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Odd H. Sælen

Studies in the Norwegian Atlantic Current
Part II: Investigations during the years 1954-59 in an area west of Stad

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STUDIES IN THE NORWEGIAN ATLANTIC CURRENT

PART II: INVESTIGATIONS DURING THE YEARS 1954-59 IN AN AREA WEST OF STAD

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TRYKT MED BIDRAG FRA NORGES ALMENVITENSKAPELIGE FORSKNINGSRÅD

CONTENTS

Summary. During the years 1954—1959, various investigations were carried out in the Norwegian Atlantic Current with the purpose of obtaining a better understanding of the structure of that current. The investigations were mostly confined to a limited area west of Stad, and were mainly concentrated on (a) detailed hydrographic surveys and (b) direct current measurements by various methods. The detailed surveys showed the frequent existence of eddies of dimensions 30—60 kilometers, and also underlined the great variability of the hydrographic conditions in the area. Current measurements were made both in the Atlantic water and in the cold deep and bottom water. Current meters suspended from an anchored ship were mostly used, but also bottom current meters and free-floating drogues. The overall movement in the core of the Atlantic current was roughly along the isobaths (NE), although in individual cases there were many exceptions to this rule. Normally, it seemed that the velocity shear within the Atlantic water was slight. In the deep water, surprisingly high velocities were often found, even close to the bottom. However, the measurements do not permit conclusions on the average flow of the deep and bottom water. Large velocity variations occurred in both water masses, but it was mostly impossible to demonstrate any tidal variations. In one case, we found variations which may have been inertial oscillations following a strong meteorological disturbance. In the investigations of 1957, an attempt was made to compare computed geostrophic currents and observed currents. Two ships were used, one of them (anchored) observing current at different depths and the other one making hydrographic observations in the area around the anchored vessel. The results clearly showed that the geostrophic approximation, when applied to small-scale surveys, will often give results which have only slight resemblance to the true currents.

Introduction. In part I of this work (SÆLEN, 1959, throughout the rest of this paper referred to as S I), observations over a series of years in the so-called Sognefjord section were analysed. It appeared to be difficult to make sense of the seemingly very irregular variations of the water transport through the section, as computed by the conventional procedure. It was felt that little more could be gained by continuing the frequent occupation of the Sognefjord section, and that a more promising line of attack would be to try to learn more about the internal structure of the current. With these considerations in mind, the work of the Geofysisk Institutt in the area of the Norwegian Atlantic Current in the years 1954—59 was concentrated on making stations in two-dimensional grids, instead of single sections, together with various kinds of cur ent measurements. The investigations were carried out within the framed area in Fig. 1. The technique of observation and the routine treatment of hydrographic data were the same as described in S I, p. 2.

1. The surveys of 1954—57. It is a well established fact that ocean currents do not, generally, flow along straight or smoothly curved lines. The old conception of e.g. the Gulf Stream as a "river in the sea" can at most be held when we are dealing with the current as a "climatological mean" phenomenon. It has been shown that, even in the strong major currents, such as the Gulf Stream off the American coast (FUGLISTER and WORTHINGTON, 1951; STOMMEL, 1958), large eddies frequently occur. In a section across the current, such eddies will give a wavelike appearance to the isolines. Examples of such wavy isolines were shown in S I, p. 22. From our investigations in 1954—57, it seems that in the Norwegian Atlantic Current the existence of such eddies is a frequent occurrence, in any case in the limited area investigated by us. This applies insofar as the observations within each grid can be considered as sufficiently synoptic and not influenced by internal waves to any appreciable degree. The results of the surveys will be discussed below on this assumption. For the presentation of the results in horizontal charts, I have chosen to draw the lines of equal depth of the isosteric surface $10^6 \Delta\alpha = 300$, as this isostere is about in the middle of the transition layer between the Atlantic water and the deep water. There is no point in presenting similar charts for temperature and salinity, as such charts would be very like those for $\Delta\alpha$, because of the close relationship between temperature, salinity and $\Delta\alpha$ in this limited area. In Fig. 2 all observations below the surface layer for the year 1955 are plotted in t — $\Delta\alpha$ and t — S diagrams. It is seen that the value 300 for $10^6 \Delta\alpha$ very closely corre-

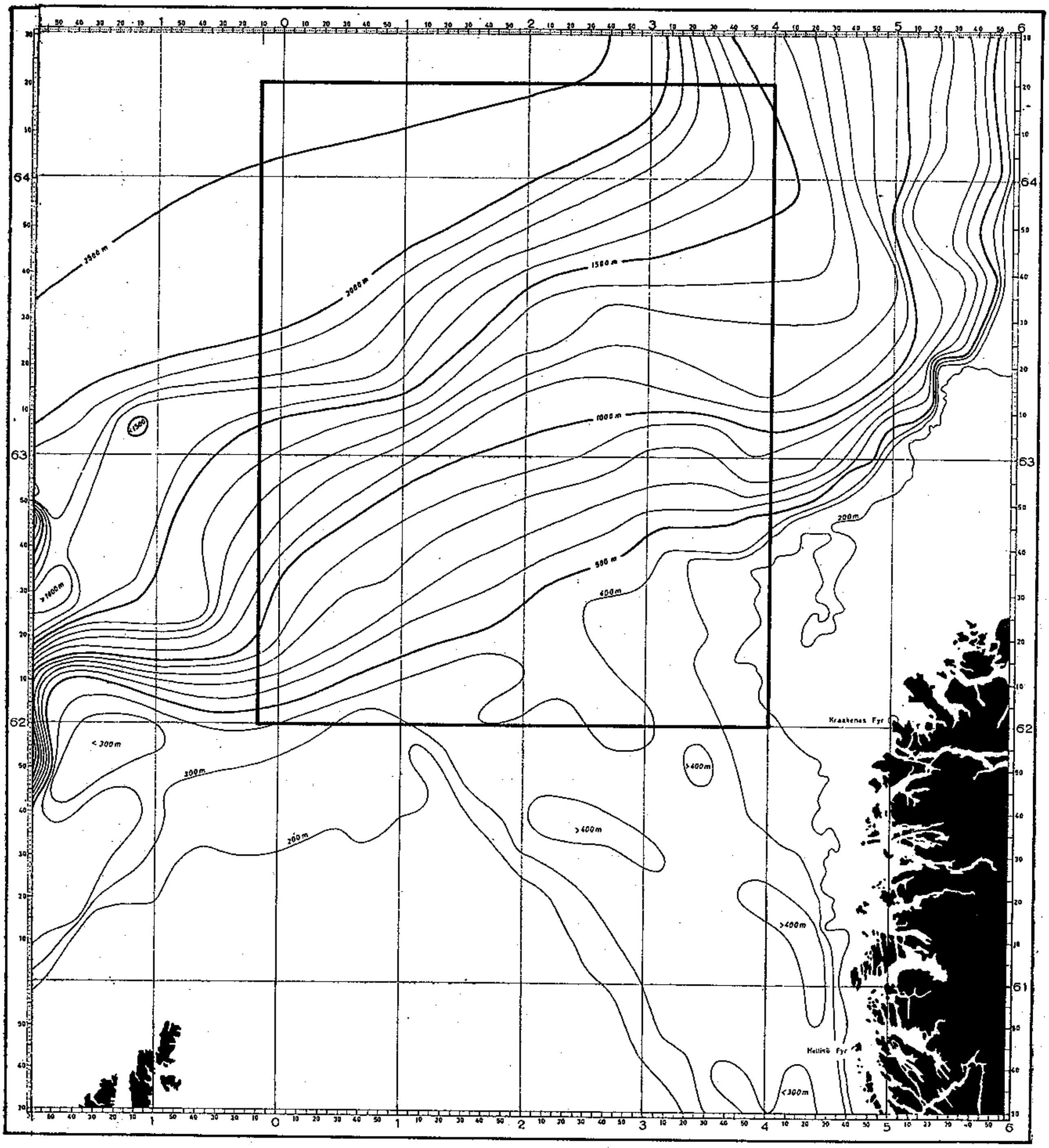


Fig. 1. The area of investigation (framed).

Table 1.

Year	Period of occupation		
1954	July 16–21	July 31–August 5	August 6–9
1955	August 2–6	August 8–10	
1956	April 26–May 3	May 19–22	
1957	June 7–12	June 27–30	

sponds to a temperature of 4°C. The depth of the chosen isosteric surface will thus give an approximate measure of the thickness of the Atlantic current.

During the years 1954—57, a total of 9 such surveys were made, as indicated in Table 1. One of these surveys is mostly outside the Atlantic current (the first one of 1957). The charts are reproduced in Figs. 3—11.

First, let us look at the 1954 surveys (Figs. 3—5). In the first survey, July 16—21, the depth lines of $10^6 \Delta\alpha = 300$ indicate a cyclonic eddy centered at about 62°50'N, 2°10'E. This is on the continental slope at a depth of about 800 meters. A corresponding anticyclonic eddy is indicated to the NW. The second survey, July 31—August 5, did not extend as far SW as the first one, but it seems reasonable to think that the depth of 240 m of the $10^6 \Delta\alpha = 300$ surface at station 107 is near the center of an anticyclonic eddy similar to that of the first survey. If this be so, the center is at about the same position as that of the first survey. The third survey of this year, August 6—9, again exhibits a similar eddy at about the same position. There are about three weeks between the first and the last survey. Thus it seems possible that a stationary eddy has existed at the above stated position during this period of time. Another explanation may be that eddies have been travelling through the area, and that our observations have by chance been timed so that different eddies have been in the said position just at times of observation. A third explanation is that there have been large internal waves, and that our observations have been taken in such phases of the wave as to produce the pictures given in Figs. 3—5. This questions will be discussed later on.

The two surveys in 1955 were carried out within a week (August 2—6 and 8—10) (Figs. 6—7). The first survey disclosed part of an elongated ridge in the topography of the $10^6 \Delta\alpha = 300$ surface. It was situated considerably farther to the west than the eddy of the previous year, but approximately over the same depth of the continental slope. A very similar "elongated eddy" was found in the same area when the survey was repeated 5 days later. Thus, also this year there is some evidence of the existence of eddies over the slope, although not so convincing as in 1954.

The cruise of 1956 was carried out at an earlier season than the previous years (April—May), and was staged in cooperation with the British research vessel "Discovery II", from which experiments were made with the Swallow neutral buoyancy float. This aspect of the cruise will be discussed later. The investigations were continually hampered by rough weather, but is was possible to obtain two surveys, the first

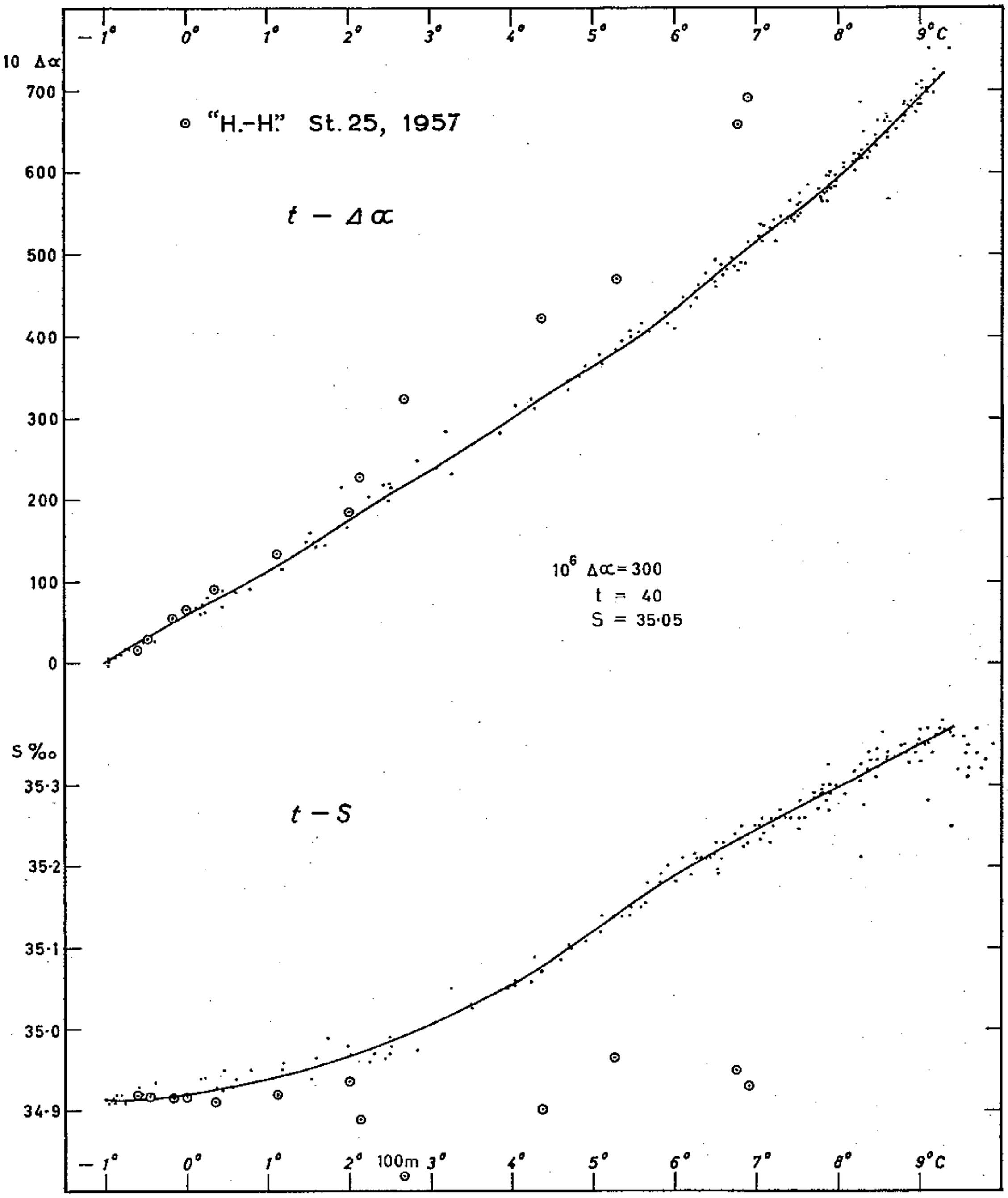


Fig. 2. t—S and t—$\Delta\alpha$ diagrams for the year 1955. The values for a station influenced by Arctic water ("H.—H." station 25, 1957) are also entered (circles).

of which was carried out from April 26 to May 3, and the second one May 19—22. (Figs. 8—9). In the first of these surveys, the depth of the $10^6\Delta\alpha = 300$ surface is in most of the area between 400 and 500 meters. A relatively small "eddy" was found on the slope at about the same position as the eddies of the previous year, the depth of the surface being a little less than 300 meters. To the west of this "eddy", the transition

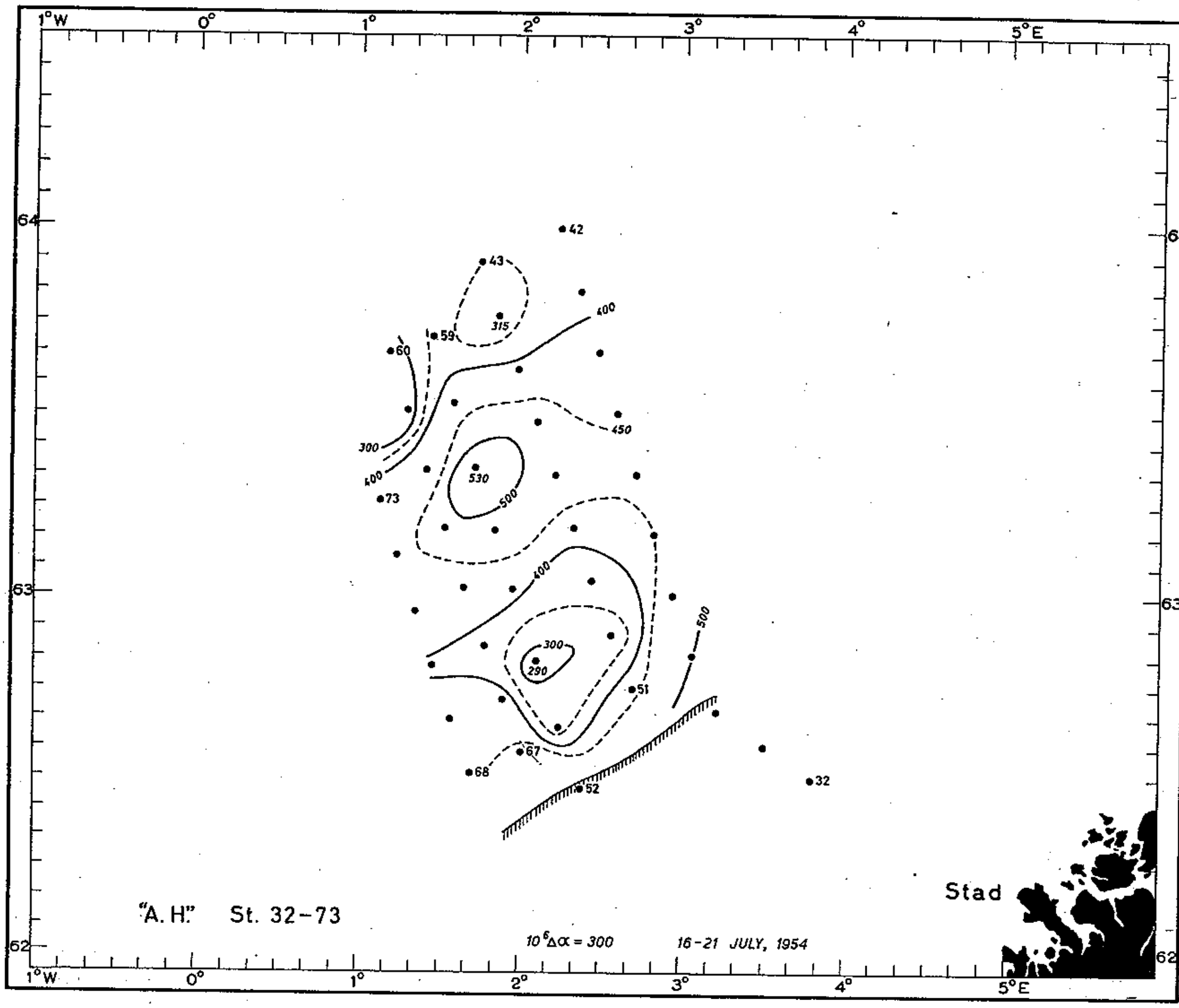


Fig. 3. Topography of $10^6\,\Delta a = 300$ surface on survey July 16—21, 1954.

layer, as defined by $10^6 \Delta a = 300$, slopes down to more than 600 meters, the deepest position recorded in these investigations. The second survey, made more than 3 weeks later, showed little or no similarity with the first. The situation was rather diffuse. There was no distinct "eddy" on the part of the slope (at about 800 meters bottom depth) were the most pronounced "eddies" were found in the previous investigations, but there was an indication of such an "eddy" farther off the coast at 1100 meters depth (station 98).

The investigations of 1957 were carried out by two ships, the "Armauer Hansen" and the "Helland-Hansen" and it was thus possible to cover a wider area (Figs. 10—11) in the two surveys made that year. The first survey covered an area farther north than previously, from 63°N to 65°N, and thus can not be directly compared with the earlier investigations. There are reasons, however, for a closer study of that survey.

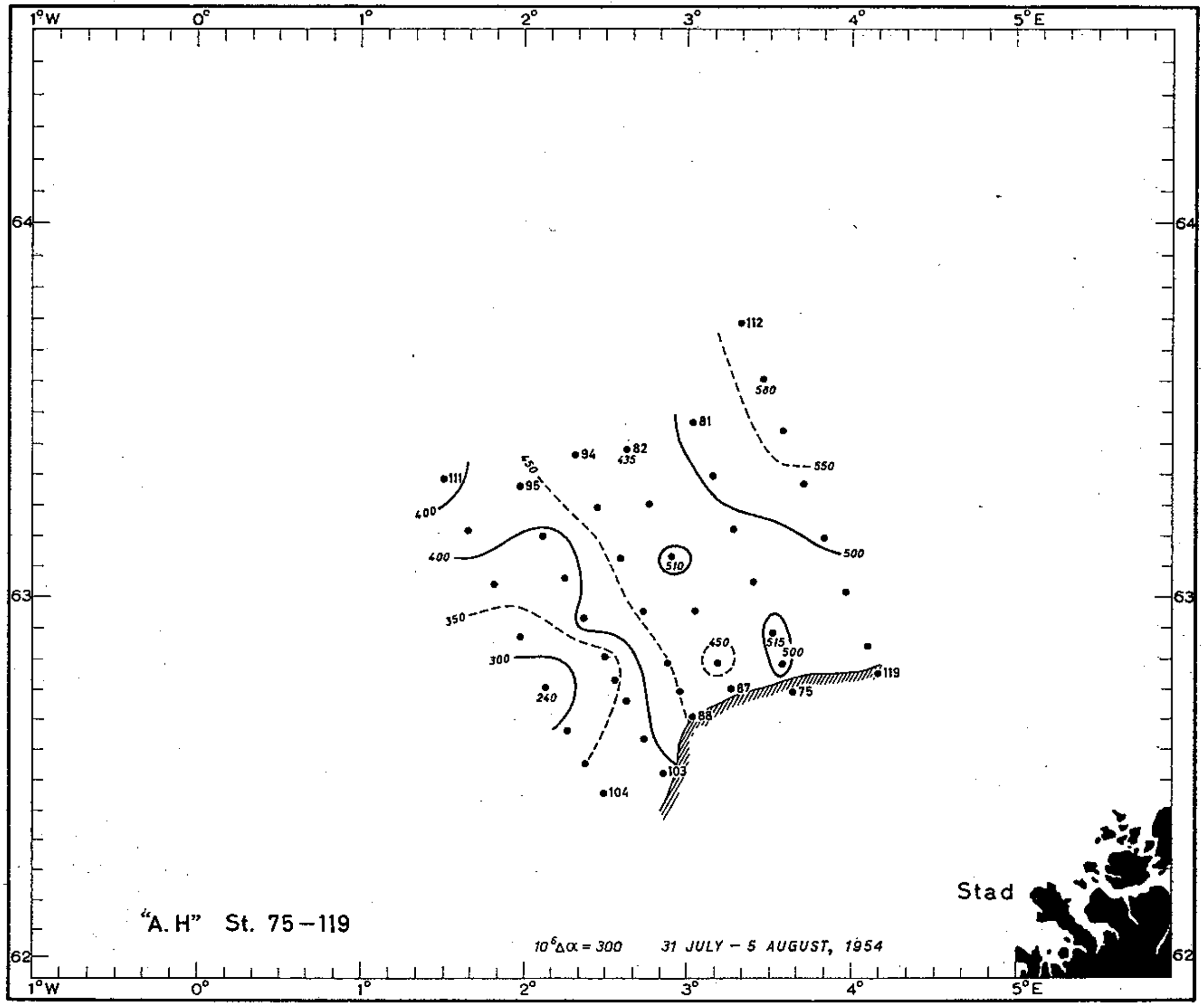


Fig. 4. Topography of $10^6 \Delta\alpha = 300$ surface on survey July 31—August 5, 1954.

The two most conspicuous features of Fig. 10 are (a) the sharp rise of the $10^6\Delta\alpha = 300$ surface at about 64°N, and (b) the sinking of that surface in the northeastern part of the area, indicating a thicker layer of Atlantic water.

The transition at 64°N is not similar to the depth variation of the $10^6\Delta\alpha = 300$ surface farther in on the continental slope, in which area the variations signify varying thickness of the upper layer of Atlantic water. When crossing the transition at 64°N, however, one enters into a different water mass, characterized by an almost complete absence of "Atlantic" water, as defined by $S > 35^0/_{00}$. On most of the stations in that region the Atlantic water in the upper layers has either completely disappeared, or exists as a thin layer of surface water with salinity only slightly above $35^0/_{00}$. This is caused by Arctic water penetrating into the region from WNW, obviously a continuation of the East Iceland Arctic current. The t — S and t —$\Delta\alpha$ relationships shown

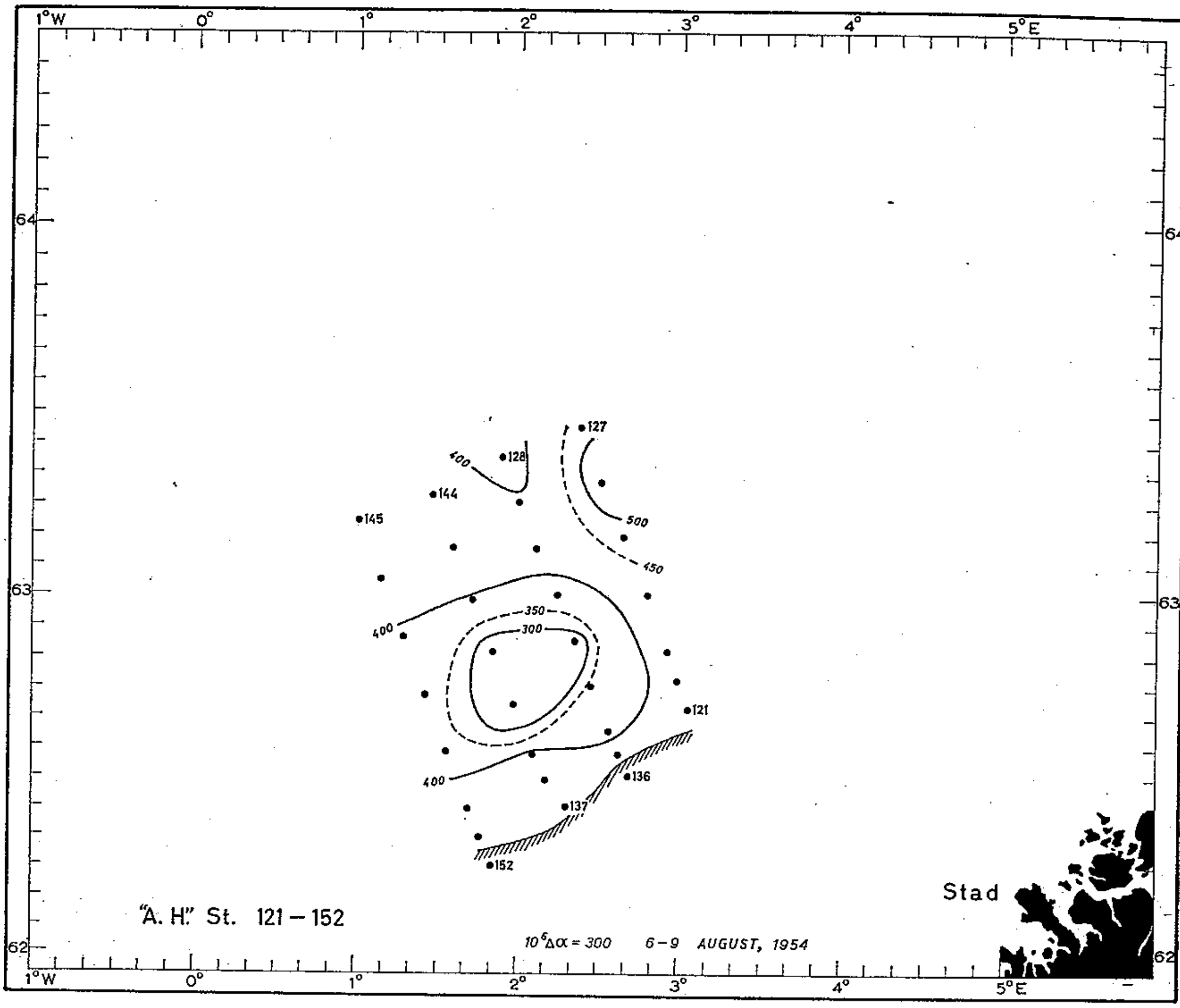


Fig. 5. Topography of $10^6\,\Delta\alpha = 300$ surface on survey August 6—9, 1954.

in Fig. 2 do not hold for water masses that are influenced by the Arctic water, and the surface $10^6\Delta\alpha = 300$ no longer indicates the transition between Atlantic and deep water. The penetration of Arctic water can very conveniently be traced by the salinity minimum that is formed at an intermediate depth when the upper layer of the Arctic water mass is mixed with Atlantic water, or when the Arctic water intrudes below an Atlantic water mass at a subsurface level. The situation is depicted in Fig. 12, which shows the areas where a salinity minimum occurs. The figure also shows the areas with stations at which the salinity is below $35^0/_{00}$. throughout the water column (heavily shaded). The Atlantic water in the northeastern part of the area is evident also in this figure. It would be misleading, however, to interpret the corresponding configuration on Fig. 10 as the one half of an "eddy" in the same sense as the ones occuring within the Atlantic water on the slope. It is obviously the last part of a branch of

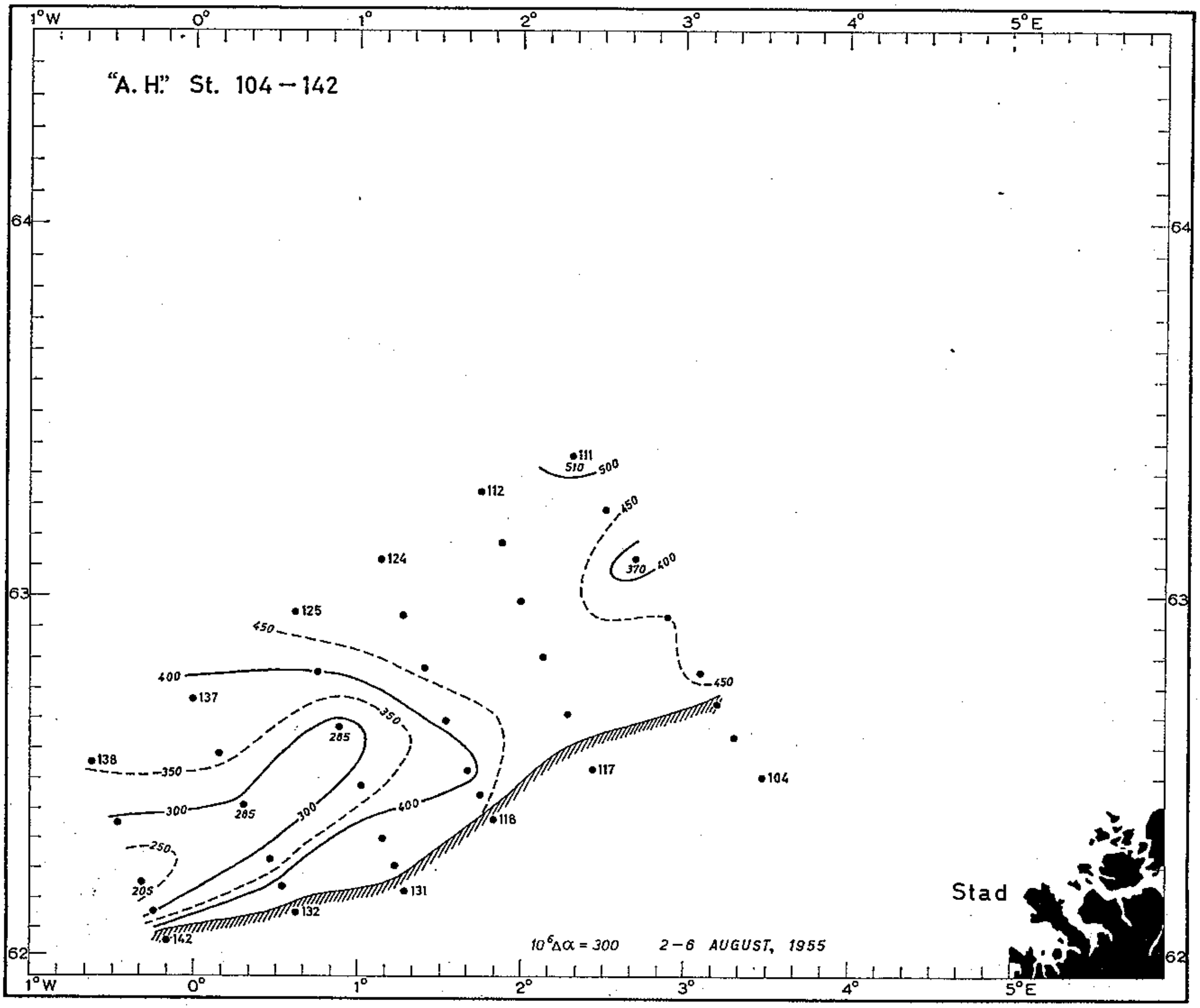


Fig. 6. Topography of $10^6\, \Delta a = 300$ surface on survey August 2—6, 1955.

Atlantic water that shoots off from the main current, making a counter-clockwise turn. This is a semi-permanent feature that was discovered already by Helland-Hansen and Nansen (1909), and which is also found from later observations in this region. This branch meets the tongue of Arctic water coming from the west in what seems to be a region of intense mixing between Atlantic and Arctic water. The transition between the Arctic water and the main Atlantic current is very sharp, as seen from Fig. 12, where the most pronounced Arctic water of the area is seen to be situated close to the main Atlantic current. Note e.g. the difference between stations "H.-H." 25 and 26, only ten miles apart (Figs 10, 12, 13). Thus, it appears that the mixing takes place mainly between the Arctic water and the branched-off Atlantic water to the north of it. In Fig. 13 some selected salinity — depth curves from different areas are reproduced, the positions of which can be seen from Fig. 12. It should be observed that the advance

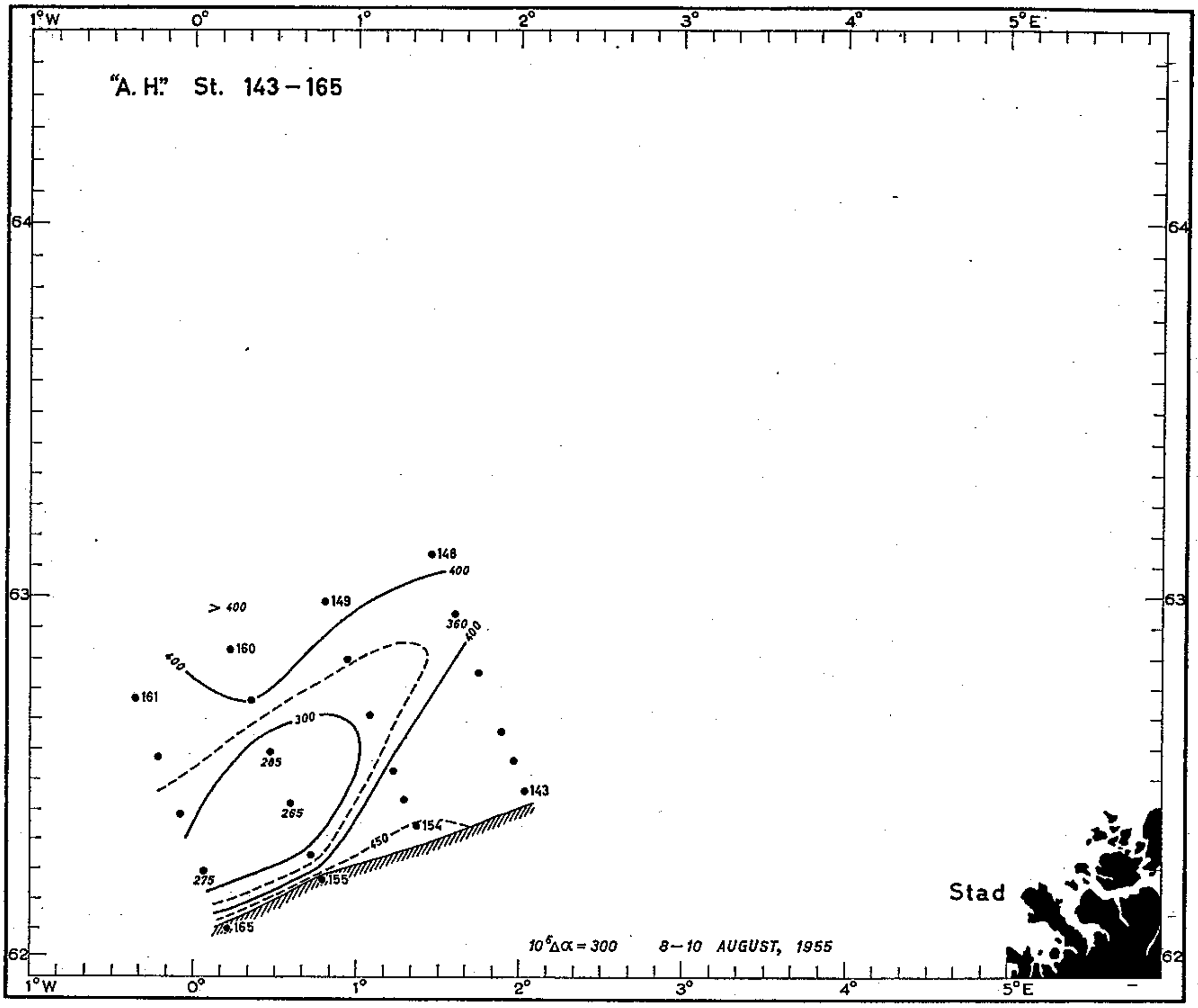


Fig. 7. Topography of $10^6\ \Delta a = 300$ surface on survey August 8—10, 1955.

of Arctic water toward the east seems to have been abnormally strong in 1957. This can be seen from Fig. 14 where the $35^0/_{00}$ isohaline at 100 meters is drawn for the years 1904, 1935, 1936 and 1957. In 1957, this isohaline continues beyond 2°E, whereas in the other three years it stops west of 0°. The salinity at the minimum is also very low in 1957, at several stations it is between 34.80 and 34.85, values that are lower than those of any of the other three years in the whole area shown in Fig. 14.

