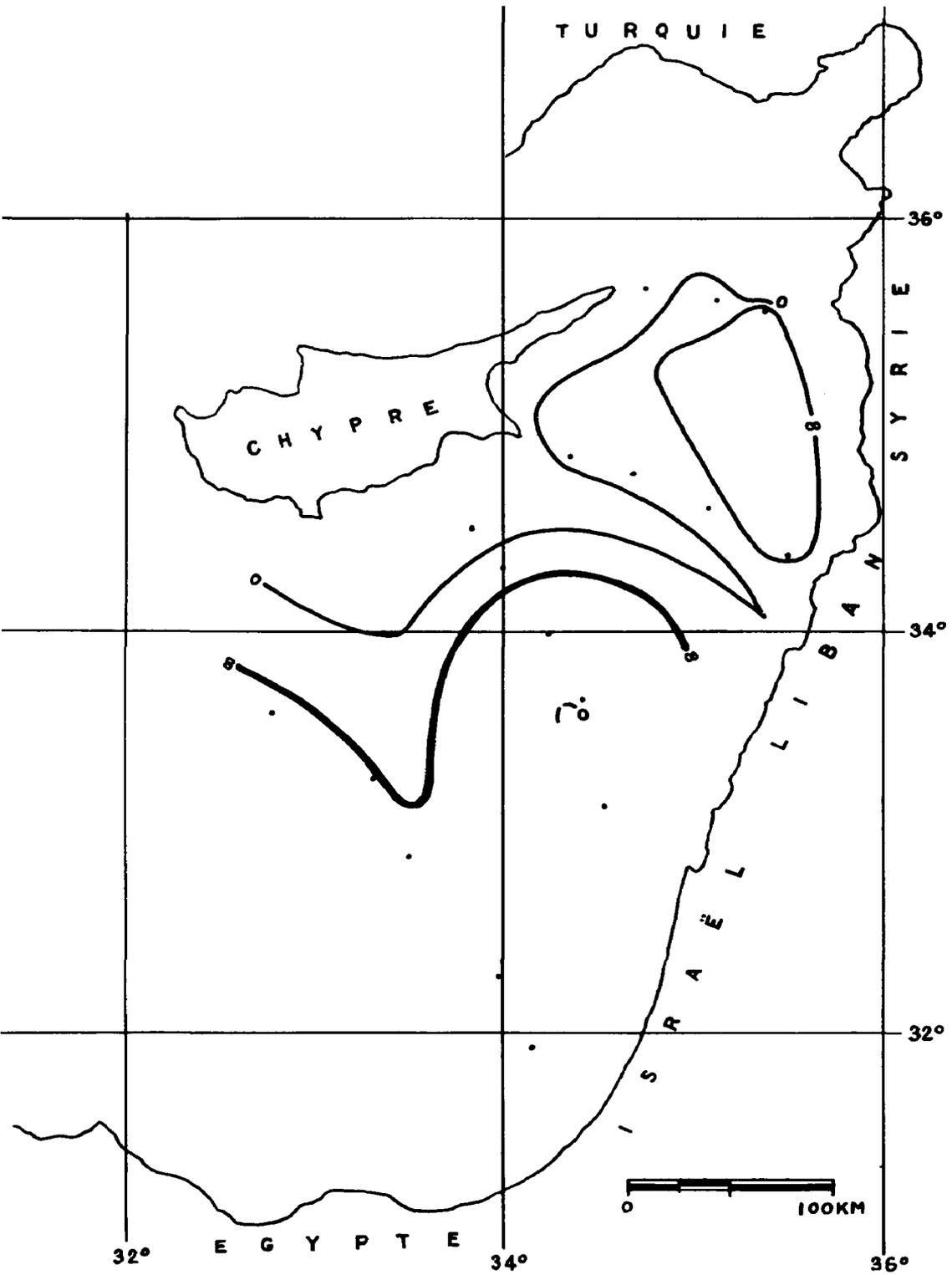


FIG. 7
Station de Recherches des Pêches maritimes à Caïffa (Israël)
(Chypre - 02)
Contours of iso-velocity lines (in cm/s) at a depth of 500 m