With these considerations in mind, let us look at the distribution of Δa in the section st. "H.-H." 38 — 46 (Fig. 15). The isosteres exhibit a very nice waveform, and one might be tempted to interpret this as an effect of internal waves. However, it is clear from the preceding discussion that the form of the isosteres is determined by a quasi-stationary distribution of the water masses in this area. The section cuts through the Arctic tongue between the two Atlantic water masses.

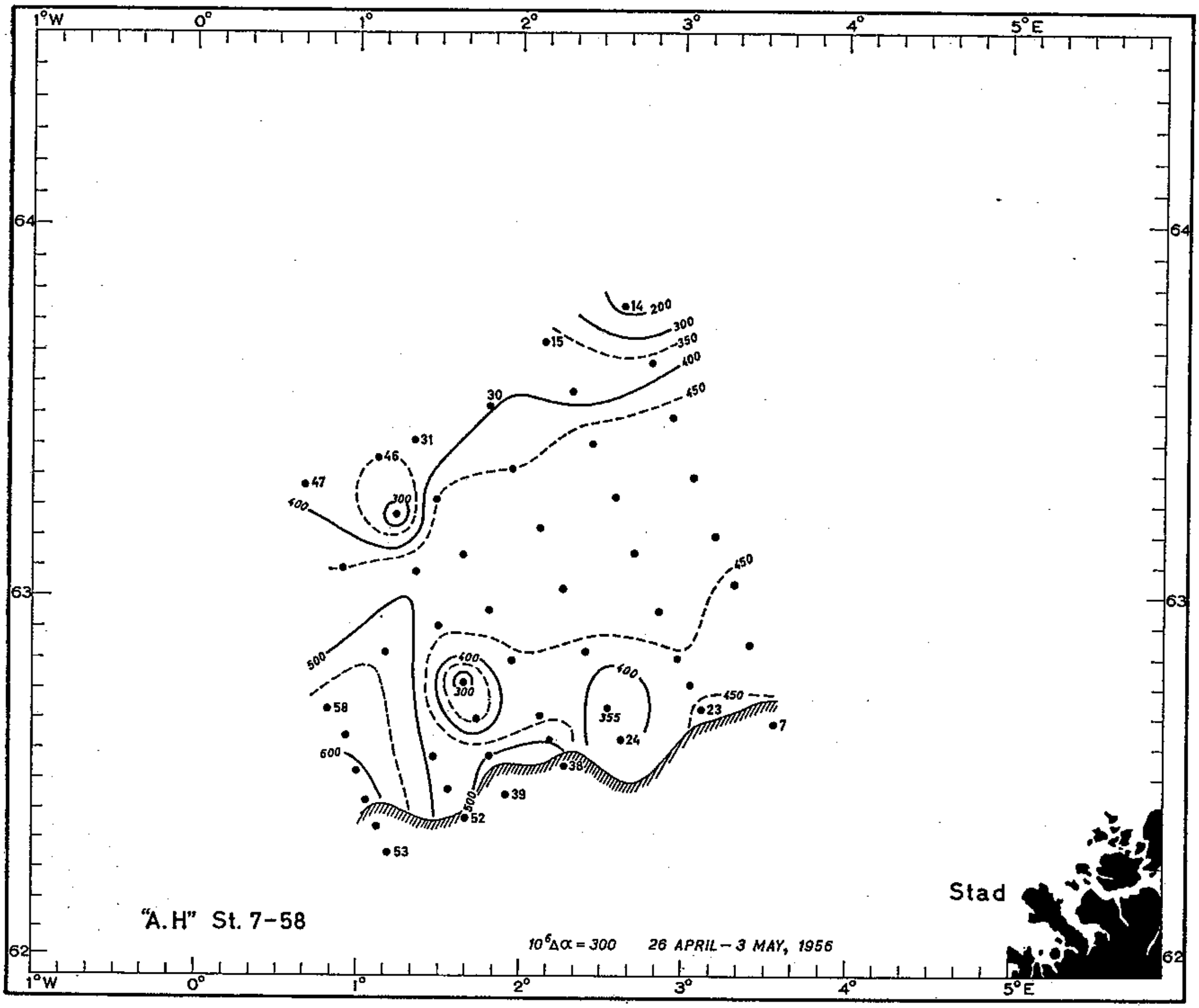


Fig. 8. Topography of $10^6\,\Delta\alpha = 300$ surface on survey April 26—May 3, 1956.

The network of stations occupied later in June 1957 (June 27—30) was in the same area as those of the previous years (Fig. 11). There is no clear-cut eddy, but we find an elongated area where the transitition layer has a depth of 300 meters, sloping down to 500 meters to the south and to 400 meters to the north. In connection with the phenomenon discussed above, i.e. the exceptionally strong influx of Arctic water in 1957, it should be mentioned that an intermediary salinity minimim is present at depths of 300 or 400 meters on many of the stations in the western part of this second grid. This is clearly a deep intrusion of water from the tongue of Arctic water, and the minimum values are at some stations as low as 34.87—88‰. Such an intrusion is not found in any of the other years 1954—56.

To sum up, we have in the four years occupied in all 8 networks of stations in more or less the same part of the Norwegian Atlantic current. With the close spacing of the

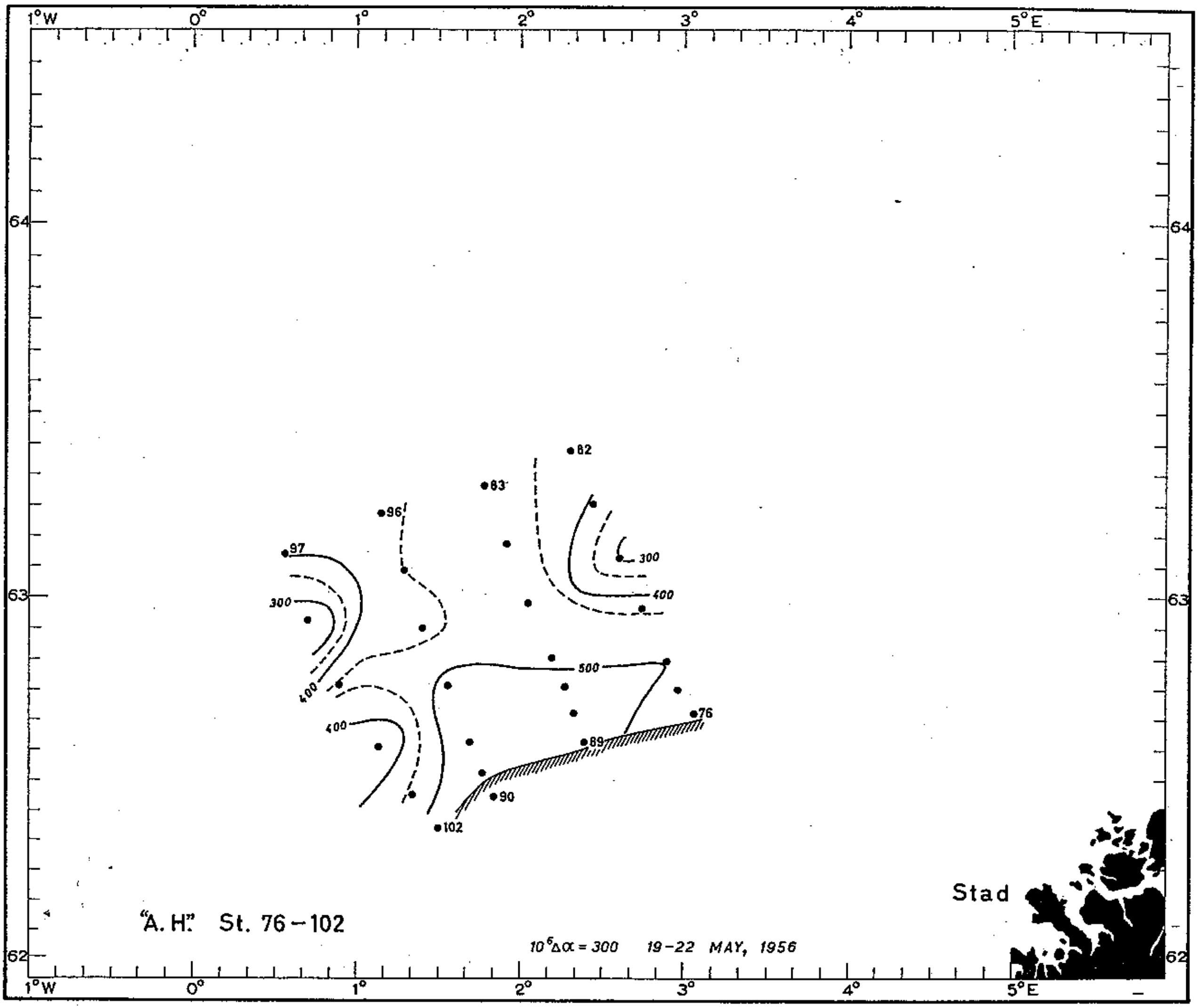


Fig. 9. Topography of $10^6\,\Delta\alpha = 300$ surface on survey May 19—22, 1956.

stations in these investigations, the features of the hydrographic situation will most often be rather irregular. As previously mentioned (p. 4), Figs. 3—11 roughly show the thickness of the Atlantic water in the current, and this is seen to vary between rather wide limits. In many cases, however, the isobaths of the $10^6\Delta\alpha = 300$ surface form the kind of pattern one would expect in an eddying motion, the diameter of the eddies being of the order 50 km. When occuring close to the continental slope, these "eddies" are always cyclonic, that is, the layer of Atlantic water is thinner in the center. In some cases, they are oblong-shaped. In a couple of the surveys, though, "eddies" are not found at all, or are only weakly developed. It is interesting to note that the area we have surveyed very nearly coincides with one of those in which Helland-Hansen and Nansen (1909, p. 144) concluded that vortices are likely to occur. They ascribe this tendency to the bottom configuration.

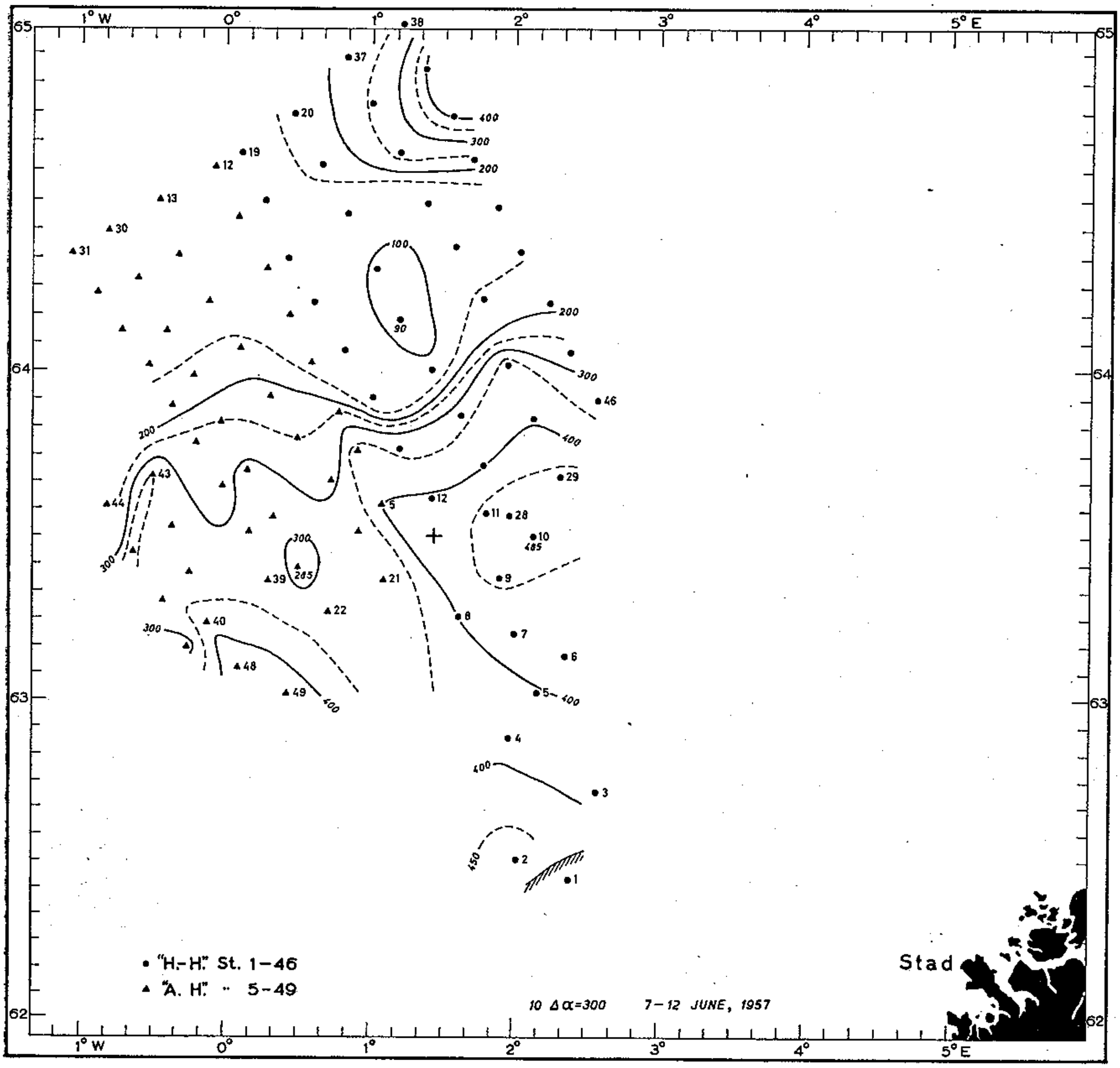


Fig. 10. Topography of $10^6\,\Delta\alpha = 300$ surface on survey June 7–12, 1957.

2. Waves or whirls? The configuration referred to above as "eddies" might, theoretically, also be a result of the influence of internal waves. The observations at the different stations might have been taken at such times (phases of the wave) that, given waves of the appropriate wavelength and period and sufficiently large amplitudes, eddy-like configurations might have been produced. In the case of oceanographic surveys off the Californian coast, DEFANT (1950a) took the view that this was possible and devised a method for the elimination of the supposed influence of internal waves from the survey charts. If the results of our surveys had been presented as vertical

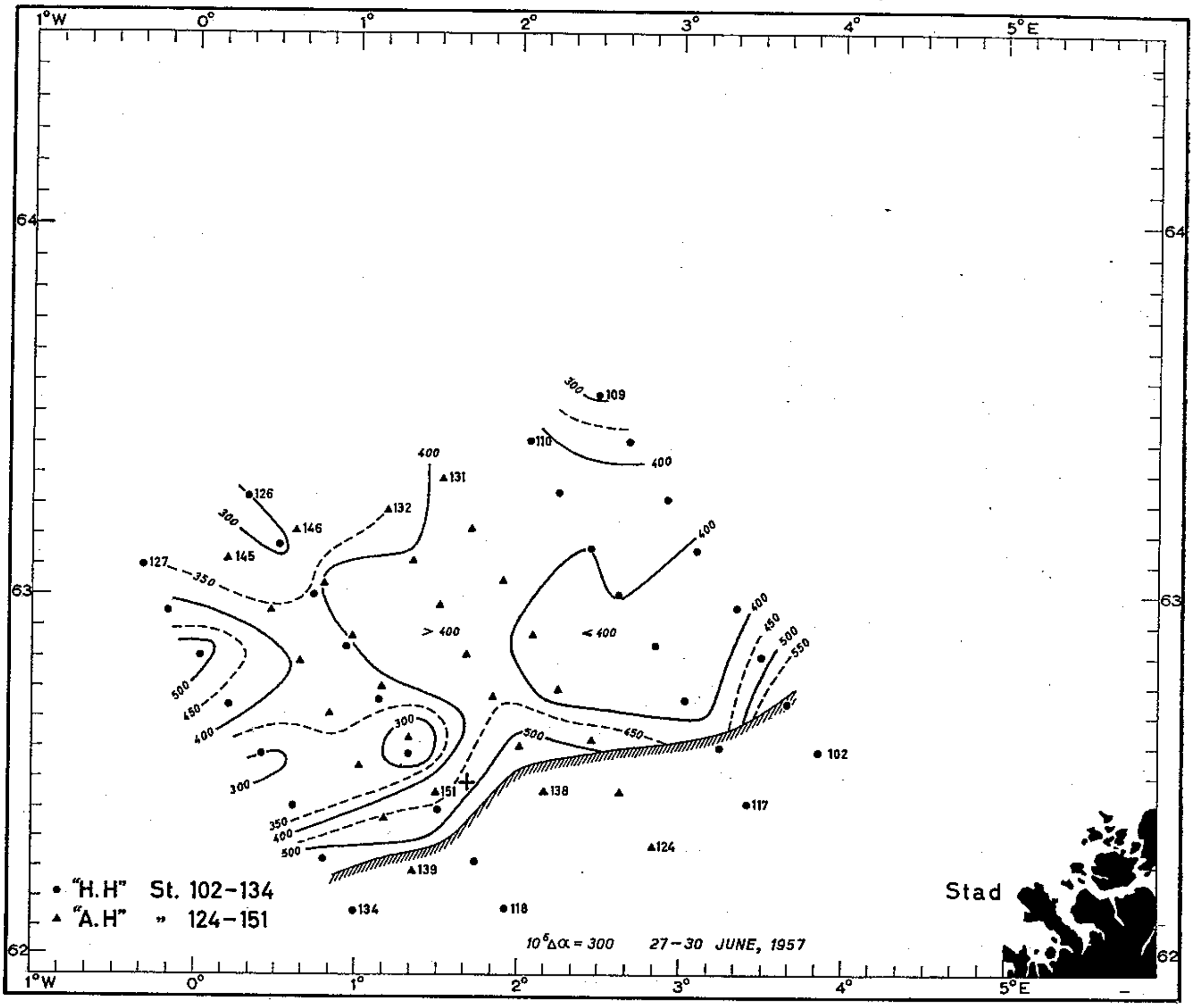


Fig. 11. Topography of $10^6\,\Delta\alpha = 300$ surface on survey June 27—30, 1957.

sections, instead of horizontal charts, the first impression would in many cases inevitably have been that of a wave (Figs. 15, 16). In one particular case, we have already shown that the wavy form of the isosteres must be ascribed to other causes (p. 10; Fig. 15). That case was an exception, however, as the section was partly outside the main Atlantic current. The discussion below will be concerned with the possibility that the eddy-like configurations of the transition layer in the main current may be due to internal waves. A final answer to the question would need a much more extensive material than that collected during these surveys. Some indications may, however, be given, but we shall first consider some relevant aspects of the theory of internal waves in the open ocean. We are here concerned only with "long" waves. The main problem in the present connection is whether, and under what conditions, such waves can attain the large amplitudes required to explain vertical displacements of 100—300 meters. The

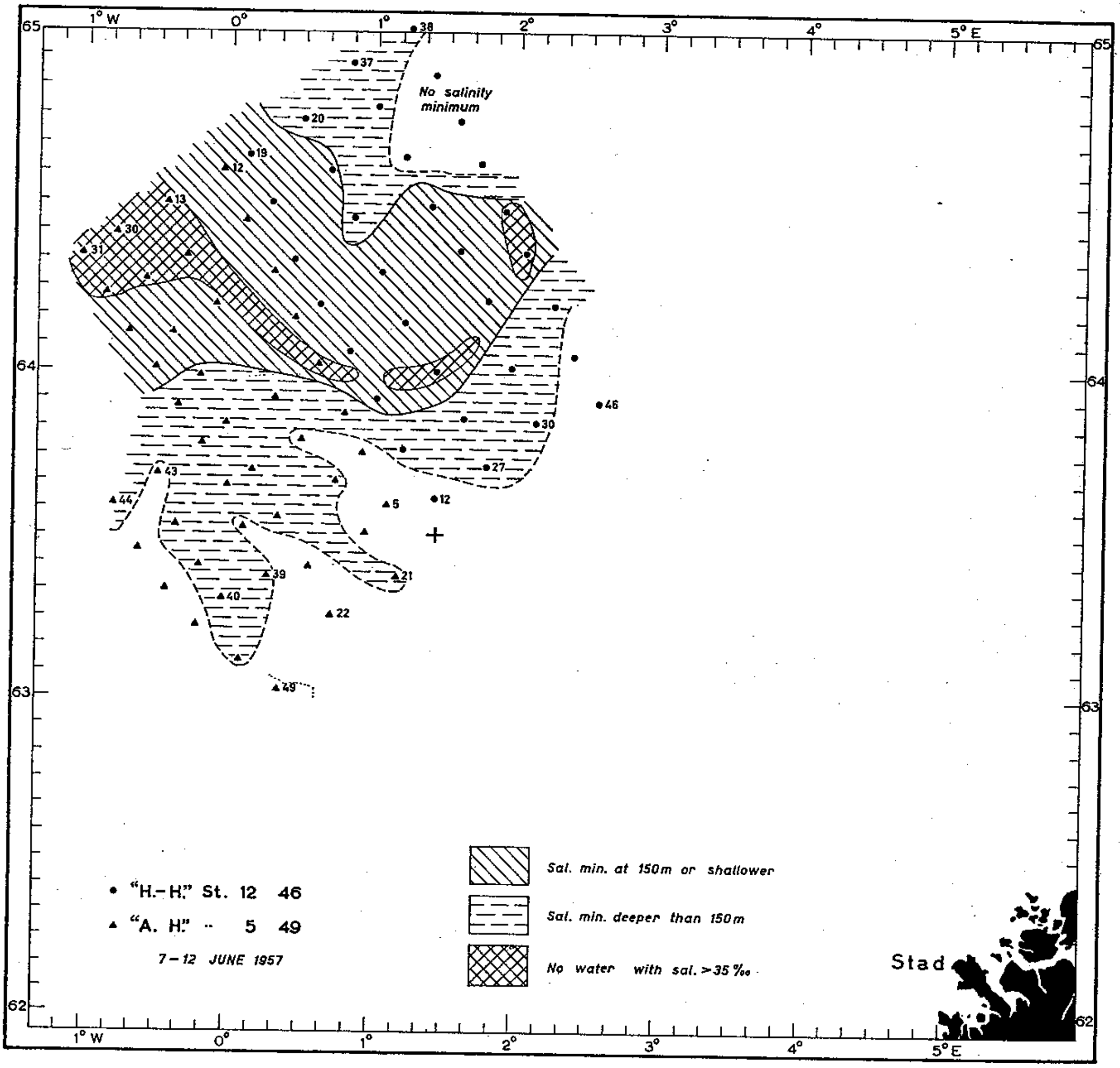


Fig. 12. Areas with salinity minimum on survey June 7—12, 1957.

most general theory of free internal waves in stratified water is still that of FJELDSTAD (1933). In this theory, the density may increase continuously with depth in an arbitrary manner, and the rotation of the earth is taken into account. He proved the possibility of the existence of internal waves (i.e. that they are compatible with the linearized equations of motion and the proper boundary conditions), and gave a numerical method for the solution of the problem. The amplitudes are, however, left undetermined, and an explicit relationship between period and wavelength can be found only in the very special cases when a density model permits an analytical solution. Later authors have attacked the problem by assuming a two-layer model of the ocean (e.g. DEFANT,

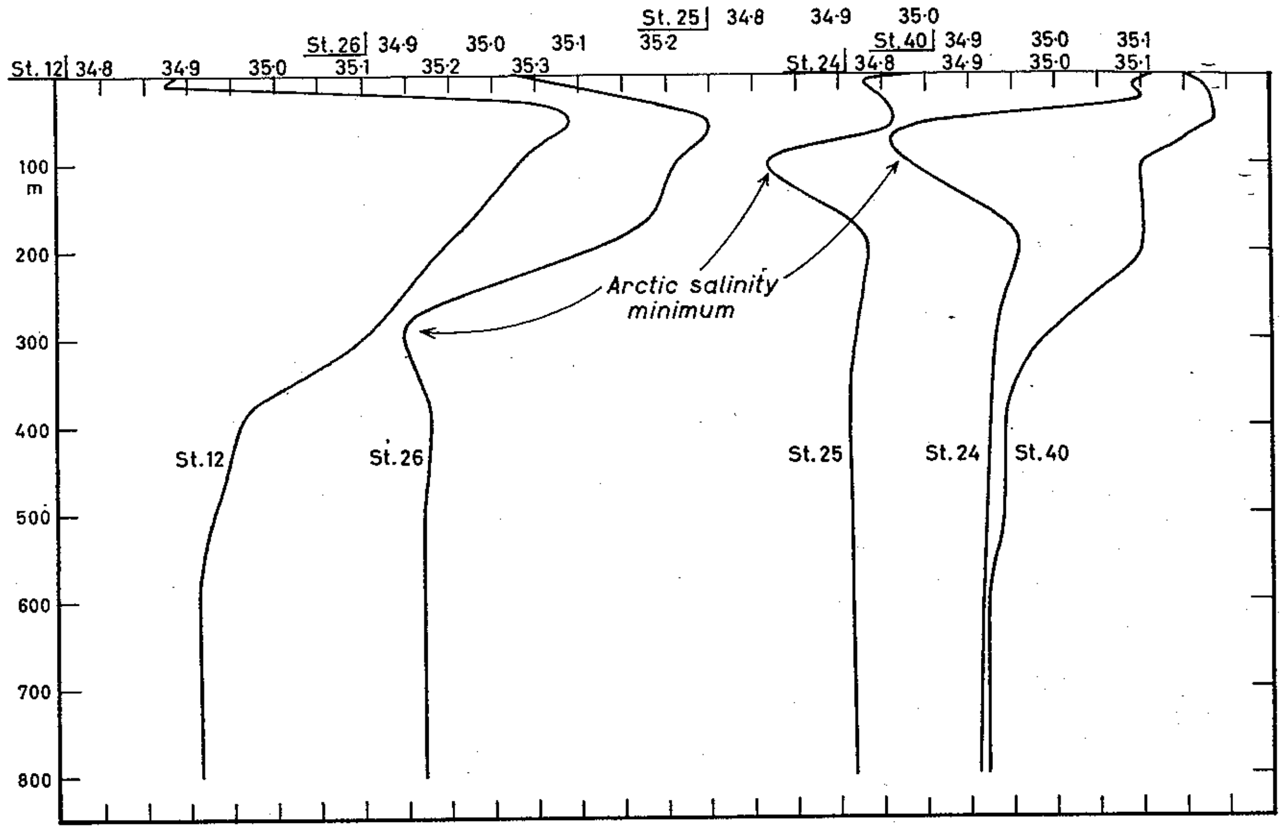


Fig. 13. Selected salinity — depth curves for "H.—H." stations on survey June 7—12, 1957. St. 12: Main Atlantic current. St. 24: Shallow salinity minimum area. St. 25: Area of strongest Arctic influence. St. 26: Deep salinity minimum area. St. 40: Branched-off Atlantic water.

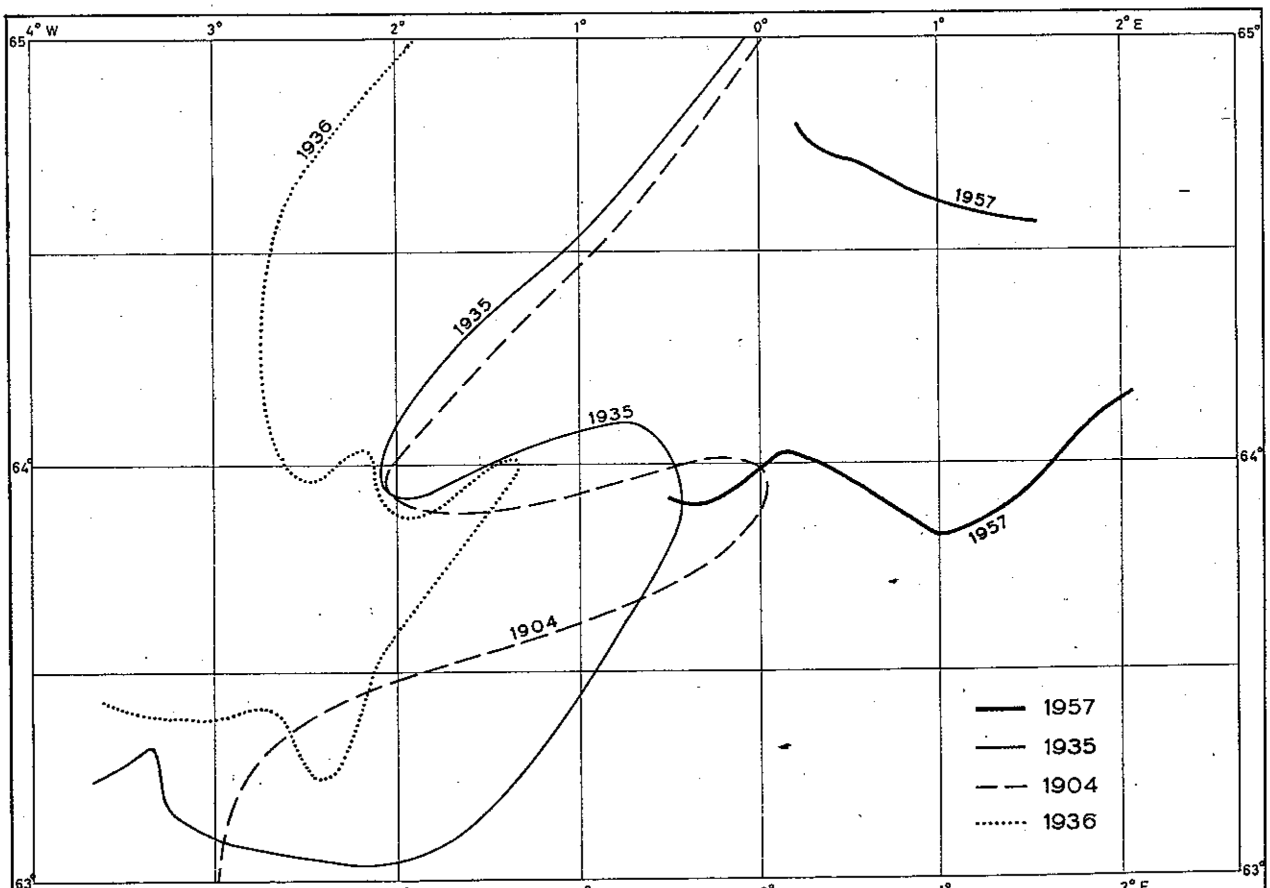


Fig. 14. Position of 35 $^0/_{00}$ isohaline at 100 meters depth in different years.

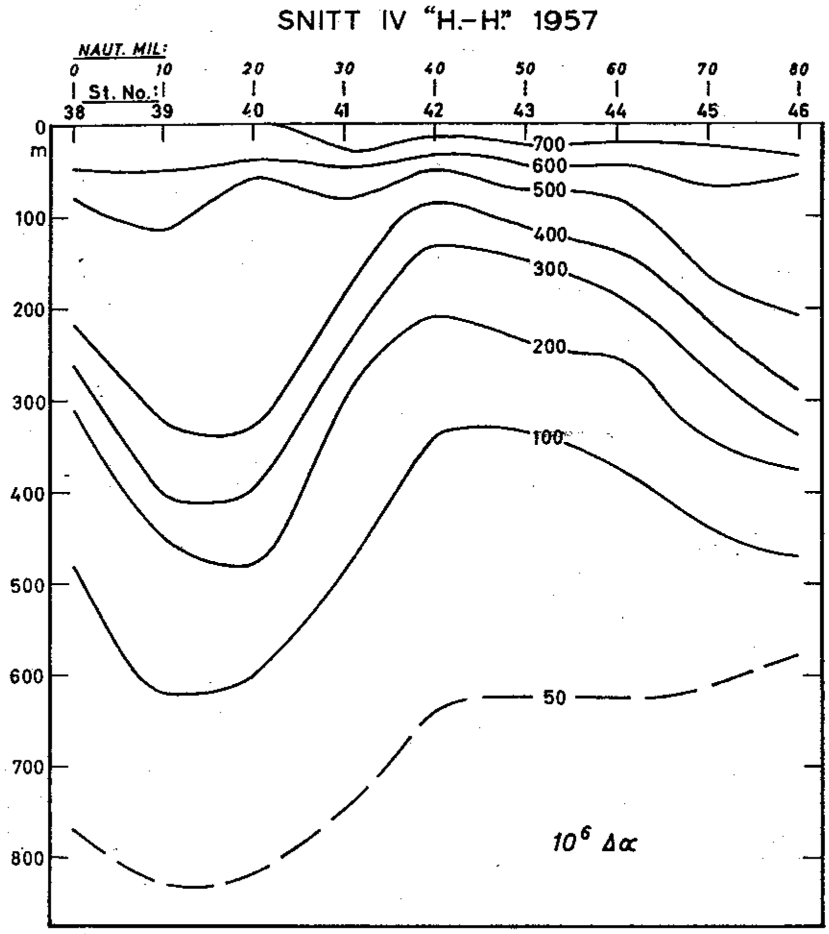


Fig. 15. Distribution of Δa in the section "H.—H." stations 38—46, 1957.

1950b; HAURWITZ, 1950; KRAUSS, 1957, 1958a, 1958b, 1959). This is of course little realistic, but it may be supposed that at least some principle features will be evident from such a model. Serial measurements of temperature, salinity and current from vessels staying at a fixed position have indicated that tidal periods predominate in the time-variation of the measurements (DEFANT, 1950b, 1952). However, severe doubts have been cast upon the significance of many of the results of the harmonic analysis (HAURWITZ, 1954; REID, 1956). There seem to be cases, however, where the conclusion is inevitable that internal waves of tidal period exist. In such cases the period of the oscillation can be deduced, and to some extent also the amplitude. The wavelengths are unknown, as are also the directions of travel. Furthermore, it cannot be determined whether the waves are progressive or standing. Granted that the period is that of a tidal constituent, it must be supposed that the waves are governed by some sort of tidal influence. The case of the direct forced internal tidal wave on the rotating earth in the two-layer model has been investigated by HAURWITZ (1950) and DEFANT (1950b). In this case, a solution is required that conforms to the tidal force with regard to both period and wavelength. It appears that the amplitudes of the internal wave will be very small except in the limiting case when the geographical latitude approaches a value such that

$$\sigma^2 = f^2 + \frac{gh_1h_2}{h_1+h_2} \frac{\varrho_2-\varrho_1}{\varrho_2} \varkappa^2$$

where σ is the frequency, f is the Coriolis parameter $= 2\,\omega \sin \varphi$, φ is the geographical latitude, $\varkappa$ is the wave number, g is the acceleration of gravity, ϱ is density and h layer thickness, indices 1 and 2 denoting upper and lower layer respectively.

It is not probable that this case of resonance should have any bearing on internal waves in the Norwegian Sea. The theory requires a very great wave length which cannot develop in this relatively small basin, where in addition the upper layer (the Atlantic water) does not extend continuously over the ocean. It has recently been indicated by KRAUSS (1959) that the predominance of tidal periods in internal waves pointed out by DEFANT (1950b, 1952) may be due to a co-oscillation in analogy to the "co-oscillating tide" in the theory of ordinary surface tides, at which the period is prescribed as a tidal period whereas the wavelength is that of the free oscillation of that period. For tides in a bay or gulf, this leads to standing waves the amplitudes of which may in some cases be increased by resonance. In the much more complicated case of internal waves, it will be extremely difficult, if at all possible, to formulate criteria which apply to natural conditions met with in cases like the Norwegian Sea. However, the possibility of resonance indicated by KRAUSS seems more likely than that envisaged by DEFANT and HAURWITZ.

It has been tacitly assumed by most of the authors mentioned that some sort of resonance is necessary to make probable the existence of internal waves of large amplitude. Indeed, if we have to accept the opinion of KRAUSS (1958a) that differences in depth of an iso-line in a section of 300 meters should be interpreted as due to internal waves of 150 meters amplitude, the intuitive and immediate reaction would be that this must be a result of resonance. The specific case considered by KRAUSS concerns sections taken across the Norwegian current at about 70°N. As frequently indicated above, similar "waves" are often found also in our sections across the same current farther south, and the apparent amplitudes are as large as those conceived by KRAUSS. The similarity of our sections and those of KRAUSS makes it probable that the large "waves" in all of them are due to the same cause. If this common cause should be resonance of internal waves, this would mean that the conditions for resonance are present fairly often and under rather varying circumstances, whereas for the surface tides the conditions for resonance occur only in rare and special cases. Before passing to the discussion of our sections, I should like to point out another circumstance in connection with the theoretical possibility (or impossibility) of internal waves of such large amplitudes. The theory of the two-layer model requires a certain relation between the particle velocity u_1 in the upper layer and the amplitude Z_2 of the discontinuity surface (see e.g. KRAUSS, 1957) which may with sufficient approximation be written:

$$u_1 = \frac{\sigma}{\varkappa h_1} Z_2 \cos(\varkappa x - \sigma t)$$

the notations being the same as on p. 18.

This relationship is derived from the equation of continuity, and thus applies to all kinds of internal waves. If forced waves are not considered, the wave length L is prescribed by

$$L^2 = \frac{(2\pi)^2}{\sigma^2 - f^2} \cdot \frac{g h_1 h_2}{h_1 + h_2} \cdot \frac{\varrho_2 - \varrho_1}{\varrho_2}$$

(EKMAN, 1931; DEFANT, 1932). For the semidiurnal period (12.4 hours) and reasonable values for h_1 (500 m), h_2 (2000 m) and $\frac{\varrho_2 - \varrho_1}{\varrho_2}$ ($5 \cdot 10^{-4}$), we obtain a wavelength of 144 km. With $Z_2 = 150$ m, we obtain an amplitude of u_1 of the order 100 cm/sec. It is utterly improbable that such strong alternating (or rotating) currents should exist in the region of our surveys. Several series of current measurements have been carried out in this region, and none of them can be interpreted as incorporating alternating currents of even far smaller amplitude. If they were as frequent as indicated by the frequency of "waves" in the sections, they would also doubtless have been noticed by fishermen and navigators.

There is also a more general point to be kept in mind: For waves of amplitudes as large as assumed by KRAUSS, the non-linear terms in the equations of motion are no longer negligible. If Z_2 is the amplitude of the discontinuity surface (in the two-layer model) and h_1 is the thickness of the upper layer, the quotient between the non-linear and the linear acceleration terms will be of the order Z_2/h_1. As h_1 may be taken to be 300—500 meters, this quotient may attain values up to 1/2, so that the approximation of the linear theory is not very good.

Now let us consider the results of our surveys (Figs. 3—11) from a practical point of view. We have mentioned above the possibility that the apparent "eddies" may be caused by the observations having been taken in different phases of an internal wave. This is entirely possible, but it requires a very special timing of the observations which is likely to happen only rarely. In point of fact, however, such "eddies" are found so frequently that it would be an extremely queer coincidence if the observations in all these cases should have been so timed, with respect to the phases of an internal wave, that the picture of an "eddy" resulted. The present author is therefore inclined to think that the eddy-like configurations in our horizontal charts, and the corresponding large "waves" in the vertical sections, really are eddies. It is not so easy to say whether the eddies are moving or stationary, but both sorts probably occur. In some of the cases where a survey has been repeated shortly after the first one was finished, eddies have been found to have disappeared from the area indicating that the eddy has moved away from the area, or they have been found in a different position. We have also a case when an eddy has not seemed to move within an period of two days. This is the case of the eddy in one of the surveys of 1957 (Fig. 11), situated at 62°35'N, 1°20'E.

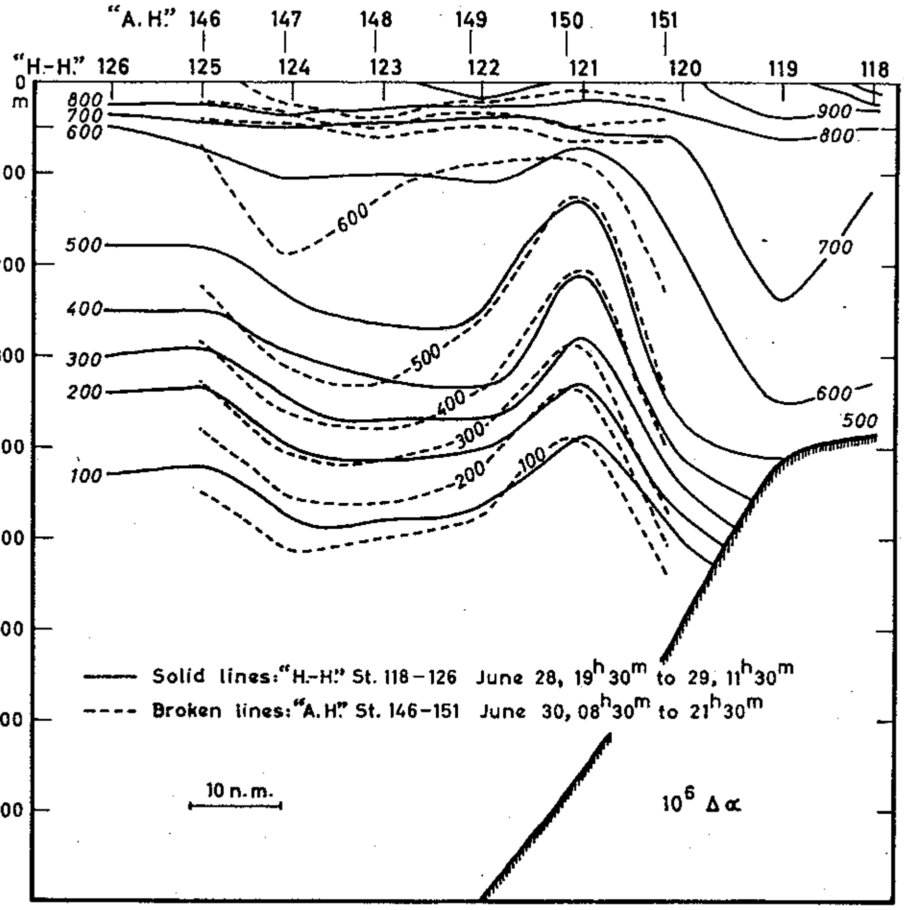


Fig. 16. Distribution of $\Delta\alpha$ in two occupations of the same section, 1957.

This eddy was crossed twice, first by the "HELLAND-HANSEN" section st. 118—126 between 19^h30^m, June 28, and 11^h30^m, June 29, and then by the "Armauer Hansen" section st. 146—151 between 08^h30^m and 21^h30^m on June 30. The two sections were very nearly along the same line, and the distributions of $\Delta\alpha$ are seen in Fig. 16. It seems as if the eddy centered around st. "H.-H." 121 and st. "A. H." 150 remained stationary for the period between the two occupations of the section. Although the form of the isolines in Fig. 16 at first sight suggests an internal wave of large amplitude, a closer study makes it hard to maintain such an interpretation. The station "H.-H." 121 was taken at 00^h30^m on June 29, and the station "A. H." 150 at about the same position was taken 42 hours later. As the station "H.-H." 121 must be at about the top of the wave, if such existed, an observation taken 42 hours later at the same position should have given a position of the iso-lines so much deeper as to be easily recognized. This applies both to the semidiurnal and the diurnal period. The feature in question will thus most probably have to be considered as a (semi-) permanent configuration. The fact that this "eddy" seems to be stationary may have some connection with the strong influence of Arctic water this year (1957), mentioned earlier. It appears that this "eddy" is connected with the easternmost extension of a tongue of Arctic water intru-

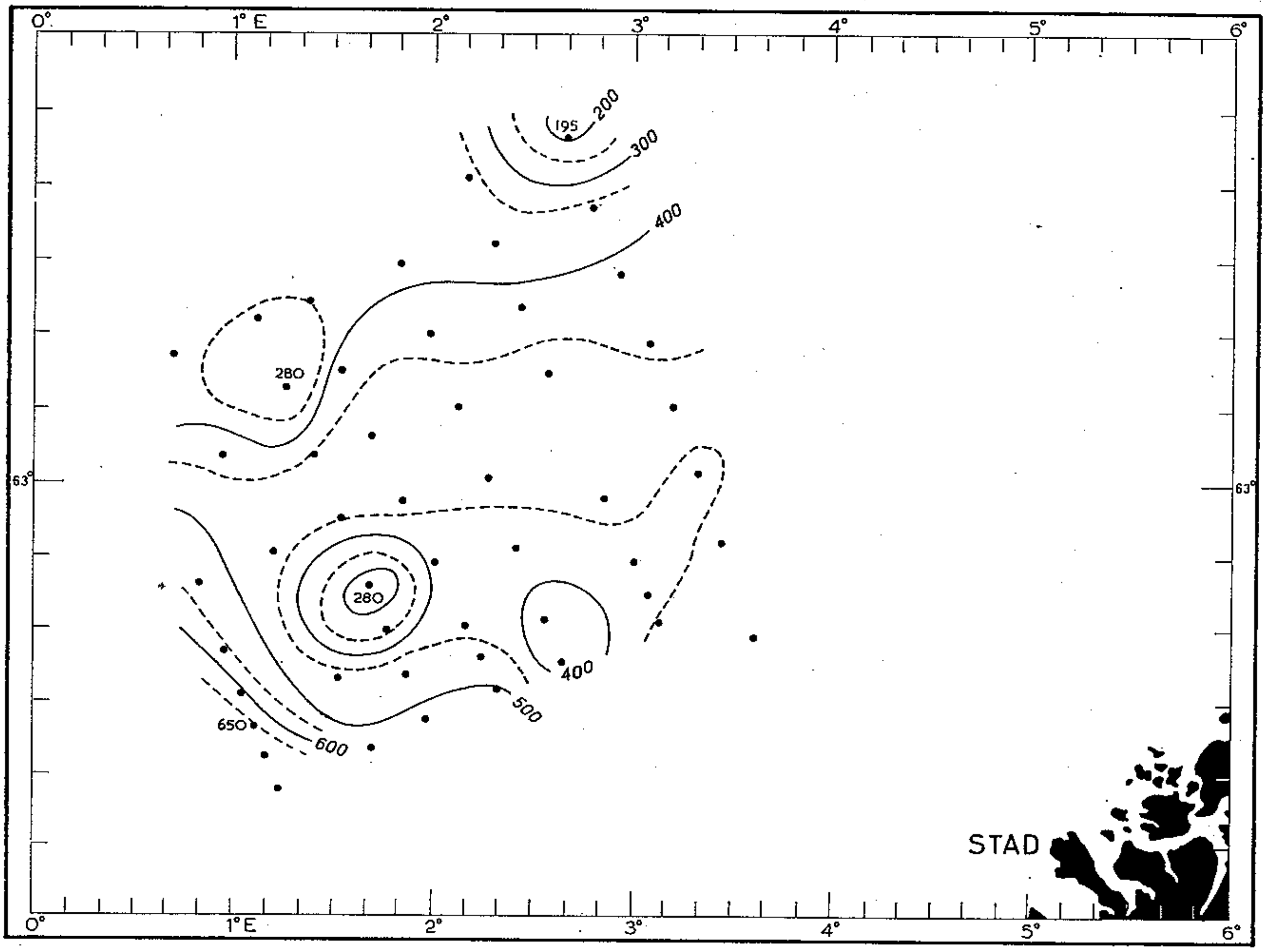


Fig. 17. Depth in meters of the 4° C isotherm on survey April 27—May 2, 1957.

ding from the west. The salinity minimum characteristic of that water appears at about 300 meters depth both at station "H.-H." 121 and station "A. H." 150, and at three stations to the west of them, namely "A. H." 141 and "H.-H." 131 and 132 (For positions see Fig. 11).

In the following sections, we shall treat in more detail a number of attempts to establish a connection between the density field and currents observed by various methods. This ambitious aim has not been reached, however, insofar as no general rules and no quantitative agreement have been found. On the other hand, a number of observations have been made which may in themselves be of considerable interest.

3. The joint cruise of "Discovery II" and "Armauer Hansen" in 1956. The first investigations of this kind were made in 1956 when the "Discovery II", using SWALLOW's (1955) neutral-buoyancy float, participated with the "Armauer Hansen" in a combined survey-current measuring program. Because of bad weather conditions, the original plan could not be carried out in full. The idea was to locate an eddy by oceanographic surveys, and to measure currents in specified parts of the eddy. Such an eddy

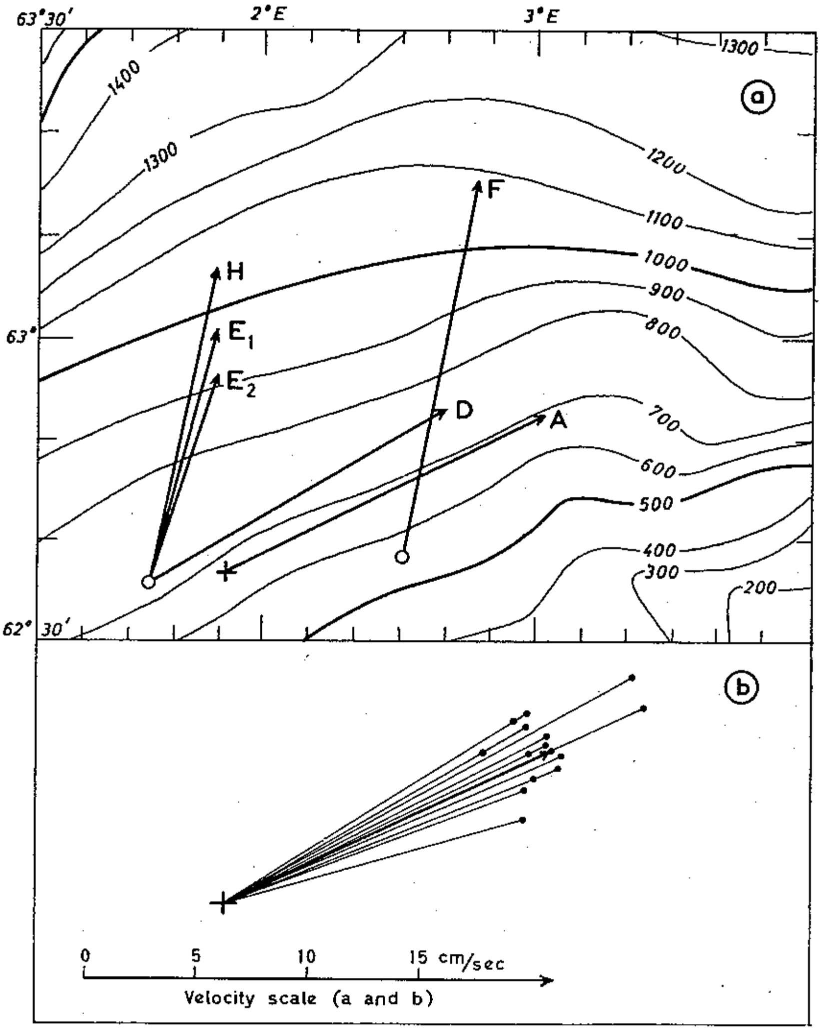


Fig. 18. a) Mean velocities from current measurements in May 1956. *A.* Bottom current, "A. H." May 10—11. *D.* "Pinger" at 440 m, "D. II" May 8—9. *E. 1.* "Pinger" at 480 m, "D. II" May 10—11. *E. 2.* Same "pinger", with later single fix included (see text). *F.* "Pinger" at 370 m, "D. II" May 13. *H.* Current cross at 450 m, "D. II" May 10—11.
b) Individual bottom current measurements, "A. H." May 10—11.

was located in the survey made by the "Armauer Hansen" during the days April 27 to May 2 (Fig. 17) in the position 62°45'N, 1°40'E, where the bottom depth was 780 meters. When, after 5 days delay, the current measurements were going to start, there was obviously no such eddy at that position. The 4° isotherm which in the center of the eddy of the first survey had been found shallower than 300 meters, was now found at about 500 meters. As time did not permit a complete re-survey of the area before the start of the current measurements, is was decided to carry out the current measurements at a position some 10—15 miles south of the position where the center of the eddy had been found during the first survey. The bottom depth was about 600 meters at that place. In spite of the continually rough weather, several series of measurements were carried out during the days 8—11 May. The "Discovery II" released and tracked

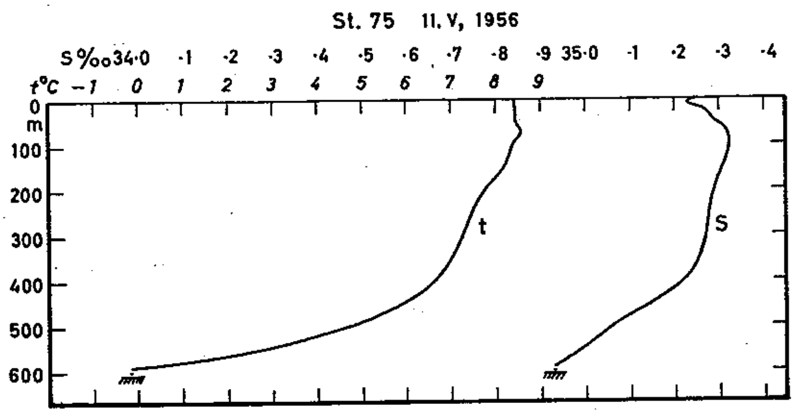


Fig. 19. Temperature and salinity curves at the bottom current measuring station.

three Swallow floats and one drogue (current cross), and the "Armauer Hansen" made a series of bottom current measurements. The results of the "Discovery II" measurements have kindly been put at my disposal by Dr. Swallow. The first "pinger" was at a depth of 440 meters and was tracked for 21 hours (from 17^h on May 8 to 14^h on May 9). The mean velocity was 15.4 cm/sec toward ENE (59° true). The second "pinger" was released at 11^h on May 10, and was tracked more or less continuously for 22 hours. Its depth was determined as 480 meters, and the mean velocity was 11.6 cm/sec toward NNE (15° true). A single fix on this "pinger" was obtained 11 hours later, it had then drifted from the previous fix with a mean velocity of about 7 cm/sec in direction 30° true, so that the overall mean (33 hours) for this "pinger" was 9.8 cm/sec toward NNE (18° true). A current cross at 450 meters drifted from 17^h on May 10 to 01^h30^m on the 11th slightly faster than the "pinger", 14.3 cm/sec in direction 12° true. The third "pinger" was released 23 miles to the east of the first two "pingers", at 01^h on May 13, and was tracked for 13 hours, giving a mean velocity of 17.4 cm/sec in direction 11° true. Its depth was determined as 370 meters.

Thus, two of the "pingers", and the current cross, all drifted in nearly the same direction, about 15° true, whereas the first "pinger" drifted in a direction about 45° to the east of the others. The scalar velocities were between 10 and 17 cm/sec (Fig. 18). It must be assumed that the "pingers" have all been in the lower part of the Atlantic water, or in the upper part of the transition layer.

The bottom current measurements made by the "Armauer Hansen", on the other hand, were made in the cold bottom water (1.5 meters above the bottom) with the Mosby bottom current meter, a picture of which is seen in S I, p. 18 (Mosby, 1953). A hydrographic station taken just after the end of the measurements is shown in Fig. 19. During the period 14^h on May 10 to 01^h on May 11, a number of 16 bottom current measurements were made. All the measurements gave currents roughly along the bottom countour lines, with velocities between 13 and 20 cm/sec. The resultant was 16 cm/sec in direction 60° true (Fig. 18). The measurements were made about 7 miles east of the first measurements of the "Discovery II", and at a depth of 610 meters. The bottom current measurements were within the drift period of the second "pinger" and the current cross, and a comparison shows that the bottom current was in fact stronger than the current at 450 meters. There was also a distinct difference in direction, the bottom current being directed about 45° more easterly than the midwater current at the same time. All the results are shown in Fig. 18. (The bottom contours in this figure may, however, in some places have to be shifted slightly). No attempts will be made to try to relate the current measurements to the hydrographic surveys made 10 days before and 10 days after the current measurements (Figs. 8, 9), as such an attempt is judged to be futile.

4.The investigations of 1957 with "Armauer Hansen" and "Helland-Hansen". The investigations of 1957 were conducted with the two ships "Armauer Hansen" and "Helland-Hansen" in June—July, again in the area off Stad on the Norwegian west coast. The investigations were carried out as a combined survey — current measurement programme. The first part of the programme is related to the survey shown in Fig. 10. After this survey had been completed, the "Helland-Hansen" anchored in 1600 m of water, at the position 63°30'N, 1°26'E, and measured the currents from June 15 to June 22 (current measuring station "A"). Simultaneously, the "Armauer Hansen" made hydrographic observations in a grid around the anchor station. The second part of the investigation is related to the survey shown in Fig. 11. After the completion of that survey, the "Helland-Hansen" dropped anchor in 600 m of water, position 62°29'N, 1°41'E and measured the currents for 10 days (July 2 to July 12, current measuring station "B") while the "Armauer Hansen" made hydrographic observations.

a. The current measurements at St. "A", 1957. This station is marked by a cross in Figs. 10 and 12. As seen from Fig. 12, the station is in the region of the Atlantic water without an intermediate salinity minimum. The observations were made either by means of Ekman singel reading currents meters, or at greater depths by Ekman repeating current meters. Observations were made at short intervals at 10 m and 25 m with the single reading instrument. At 200 meters, masurements were made with the same instrument at long and irregular intervals. At the 500 meter level, observations were made with the Ekman repeating current meter (EKMAN, 1926), fitted with a clockwork intended to release the balls at regular intervals within a measuring period between 2½ and 3 hours. Between these periods there were breaks in the record with a duration of about 1 hour, or a little less, with a few breaks of longer duration.

The anchoring was made from the bow at a depth of 1600 m, with 1880 m of anchor wire paid out. This gives a maximum radius of swing of about 1000 m. In reality, this maximum will not be obtained, as the anchor wire will probably always be curved even in strong currents.

The current measurements at this anchor station will not be treated in full, but only to the extent that they are of value to a comparison between the observed currents and

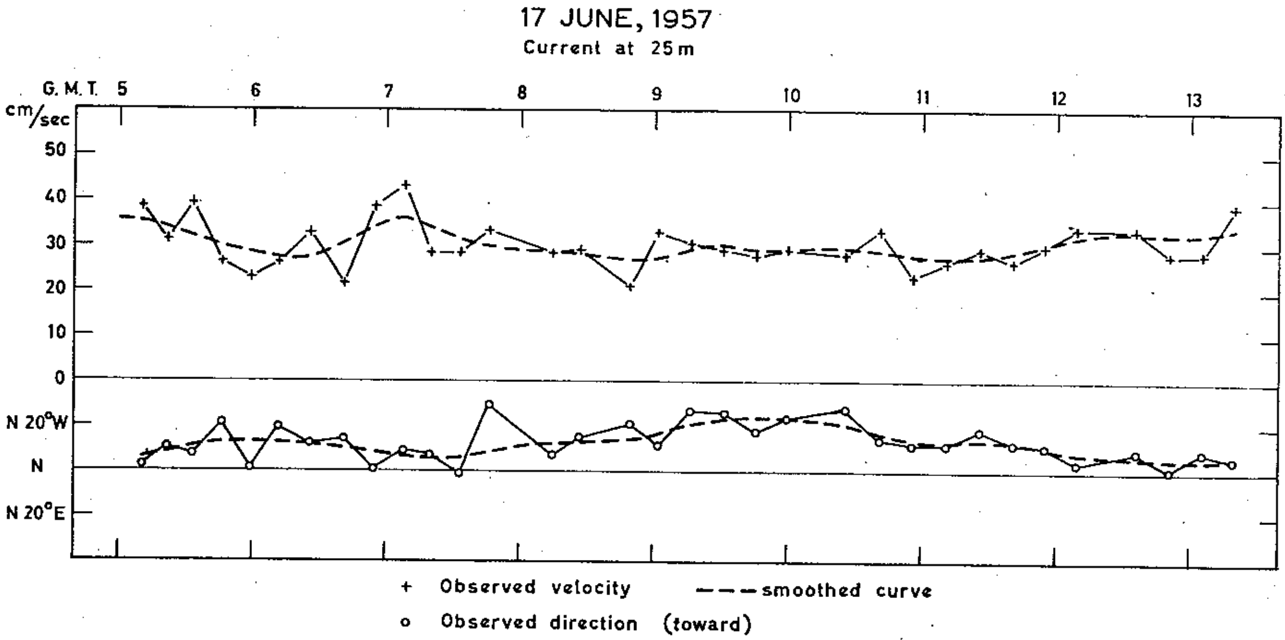


Fig. 20. Anchor station "A", 1957. Part of the original plot of the current measurements at 25 meters, together with the smoothed curves.

the computed geostrophic current. In the grid of hydrographic stations made by the "Armauer Hansen", the maximum depth of observation was 500 meters. The geostrophic currents are therefore computed relative to that level. In order to exclude as far as possible the influence of the wind, we have chosen to use the current measurements at 25 meters together with the measurements at 500 meters, and the difference in observed current between those two levels will be compared to the geostrophically computed current at 25 meters (dbar) relative to 500 meters (dbar).

The measurements at 25 meters were made as frequently as possible. Each measurement with the Ekman current meter lasted as a rule between 5 and 10 minutes, depending on the strength of the current. On an average, between 4 and 5 measurements were made each hour. On a few occasions, measurements had to be suspended for some time (up to 70 minutes) due to instrument failure. But on the whole, the series at 25 meters may be regarded as a fairly continuous one. A plot of the original observations gives a rather irregular picture (as an illustration, part of this plot is reproduced in Fig. 20). This irregularity inevitably leads to the old and much-discussed question of the influence of the ship's movements on current measurements from an anchored vessel. The investigations under discussion cannot add much to this discussion, as we had no means of detecting the movements of the ship. From works by different authors, it emerges that the movements of a ship at anchor may easily reach velocities of a magnitude high enough to explain the apparant irregularities in the current velocity observed at our anchor station. WITTING (1905) and NANSEN (1906) were aware of this, and Nansen concluded that current measurements from a ship at anchor should be regarded with the utmost scepticism. An extensive review of the question is given by THORADE

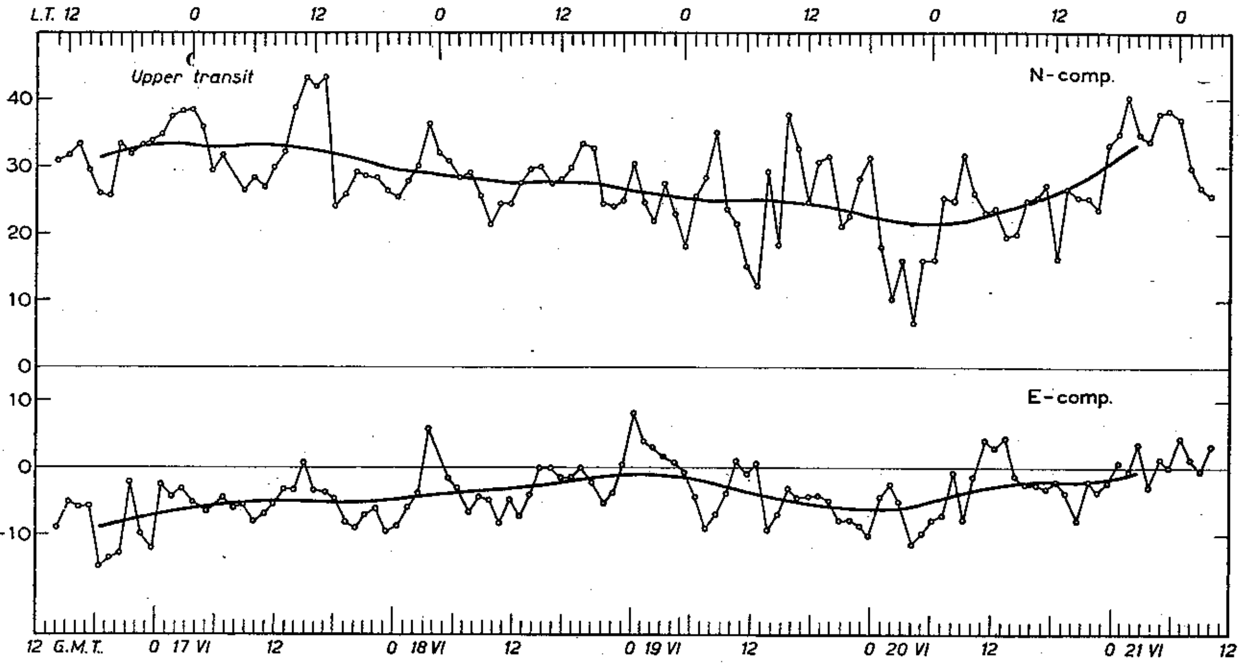


Fig. 21. Anchor station "A", 1957. Hourly values (lunar hours) of the velocity components at 25 meters.

(1933). A certain amount of information on the magnitudes of the ship's velocity may be attained from the cases where means have been available to ascertain the ship's position with sufficient accuracy. WITTING (1905) and NANSEN (1906) both concluded from their measurements that the velocity of the ship's movements might at times be of the order 10 cm/s, as was also found by the "Poseidon" in the Baltic (se THORADE, 1933, p. 2870). In 1953, the present author conducted current measurements in the Trondheimsfjord and made very accurate and continuous determinations of the ship's position by means of two theodolites on shore. Although the small vessel was anchored fore and aft, and the depth was moderate (about 100 meters), ship movements of velocity 5—10 cm/s were frequent (unpublished report). We have also in a few cases information on the movement of ships anchored at greater depths. DEFANT (1932) tried from astronomical observations to infer the magnitude of the velocity of the movement of the "Meteor" at several anchor stations at great depths, and arrived at values up to 10 cm/s. On the 15th cruise of the 5000 tons "Vitjaz" in 1953, the ship's movements were determined at deep sea anchor stations by a radiolocation method (SISOJEV, 1956; BURKOV, 1957), and ship's velocities were found which in several cases considerably exceeded 10 cm/s. Thus, at anchor station 2227 (depth 3300 m, with 4800 m wire paid out), mean ship's velocities in excess of 10 cm/s were recorded during periods totalling more than 6 hours. In several cases, the velocity was higher than 20 cm/s.

In view of these experiences, there seems to be little point in trying to analyse the short-time variations (from one measurement to the next) in the measurements, as they may as well be caused by the ship's movements as by real variations in the current. All the measurements at 25 m were, therefore, plotted in a diagram (velocity and direction separately) on a large scale, and a reasonably smooth curve was drawn through the points. This method was used instead of the more objective one of forming mean values of, say, 3 or 5 measurements, because the interval between measurements was rather variable. Also, the smoothing has been carried out on the directly observed values (velocity and direction) instead of the components of the current. The latter method is usually considered to be the formally more correct one, but the difference will in this case be insignificant, as the variations in direction are rather small. These diagrams will not be reproduced in full, but a representative part of them is shown in Fig. 20. Hourly values (lunar hours) were read from the smoothed curves, and N and E components computed and plotted (Fig. 21). Running means for 12 hours and for 24 hours were computed. There was little difference between the two mean curves, and the mean curves in Fig. 21 were drawn intermediate between them.

As previously mentioned, the measurements at 500 m were made by the Ekman repeating current meter in series each of approximately three hours duration. As a rule, there was a break of between 1/2 and 1 hour between the series, and in some cases (four) a whole series is missing. In some of the series, there are parts where the directions seem to be questionable for some unknown reason. The measurements at 500 meters are, therefore, not fit for a representation in a continuous time series (curve) similar to that given for the measurements at 25 meters in Fig. 22. The different mode of working of the two types of current meters, with respect to the indication of direction, should also be made clear: In the single-reading Ekman current meter, several balls (three for each 100 revolutions of the propeller) will fall into the compass box during each measurement, so that a mean can be worked out, and there is no possibility that the direction should be determined by a "stray" ball indicating a direction very different from the average direction of the current meter during the measuring period. In the Ekman repeating current meter, on the other hand, the direction is indicated by balls falling at the same time as the balls giving the number of revolutions, in our case only every four minutes (for details, see Ekman, 1926 and 1953). In addition, in a few of the series the spread of the direction is so great and inconsistent that it is difficult even to define a mean direction for a reasonable period (we have used 20 minutes). In these cases, the spread cannot be attributed to the ship's movements, as the spread of the balls of the single-reading instrument at 25 m was not greater than usual, and much less than at 500 m. This applies especially to the series 2, 4, 9, 15, 16, 18, 21, 22 (Fig. 22). It must be assumed that an instrumental error of unknown origin has affected the recording of direction in these cases. Concerning the magnitude of the velocity, the construction of the instrument is such that sometimes an intelligent guess has to be made for the number of revolutions. However, an error in this respect will be smoothed out in the long run, as revolutions missing at one reading will be added to the next reading.