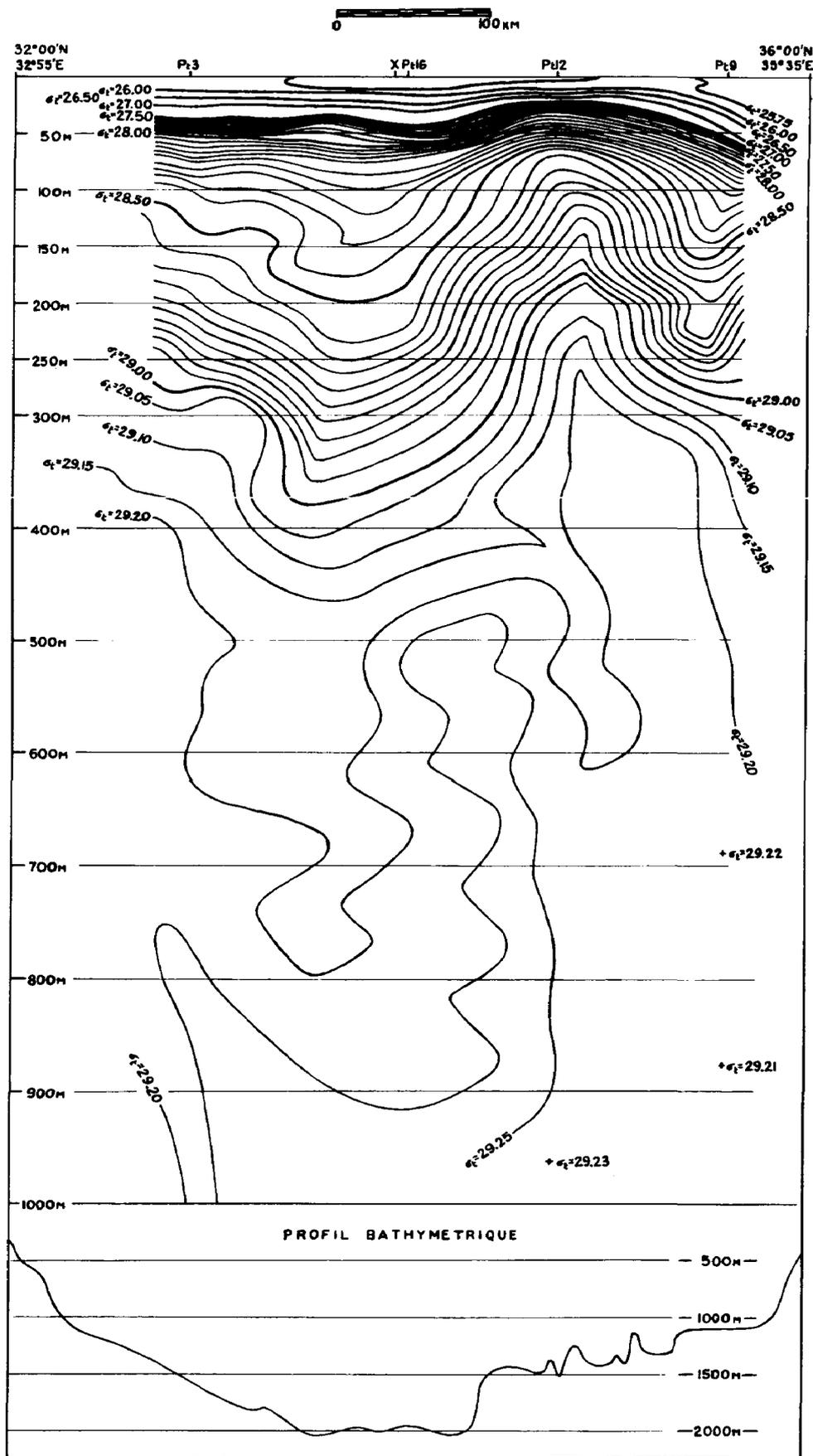


FIG. 8
 Station de Recherches des Pêches maritimes à Caiffa (Israël)
 Study of the Levantine Basin (Chypre -02)
 Vertical section of isopycnal lines

Thus the procedure for, and the validity of, the above computations have both been proved.