For the reasons outlined above, the observations at 500 meters are presented as means of 5 and 5 individual readings, i.e. 20 minutes' means. For an easier comparison between the current at 25 and 500 meters, velocity and direction at 500 meters are

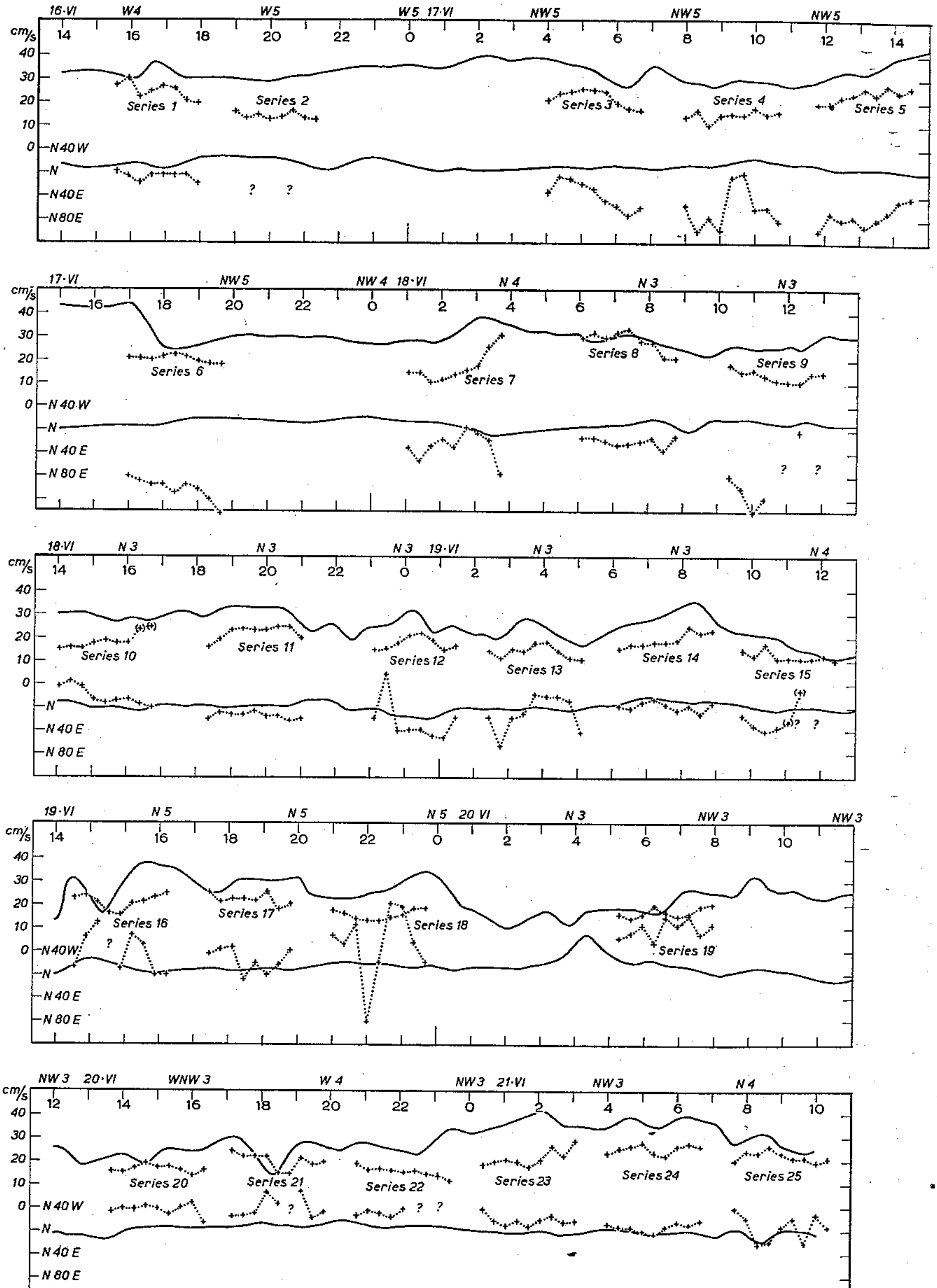


Fig. 22. Anchor station "A", 1957. Velocity and direction of the current at 25 meters (smoothed curves) and at 500 meters (20 minutes' mean values).

plotted together with the smoothed curves at 25 meters in Fig. 22. The current measurements started at about 20^h on the 15th, but the anchor probably did not hold before 14^h on the 16th. After that time, there was no indication that the ship was dragging the anchor. Accordingly, only the measurements made after 14^h on the 16th have been considered. The wind direction and strength (in Beaufort) are entered just above the time scale in Fig. 22.

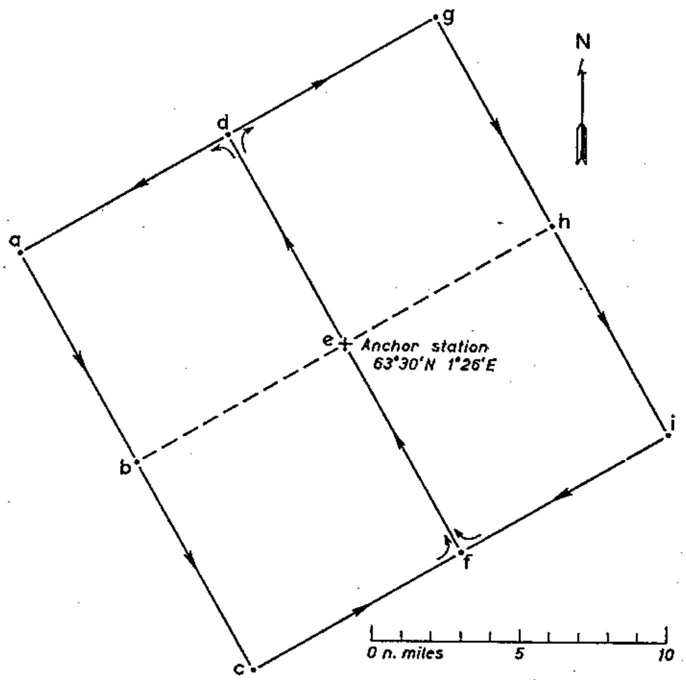


Fig. 23. Pattern of hydrographic stations around the anchor station "A", 1957. Arrows indicate the course of the "Armauer Hansen".

b. Comparison between observed and computed currents. The hydrographical observations to be compared with the current measurements were made by the "Armauer Hansen" in a pattern intended to determine the dynamical topography of a limited area around the anchor station. Stations were occupied at the positions denoted by a, b, . . ., i in Fig. 23, (e is the anchor station of the "H.-H."). The stations were positioned on lines running N 30°W (true), spaced 8 nautical miles apart. The sequence of the stations was *e* d a b c f *e* d g h i f *e*, and a total number of five and a half rounds was made. The positions are subject to considerable uncertainty, as most of the time the weather did not permit astronomical observations to be made. A check on the positions was obtained each time the ship went up to the anchor station of the "Helland-Hansen". It appeared from these checks that the ship had often been displaced toward the NNW, in accordance with the surface current measured at the anchor station. The ship's course was nearly always headed almost directly on the anchor station, but the logged distance was in error by an amount varying between 0 and 4 miles, although precautions were taken to correct for the current set. Generally, the positional error should be smallest at the station taken just after the anchor station, (that is, station d), and then gradually increase until the station taken just before the next occupation of the anchor station (that is, station f). No practicable method has been found, however, for the correction of the positions so as to give more correct values of the dynamic depths in the points of reference a, b. . .i. When reading the discussion of the hydrographic observations as given below, one should therefore keep in mind this limitation on the accuracy of the presentation.

For the comparison of measured and observed currents the following procedure was adopted. The difference in dynamic depth between 25 and 500 dbar was computed

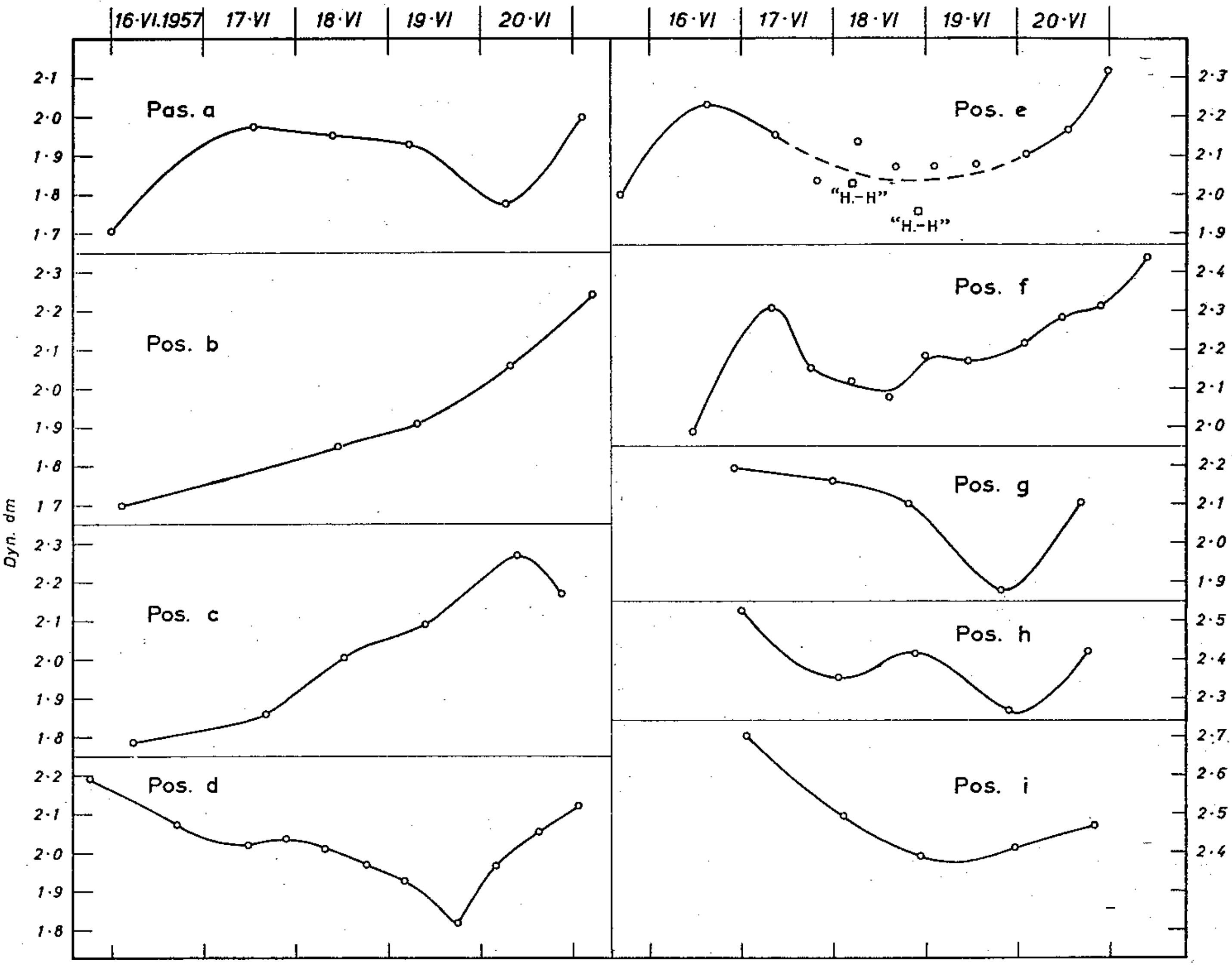


Fig. 24. Dynamic depth difference between 500 and 25 db plotted against time for the different positions in the station grid.

for each occupation of the different stations. As observations were not made at 25 m, but at 0 and 50 m, this involved some interpolation. For each of the stations a, b...i, this computed difference was plotted against time, and a reasonably smooth curve was drawn between the points (Fig. 24). The dynamic depths to be used for the comparison were taken from these curves at certain points of time, the selection of which will be explained below. It is evident from the curves Fig. 24 that, at each of the 9 positions, the dynamic depth difference varied between rather wide limits during the period of investigation. It is also possible that conditions at a given place may have been more rapidly variable than indicated by the curves drawn in the figure. This is exemplified at the anchor station (e), at which position stations were taken most frequently, a couple of times also by the "Helland-Hansen". Observational errors, especially in the salinity, may account for variations up to 0.03 dyn.dm, roughly. Internal waves, if present, could also affect the values. The general trend of the curves, however, as well as the more pronounced differences between the various positions, are considered to be connected with real differences in the properties of the water masses. Thus, the conspicuous notch on the 19th in the curve for position d can be shown to have been caused by an intrusion of arctic water, with a salinity minimum of 34.85‰ at 300 meters. The same is true for the corresponding feature in the curves for positions a and g. In spite of the possibilities of errors mentioned above, there is a significant general trend in the dynamic topography of the area, persisting throughout the period of investigation. There is a "High" in the eastern part of the area, especially at positions g and h, and a "Low" in the western part of the area. This configuration gives a geostrophic current in a northerly direction. This is found in all the cases used for comparison with the current measurements (Figs. 25—26), and is also evident in the topography constructed from the average values of $\Delta D_{25\text{-}500}$ (Fig. 27). The geostrophic currents in Figs. 25—26 (thick arrows) were constructed by means of the components normal to lines b—h and d—f, using the differences in dynamic depth between positions b and h and between d and f. This means a certain averaging, this time with respect to area, but it has been preferred as a more objective and reproducible method than an evaluation on the basis of the contour lines would have been. It is evident from the above discussion that there may be a considerable difference between the velocities in Figs. 25—26 and the "real" values of the geostrophic current at position e, for a multitude of reasons: inaccurate positions; observational errors; internal waves; the interpolation in time of the dynamic depths; the averaging in space.

Now let us turn again to the current as measured at the anchor station (e). As mentioned before, the proper quantity to be compared with the geostrophic currents will be the vectorial difference between the currents at 25 and 500 meters. It is fairly obvious that it is not appropriate to use single current measurements for this purpose, or even averages over a period of, say, 20 minutes. The measurements to be used should as far as possible be free from random variations, and this requires the use of averages over somewhat longer periods. Furthermore, the variations in the current within the period of averaging should be as small as possible, so as to be compatible with the conditions for the geostrophic currents (steady state). In this connection, attention is directed to the previously mentioned fact that, at 500 meters, the reality of the measured directions are, in some of the series, subject to doubt. With these considerations in mind, it has been assumed (1) that averages for a complete series at 500 meters (about 3 hours) should be used in connection with the corresponding averages at 25 meters, and (2) that only those series should be used in which the variations are fairly small and steady. The following series have been selected (see Fig. 22): Nos. 3, 5, 6, 8, 10, 11, 14, 20. Of these series, the Nos. 5, 8, 10, 11, 14 and 20 are those which best conform to the above requirements.

Before we proceed to the comparison between observed and computed currents, it is necessary to consider the influence of tidal variations on the observed currents. The

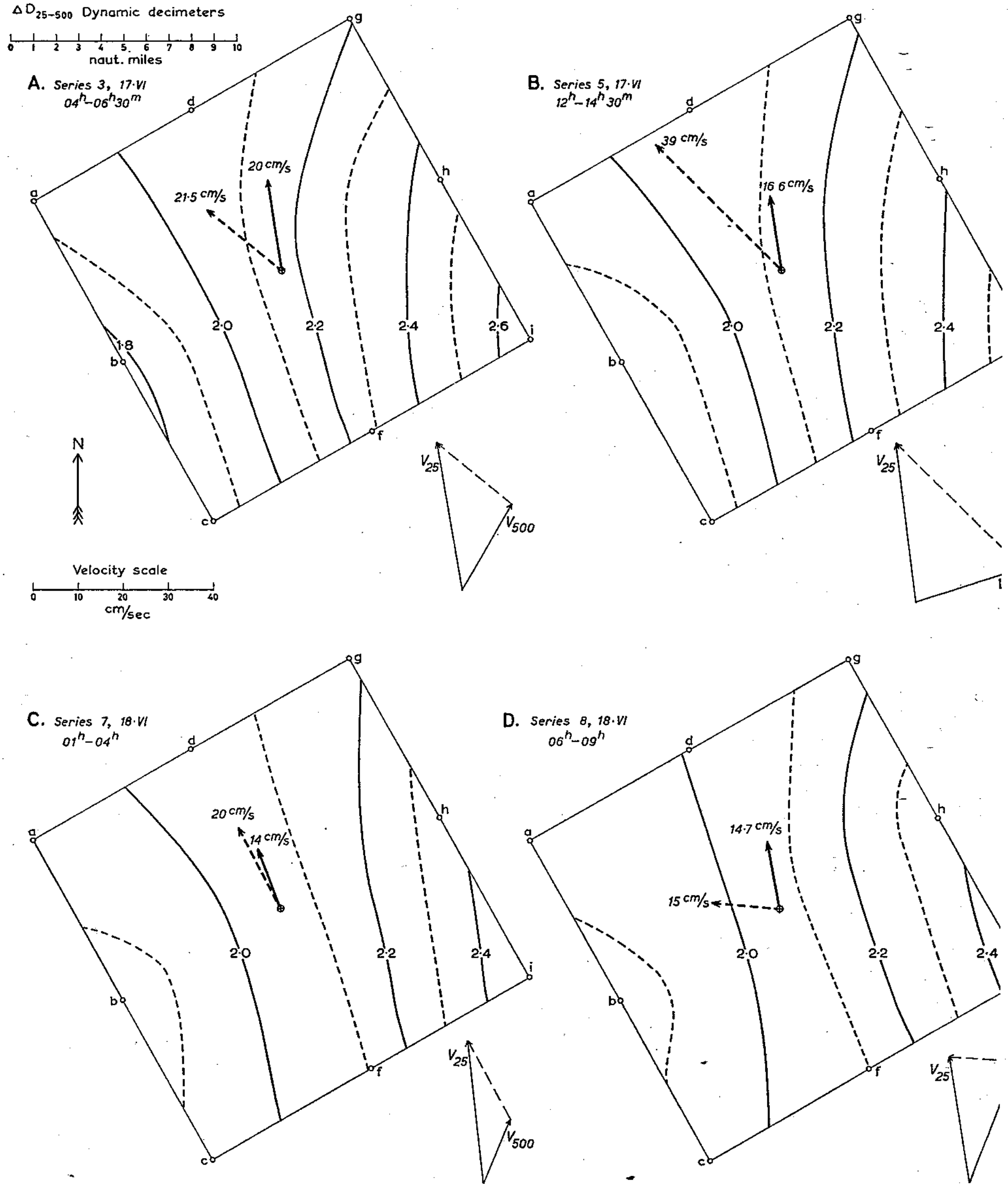


Fig. 25. Anchor station "A", 1957. Dynamic topography of 25 db relative to 500 db (dyn. dm) and observed currents at 25 and 500 meters. Thick vectors: geostrophic velocity. Thin vectors: observed velocity. Broken vectors: vector difference between observed currents at 25 and 500 meters. Series 3, 5, 7, and 8.

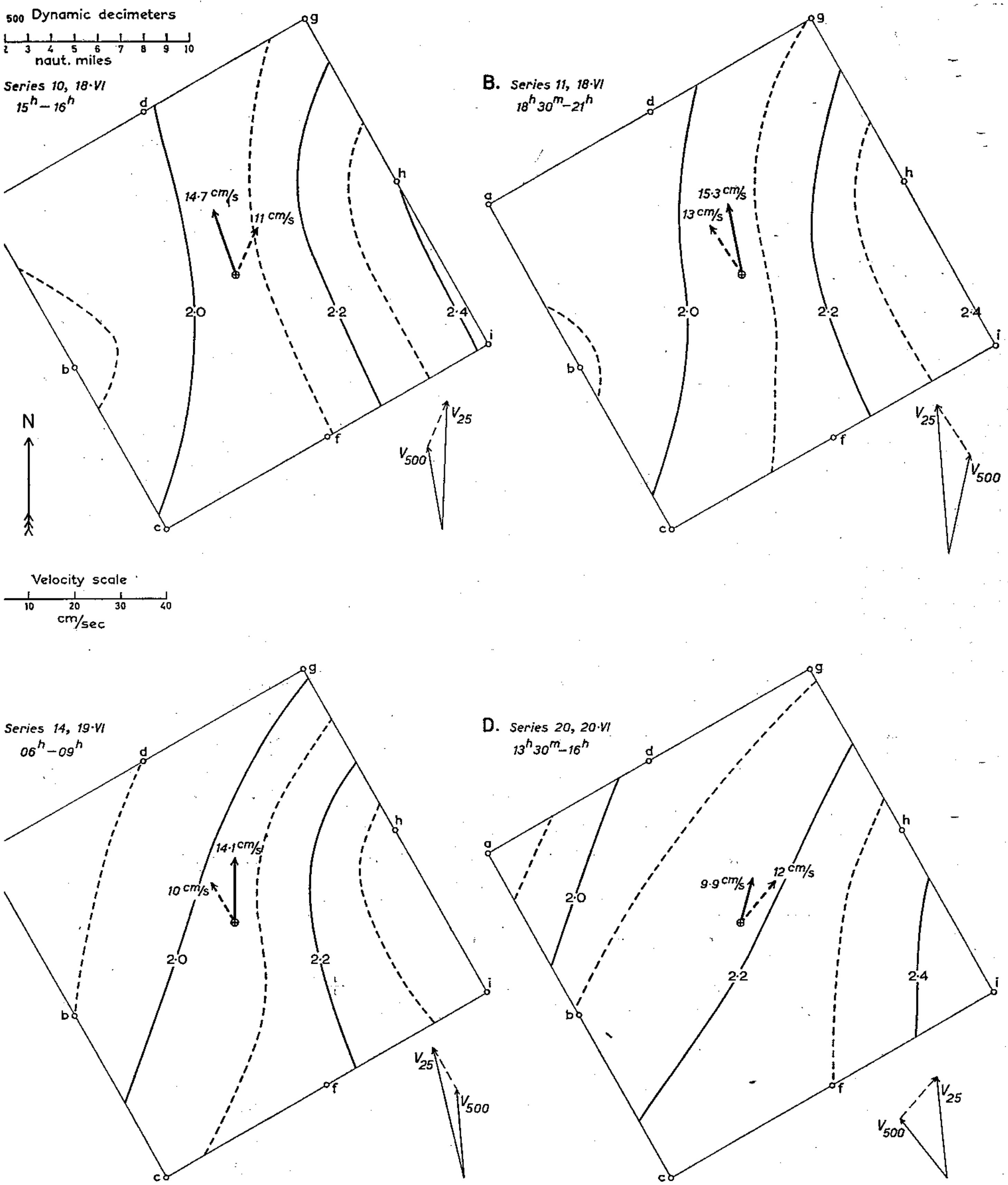


Fig. 26. Anchor station "A", 1957. Dynamic topography of 25 db relative to 500 db (dyn. dm) and observed currents at 25 and 500 meters. Thick vectors: geostrophic velocity. Thin vectors: observed velocity. Broken vectors: vector difference between observed currents at 25 and 500 meters. Series 10, 11, 14, and 20.

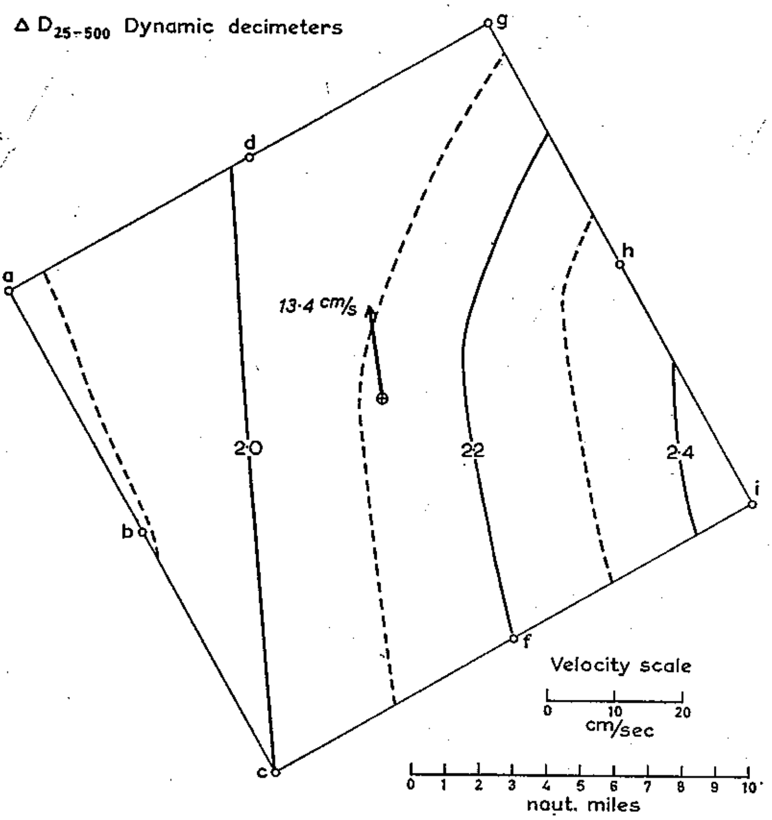


Fig. 27. Anchor station "A", 1957. Average dynamic topography (25—500 db) of surrounding area.

observations at 25 meters are sufficiently continuous to allow a harmonic analysis. This is not the case, however, for the observations at 500 meters. Thus, a procedure, common to both levels, for the elimination of the tidal part of the current, cannot be applied. This circumstance need not be a serious drawback, however, if the regular (as contrasted to internal) tidal wave obeys the ideal law for the distribution of velocity in a "long" wave, *viz.* that the velocity should be the same at all levels. If this is the case, the tidal part of the velocity should disappear in the difference $V_{25}—V_{500}$. Also, it is very difficult to find any significant tidal variations in the velocity components at 25 meters (Fig. 21). A harmonic analysis has been carried out. The analysed quantity was the hourly value (lunar hours) minus the variable long-time mean value (Fig. 21). The length of the analyzed period was 96 lunar hours. With an arbitrary time zero (2130 GMT on June 16, that is, 6 lunar hours before the moon's upper transit at Greenwich on June 17), the result was for the two components N and E:

$$V_N = -0.3 - 1.0 \sin \sigma t + 0.0 \cos \sigma t - 2.6 \cos 2\sigma t - 0.8 \sin 2\sigma t + 2.1 \cos 3\sigma t + 0.8 \sin 3\sigma t - 0.3 \cos 4\sigma t - 0.7 \sin 4\sigma t.$$

$$V_E = 0.0 - 0.4 \cos \sigma t + 0.2 \sin \sigma t - 0.3 \cos 2\sigma t + 0.9 \sin 2\sigma t - 0.3 \cos 3\sigma t + 0.7 \sin 3\sigma t + 1.0 \cos 4\sigma t - 0.3 \sin 4\sigma t$$

where $\sigma = 2\,\pi/24$, t in lunar hours.

As the semidiurnal lunar period is the dominating period in the tides in this region the high value of the eight-hourly term (in the N-component) indicates that the reality of the results may be open to doubt. This suspicion is strengthened by the result of the application of a significance test to the 12 hours period in the N-component. Similar tests have previously been applied to oceanographic time series by HAURWITZ (1954) and REID (1956). The test is based on a method due to SCHUSTER (see e.g.

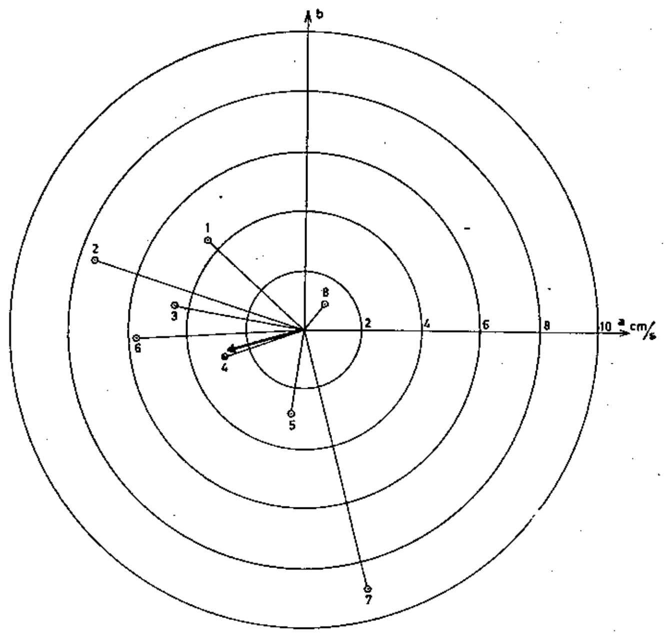


Fig. 28. Anchor station "A", 1957. Harmonic dial for N — component at 25 meters, semidiurnal lunar period. Abscissa: coefficient (a) of cosine-term. Ordinate: coefficient (b) of sine-term.

CONRAD and POLLAK, 1950). The basic idea is to compare the results of the harmonic analysis with the results of a similar analysis of a set of random numbers. The whole analysis interval is subdivided into partial intervals equal to the period under investigation, and each of these are analysed separately. An expression can then be derived for the probability that the amplitude obtained in the harmonic analysis might have been obtained from random data. Details will not be given here; readers are referred to one of the works mentioned above. In our case, a total of eight individual periods can be used for the test. The results of the analysis of the N-components of those eight single periods are presented as vectors (end-points numbered 1,2...,8) on a harmonic dial (Fig. 28). In Fig. 28, the thick arrow gives the vectorial mean (equivalent to the result of a harmonic analysis of the mean values for all 8 periods). Qualitatively, the dispersion of the points on the harmonic dial gives an impression of the degree of randomness in the results. In our case, the calculation shows that the probability that the amplitude found by the analysis is due to random data is 0.14, that is, the amplitude would occur in one out of seven cases from random data. Although there may be certain inadequacies in the test method, and although the choice of the "significance level" will always be subjective, the calculated probability is so large that it is not safe to regard the reality of the oscillation as established. Thus, there seems to be little point in attempting any form of correction for tidal currents, especially as a similar harmonic analysis cannot be made for the observations at 500 meters.

The results of the comparison between observed and computed currents for the periods selected in accordance with the previously stated principles (p. 31) are shown in Figs. 25—26. The computed geostrophic currents are entered as thick arrows with the point e as origin. From the same origin, the vector differences V_{25}—V_{500} are entered as broken arrows. The vectors V_{25} and V_{500} are as a rule averages for a whole period of measurements at 500 m, but in a few cases only part of the period has been used. The average relative topography, constructed by means of the curves in Fig. 24, is presented in Fig. 27. It is obvious from these diagrams that the agreement between observed and computed currents is not too good. However, in view of the uncertainties both in the measured and the computed values, a full agreement could no be expected; but the deviations sometimes seem to be too great to be explained by errors in observations and in methods of computation. Part of the deviations, sometimes probably the greater part, must be due to the fact that the basic assumptions in the comparison are not fulfilled. It would seem appropriate, then, to try to review the sources of error and the failings of the assumptions.

1) Errors in the computed values. The dynamic depths are computed by means of observed values of depth (pressure), temperature and salinity. Of these quantities, the salinity is the one most likely to be in error by any significant amount. If the salinity is in error by $0.02^0/_{00}$ at all depths at a station down to 500 meters, an error in the dynamic depth difference 0—500 m of 0.04 dyn.dm will occur. If, furthermore, there is an error of this magnitude, and in opposite direction, at both the positions used for the calculation of one of the components (c and h, or d and f), an error in velocity of about 2 cm/s will arise. Such an error presupposes a rather unlikely combination of the most adverse circumstances, and it is probable that the error in velocity caused by erroneous salinitities will not exceed 1 cm/s.

A second source of error in the computed values is the uncertainties of the positions. As previously mentioned, the errors in positions are always approximately in the direction N 30° W, and may at the end of a "semiround" attain values up to 4 nautical miles. The resulting error in velocity will thus mainly affect the component perpendicular to the line d—f. Because of the interpolation made to obtain a synoptic picture, it is difficult to make an accurate estimate of this error. If we take the worst possible case, the point f would be 4 miles nearer to the anchor station, whereas point d would not be shifted much. The effect on the velocity component normal to d—f would be an increase by 1/3, that is, up to about 2 cm/s. As this component is nearly always the smaller one, the effect on the magnitude of the velocity will as a rule be small, and the direction will be deflected a small angle cum sole. Only in a few cases, however, is there any possibility that this error can reach such a size as in the above example.

As previously mentioned, the occurrence of internal waves would influence the dynamic heights. This possibility cannot be dealt with here, as the data on hand give no means for detecting such waves.

To sum up, it seems that the greater time-variations in the geostrophic current, such as computed by Helland-Hansen's formula, are probably real. Thus, the difference between the computed currents in the first few series (16—20 cm/s) and those computed for the last two series (10 cm/s) is almost certainly not due to errors in observation and computation. On the other hand, the differences between series 7, 8, 10, 11, and 14, may be due to such errors.

2) Errors in the observed currents. The most obvious of such errors, i.e. the movements of the ship, has been discussed before (p. 25). It is to be hoped that in any case the short-periodic part of such movements has been eliminated by smoothing the curves. Another source of "error" is the tidal currents. Even if the result of the harmonic analysis cannot be considered significant, it cannot be concluded that such currents do not exist. However, the tidal current will probably cancel out in the differences formed to obtain the relative current. The internal waves, if present, again constitute

an unaccountable source of error. The accompanying horizontal currents will not necessarily be the same at all levels, and will thus not be eliminated in the relative currents.

3) Deviation from geostrophic conditions. If we look at the diagrams Figs. 25—26, it appears that there are large differences between the different series with regard to the agreement between computed and observed currents. In view of the above considerations, no very close agreement should be expected, however. In a few of the series, the difference is no greater than can be explained by the combined effects of the errors mentioned above. This applies to the series 11, 14, 20, and perhaps, also series 7. In other series, the difference is so great that it cannot be attributed to mere errors, e.g. series 5. The most probable reason for this fact is that the conditions for geostrophic currents are not fulfilled. One of the conditions is that there should be no acceleration. In such a small-scale experiment as the present one, it seems very likely that water movements would occur, partly in the form of moving eddies, which may not be in geostrophic balance. These are effects that would, in investigations on a larger scale, have to be classified as macro-turbulence. The accelerations may be divided into (1) local accelerations and (2) convective accelerations. If the local accelerations are judged by the curves Fig. 22, we may in some cases obtain values that are close to the value of the Coriolis acceleration (e.g. series 7, Fig. 22). In such cases, however, the velocity curves at 25 and at 500 meters usually have a similar trend, so that the effect on the relative current will be reduced (S I, p. 12). The convective acceleration may be judged roughly from the curvature of the dynamic isobaths in Fig. 25—27. However, the drawing of these lines is so subjective that no reliable estimate can be made. It should only be said that the curvatures must be greater than those found in Figs. 25—27 if the centrifugal acceleration is to exceed 1/10 of the Coriolis acceleration.

Another condition for the geostrophic computations is that there should be no friction. This condition excludes the Ekman wind current in the upper layers. As the current measurements were made at 25 meters, the influence of the Ekman current would be reduced to about 1/4 of the surface wind current, if the "frictional depth" is taken to be 60 meters. The wind force during the investigation was between 3 and 5 Beaufort from directions between N and W. With the usual assumptions on the magnitude of the wind stress (see e.g. SVERDRUP: The Oceans) this would mean a surface wind current between 10 and 15 cm/s, and a current of 3—4 cm/s at 25 meters towards directions roughly between west and south. In a few cases, this would suffice to explain the discrepancy between observed and computed currents. In the majority of the cases, however, the discrepancy is either too large or in a wrong direction to be explained by this factor. Thus, the effect of friction can only be regarded as another of the factors that may, or may not, contribute to the establishment of the said discrepancy.

In the diagrams Figs. 25—26, it is noticeable that the difference in direction between the observed current at 25 meters and at 500 meters is often so great that it is definitely outside the limit of errors. This fact would, in itself, indicate that non-geostrophic factors are of importance. It can be shown that for a horizontal geostrophic

Table 2. *Comparison between observed and computed currents for selected series.*

Series No.	Deviation of observed current direction from that of the computed relative current (degrees)			Scalar velocities in cm/s		
				Observed		Computed
	Current at 25 m	Current at 500 m	Vector diff. 25–500 m	Diff. between scalars 25–500 m	Scalar value of vector diff.	Relative current 25–500 m
3	< 5	47 R	45 L	11	21.5	20
5	< 5	70 R	35 L	10	39	16.5
7	15 R	45 R	5 L	16.5	20	14
8	< 5	30 R	70 L	0	15	14.5
10	20 R	10 R	40 R	10	11	14.5
11	7 R	24 R	22 L	10	13	15.5
14	15 L	5 L	30 L	10	10	14
20	21 L	53 L	25 R	5.5	12	10

R: To the right of the computed current.
L: To the left of the computed current.

current, if the divergence of the velocity is zero and the Coriolis parameter is a constant, the velocity will have the same (or opposite) direction at all levels. This is the "law of parallel fields" (Ekman, 1923, see also Defant, 1961, p. 477). If this law is valid in our case, it is obvious that the vector difference V_{25}—V_{500} should not be expected to agree with the computed geostrophic current in the cases where the difference in direction between the observed currents at 25 and 500 meters is large, This applies especially to the series 3, 5, 8, 20.

Now, we have several times stressed the wellknown fact that the geostrophic computations can only give relative currents. As regards the direction of the current, however, it follows from the above-mentioned theorem that the direction of the absolute geostrophic current will be the same as that of the relative current. In our case, this means that the currents at 25 m and at 500 m, as well as their vectorial difference, should be in the direction of the computed relative geostrophic current. This requirement is best fulfilled for the currents at 25 meters, as will be seen from Table 2.

At this depth the observed current has a direction that does not differ greatly from that of the computed relative current, at most 20 degrees, and several times less than 5 degrees. As a further illustration of this fact, Fig. 29 shows the two directions plotted as functions of time. Although the variations of the direction at 25 meters are irregular, they are small, and the direction rarely differs by more than 20° from the direction of the computed relative current. For the current at 500 meters, as well as for the vector difference of the observed currents, the corresponding differences are much larger, up to 70°, and very irregular (Table 2). In the last three columns of Table 2, we have entered the differences between the scalar values of the velocities at

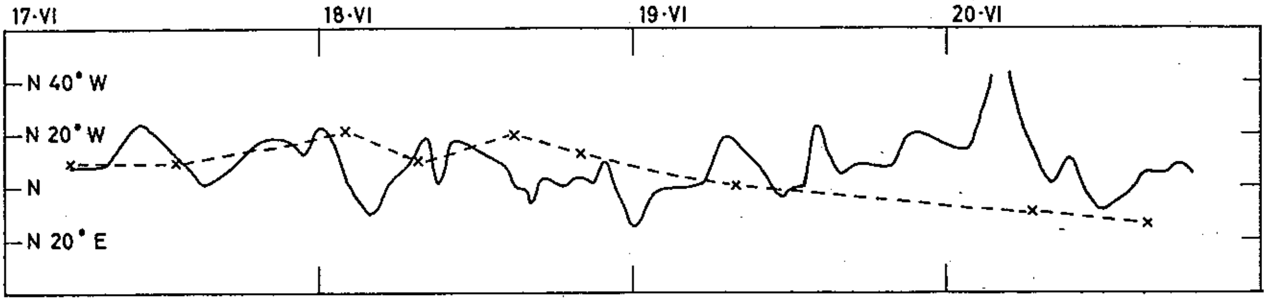


Fig. 29. Anchor station "A", 1957. Direction of computed relative current (x — — — x) and of observed current at 25 meters (——————).

25 and at 500 meters; the scalar values of the vector difference 25—500 meters; and lastly the scalar value of the computed relative current 25—500 meters. It is seen that, on the whole, the differences between the scalars give no better agreement with the computed current than do the scalar values of the vector difference, although one wildly discordant value of 39 cm/s in series 5 is reduced to a more reasonable value.

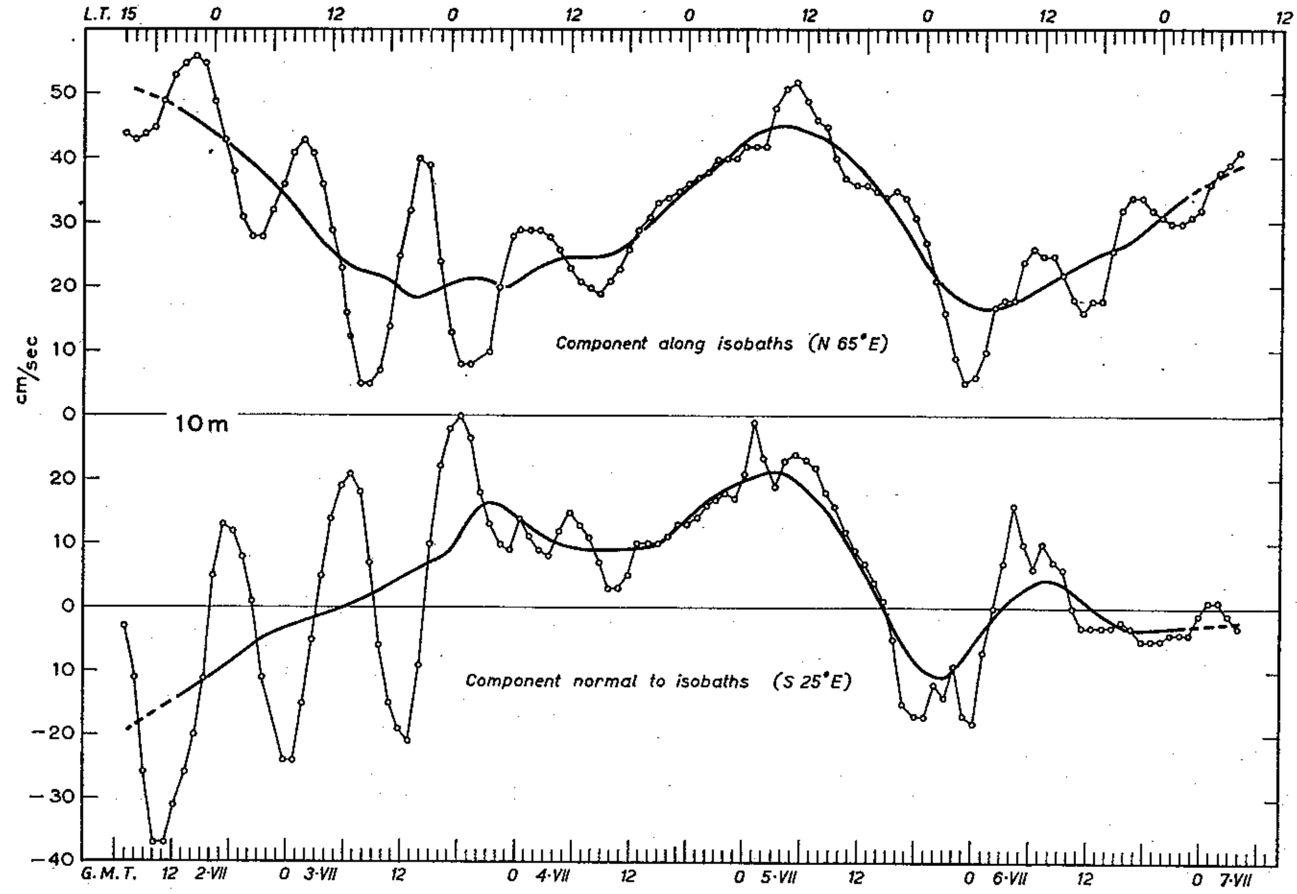
The preceding discussion seems to indicate that the deviations from geostrophic conditions are most likely to be caused by a deviation from such conditions at the 500-meters level. It is very difficult to perceive why this should be so: in fact, one would *a priori* think that the movements at the 25-meters level would be more exposed to the influence of external factors such as wind. It has been mentioned before (p. 27) that, in some of the series at 500 meters, the registration of direction seemed to have been erroneous; and it may justly be asked whether a similar error may not have influenced the direction in the series that have been used in the preceding comparison. However, all the series we have used were free of such irregular direction variations as were present in the series that were discarded, and there seems to be no reason to suspect the directions to be in error. As an example, we give the distribution of balls in series 8: Out of 46 balls, 1 ball was in compartment No. 30, 1 in No. 31, 7 in No. 32, 18 in No. 33, 17 in No. 34 and 2 in No. 35.

One further point should be mentioned. It has been stated before (p. 34) that a tidal current, independent of depth, would cancel out in the differences formed to obtain the relative current. This is true when vector differences are used. However, even if the geostrophic current conforms to the "law of parallel fields", the resultant of geostrophic current and tidal current will as a rule not have the same direction at all depths. Consequently, even if the observed currents at two different depths are not parallel, their vector difference will give the relative geostrophic current, provided that the non-geostrophic part of the current (the tidal current) is independent of depth. In our case, however, the results do not indicate that such circumstances are responsible for the difference in direction between the currents at 25 and 500 meters.

As a conclusion to this section, it can be said that this experiment once more confirms the need for caution, when working on a small scale, in identifying geostrophically computed currents with observed currents (see e.g. DEFANT, 1961, pp. 504—

508). It is true that the observed currents were of the same order of magnitude as the computed ones, and that the component of the observed current in the direction of the computed geostrophic current was always positive; but the differences were in many cases so great that a conclusion on real currents from the computed ones would be seriously in error.

c. The current measurements at station "B", 1957, first period. The anchor station "B" was occupied by the "Helland-Hansen" in position N 62°29', E 01°44' at a depth of 615 meters with 810 meters of wire paid out. The measurements started on July 2. Before that date, the two ships had made a hydrographic survey of the surrounding area (July 27—30), the main features of which are seen in Fig. 11, the anchor station being indicated by a cross. The "Armauer Hansen" continued to make hydrographic observations in the vicinity of the anchor station until June 5, when she had to go in for refueling. The current measurements continued until July 12, but a break had to be made from the 7th to the 9th, because of a spell of rough weather. The "Armauer Hansen" resumed the hydrographic work in the vicinity of the anchor station on the 9th, and continued throughout the duration of the anchor station. The two parts of the investigation (before and after the break in the current measurements) will be treated separately, and we shall first consider the current measurements made in the first period (July 2—7). The measurements started at 07^h on the 2nd, and were carried on until about 16^h on the 7th, when the sea became so rough that the work had to be discontinued. However, it was obvious from the Loran readings and the bottom depth recordings that the ship had started dragging the anchor earlier that day, so that the measurements made after 06^h on the 7th have to be discarded. Measurements were made at 10 and at 25 meters with the Ekman single reading current meter. The series at 10 meters was the most continuous one, with an average number of measurements per hour of 4.7, as contrasted with 3.4 at 25 meters. The series at 10 meters also lasted longer than that at 25 meters, as the latter series was discontinued at 09^h on July 6. A number of single measurements were also made at 50 meters with variable intervals. The results have shown, however, that such haphazard measurements at an anchor station are of little value because of the large variations that may, at any depth, occur between successive measurements with a few minutes' time difference. The Ekman repeating current meter was used to obtain series of 2—3 hours duration at greater depths (200—300—400—500—600 meters). The observations at 10 and 25 meters will be treated first. Because there seemed to be a distinct tidal variation, at least in a part of the series, both depths have been considered at this station, so that a comparison could be made between the oscillations at the two levels. The measurements have been subjected to averaging in a manner similar to that used for anchor station "A". The only difference is that the hourly values (lunar hours) have been taken from curves for the velocity components of the single measurements rather than curves for speed and direction. This was done because the directions seemed to be more variable at this station than at station "A". It appeared, however, that it made little difference which of the methods was used. As this station was situated on the steepest part of the



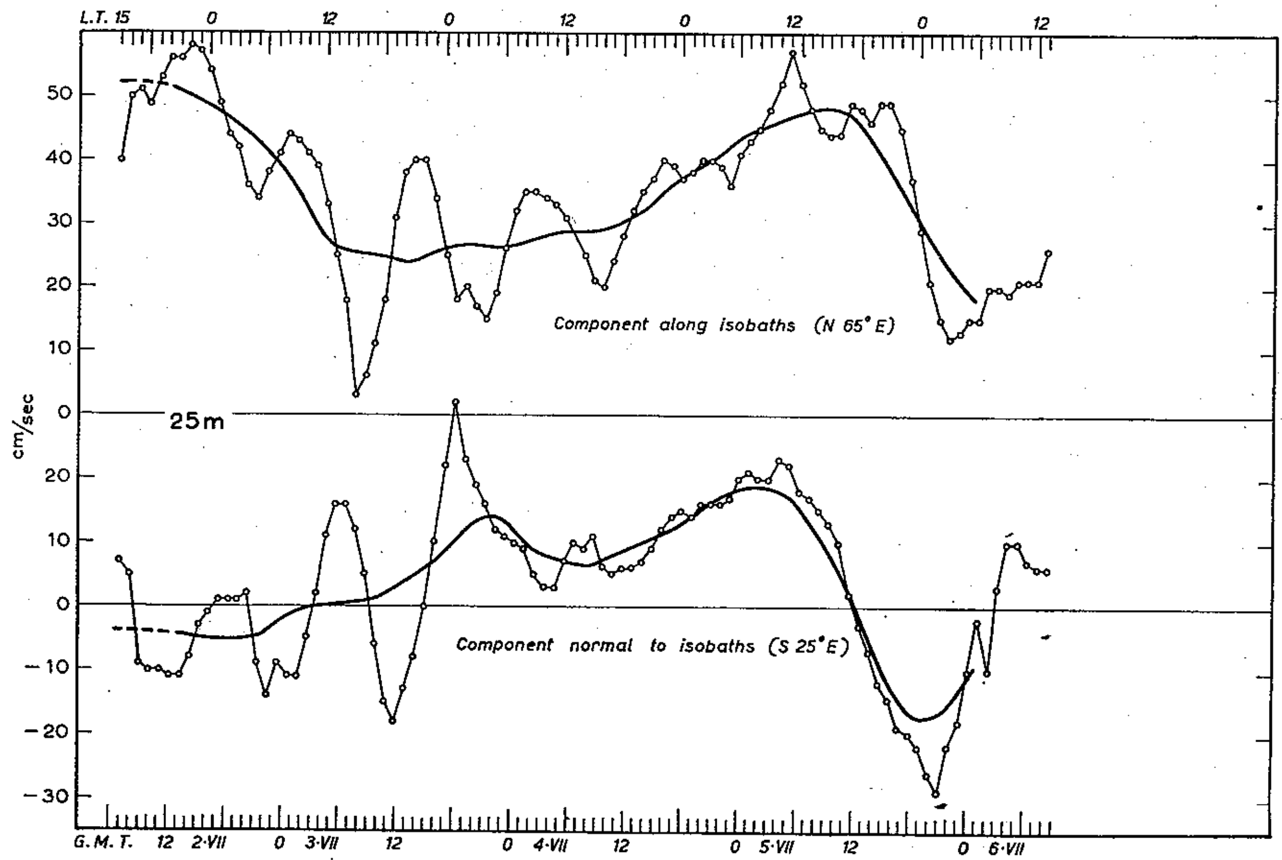


Fig. 30. Anchor station "B", 1957. Hourly values (lunar hours) of velocity components at 10 and at 25 meters, July 2—7.

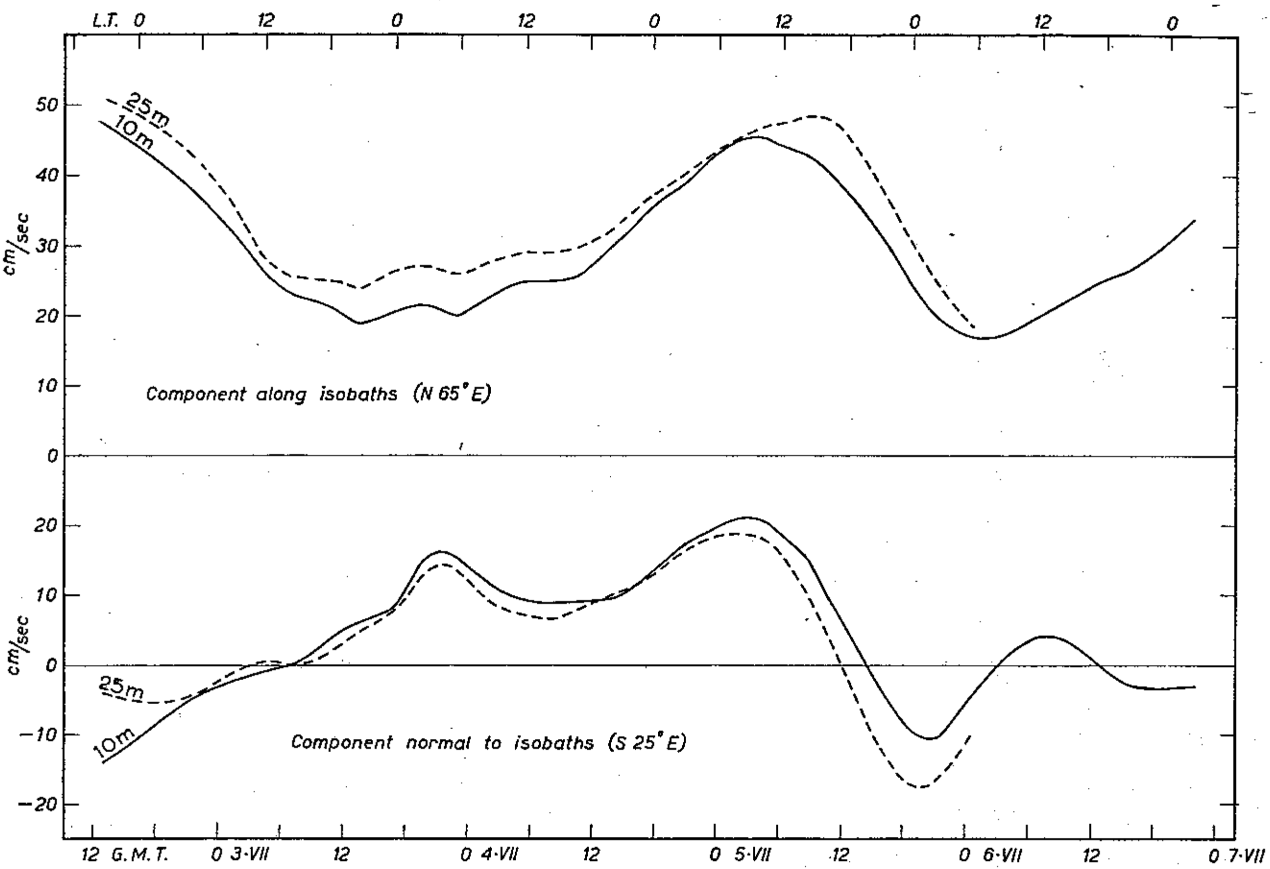


Fig. 31. Anchor station "B", 1957. 12 — hourly running means (lunar hours) of the velocity components at 10 and 25 meters, July 2—6.

continental slope, and in relatively shallow water, it was preferred to use component directions roughly along the isobaths and normal to them, instead of the usual N—E decomposition. The hourly (lunar hours) values of the components for 10 meters and 25 meters are represented in Fig. 30. It will be noticed that the curves are, in one sense, much more regular than the corresponding curves for station "A" (Fig. 21). On the other hand, the gross-feature variations are much larger than at station "A". These variations are of two kinds: (1) tidal variations, and (2) variations in the 12-hourly mean values. The variations of tidal period are very conspicuous at both depths, especially at 10 meters, in the first part of the series. There is no doubt about the reality of a variation with a period of about 12 hours during

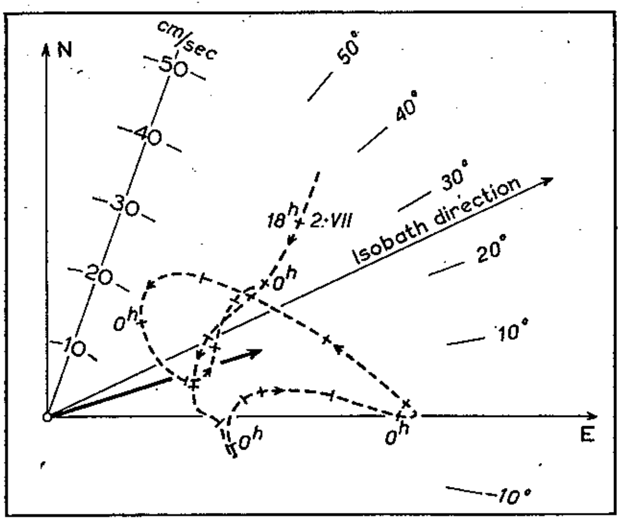


Fig. 32. Anchor station "B", 1957. Hodograph of velocity vector (12-hourly running means) at 10 meters, July 2—6.

the first 36 hours (3 periods). After that, however, there appears to be no tidal period in the variation. The tidal variation is superimposed on a "trend" which is also strongly variable. The trend has been taken as the 12-hourly running mean (lunar hours), and it is entered into the diagrams Fig. 30. For an easier comparison between 10 meters and 25 meters, the components of the trend at both depths are pictured in Fig. 31. It is seen that the trend is very similar at the two depths. Although the values at a certain time may differ by 5—10 cm/s, the general variation is very much the same. The 12-hourly mean at 10 meters is also drawn in a central vector diagram (Fig. 32) in which the curve connecting the end-points of the vectors is drawn. The time (GMT) is indicated by ticks across the curve every 6 hours. The vectors are confined to a sector of 64°. The vectorial mean (thick arrow) is 30 cm/s, directed only 8° south of the direction of the isobaths. In this connection, it should be pointed out that the direction of the isobaths is not very accurately determined, and may be wrong by a few degrees. At the 25 meters level, the vectorial mean is 35 cm/s in a direction 6° south of the isobaths. The series at 25 meters is shorter than that at 10 meters, so that for a comparison between the currents at the two depths one should use only that part of the series at 10 meters which corresponds to the series at 25 meters. If this is done, a speed of 31 cm/s in a direction 9° south of the isobaths is obtained for the vectorial mean of the residual current at 10 meters. The difference between the two depths is thus not great, but it is probably significant. At both depths, the total mean current flows very nearly along the isobaths, but it is seen from Fig. 32 that the mean current for the individual 12-hourly periods may differ by more than 30° from the isobath direction.

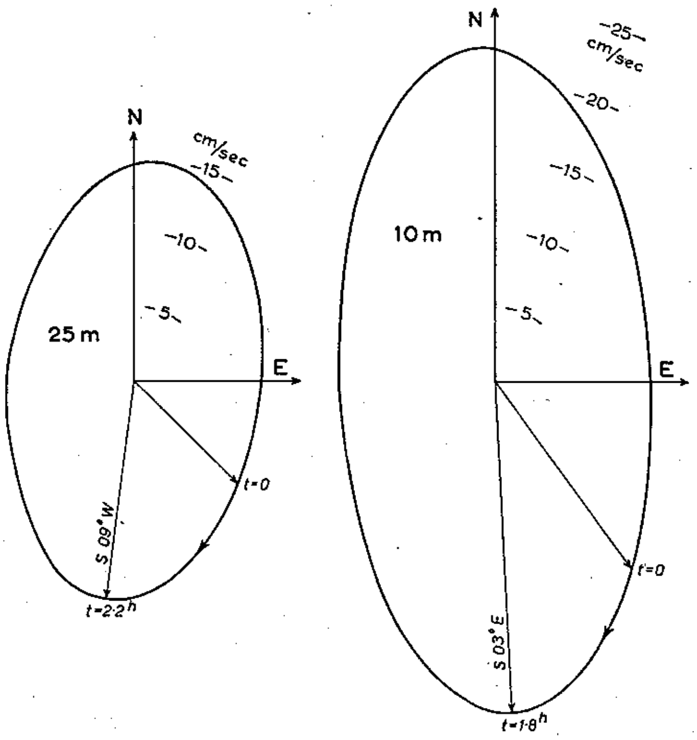


Fig. 33. Anchor station "B", 1957. Current ellipses (semidiurnal lunar period) at 10 and at 25 meters, July 2—6.

The tidal variation is, as mentioned, very distinct during the first 36 hours. For the harmonic analysis of the tidal variation, the trend was first removed by subtracting the 12-hourly running mean. The results showed, as could be expected, that the amplitudes were very small and irregular beyond the first 36 hours. It was obvious that they were not significant, so a significance analysis was not carried out. For the first 36 hours, the analysis for the semidiurnal lunar period was carried out separately for the three periods. The results are summarized in Table 3, and the current ellipse at 10

Table 3. *Results of harmonic analysis of current for the first three semidiurnal lunar periods of observation series starting 2. VII. 1957. Time in lunar hours, reckoned from moon's upper transit at Greenwich.*

	10 meters				25 meters			
	Maximum current			Min. current	Maximum current			Min. current
	cm/s	Direction	Time (h)	cm/s	cm/s	Direction	Time (h)	cm/s
1st period	21.2	S 17° E	1.6	8.6	9.2	S 12° W	2.7	5.0
2nd period	21.6	S 03° E	2.0	9.6	13.6	S 06° E	1.9	8.5
3rd period	25.7	S 09° W	1.8	13.9	22.3	S 16° W	2.3	13.0
All 36 hours	22.4	S 03° E	1.8	10.6	14.8	S 09° W	2.3	8.8

meters and at 25 meters for the mean of the first 36 hours are shown in Fig. 33. It is seen from the table that the orientation of the current ellipses, and the time of maximum current, vary within rather narrow limits. In all cases, the rotation of the current vector was *cum sole*. As regards the strength of the current, however, there were considerable differences between the three periods and between the two depths. The differences between the periods were specially great at 25 meters. When all the three periods were analysed together, the maximum current at 10 meters was 22.4 cm/s and at 25 meters 14.8 cm/s. It should be noticed, however, that the third period at 25 meters gave about the same maximum current as that found for the average tidal ellipse at 10 meters. The apparent decay of the tidal variation after the first 36 hours might at first be thought to be connected with the spring-neap variation of the tide. However, on a closer examination this seem to be unlikely. At the nearest standard port for sea level observation on the Norwegian coast — Kristiansund — there is spring tide on June 30 and neap tide on June 6—7 (Fig. 34). If the current is supposed to be roughly proportional to the tidal range, it is true that one should expect a certain reduction of the current from July 2 onward. This reduction should, however, be gradual, and even at neap July 6—7 the current strength should be more than half the current on July 2. Furthermore, there ought to be a decrease in the tidal current during the three periods shown in Table 3. In fact, there is an increase. It may be concluded, then, that this effect can only be a contributary cause of the "disappearance" of the tidal variation in the current. It seems more likely that the tidal variation has been masked by irregular variations in the non-tidal part of the current. Another possibility, which should not be overlooked,

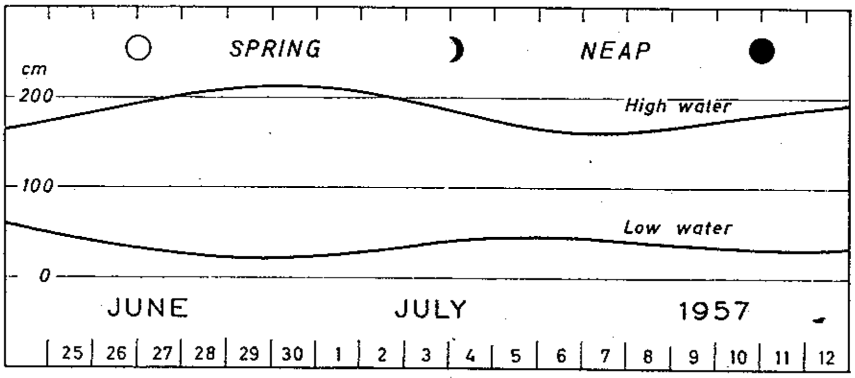


Fig. 34. Tidal range at Kristiansund.

is that the variations of tidal period in the first 36 hours are not due to the normal tidal current, but to an internal wave. Such an assumption might account for the difference between 10 m and 25 m in the magnitude of the "tidal" current. In the normal tide current, no such difference should occur. The material is too sparse, though, really to ascertain whether this difference is more than a coincidence. Thus, as mentioned above, in the 3rd period the current at 25 meters was as strong as the average for all three periods at 10 meters.

Current measurements at greater depths. As previously mentioned (p. 40), with intervals up to 4 hours between each measurement, the observations at 50 meters are mostly too scattered to form the basis of any sensible analysis. Only during the time from 19^h on July 2 to 08^h on July 3 were the measurements made so frequently (two or three per hour) that a comparison with the measurements at the shallower levels may be justified. This period coincides with the second analysis period of Table 3. A harmonic analysis gave for the semidiurnal lunar term at 50 meters a maximum current for 11.1 cm/s directed toward S 06°E at a time 2.5 LH after the moon's upper transit. The minimum current was 4.1 cm/s, and the rotation of the current vector was *cum sole*. It should be noticed that, since continuous 12-hourly means were not available, a Lamont correction was used to remove the trend. A comparison with the values for the 2nd period in Table 3 shows that there is a close resemblance to the current at 25 meters, but that the tidal current at 50 meters is a little weaker than at 25 meters. The mean values of the velocity components at 50 meters for this period were 34 cm/s along the isobaths and ÷ 1 cm/s normal to them, that is, very close to the corresponding mean values at 25 meters (35 and ÷ 1 cm/s).

The measurements with the Ekman repeating current meter were made mostly at 200 and 600 meters. During the period under consideration, 15 series, mostly of a duration of about 2 hours, were made at 200 meters and 9 series at 600 meters. In addition, a few series were made at 300 (2), 400 (3) and 500 meters (2). No harmonic analysis can be made for any of these depths. In Fig. 35, the results have been plotted as current strength and direction, and combined with the smoothed curves for 10 meters for a comparison with the currents in the surface layer, as in Fig. 22. As in that figure, the results from the Ekman repeating current meter are presented as mean values for 5 single measurements, that is, for a period of about 20 minutes. The layering of the water masses is indicated in Fig. 36. The thick lines in that figure are averages of seven stations, all very similar, taken during the period 03^h on July 2 to 02^h on July 5. A single station taken near midnight on July 5, which differs somewhat from the others, is also shown. It appears that the 200 and 300 meter levels are both in the warm and salt Atlantic water, and the 600 meter level is in the cold bottom water, a definitely different water mass. The 400 and 500 meter levels may be said to be in the upper and lower part, respectively, of the transition between the two aforesaid water masses. As far as can be deduced from the scattered series at 200 meters, the water at that level takes part only to a limited degree, or not at all, in the oscillatory movements that

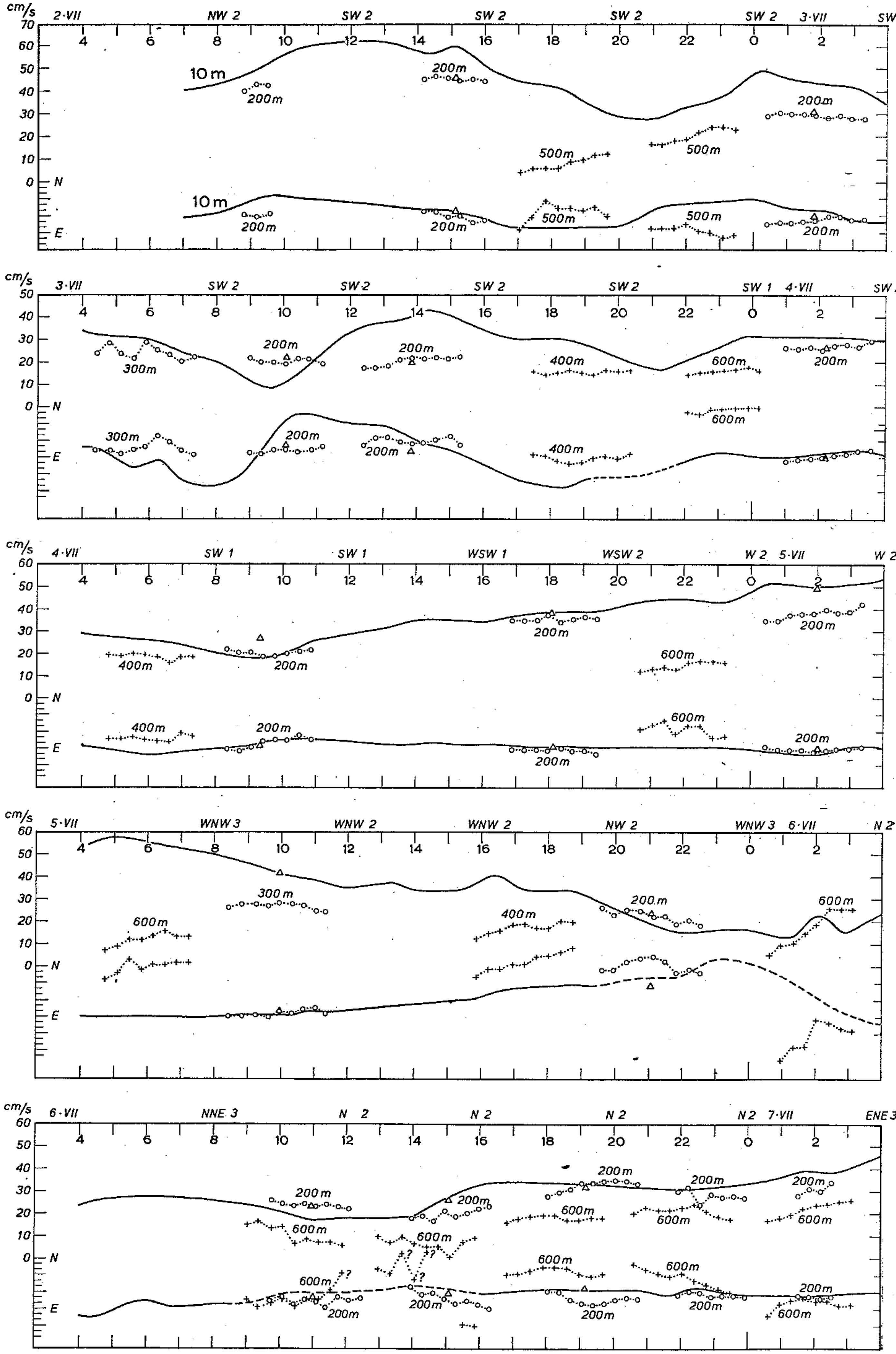


Fig. 35. Anchor station "B", 1957. Velocity and direction of the current at 10 meters (smoothed curves) and at 200 meters and deeper levels (20 minutes' mean values). July 2—7.

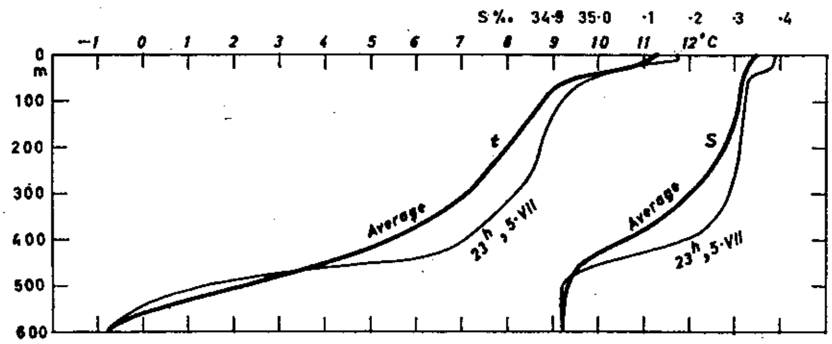


Fig. 36. Anchor station "B", 1957. Average temperature and salinity curves, July 2—5 (7 stations).

have been shown to exist in the surface layers during the first couple of days. There are great differences between the currents at 10 meters and at 200 meters. These are most conspicuous in the current strength, but sometimes also in the direction. There is, in fact, a much closer agreement between the currents at the two levels when at the 10 meter level, instead of the smoothed curve of the observed values, the 12-hourly means are considered. For comparison, the 12-hourly means at 10 meter have been entered as triangular marks in Fig. 35 for each of the series at 200 meters. After the first two days or so, the agreement between the observed currents at 10 and 200 meters is on the whole much better, especially as regards the direction. In some cases, e.g. for all the series on July 4, the agreement can be said to be complete, so that the bulk of the Atlantic water is probably moving as a single body. As we have just seen, this was also approximately the case for the first two days when the oscillatory component of the water movement at 10 meters had been subtracted. This leads us to believe that such a state is the normal one, and that deviations from this state may be caused by oscillatory movements that affect only part of the Atlantic water mass, in the present case the upper layers. In the present investigation, the oscillations were of tidal period. As mentioned before, the "normal" tide should, under ideal conditions, be the same at all levels. The possibility of internal waves has also been mentioned. It is evident that this possibility cannot be realized with the simple two-layer model, in which the velocities must be the same at all levels above the transition layer between the two main water masses (Atlantic water and deep water). The fact that the velocities in the oscillation were decreasing with depth and probably vanishing at 200 meters, would indicate that the more realistic theory of FJELDSTAD (1933) would be more appropriate. On the basis of the observations at hand, however, it is not possible to test whether the oscillations can be ascribed to internal waves. Inertial oscillations can probably be ruled out, as it is difficult to make the curves (Fig. 30) fit in with the inertial period of 13.5 hours.