It can be seen from figure 7 that the region can be divided into two separate sectors, at a depth of 500 m, by a tongue of "slack waters".

In the sector North of the line Beirut to Cape Greco in Cyprus the distribution of lines of equal velocity of current conforms without further accurate information to the already stated bottom conditions. The lines of equal velocity of current are distributed according to bottom relief, the lines for the highest velocities covering the greatest depths.

In the southern sector, on the contrary, the distribution of these lines of equal velocity of current shows a clear analogy with the water motion when it is a case of a "wash-trough bottom". In this case, as is well known, the heaviest elements converge towards the axis of motion, i.e. towards the places where the velocity is cancelled out. On the vertical sections showing isopycnic lines given in figure 8 (ENGEL, 1965) this is proved by the existence of a column enclosing the densest waters ($\sigma_t \geq 29.25$).

It remains to be stated why we started by choosing a reference level at the sea surface. The actual outcome is the definition in depth of "slack waters" having no motion, as the resultant of high gradients in hydrostatic pressure and consequently of velocities cancelling themselves out through turbulence friction.

For the case of a measurement coverage capable of determining the depth of the reference plane it would only be necessary to start from this plane in the reverse direction in order to compute the velocity of surface currents.

III. — CONCLUSION

The picture of geostrophic currents at the sea surface, as shown in figure 2, agrees with the observations that are likely to change the NIELSEN picture.

Furthermore, the computation data plotted on this figure enable us to explain the cause of phenomena noted by LIEBMAN, OREN, EMERY and BENTOR, EMERY and NEEV, EMERY and GEORGE, whilst granting the possibility of ROUCH's point of view. At the same time the AKYÜZ eddy units in the Gulf of Iskenderun make the representation of such an eddy unit in figure 3 plausible for a depth of 20 m near the Syrian coast. The existence of such an eddy unit is, furthermore, confirmed by North-South currents, as surveyed by EMERY and GEORGE, along the North Lebanese coast. Its connection to the eddy units in the Gulf of Iskenderun is probable, and its extension towards the South at a definite epoch, provides an explanation for the Southwestern currents disclosed by LACOMBE and TCHERNIA.

In the fairly stable atmospheric conditions of summer the geostrophic circulation thus constitutes the principal component of the total circulation.

According to available data, current velocities at the surface are small. In this connection it has not been possible to obtain accurate computed values. However, for reference purposes, we may say that the mean value of permanent currents for the port of Ashdod is about 10-12 cm/s NNE.

Under the surface layer the virtual viscosity would seem to limit the water motion to about 8 cm/s. This same cause would also give rise in deep waters to "slack waters" of nil motion, resulting from high gradients of hydrostatic pressure, and consequently velocities that become nil through turbulence friction.

Beyond a depth of 550 m all possibility of circulation towards the North stops, at approximately the latitude of Cape Andreas in Cyprus on account of the existence of a sill.

Consequently at great depths the water motion could only occur in the measure that the balance of internal forces allows it, and only on paths following the bottom topography. This fact would entail currents following ring-shaped paths, giving rise to eddy units which are likely to be reflected right up to the sea surface.

On the other hand we may deduce the existence in deep sea of local circulation differing from the regional circulation at the surface. This is the effect of bottom conditions and is already noticeable at the 500 m depth, thus making water motion and the density distribution dependent upon the bottom topography.

To sum up, the pressure distribution and the development of the resulting currents occur according to layers moulded on the bottom topography. The highest velocities are found at places where the depth is greatest, whereas the "slack water" areas are related to lesser depths.

Conversely one might seek a relationship between oceanographic data and geology. Slack water areas, giving rise to undisturbed sedimentation, are related to places where sedimentation is the most intense, and consequently concern the lesser depths. On the other hand the directions of the geostrophic current lines coincide in many places with the aspect of isobaths of the bottom, with the Nile's alluvial deposits and, in a general way, with the coastal relief.

The present picture of the coastal regions and the bathymetry of the Eastern Mediterranean would seem to have as origin the action, on a solid and pre-existing shelf, of the geostrophic circulation as the principal component of the total circulation.

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CHART :

British Admiralty : Mediterranean, Eastern Portion. — Crète to Alexandria : Chart No. 2606 reduced.