The observations at levels deeper than 200 meters were mostly made at 600 meters, in the cold bottom water, and only 15—20 meters above the bottom. On July 6 and 7, two Ekman repeating current meters were used simultaneously, one at 200 and one at 600 meters, in spite of the risk of entangling the suspension wires. On the whole, the velocities were markedly lower at 600 meters than at 200 meters (see Fig. 35). This is most clearly brought out during the time when measurements were made simultaneously at both levels (last section of Fig. 35). However, even so close to the bottom, the velocities were often as high as 20—25 cm/s. Velocities lower than 10 cm/s were also found in several cases. The directions seem to deviate more from the directions at 10 meters than those at 200 meters do; but they are all confined to the sector between N and E. It should be kept in mind that at low velocities (below 10 cm/s, roughly) the directions are often poorly defined. This shortcoming is, most likely, due to the instrument used. In two of the series, the directions were due north, that is, at an angle of more than 60° with the bottom contour lines. The observations are too few to determine whether the directions will, on an average, be along the countour lines, as was the case for the 10 meter level.

The hydrographic stations made by the "Armauer Hansen" in the vicinity of the anchor station "B" were intended to form a pattern similar to that at the anchor station "A", so that geostrophic currents could be computed and compared with the observed currents. Several factors combined to obstruct a satisfactory execution of this plan. Very frequently, a dense fog made astronomical positioning impossible, and the strong and probably variable current set could thus not be corrected for with the desired accuracy. In addition, the hydrographic situation in the vicinity of the anchor station was such that small displacements of the positions of some of the stations would produce great effects on the computed currents. As an example, the position of "A. H." station 154 should, according to the dead-reckoning, be only 2 n. miles ENE of the anchor station position. The difference in dynamic depth (surface relative to 500 db) between station 154 and the anchor station was 0.8 dyn.dm, giving a geostrophic current of 163 cm/s. This is an impossibly high figure. A current of reasonable magnitude (1 knot or less) would be obtained under the assumption that the real position of station 154 was at least 6—7 n.miles away from the anchor station, probably in direction ENE. Under these circumstances, no attempt can be made to make a comparison between computed and observed current similar to that made at anchor station "A". The picture given in Fig. 37 of the dynamic topography of the surface relative to 500 db should be considered only as an illustration of the gross features of the hydrographic situation. The first station (No. 154) was taken 10 hours before the "Helland-Hansen" anchored in position "B", and the stations up to No. 164 were taken without any contact with the anchored vessel. After that, the "Armauer Hansen" could, with intervals, correct her position with respect to the anchored vessel, so that the positions for the stations taken after that time are probably more reliable. The last of the stations in the pattern (No. 197) was taken at 5^h on July 5. The most conspicuous feature of Fig. 37 is the "Low" NNE of the anchor station, contrasted with a "High" south of the anchor station. The direction of the relative geostrophic current should thus be roughly toward the east. For the reasons just mentioned, it is not possible to give any estimate of the speed. The extreme crowding of the dynamic isobaths south of the "Low" may in part be due to

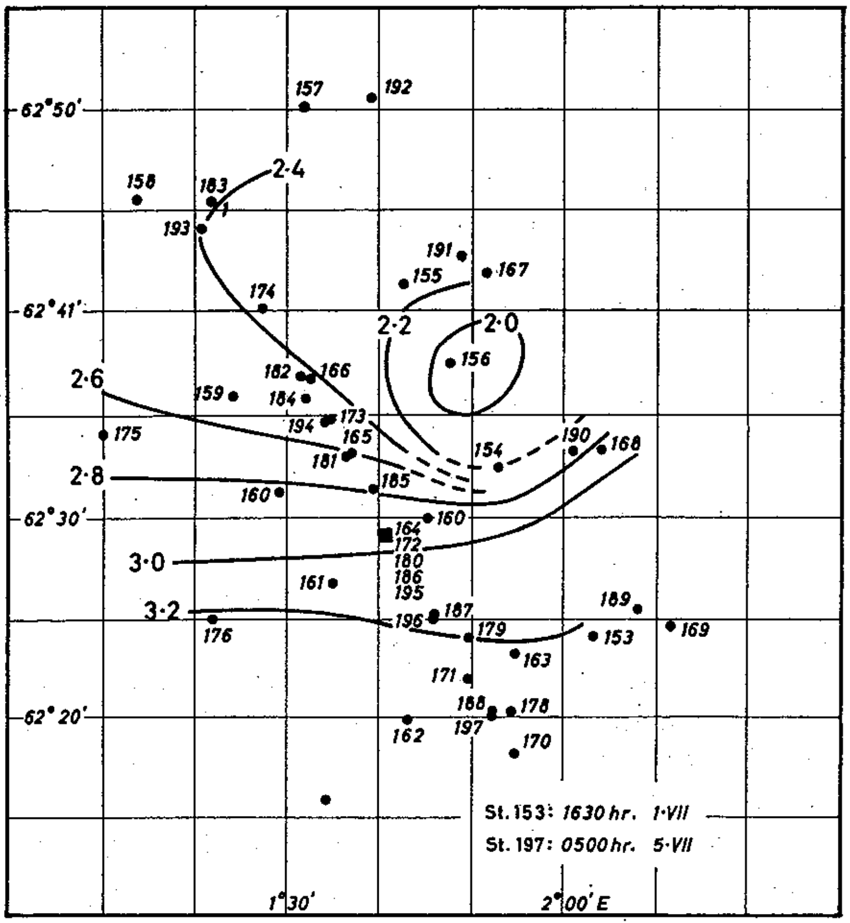


Fig. 37. Dynamic topography (surface relative to 500 db, ΔD in dyn. dm) in the vicinity of anchor station "B", 1957, July 1—5.

erroneous positioning. The position error of the geostrophically computed current will be least for the sections running through the anchor station. In all, five sections were made through the anchor stations. For each of these sections, we have computed the geostrophic current between the two stations nearest to the anchor station on each side of it (surface relative to 500 db, or 450 db if the coastward station was too shallow). The results are compiled in Table 4, together with the observed values at 10 meters.

Table 4. *Currents at anchor station "B".*

Geostrophic			Observed at 10 m		Period
Between stations	Between levels	Velocity cm/sec	Smoothed curve	12-hours mean curve	
163—165	0—450 db	19.5	56	47	2. VII, 14—17^{h}
171—173	0—450 db	14	31	21	3. VII, 4— 7^{h}
179—181	0—500 db	8	22	26	3. VII, 20—23^{h}
187—185	0—500 db	13.5	20	27	4. VII, 8—12^{h}
196—194	0—500 db	26	50	48	5. VII, 0— 3^{h}

It is seen that the computed relative currents are much smaller than the observed currents at 10 meters whether the latter are taken as the smoothed values from Fig. 35 or from the curve for the 12-hourly running mean value in Fig. 31. However, the computed and observed values in Table 4 are not directly comparable. Firstly, the computed values are only components normal to the section. As the normal direction is probably most often less than 30° from the true current direction, this will only account for a difference of less than 15%. Secondly, it may be inferred from the measurements made at 500 meters and 600 meters that the current at 500 (450) meters may at times be rather strong, up to 20—25 cm/sec, and may thus account for most of the differences between computed and observed currents in Table 4. Even with these restrictions on the comparability of computed and observed currents, it seems that there is a certain correlation between them, as the highest values of the computed velocity in Table 4 correspond to the highest values of the observed velocities.

d. The current measurements at station "B", 1957, second period. As mentioned before, the current measurements had to be discontinued for a couple of days from July 7. A small, but concentrated atmospheric low between Scotland and Norway gave winds from E or NE of force 6—7. As the wind was right against the current, the sea became very rough and irregular. When the measurements were resumed on July 9, the wind had abated to NE 4, but the sea was still irregular. Observations were made continuously at 10 meters from 11^h on the 9th to 9^h on the 12th. Measurements at 25 meters were made from 16^h on the 10th, but less frequently than at 10 meters. With the Ekman repeating current meter, series of measurements were made mostly at 200 meters. It turned out that the ship had started to drag the anchor on the morning of July 10. This could be detected because the drift had been in a southerly direction, and thus at a large angle both with the bottom contour lines and the single set of Loran lines which could be received in the area. It appeared from the Loran readings that the ship must have been drifting about two nautical miles between 7^h and 14^h. After that time, the Loran readings stayed constant, which, of course, does not exclude the possibility that the ship may have drifted along the Loran lines. There was, however, no indication that such a drift had taken place. First, let us look at the period before the ship started dragging the anchor. The measurements (smoothed) are presented in Fig. 38 in a like manner as in Figs. 22 and 23. During this period, the current was at times very strong, up to 75 cm/sec, a value very seldom reached in this branch of the current. The direction was first to the SE, slowly veering nearly due S. This is in striking contrast to the conditions during the first period at this anchor station, when the directions were always between E and N, centering around N 65° E, the direction which should be expected in this area. The velocities at 200 meters were also very high, and roughly in the same direction. As far as can be judged from the few series at 200 meters, the maximum found at 10 meters is not found at 200 meters, but the three series at 200 meters would fit, roughly, with the average 10 meter-curve. A single series at 575 meters also gave high values, about 30 cm/sec, but at that depth the direction was almost due east,

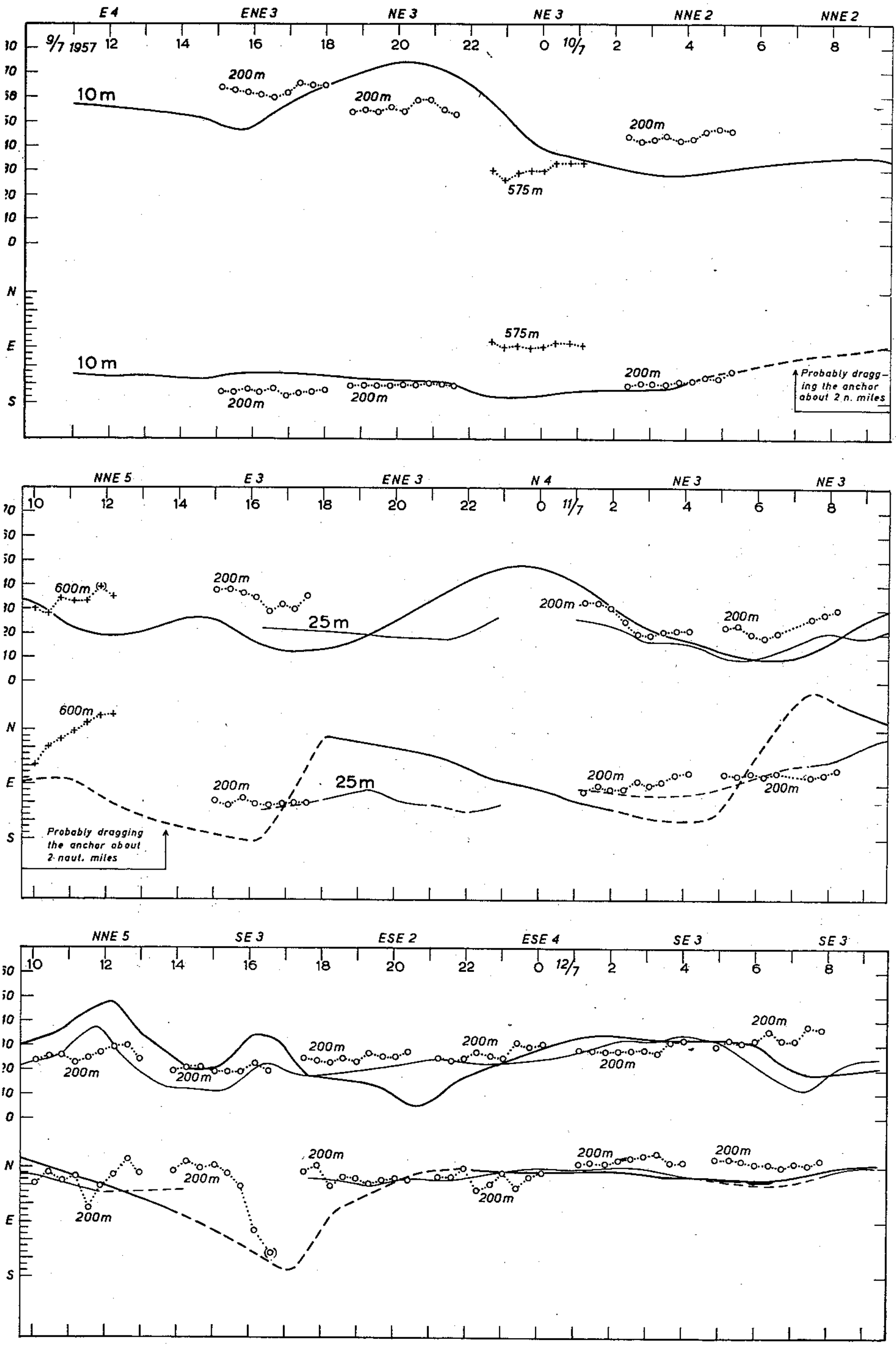


Fig. 38. Anchor station "B", 1957. Velocity and direction of the current at 10 and 25 meters (smoothed curves) and at 200 meters and deeper levels (20 minutes' mean values). July 9—12.

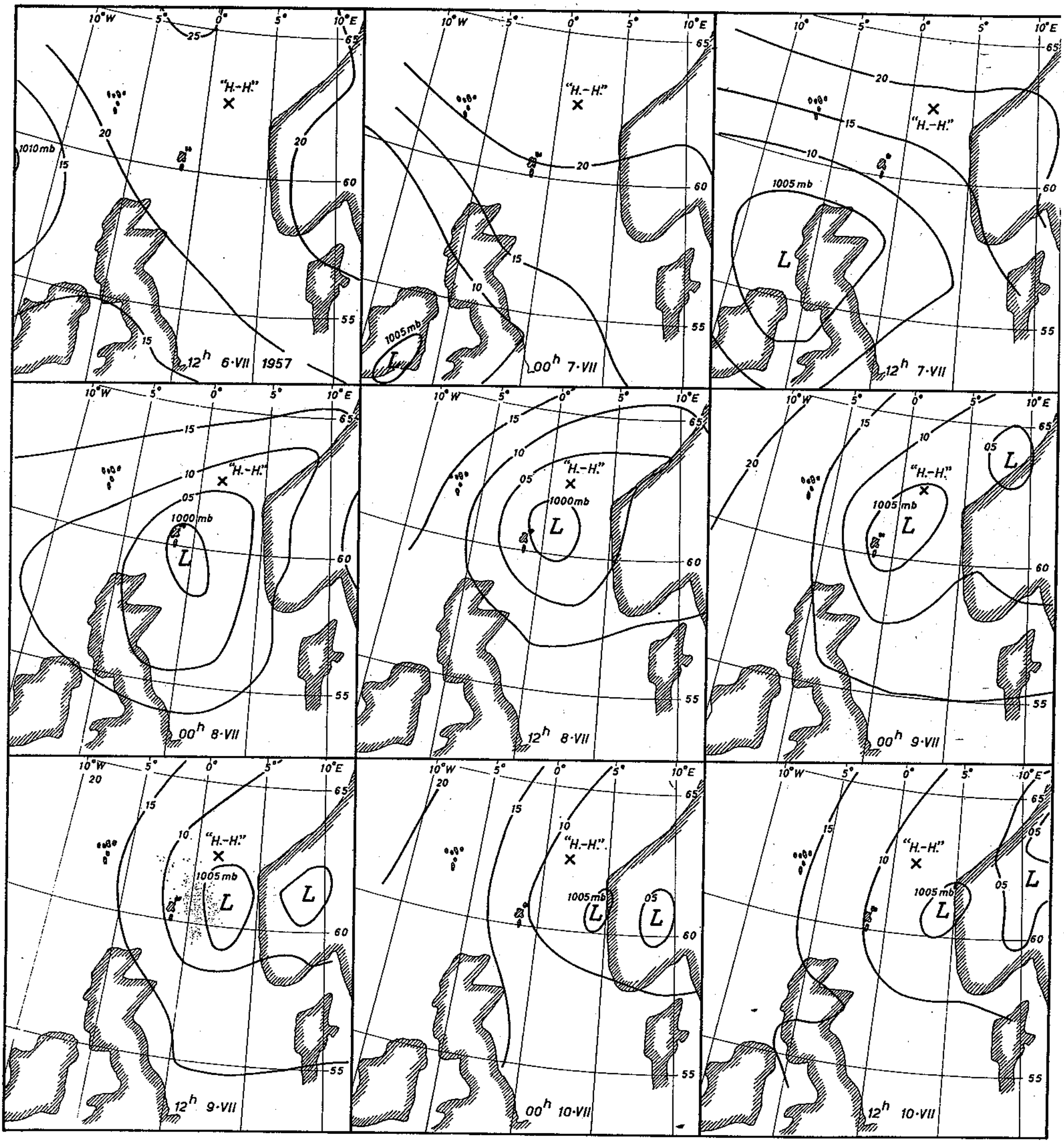


Fig. 39. Atmospheric pressure distribution, July 7—10, 1957.

that is, about 80° from the direction at 10 meters. The strong southerly current at 10 meters is an exceptional occurrence in this area. Very probably, it is in some way connected with the atmospheric low which interrupted the measurements from July 7 to July 9. While deepening, the center of this low moved quickly from Scotland to a position south of the anchor station. At that position, it stopped and its intensity gradually decreased. Some stages in the development of the depression is shown in Fig. 39. It is very difficult to say which factors may be responsible for the strong southerly current. While the wind was at its strongest, during the night between the 7th and the 8th, the surface current was still setting toward the east or north-east, against the wind. It may be guessed that the easterly wind surged the water away from the coast and thus created a pressure field consistent with the southerly direction. It is difficult, thought, too see how the water could have piled up in the open sea. A possibility is that this may have been effected by the convergence between the generally easterly flow and the wind-drift in the surface layer. However, this can at best only be part of the explanation, as such a "slope current" should be the same at all depths, and we have near the bottom (at 575 m) observed 30 cm/sec toward the east. There may have been a baroclinic effect, and it is also possible that the currents have not been in geostrophic balance. The hydrographic stations taken by the "Armauer Hansen" during the period of the south-south-easterly currents (sts. 200—206) were all along a line directed N 30°W—S 30° E through the position of the anchor station, that is, almost in the direction of the observed current in the upper layer, and can thus only be of limited interest in this connection. Between st. 200 (the anchor station, 13^h on the 9th) and st. 201 (16^h on the 9th), taken 10 n.miles N 30° W off the anchor station, the geostrophic current component normal to the section (0 relative to 500 db) was only 4 cm/sec, and directed toward N 60° E. Between st. 205 (8 n.m. N 30° W of the anchor station, 04^h on the 10th) and st. 206 (the anchor station, 07^h on the 10th) the said geostrophic current component was 13 cm/sec in the same direction. As the component of the observed current was in the opposite direction, it seems that there can have been no connection between the observed and the relative geostrophic current, as computed by Helland-Hansen's formula. It can be seen from Fig. 39 that the air pressure in the area south of the anchor station decreased very rapidly, about 15 mb in 12 hours, prior to the measurements under discussion. It is known from coastal sea level observations that such air pressure variations are as a rule very quickly compensated by a corresponding rise of the sea level. (See e.g. Unoki, 1950; Lisitzin, 1943; Rouch, 1944; Imbert, 1956). If such a rise may be assumed to take place also in the open sea, this would call for an inflow into the area, which would be non-geostrophic. Admittedly, the velocity of the flow would be very small if the flow is assumed to be more or less evenly distributed in all directions leading into the area. Only if it were in some way concentrated, e.g. by the influence of the coast, could this flow be of any importance in our case. In all, it must be admitted that none of the possible explanations sketched above are obvious or directly convincing. The complexity of the hydrography of this region adds to the difficulties in explaining surprising features such as the strong south-south-east current. As a further example of the intricacy of the current system in the region, the following incident may be of interest. On July 7, a life-boat was lost from the anchored "Helland-Hansen". This happened at 2330 hrs, when the wind was strongest (force 7 or 8 Beaufort). The life-boat was floating on the tanks, and just visible above the water. It was seen to be carried with the current right against the wind, toward the east or north-east. On July 25 it was found by a fishing vessel, not on or near the Norwegian coast, but 40 n.miles NE of Shetland (pos. about 61°30'N, 0°E), that is, 60 n.miles away from the anchor station in the highly unexpected direction S 40° W. It is probably futile to speculate on the route it may have taken between the two positions during the intervening 18 days.

As mentioned above, the ship had been dragging the anchor in the morning of July 10. At about 14^h, it seems certain that the anchor had got hold again, and we shall now discuss the measurements made after that time and until the end of the anchor station at 10^h on July 12. During this period, the quality of the measurements with the single-reading Ekman current meter at 10 and 25 meters was often very poor. In some periods, the balls indicating the directions were spread more or less evenly over a large sector which was at times more than 100 degrees. This occurred mostly when the current was decreasing and in periods of weak current, and obviously must have been connected with the ship's movements. In addition to the three types of ship movement described e.g. by DEFANT (1932), "false" directions may in this case also have been induced by the ship's jerking in the irregular sea. In the measurements with poorly defined directions, it must be assumed that the speed, too, was to a great extent influenced by the ship's movements. Under such circumstances, a more refined analysis would be of little value. Nevertheless, several interesting features seem to be fairly well established, at least qualitatively, as may be seen from Fig. 38, where broken lines indicate the periods of poorly defined directions with the single-reading Ekman. The curves have been smoothed in order to eliminate the short-period fluctuations. In Fig. 40 the smoothed values at 10 meters are shown as the hodograph of the velocity vector. The series at 10 meters was the most continuous one, with an average of 4.2 measurements per hour. At 25 meters, the measurements were made less often and there were some long breaks in the series. At 200 meters, some series were obtained with the repeating current meter. The measurements at 10 meters are characterized by variations in direction much larger than during the previous period. Even if we consider only the measurements with well defined direction (the unbroken part of the line), there is a span of about 180°. Assuming that the main trend of the direction is represented by the curve also where the line in Fig. 38 is broken, the curves indicate a characteristic variation which covers a couple of periods. The period is semidiurnal, but it cannot be very precisely defined from the observations. In the curves for the components (not reproduced here), the time difference between the maxima varies between 11 and 15 hours. As the inertial period is 13.5 hours, there is, in this short series, no means of deciding whether the oscillation is of the semidiurnal tidal period or connected with an inertial phenomenon. A harmonical analysis of 24 hours (lunar or inertial hours) gives for the inertial period slightly larger values of the amplitudes (Major axis 25.4 cm/s, minor axis 17.8 cm/s) than for the semidiurnal lunar period (Major axis 24.4 cm/s, minor axis 16.4 cm/s). However, in view of the many irregularities in the observations, the difference is certainly not significant. Generally, the oscillations of inertial period may be classified conveniently as: (1) oscillations with purely horizontal velocity and (2) oscillations connected with wave motion. The oscillations under (1) are commonly termed "pure inertial motion", and are generally considered as a generalization of the movement of a particle of mass on a rotating disc in the "circle of inertia", so that the water at any particular level moves as a rigid sheet (see e.g. PROUDMAN, 1953, p. 75). However, a more general horizontal inertial motion is possible, but also in the general case will the characteristic feature of the inertial motion occur: the oscillating part of the velocity vector rotates uniformly *cum sole* in a circle. This can be shown on the following assumptions: (1) The water is homogeneous, (2) the friction can be disregarded, (3) the non-linear terms of the acceleration can be disregarded, (4) the Coriolis parameter is treated as a constant. The horizontal equations of motions are then

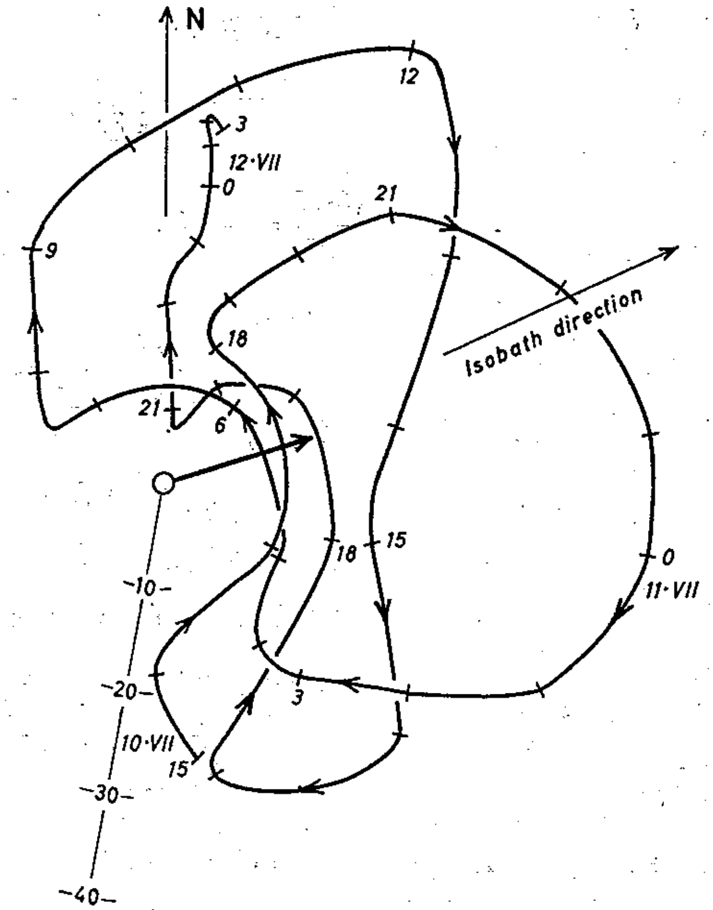


Fig. 40. Anchor station "B", 1957. Hodograph of velocity vector (smoothed values) at 10 meters during the period July 10—12.

$$\frac{\partial u}{\partial t} = \lambda v - g\frac{\partial \xi}{\partial x} \tag{1a}$$

$$\frac{\partial v}{\partial t} = -\lambda u - g\frac{\partial \xi}{\partial y} \tag{1b}$$

u, v are the horizontal components of the velocity, with the y-axis *contra solem* from the x-axis, λ is the Coriolis parameter, and ξ the elevation of the surface above a level surface. In the case of purely horizontal motion, the equation of continuity is

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0 \tag{2}$$

and ξ is a function of x and y only. With

$$U = u + \frac{g}{\lambda}\frac{\partial \xi}{\partial y} \; ; \; V = v - \frac{g}{\lambda}\frac{\partial \xi}{\partial x} \tag{3a, b}$$

we obtain

$$\frac{\partial^2 U}{\partial t^2} = - \lambda^2 U \quad \text{and} \quad \frac{\partial^2 V}{\partial t^2} = - \lambda^2 V \tag{4a, b}$$

with the equation of continuity

$$\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0 \tag{5}$$

The general solution of (4a, b) is

$$U = A\,(x, y) \cos \lambda t + B\,(x, y) \sin \lambda t \tag{6a}$$

$$V = B\,(x, y) \cos \lambda t - A\,(x, y) \sin \lambda t \tag{6b}$$

The equation of continuity (5) is satisfied provided that

$$\frac{\partial B}{\partial x} = \frac{\partial A}{\partial y} \quad \text{and} \quad \frac{\partial B}{\partial y} = - \frac{\partial A}{\partial x} \tag{7a, b}$$

The expressions U and V are the deviations of the velocity components from the stationary unaccelerated state. If, at any time, U and V are zero (geostrophic balance), we must have $A = B = O$, and there will be no inertial oscillations. If at e.g. $t = O$ there is not geostrophic balance, then $A = U_0$ and $B = V_0$, and the "deviation vector" (U, V) will rotate *cum sole* in a circle. Thus, an initial disturbance (U_0, V_0) of the velocity field may give rise to a horizontal "pure inertial current". However, equations (7a, b) place restrictions on the field of the initial velocity disturbance: the field must be irrotational in order that such a motion shall be possible, with a velocity potential ψ such that

$$A = \frac{\partial \psi}{\partial x}, \quad B = \frac{\partial \psi}{\partial y} \quad \text{and} \quad \frac{\partial^2 \psi}{\partial x^2} + \frac{\partial^2 \psi}{\partial y^2} = 0 \tag{8}$$

As pointed out by EKMAN (1941), this would mean, formally, that a limited disturbance in the open sea cannot give rise to an oscillatory motion such as described above, because if the function ψ is constant on a closed curve it must also be constant in the area enclosed by that curve.

Thus, if a horizontal oscillation shall take place as a result of an initial unbalance (with regard to geostrophic conditions) between current and pressure fields, this unbalance cannot be arbitrarily distributed. It may, however, be assumed that the conditions may often be such that approximately similar oscillations are possible. On the other hand, it may be said that if such oscillations occur, they must have the inertial period and must be circular and *cum sole*. Formally, the reason for the restrictions on the initial velocity fields is that, when ξ is considered as a given function of x and y only, there are three equations (1a, b and 2) to be satisfied by the two variables u and v.

It lies near at hand, then, to assume that at least part of the movements invoked by a deviation from the geostrophic balance will involve vertical components, so that ξ must be considered as a time-variable. Under similar assumptions as above, and the added assumption of a constant depth h, the equation of continuity takes the form

$$h\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = -\frac{\partial \xi}{\partial t} \tag{9}$$

Equation (9) together with (1a, b) are the equations commonly used for describing wave motion. Simple wave solutions of the system are given in all textbooks (see e.g. PROUDMAN, 1953, p. 265), and they will not be reproduced here. The point to keep in mind is that the wave period can take any value, and there is no obvious reason why the inertial period should be preferred (see e.g. EKMAN, 1941). A wave with the inertial period will not exhibit the same features as the "pure inertial motion": the horizontal particle velocity vector may rotate in an ellipse, and *cum sole* or *contra solem* according to circumstances. It can be shown, however (EKMAN, 1941), that in the case of a *cum sole* rotation, the wave length must be very large, of the order of several thousand kilometers, so that it seems unlikely that such a wave could be initiated by a local disturbance. In the case of stratified water, the internal wave theory of FJELDSTAD (1933) shows the possibility of internal waves with a factor $e^{\alpha y}$ in the variables, and with an arbitrary period. If the period is taken to be the inertial period, it can be shown (from the equations p. 14 in the mentioned paper) that internal waves with an elliptical hodograph are possible if the amplitudes are decreasing to the left (northern hemisphere) of the direction of propagation ($\alpha < O$). If the decrease of the amplitudes is in the opposite direction ($\alpha > O$), the only possible motion is a special case of the previously described horizontal inertial movement. This is the same feature as found in surface waves in homogeneous water (see e.g. EKMAN, 1941). There is still no reason, however, to prefer the inertial period.

Such waves as referred to above are, however, only a special case of motion compatible with equations (1a, b) and (9). A solution for a special case of the problem of describing the motion following an initial unbalance (with regard to geostrophic conditions) has been given by CAHN (1945). He considers a homogeneous ocean with an initially horizontal surface, and an initial current in a strip parallel to the x-axis. The resulting motion was shown to be oscillatory, with a period approximately equal to the inertial period. An interesting feature of CAHN's solution is that the oscillations will be damped out fairly quickly, as a result of geometric dispersion. This may explain the relatively short duration of the oscillations found on our anchor station. BOLIN (1953) has extended CAHN's theory to the case of a stratified ocean with a vertical velocity shear in the initial state. In addition to the motion described by CAHN, he found a system of internal inertial oscillations. Similar results, with an arbitrary density distribution, were found by FJELDSTAD (1958). In this connection it should also be mentioned that ROSSBY (1938) studied the final state of equilibrium of a stratified ocean after a disturbance of the geostrophic balance of an upper layer, and found that

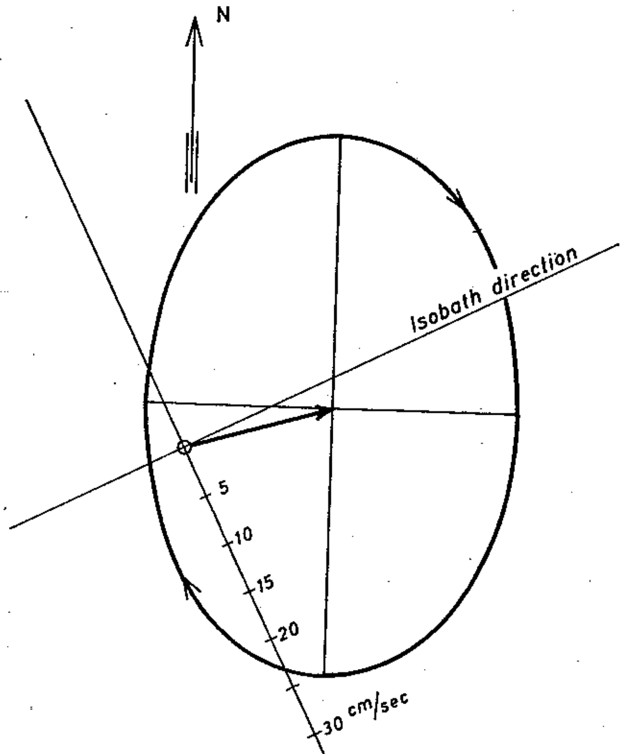


Fig. 41. Anchor station "B", 1957. The harmonic constituent of inertial period (13.5 hours), July 10—12.

currents of considerable strength would be induced also in the deeper layer. This is apparent also from the numerical examples given by BOLIN (1953) and FJELDSTAD (1958). It is probable that these findings have a bearing on the strong and variable bottom currents discussed on several occasions in the present paper.

In the present case, the harmonic constituent of the inertial period (13.5 hours) was as shown in Fig. 41. The vector rotates *cum sole*, but the ratio between the axes is 0.7 instead of 1.0. This should not, however, be taken as a sufficient reason for excluding the possibility of an inertial oscillation. The quality of the observations was such that much of the difference may be attributable to errors. Furthermore, the existence of a semidiurnal lunar component could easily have modified a dominating inertial oscillation in such a way that the resulting oscillation, as found by the harmonic analysis, would be elliptical.[1] In itself, this is of course no proof of the existence of an inertial oscillation. However, the circumstances seem to favour such a guess. First, the sudden arrival of the atmospheric low has obviously led to an abnormal hydrographic situation with a strong southward current (until July 10). Normal conditions are found again on July 12 (see Fig. 38), when the current in the whole water mass down to 200 meters is again directed toward the north with a speed of about 30 cm/sec and without oscillations. It would seem very probable, then, that oscillations found in the intervening period should be interpreted as inertial oscillations accompanying the transition to "normal" conditions.

In the derivation above, friction was excluded. It has been shown (see e.g. EKMAN, 1905; FJELDSTAD, 1930; F. DEFANT, 1940) that also in the presence of friction, transition phenomenae are of the nature of inertial oscillations.

In Fig. 38, the observations at 25 meters are also shown. Although the series is not continuous at that depth, there are enough observations to show that the water at 25 meters took part only to a small extent in the oscillations between July 10 and July 12. In the case of a long wave, no such difference should occur. Even in an internal

[1] Moreover, the condition that an inertial oscillation should have a circular hodograph does not generally apply to non-horizontal oscillations, as this condition is a consequence of the constancy of ξ in equ. (1a, b).

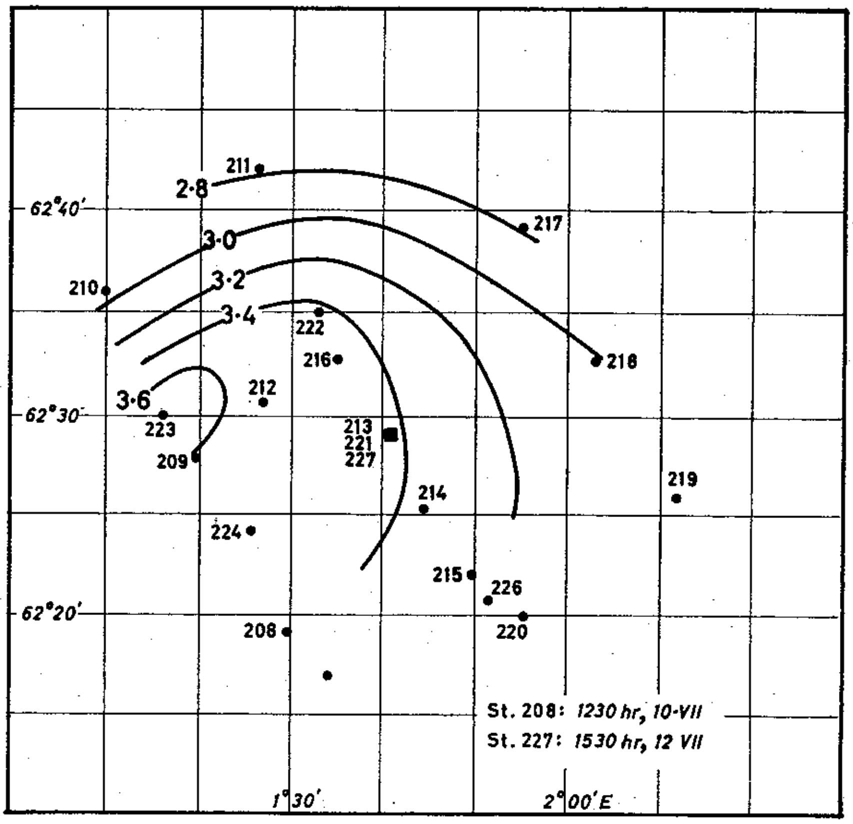


Fig. 42. Dynamic topography (surface relative to 500 db, ΔD in dyn. dm) in the vicinity of anchor station "B", 1957, July 10—12.

wave, it would seem unlikely that as great differences as appear from Fig. 38 would occur over a distance as short as 15 meters. In the case of "pure inertial oscillations", however, such a difference would be easier to understand.

Before we leave the subject of these current measurements, a comparison between Fig. 38 and Fig. 41 may explain the poor quality of the directions previously remarked on. Assuming for a moment that the current may be represented by the hodograph in Fig. 41, it is seen that with the center of the hodograph situated so close to the ellipse, there will be a very sudden change by 180° from S to N, as appears from Fig. 38. During this change, the velocities are small, so that irregular movements will have a great influence, leading to a large dispersion of the balls indicating the direction, as observed. If the current were faithfully represented by Fig. 41, we should also have expected a short period with westerly currents, at variance with the picture in Fig. 38. It should then be remembered that the velocity also contains a rather large residual of terms of different periods in addition to the velocity represented by the hodograph of Fig. 41, and that a small shift of the ellipse would make the center of the hodograph fall outside the ellipse so that no westerly current could occur (see Fig. 40). Lastly, it is seen that the constant term of the harmonic development, represented by the thick arrow in Fig. 41, is directed roughly along the isobaths, as is also the mean vector in Fig. 40. This may be a coincidence, however.

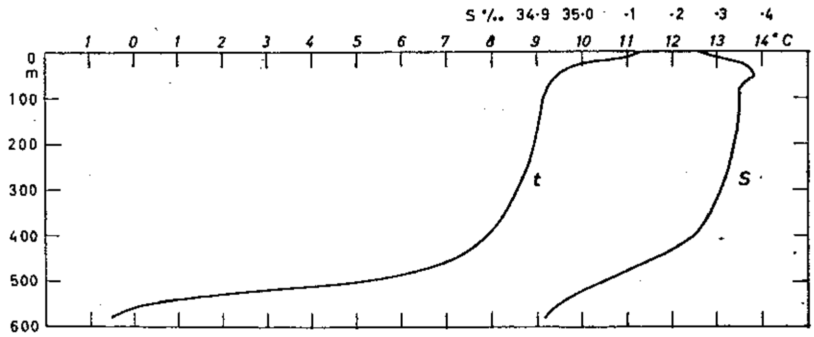


Fig. 43. Anchor station "B", 1957. Average temperature and salinity curves, July 10—12 (3 stations).

Some hydrographic stations, the positions of which appear from Fig. 42, were taken by the "Armauer Hansen" during the period July 10—12. In the figure, the dynamic height anomalies (0 db relative to 500 db) are shown. There is obviously no point in trying to compare geostrophic currents computed from these stations with the rapidly varying currents measured at the anchor station during this period. A comparison with the corresponding picture (Fig. 37) of the dynamic topography one week earlier reveals a striking difference, however, and gives an excellent illustration of how rapidly a radical change in the hydrographic situation may take place in this region. The survey of July 1—5 was dominated by a "low" with values of Δ D (0—500 db) down to 2.0 dyn.dm, whereas in the survey of July 10—12 we find markedly higher values with a maximum of more than 3.7 dyn.dm. Such high values are very uncommon in this area, although not wholly unprecedented. They are due to a deep accumulation of Atlantic water, such as shown in Fig. 43. This figure shows the mean values of temperature and salinity, plotted against depth, for the three occupations of the anchor station during the period July 10—12 (sts. 213, 221 and 227), and should be compared with Fig. 36. It is seen that the "Atlantic" water (S > $35^0/_{00}$) was found deeper than 500 meters, and the typical deep water was found only at the deepest observation level (580 meters).

After the termination of the anchor station, the vessels worked a short hydrographic section across the continental slope 4 times in rapid succession, together with bottom current measurements at each station. These sections, and the corresponding bottom current measurements, clearly show the difficulties in estimating the absolute current from geostrophic calculations. They have been discussed in S I (pp. 18—20), to which the reader is referred.

5. Investigations in 1959. *a. Sections combined with bottom current measurements.* During the days May 10—14, a programme with short sections across the continental slope, with bottom current measurements at each station, was carried out. The programme was similar to that made in 1957 (S I, p. 18—20). A section about 50 naut. miles long was occupied four times in rapid succession, and two bottom current measurements were made at each station by means of the MOSBY bottom-current meter (MOSBY, 1953). This time, we had only one ship (the "Helland-Hansen") at disposal, and the four occupations of the section therefore took more time than in 1957 (88 hours as

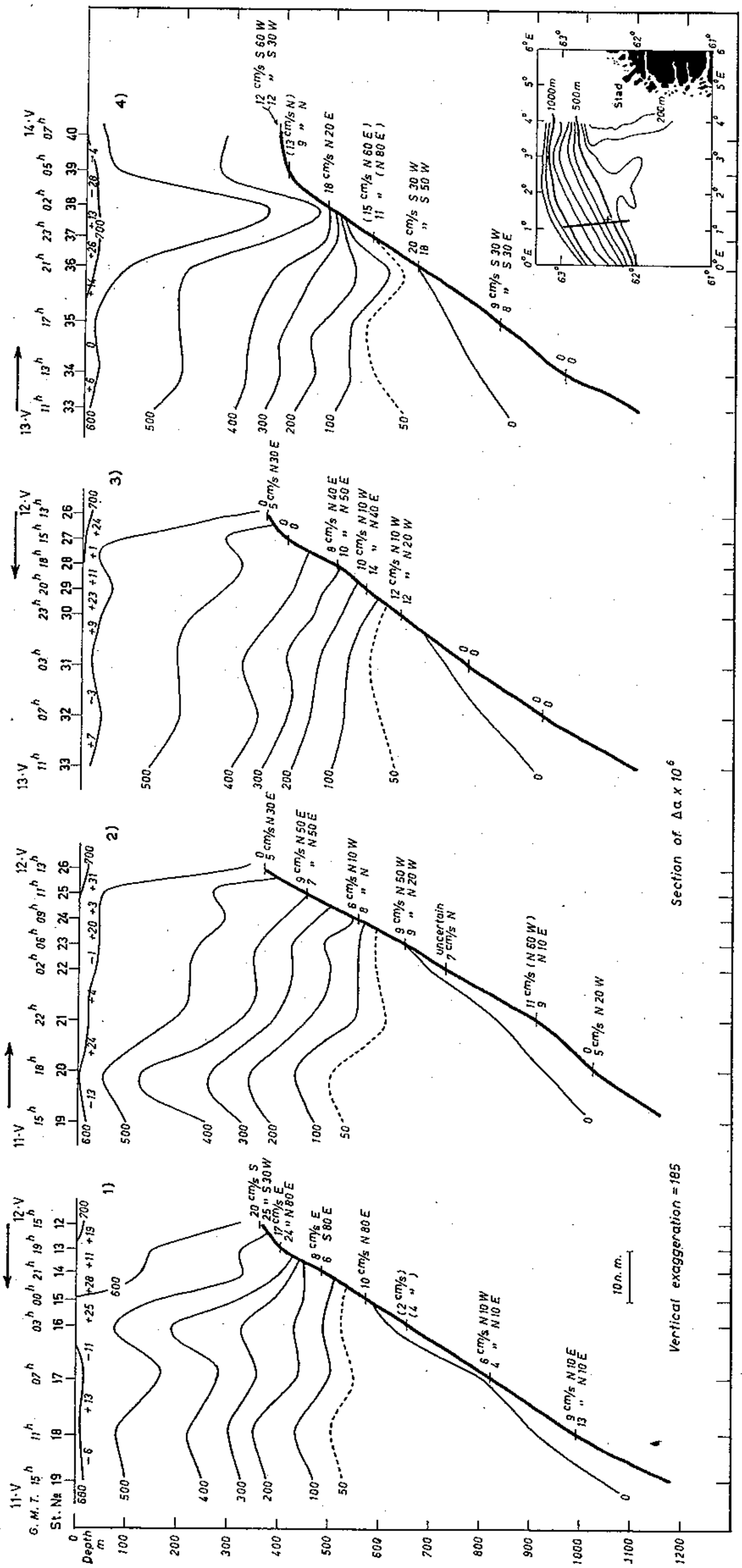


Fig. 44. Distribution of anomaly of specific volume in a section occupied four times, May 10—14, 1959. Map insert shows location of section.

Table 5. *Bottom current measurements on sections May 10—14, 1959.*

Depth m	Section No	Date (May)	GMT h m	Velocity cm/sec	Direction (toward)
370	1	10	16 45	20.0	S
370	1	10	17 20	24.7	S 30°W
365	2,3	12	13 54	no current	
365	2,3	12	14 20	5.0	N 30°E
380	4	14	07 55	11.7	S 60°W
380	4	14	08 24	11.5	S 30°W
405	1	10	19 14	16.7	E
405	1	10	20 16	24.4	N 80°E
450	2	12	11 45	9.4	N 50°E
450	2	12	12 18	6.5	N 50°E
410	3	12	15 50	no current	
410	3	12	16 40	no current	
400	4	14	05 24	(13.3	N)
400	4	14	05 52	9.0	N
485	1	10	22 20	8.0	E
485	1	10	22 57	5.5	S 80°E
555	2	12	09 18	6.0	N 10°W
555	2	12	09 55	8.0	N
510	3	12	18 30	7.5	N 40°E
510	3	12	19 07	10.2	N 50°E
470	4	14	03 37	18.2	N 20°E
575	1	11	02 23	10.3	N 80°E
650	2	12	06 35	8.6	N 50°W
650	2	12	07 20	8.8	N 20°W
570	3	12	21 23	10.2	N 10°W
570	3	12	22 10	13.6	N 40°E
575	4	13	23 45	(15.0	N 60°E)
575	4	14	00 20	11.2	(N 80°E)
660	1	11	04 47	(2.0 uncertain)	
660	1	11	05 30	(4.0 uncertain)	
730	2	12	03 38	uncertain	
730	2	12	04 20	6.5	N
635	3	13	00 30	12.3	N 10°W
635	3	13	01 10	11.5	N 20°W
675	4	13	21 18	19.6	S 30°W
675	4	13	21 55	18.0	S 50°W
820	1	11	08 33	6.1	N 10°W
820	1	11	09 20	4.0	N 10°E
880	2	11	23 20	11.4	(N 60°W)
880	2	12	00 22	9.0	N 10°E
770	3	13	03 48	no current	
770	3	13	04 43	no current	
830	4	13	17 52	9.0	S 30°W
830	4	13	18 46	7.5	S 30°E
1000	1	11	12 18	8.5	N 10°E
1000	1	11	13 07	13.0	N 10°E
1025	2	11	18 46	no current	
1025	2	11	19 50	5.2	N 20°W
925	3	13	07 58	no current	
925	3	13	08 51	no current	
980	4	13	14 05	no current	
980	4	13	15 10	no current	

compared to 48 hours in 1957). The section was very near the 1957 section, but slightly more to the north (see map insert, Fig. 44). The distribution of $\Delta\alpha$ in the section was as shown in Fig. 44. In that figure, the bottom velocities measured at the stations are also shown. To make comparison easier, the bottom current measurements are also summarized in Table 5. Fig. 44 shows that the distribution of $\Delta\alpha$ varies distinctly from one occupation of the section to the next, although the whole programme was completed in less than four days. The differences are again mostly connected with "waves" on the isolines. This wavy form is especially conspicuous on the 1st, 2nd and 4th sections; and it is of the same kind and the same dimensions as found on several previous occasions (Figs. 15, 16, see also S I, Figs. 7 and 9). The geostrophic velocities at the surface (cm/sec), relative to 1000 db or bottom, are entered just below the surface line in Fig.

44. A positive sign means current toward the eastern side of the section. Although the relative velocity will be negative in some parts of the sections, because of the wave-form of the isosteres, the total relative transport through a section will always be positive, because the scale of the "waves" is so small that the general trend of the isosteres prevails even in so short sections. The bottom current measurements gave velocities of the same order of magnitude as the measurements in 1957. In other respects, however, there are differences. In 1957, all the measured velocities, with the exception of those at the 800 meter level, had directions between north and east. In 1959, quite a number of the velocities had southerly or westerly components. It is seen from Table 5 that only 18 out of 35 single measurements with well-defined direction gave directions towards N, E or between N and E. Seven of the measurements gave directions S, W or between S and W, eight between N and W, and two between S and E. Although the majority of the directions was still in the N—E bracket, there was thus a much larger proportion of the velocities in other directions in 1959 than in 1957. It is difficult to find any system in the distribution of the directions over the different depth levels. Currents with westerly or southerly components were found at nearly all levels. Thus, strong currents between S and W were measured in the Atlantic water at the 370—380 meter levels as well as in the sub-zero temperature water at 675 meters (the isosteric line $\Delta\alpha = 0$ very nearly coincides with the isotherm $t = \div 0.9°C$). The current strength was also irregularly distributed over the depth levels. Although it can be said that the current was generally weakest at the deepest level (about 1000 meters), cases of weak or missing current were also found at the shallowest levels, and at 1000 meters on section 1 a current of about 10 cm/sec was observed. Thus, the bottom current measurements on these sections give a rather confused picture. Great variations in current strength and directions are found within all depth brackets, and also from level to level in the different sections. The series of bottom current measurements to be described in the following paragraph will show, however, that such variability should not be unexpected.

b. Current measurements at anchor station "A", 1959. The purpose of this anchor station was to obtain a series of bottom current measurements in the cold bottom water which ascends on the slope to depths of 500—600 meters (S I, p. 24). The hydrographic conditions at the station are seen from Fig. 45, showing the temperature-depth and salinity-depth curves from four hydrographic serial observations made at the position. At the deepest observation level, 580 meter (15 m above the bottom), the temperature varied between — 0.45 and — 0.89°C. The first three observations, taken on May 14, 15 and 16, showed a very sharp temperature gradient extending right to the bottom, whereas at the last observation taken in the morning of the 17th after the termination of the current measurements, a thicker layer of cold bottom water is apparent. The anchor was down at about 18^h on May 14 at a depth of 595 meters, the position being 62°21'N and 1°13'E. The position is indicated by a cross in the map insert of Fig. 44. The measurements lasted about 48 hours. A total of 60 measurements

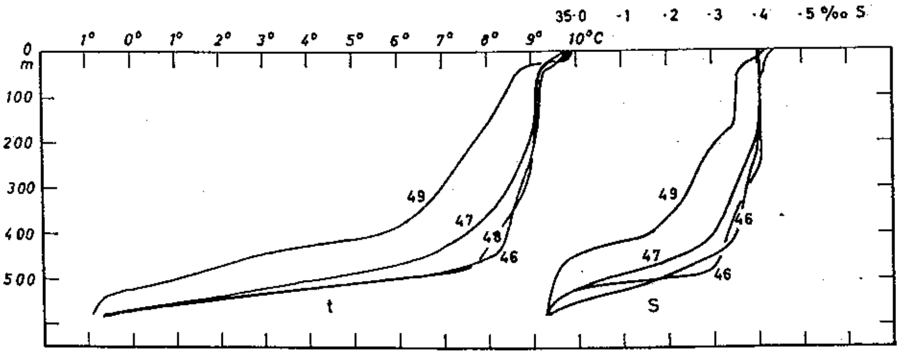


Fig. 45. Anchor station "A", 1959. Temperature and salinity curves, May 14—17.

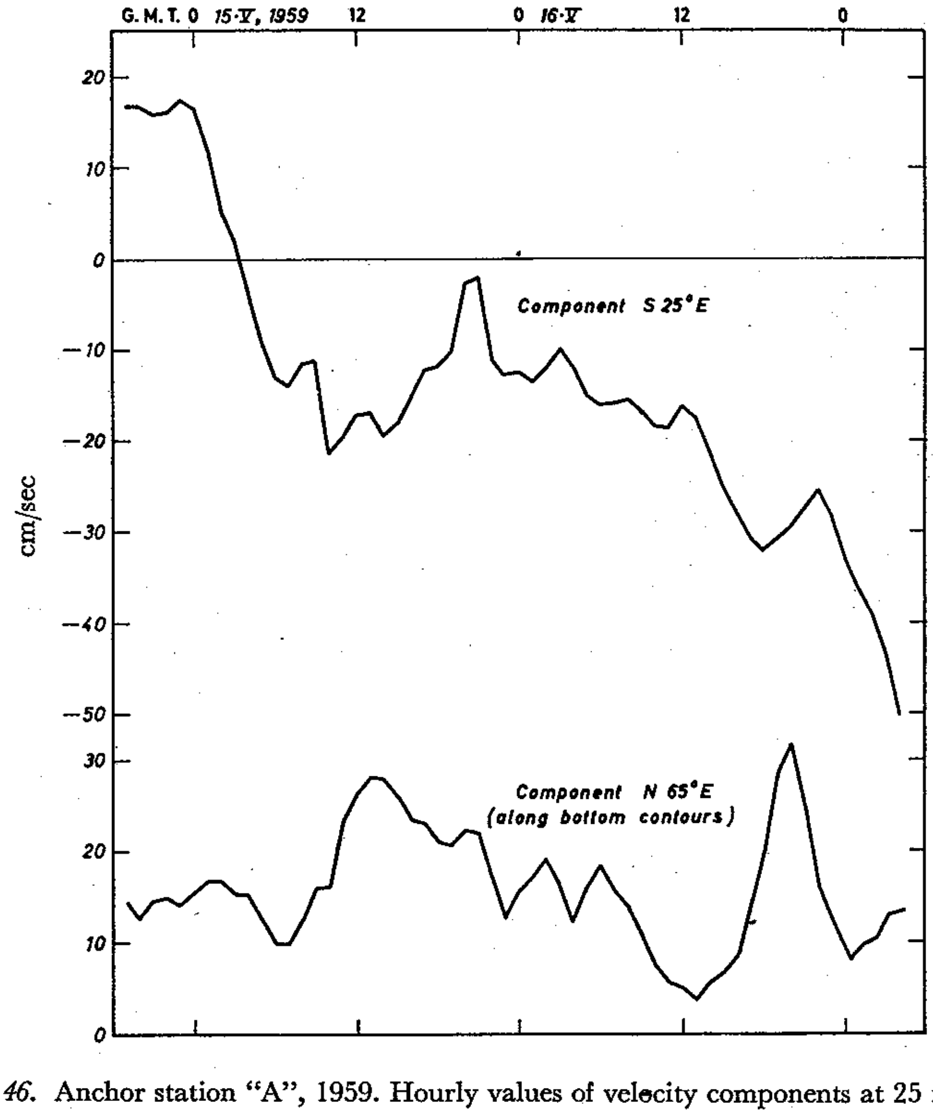


Fig. 46. Anchor station "A", 1959. Hourly values of velocity components at 25 meters.

was made with the MOSBY bottom current meter. In addition, measurements were made at 25 meters with the Ekman single reading current meter, on an average 4 measurements per hour. From the smoothed curves of speed and direction, hourly values were taken and decomposed along the bottom contour lines (N 65°E) and at right angles to them (S 25°E). The components are shown in Figs. 46 and 47. The hodographs of the velocity vectors at 25 m and at the bottom are plotted in Fig. 48. The currents at 25 m are more or less as could be expected, with directions mainly between N and E, and current strength varying between 15 and 50 cm/sec. There is no obvious periodicity in the variations, though. The series of measurements at the bottom is the longest one made on the slope bottom thus far. It started at about 19 GMT on May 14 and lasted until 18 GMT on the 16th. The last measurement had to be discarded because the ship had started dragging the anchor and obviously also dragging the apparatus along the bottom, and this led to a distribution of the balls over a large number of widely different compartments. In all the other measurements (of course except those with the very weak current strengths) the balls were mostly concentrated in a few neighbouring compartments, and the measurements seemed on the whole to be of a very satisfactory quality. This series of bottom current measurements strongly contrasts with the series made in 1956 at a similar location on the slope (p. 19). In that case, all the measurements in a series of 11 hours gave a current roughly along the isobaths with the strength varying within reasonable limits and a mean of 16 cm/sec. Intuitively, this would seem to be a reasonable and normal situation. The series of 1959, however, does not at all conform to this simple pattern. During the 45 hours of the series, the velocity vector rotated around the whole compass, and the strength ranged from below the measurable limit to nearly 25 cm/sec. At the time of the strongest current, the speed was greater at the bottom than at 25 meters. Judging from the curves in Fig. 47, there is no indication of a variation that can be fitted to any tidal period,

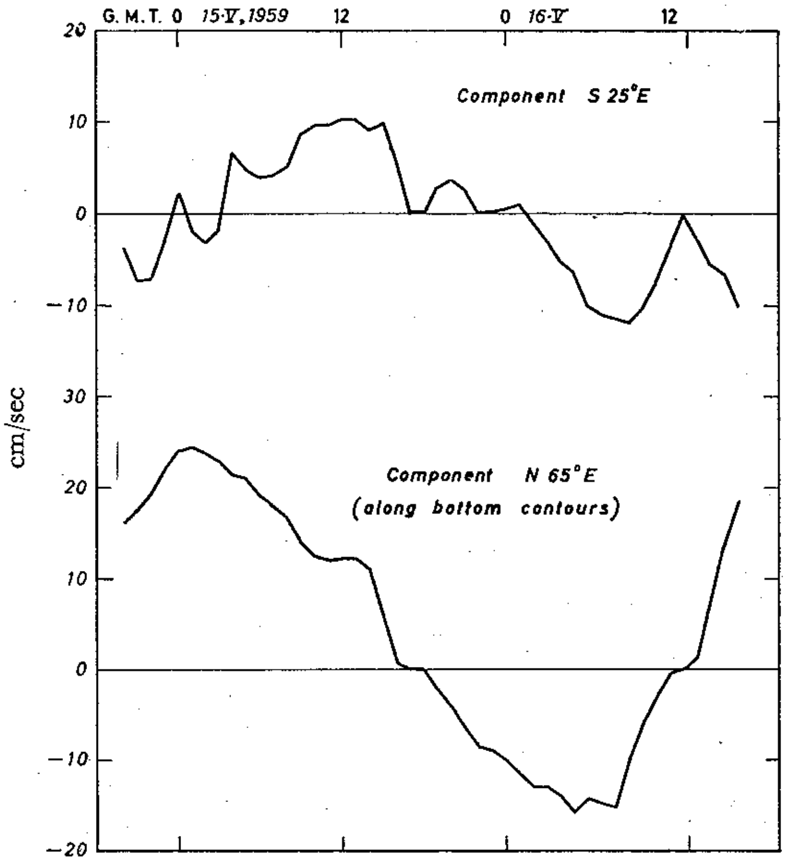


Fig. 47. Anchor station "A", 1959. Hourly values of velocity components 1.5 meters above the bottom.

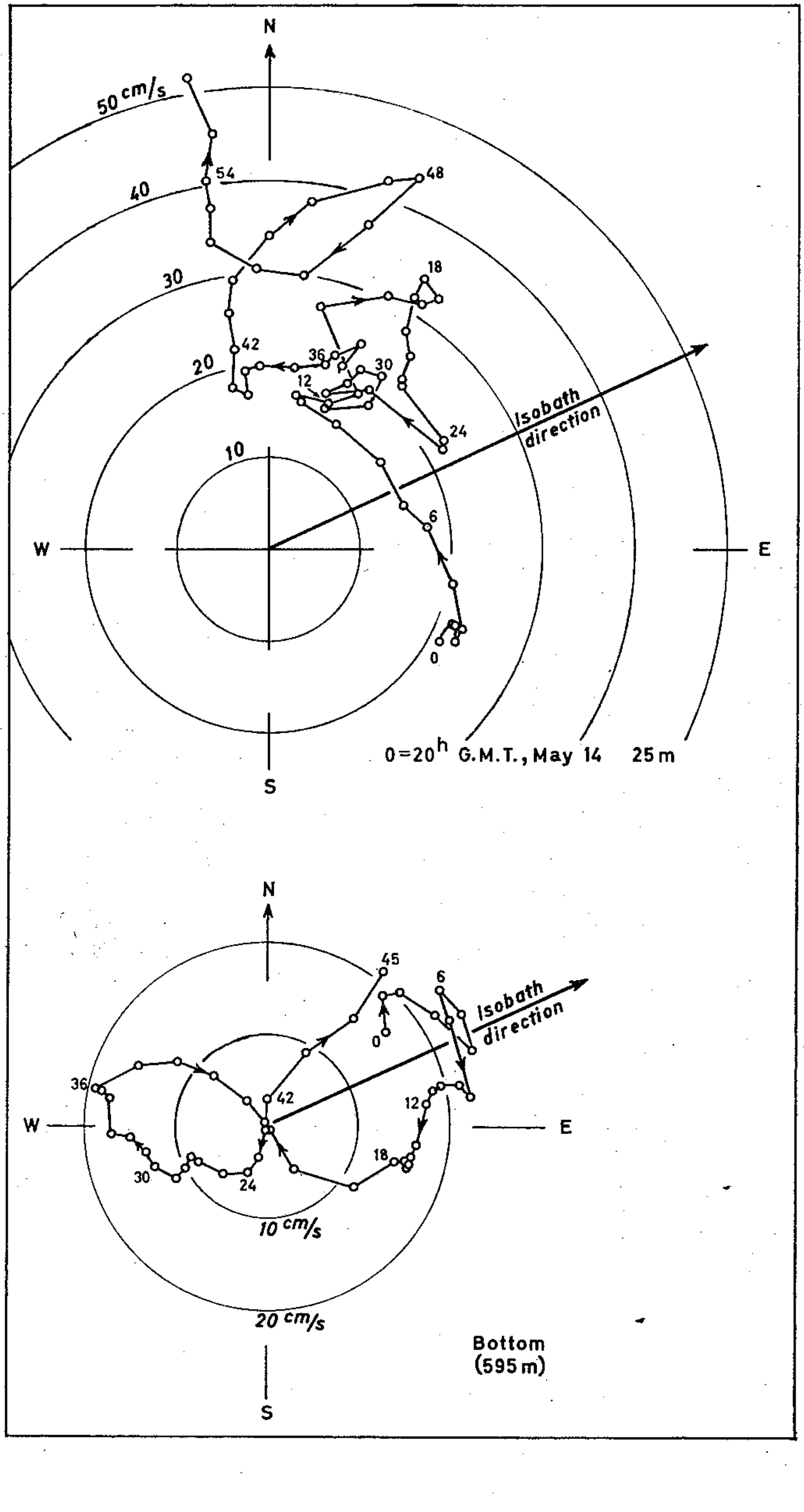


Fig. 48.

Anchor station "A", 1959. Hodographs of velocity vectors at 25 meters, and at 1.5 meters above the bottom.

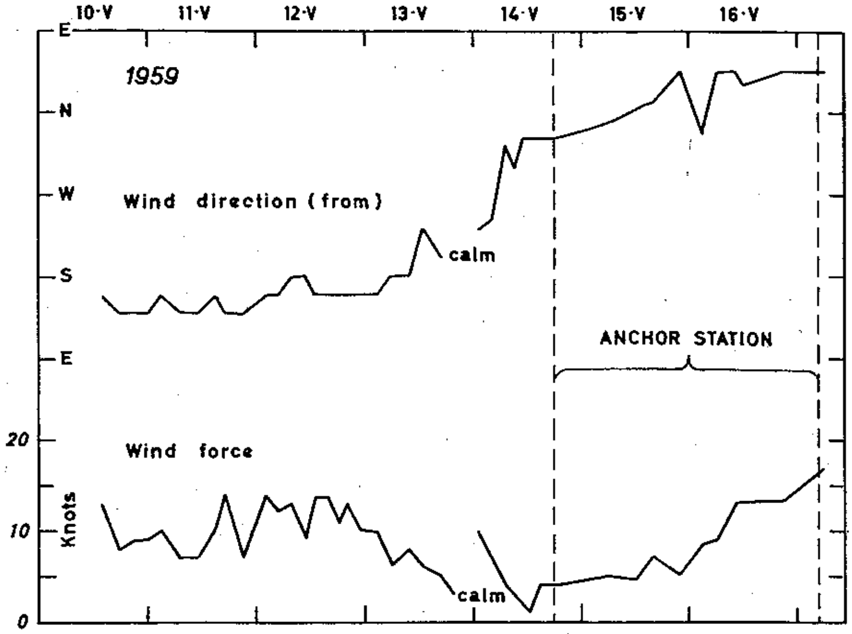


Fig. 49. Wind conditions before and during anchor station "A", 1959.

and the same applies to the inertial period (13.5 hrs). The apparent period of between 40 and 45 hours does not conform to any period of physical significance, known to the present writer. It is seen that the component along the bottom contours is by far the strongest and the most regularly varying, the amplitude of the transverse component being only one half of that of the other component. Thus, it may be said that a dominant feature of the movement is a flow back and forth along the isobaths. If the current may be regarded as approximately geostrophic, this means that the pressure gradient at the bottom was mainly perpendicular to the bottom contour, and varying in the same manner as the current. The pressure variation could be due to water piling up against the slope and then receding and swinging back, giving way to a reversed pressure gradient. As the current at 25 meters did not exhibit the same variation as the bottom current, such an explanation would presuppose that a redistribution of the field of mass was taking place during the variation, so that the current variations at the bottom were connected with an internal mode of oscillations. During the last three days before the anchoring, the wind was blowing steadily from SSE with an average force of 12 knots (Beaufort 3—4). Approximately a day before the measurements started, the wind decreased to almost calm and veered to W and NW (Fig. 49). Later on, it veered almost due N and increased in strength until it reached more than 15 knots on May 16—17; so that the measurements had to be discontinued. The air pressure field is shown in Fig. 50. The wind variations are to some extent compatible with the tentative explanation sketched above. If the persistent SSE winds in the days prior to the anchor station had carried water away from the slope and thus created a pressure gradient giving a current along the slope in south-westerly direction, it seems possible that the abatement of the wind on the last day before the anchor station may have initiated

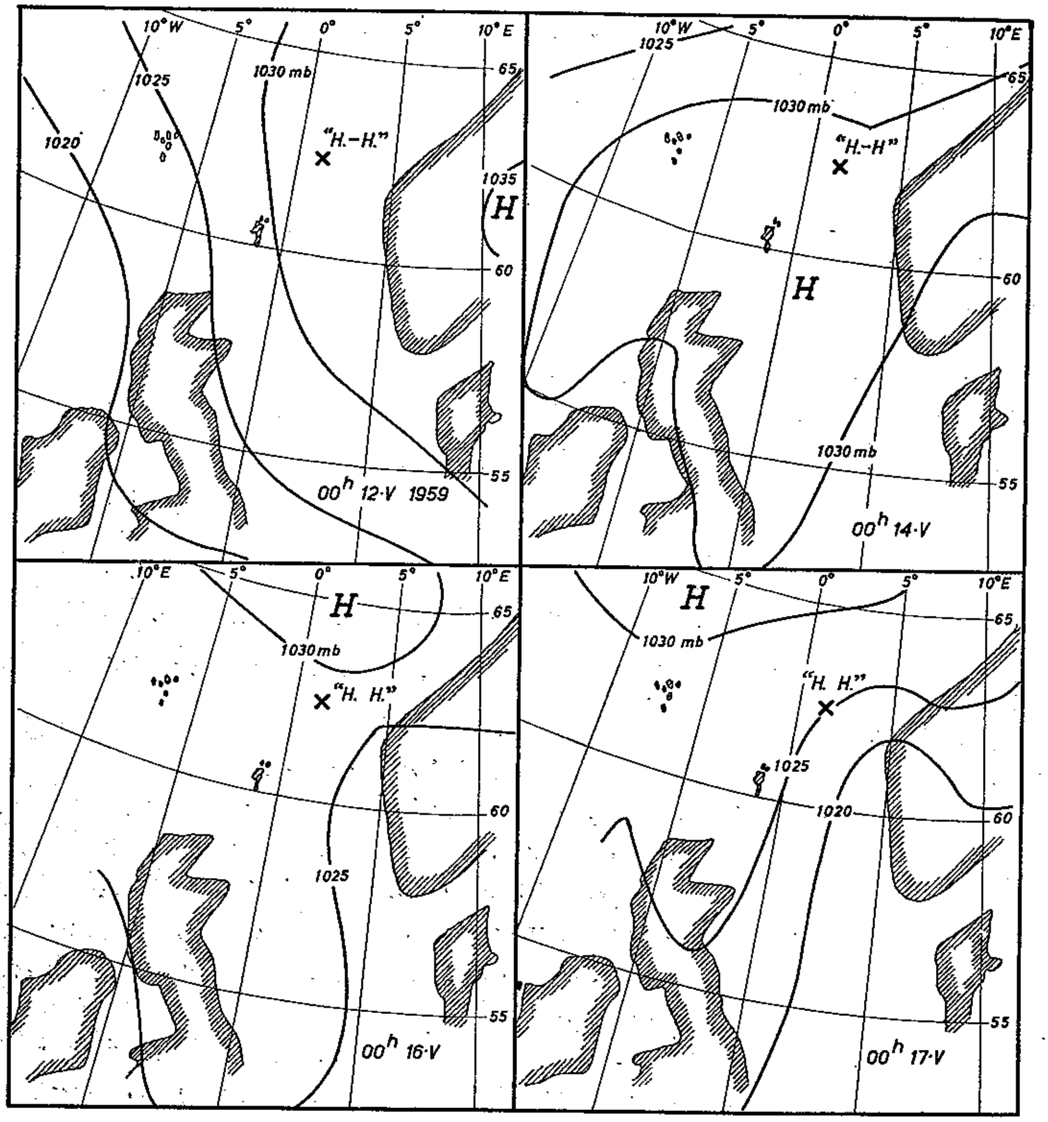


Fig. 50. Atmospheric pressure distribution, May 12—17, 1959.

an oscillation giving a current in the opposite direction on the first day of the anchor station. The reversal of the current in the second half of the measuring period (Fig. 47) would then correspond to the next phase of the oscillation. Several objections may be raised to such an explanation. Firstly, it seems that the winds during the days prior to the anchoring, although persistent, were rather too weak to cause such a marked oscillation as that observed. Secondly, one would expect the period of oscillation to be the inertial period. The explanation must therefore remain highly speculative.

c. Current measurements with drogues (current crosses). In order to obtain current measurements in the Norwegian Atlantic Current free from the disturbing influence of the ship's motion, a series of current measurements using free floating drogues was planned. Preparatory to this experiment, a few short sections were made to obtain an overall

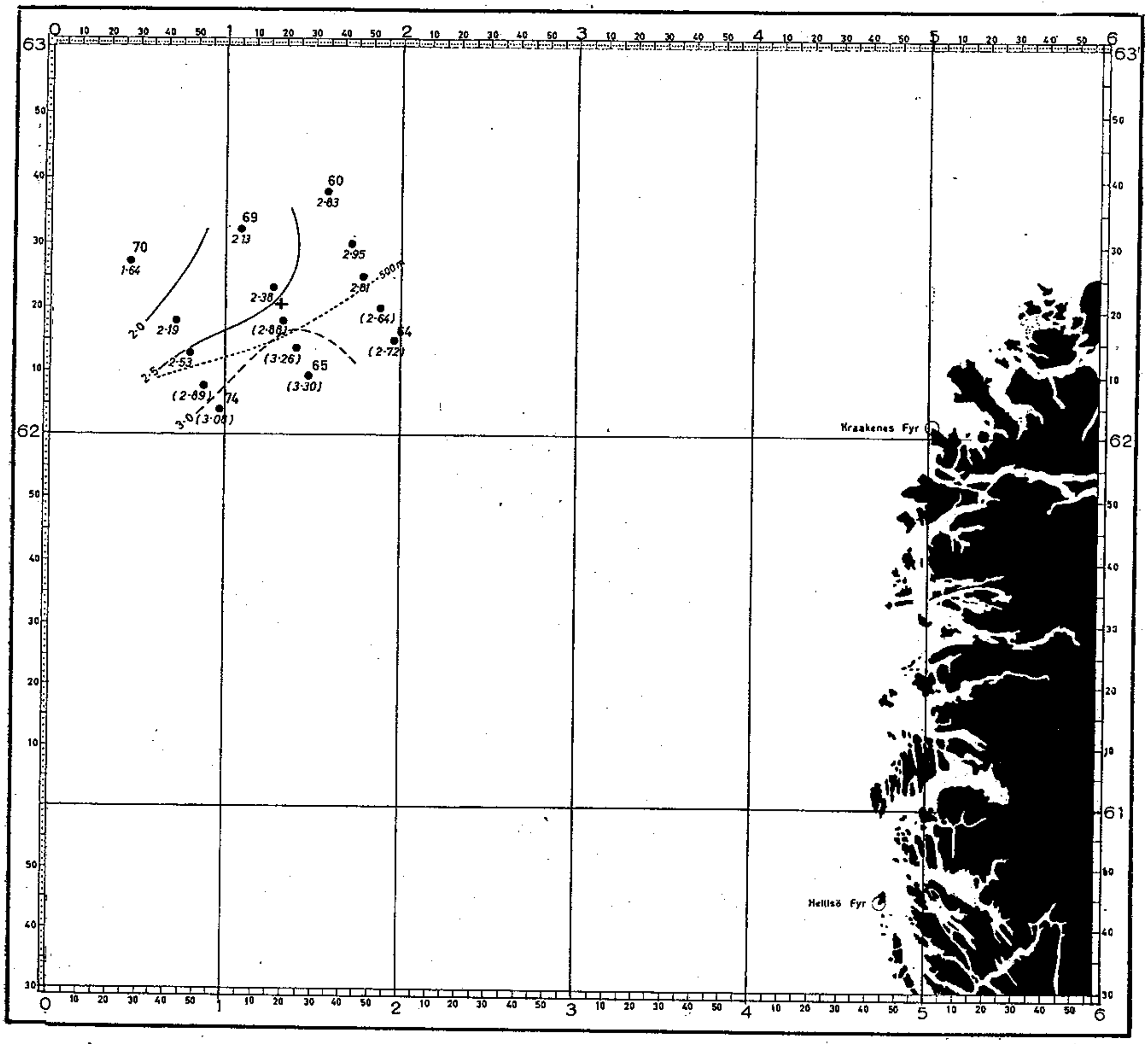


Fig. 51. Dynamic topography (surface relative to 500 db. ΔD in dyn. dm) of area of drogue measurements, May 18—19, 1959.

picture of the hydrography of the area (Fig. 51). The survey was completed in 24 hours (13^h on May 18 to 13^h on May 19). As is evident from the results of the previous investigations in this area, the picture given by such a small scale survey may change fairly rapidly. In Fig. 51, we have therefore outlined only the main features of the dynamic topography of the surface relative to 500 db.

The most distinct feature is seen to be a depression in the NW-corner of the area, contrasted with high values of ΔD in the western and southern part. Thus, the relative geostrophic current will on the whole be toward the northeast. Although the computed dynamic heights are almost certainly subject to fluctuations, this feature is so marked that it can probably be taken as representative for the situation before the start of the

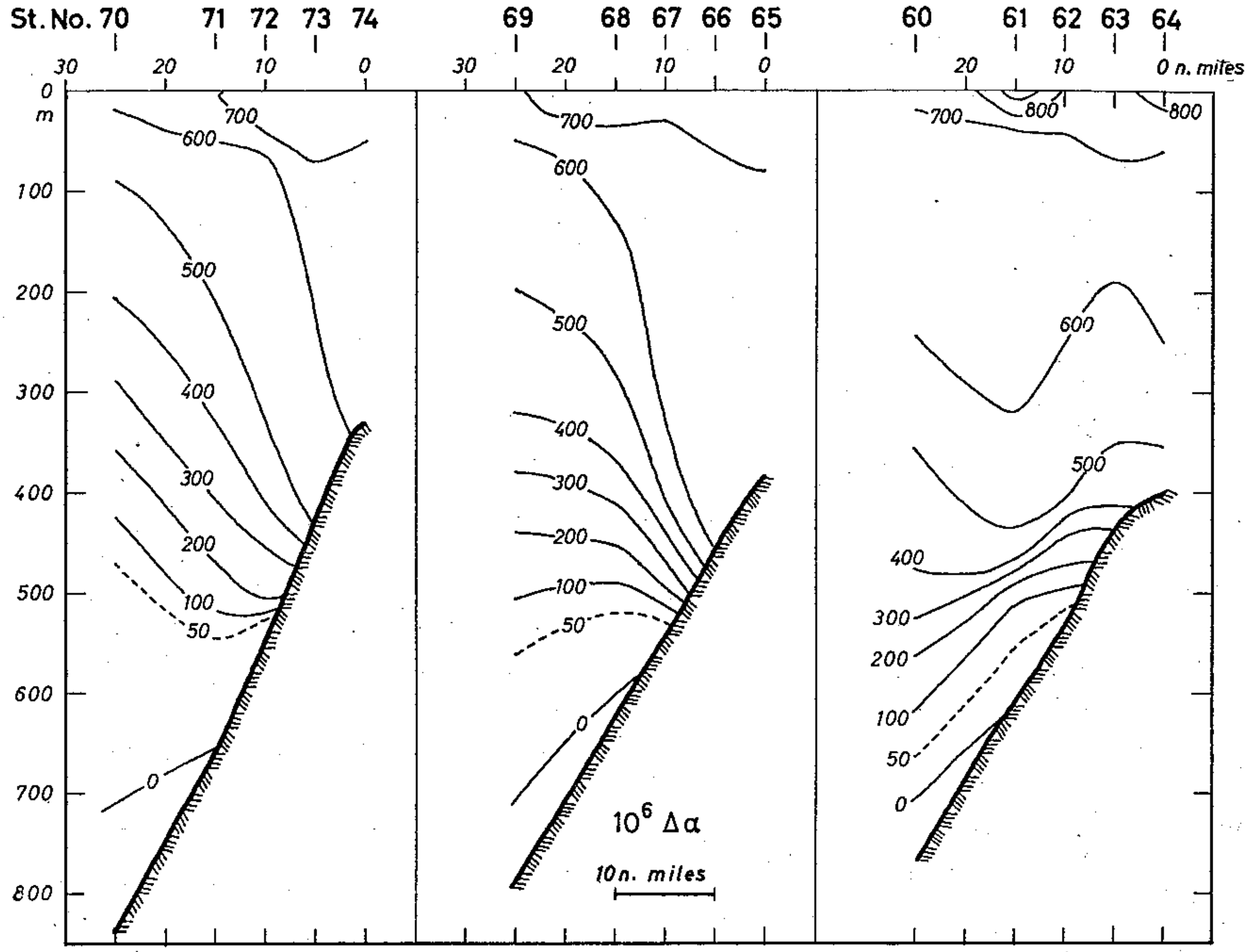


Fig. 52. Distribution of anomaly of specific volume in three sections, May 18—19, 1959.

current measurements. The current cross measurements were centered in the middle of the surveyed area (between stations 67 and 68). The distribution of $\Delta\alpha$ in the three sections is reproduced in Fig. 52, and the geostrophic current components normal to the section between successive stations are given in Table 6 for some selected depths. The current is relative to 500 meters, or to the bottom at shallower depths. The positive direction is toward the ENE side of the section.

It is seen that there is a marked difference between the two neighbouring sections, 60—64 and 65—69. This indicates the possibility that a relatively slight shift in the position of the water masses could bring about a considerable change in the current measured e.g. at a position between sts. 67 and 68, where the drogue measurements took place. Such changes were, in fact, found on a few occasions during the measurements (Fig. 54).

The measurements with free-floating drogues (current crosses) were carried out during the period May 19—May 24. The current crosses were made of two thin iron sheets at right angles to each other, with an area of 1 m² each. The cross was suspended by a thin nylon gut from a plastic float on the surface, equipped with a radar target

Table 6. *Geostrophic current (cm/sec) normal to sections st. 60—64, 65—69, and 70—74. Reference line: 500 db (or bottom).*

Depth m	Between stations				Between stations				Between stations			
	60-61	61-62	62-63	63-64	65-66	66-67	67-68	68-69	70-71	71-72	72-73	73-74
0	+ 5	— 11	— 12	+ 6	+ 3	+ 34	+ 40	+ 10	+ 24	+ 26	+ 27	+ 17
100	+ 3	— 9	— 13	+ 5	0	+ 33	+ 39	+ 8	+ 22	+ 21	+ 23	+ 19
200	+ 2	— 8	— 10	+ 3	0	+ 32	+ 32	+ 6	+ 18	+ 18	+ 16	+ 16
300	0	— 6	— 7	+ 1	0	+ 27	+ 23	+ 4	+ 13	+ 14	+ 12	+ 6
400	— 3	— 4	— 3	0	0	+ 17	+ 10	+ 2	+ 6	+ 7	+ 7	0

on the top of a pole about 4 meters above the surface of the sea (Fig. 53). The error introduced by the drag of the surface current on the float was probably very small, since the area of the float was small compared with that of the current cross and, as indicated by the measurements with drogues at different depths, there was probably very little difference between the currents at the surface and at the level of the current cross. As a fix point an anchored buoy with a radar reflector was used. This buoy was anchored in the position 62°21'N, 01°18'E, where the depth was 580 meters (cross in Fig. 51). The length of the 2 mm anchor wire was 613 meters, giving a maximum radius of swing for the position buoy of 200 meters. The drifting buoy was launched close to the anchored buoy, and the distance and direction between them during the drift were measured by radar. As a rule, the ship stayed close to one of the buoys, most often the drifting buoy, and observed the radar echo from the other buoy. When more than one drifting buoy was used simultaneously, the ship stayed close to one of them or, in some cases, moved successively from one buoy to another. When the distance between the buoys had become greater than the radar range, the drifting bouy was taken on board the ship and launched again close to the anchored buoy. In this manner a total of 12 drifts were measured. In the first eight of them we used a single drogue with the cross at 50 meters. In the next two series, an additional drogue was used with the cross as close to the surface as possible (4 meters below the surface). In the last two series, three drogues were tracked simultaneously, a drogue with the cross at 200 meters being added to the aforementioned two drogues. As a rule, the targets we used could not be seen on our radar (Decca 212) at distances greater than about 3 nautical miles. The distance of nearly 5 miles in one of the series (No. 9, Fig. 56) was obtained by sighting the fixed buoy and the

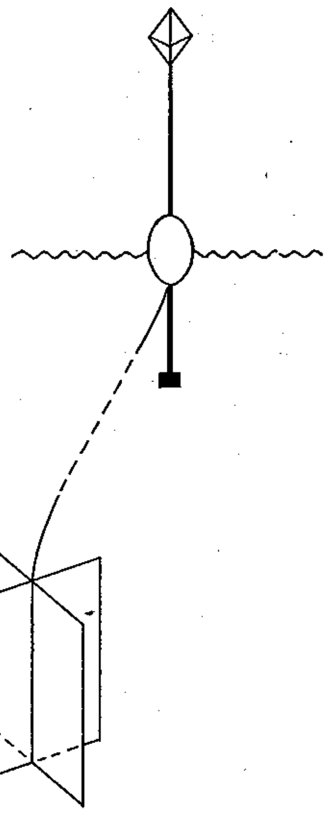


Fig. 53. Current cross used for measurements May 19—24, 1959 (not drawn to scale).

drifting buoy simultaneously from a position midway between the buoys. It should be mentioned that the programme was carried out under exceptionally favourable weather conditions. The sea was smooth, at times completely calm, and the wind force did not exceed 2 Beaufort except during the first series, when we had force 3—4. This was very lucky indeed, for it turned out that it would have been very difficult, if at all possible, to carry out the measurements at winds of 4 or 5 Beaufort. Experience showed that it would then be nearly impossible to distinguish the radar echo from the "noise" made by the waves on the radar screen.

Estimate of errors. As mentioned before, the anchored buoy had a maximum radius of swing of about 200 meters, a value which would, however, be reached only if the wire rope formed a straight line. The mean velocity of a drogue is determined by observing its position vector relative to the anchored buoy at a time t and at a later time $(t + \Delta t)$. If the anchored buoy has changed its position during this time by a vector $\mathbf{r}_0$, this will cause an error $\mathbf{r}_0/\Delta t$ in the velocity. This error will influence the measurement of the direction or strength of the current, or both, depending on the direction of $\mathbf{r}_0$ relative to the true current. In our case, the biggest conceivable error from this source would be $400/\Delta t$ m/sec. It is utterly improbable, though, that errors of such magnitude should have occurred in our observations. The position of the anchored buoy will be determined by the direction and strength of the current, barring dragging of the anchor. In all our series, with one exception, the direction of the current changed very little during the drift period of a drogue, so that the successive positions of the anchored buoy during such a drift period must have been approximately on a straight line. There are also reasons to believe that, with the mentioned exception, the distances between the different positions of the buoy in any one series must have been fairly small. If all the relevant parameters were known, the buoy's distance from the point at the surface vertically above the anchor (this distance will be called "the radius of swing") could be computed by means of methods similar to that presented by Mosby (1952, 1955) and Kullenberg (1951) (see also Watson, 1953). Approximate calculations show that the thin 2 mm wire will be so thoroughly stretched even at currents of about 20 cm/sec, that a further increase of the velocity can give only a relatively small addition to the radius of swing. For example, with a vertically constant velocity of 20.8 cm/s and a reasonable value of the current drag on the buoy, the radius of swing will be only 39 meters less than the maximum radius of swing. If we assume 20.8 cm/s in the upper 300 meters, and half that value from 300 meters to the bottom, the result is virtually the same. Even if the deep water should move in a direction different from that of the Atlantic water, the result would probably not be much different. Because the parameters entering into the calculation are not accurately known, there is room for some latitude in the calculated values; but it seems safe to conclude that the deviation from the maximum radius of swing is in any case less than 50 meters at the stated value of the velocity. With the exception of series No. 4, the velocity was always higher than 21 cm/sec. Thus, we may conclude that the radial differences between successive sightings of the buoy must always have been less than 50 meters, and probably most often

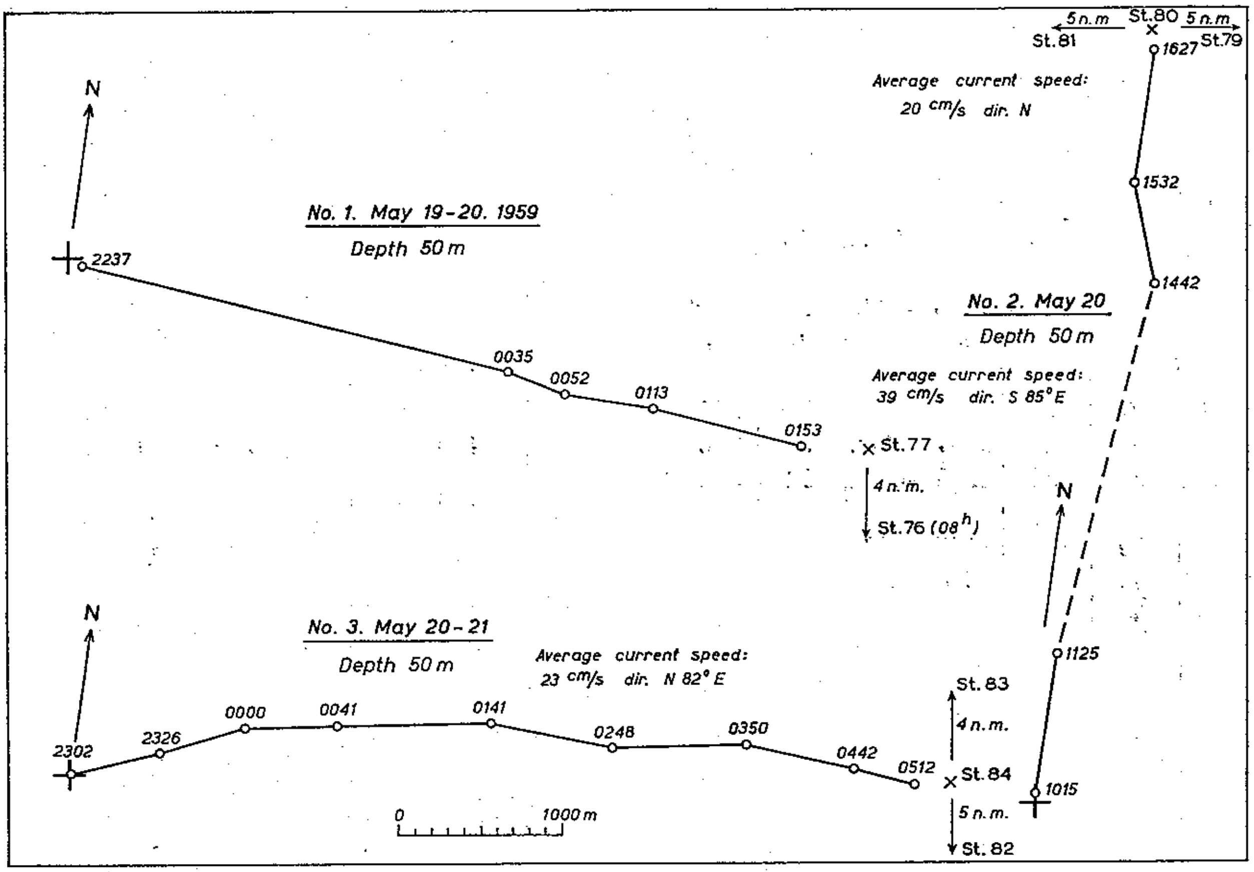


Fig. 54. Paths of drogues, series 1—3.

considerably less. The error in the mean value of the speed taken over 1—2 hours will thus be of the order 1 cm/sec, at most. Also the error connected with a change of current direction will be slight, as the directions within any single series with the mentioned exception never changed more than slightly. Thus, a change of 10° in the current direction will cause a displacement of the buoy of the order of 35 metres, corresponding to an error of less than 1 cm/sec in a mean velocity taken over 1—2 hours. As the vector of the velocity error will be nearly normal to the velocity, the error in the magnitude of the velocity will be negligible. With a current of 25 cm/sec, the error in the direction will be less than $\frac{1}{25} \cdot \frac{180}{\pi} = 2°$. Another source of error is in the determination of distance and direction on the radar screen. The directions may be determined within 1 or 2 degrees, and at a distance of 2 miles this will give errors of 65—130 meters. The distances are, roughly, correct within $\pm$ 2%, so that, in extreme cases, an error of 3 cm/sec may occur in mean values taken over 1—2 hours. The errors from this source may thus be considerably greater than those associated with the movement of the buoy. As a conclusion, one may say that a combination of the worst possible circumstances may give errors up to about 6 cm/sec, but that the majority of mean values taken over 1—2 hours will probably be correct within 2—3 cm/sec.

Table 7. *Velocities deduced from drifting current crosses.*

Series No	Depth m	Date (May)	Time	Velocity cm/sec	Direction (toward)
1	50	19	2237—0035	38	S 85°E
1	50	20	0035—0153	40	S 84°E
2	50	20	1015—1442	20	N 05°E
2	50	20	1442—1627	22	N 09°W
3	50	20	2302—0041	28	N 70°E
3	50	21	0041—0248	22	N 85°E
3	50	21	0248—0512	22	N 88°E
4	50	21	1106—1526	08	N 85°E
4	50	21	1526—1934	—	—
4	50	21	1934—2300	15	N 66°E
4	50	21	2300—0110	17	N 65°E
5	50	22	0839—1035	33	N 82°E
5	50	22	1035—1204	33	N 84°E
5	50	22	1204—1320	34	N 77°E
6	50	22	1524—1723	30	N 88°E
6	50	22	1723—2009	28	N 88°E
7	50	22	2130—2333	36	N 66°E
7	50	22	2333—0114	34	N 72°E
7	50	23	0114—0245	33	N 74°E
8	50	23	0334—0533	26	N 67°E
8	50	23	0533—0703	24	N 65°E
8	50	23	0703—0826	30	N 66°E
8	50	23	0826—0936	39	N 66°E
9	4	23	1053—1318	37	N 63°E
9	4	23	1318—1508	38	N 58°E
9	4	23	1508—1708	42	N 57°E
9	50	23	1035—1148	37	N 58°E
9	50	23	1148—1318	38	N 57°E
9	50	23	1318—1508	35	N 59°E
9	50	23	1508—1708	35	N 58°E
10	4	23	1815—1941	36	N 58°E
10	4	23	1941—2123	35	N 65°E
10	4	23	2123—2258	31	N 69°E
10	50	23	1840—2123	30	N 60°E
10	50	23	2123—2258	34	N 60°E
11	4	24	0020—0248	22	N 48°E
11	4	24	0248—0507	25	N 38°E
11	50	24	0047—0210	30	N 55°E
11	50	24	0210—0344	28	N 51°E
11	50	24	0344—0645	22	N 55°E
11	200	24	0103—0248	28	N 60°E
11	200	24	0248—0445	22	N 60°E
11	200	24	0445—0715	21	N 50°E
12	4	24	1110—1234	38	S 85°E
12	4	24	1234—1412	23	S 81°E
12	4	24	1412—1618	30	N 68°E
12	50	24	1129—1307	38	N 87°E
12	50	24	1307—1440	32	N 75°E
12	50	24	1440—1605	33	N 82°E
12	200	24	1243—1424	33	N 80°E
12	200	24	1424—1626	32	N 78°E

The results of the measurements are presented in Figs. 54—57, where the motion of the drogues is shown by dots representing the successive "fixes", with the corresponding times (Norwegian summer time = GMT + 2 hours) attached. The hydrographic stations are also indicated in the figures (see below). The velocities that can be deduced from Figs. 54—57 are summarized in Table 7. For reasons given above, we have for this purpose used periods longer than 1 hour, sometimes up to 3 hours, and the velocities given in the table are to be considered as mean values over the periods stated therein. It is seen from Figs. 54—57 that the current was predominantly toward

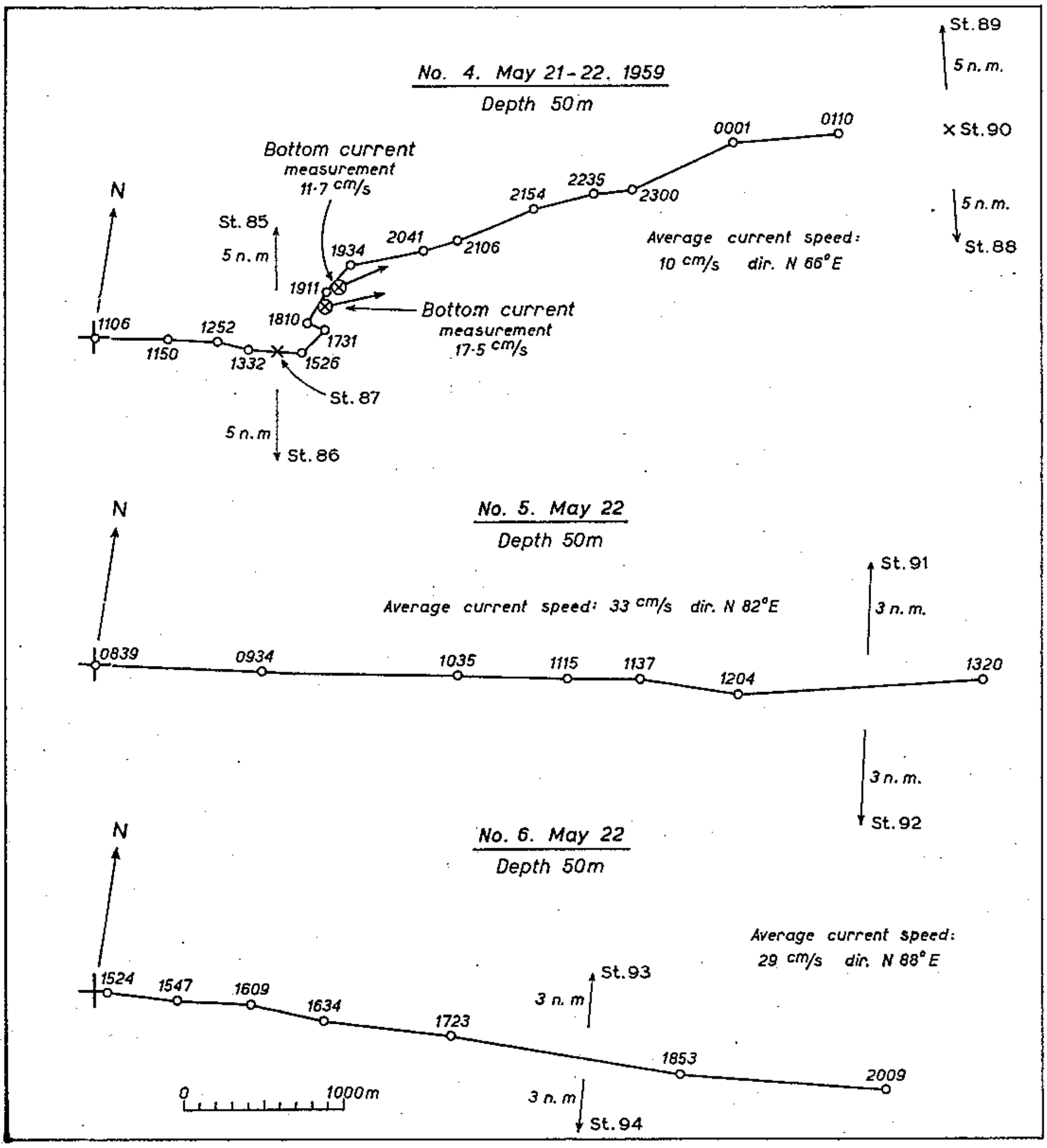


Fig. 55. Paths of drogues, series 4—6.

directions between N and NE, and that the changes in direction took place gradually. There is one notable exception to this rule: In the second series, the current (at 50 meters) was directed almost due north, whereas in the series just before and just after it was directed almost due east. In these early series, we had not yet established a routine so that the series could follow each other with short interruption, and it is therefore not possible to ascertain just how quickly the change occurred. There is little in the way of external factors to explain the change, and it must probably be classified as a case of macroturbulence, perhaps related to a large eddy similar to those previously described in this paper. It is also seen that only in a few cases (series 4, 8, 12) do the velocities within a series vary much more than can be explained by observational errors. Series 4 is the most remarkable of these cases (Fig. 55). After 5 hours' slow drift toward

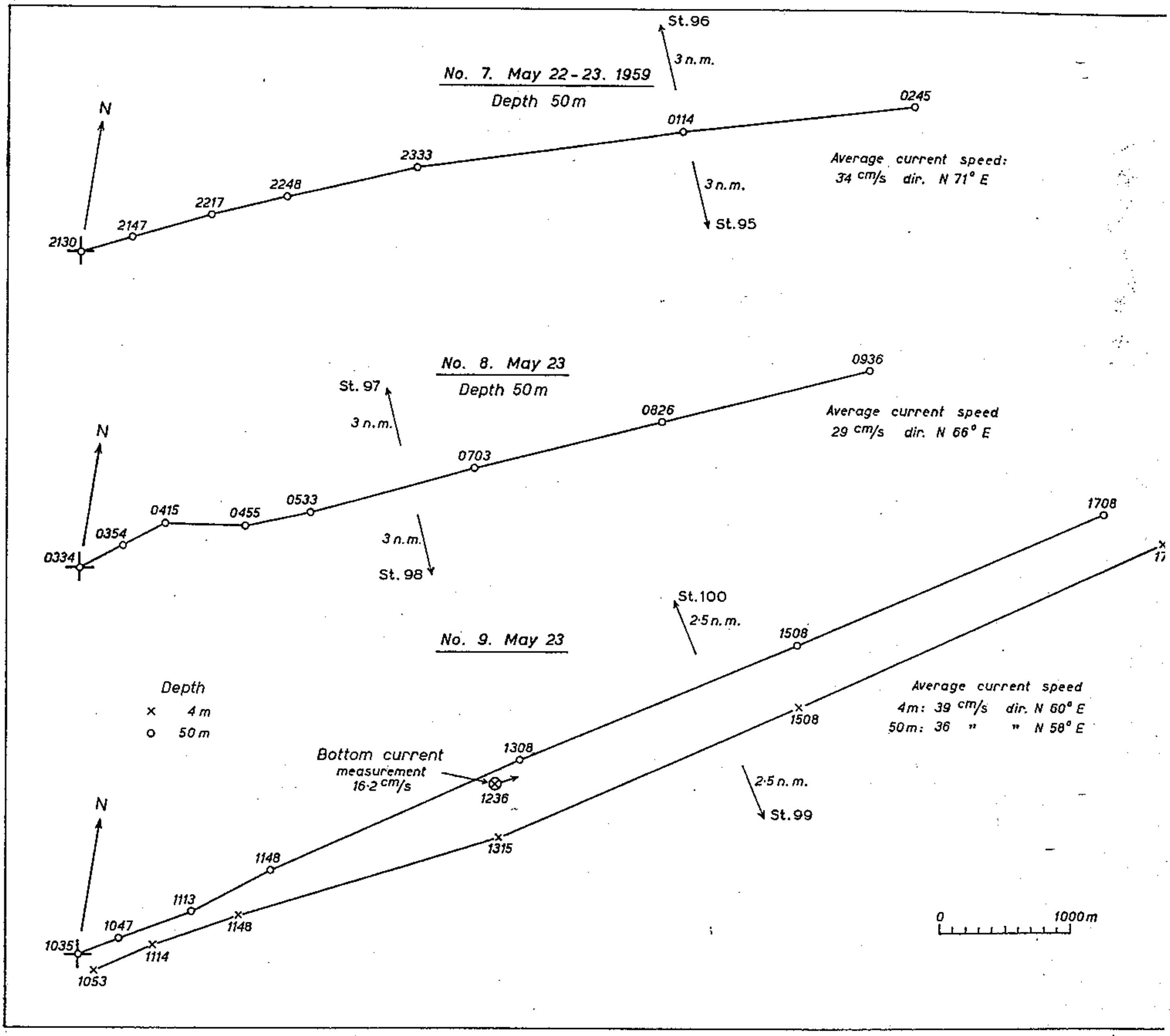


Fig. 56. Paths of drogues, series 7—9.

the east, the drogue came to a stop and moved very little during the next 4 hours. This is the only case when the movement of the anchored marker buoy can have had any appreciable influence on the velocities computed from the drift of the drogue. In any case, the current must have been insignificant during at least 4 hours. After that, the drogue gained speed again and ultimately reached 17 cm/sec in a direction which differed about 20 degrees from the direction before the stop. While the drogue lay motionless, we took the opportunity to make two bottom current measurements in the vicinity of the drogue (at 590 meters). The first was made at 19^h10^m, and gave

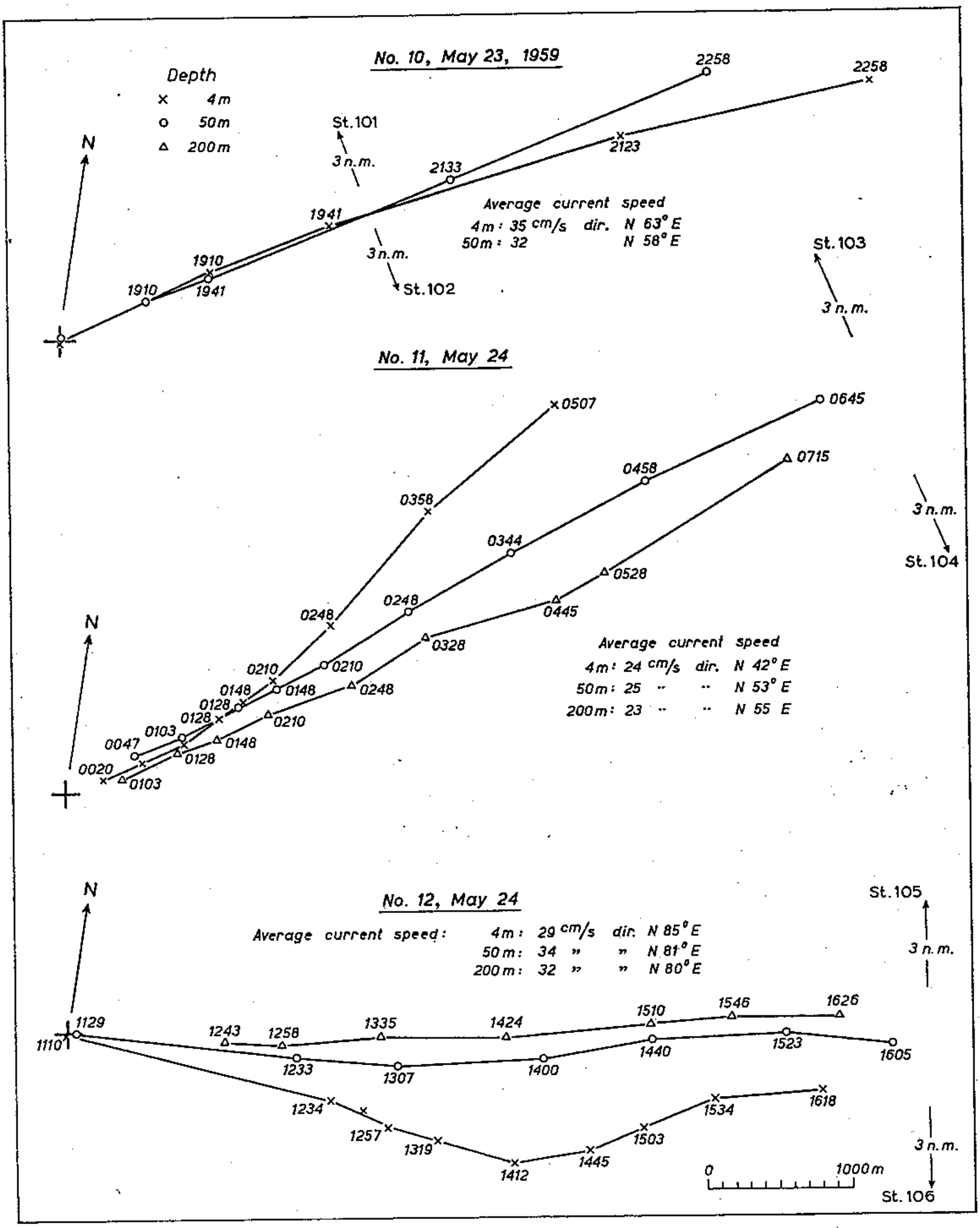


Fig. 57. Paths of drogues, series 10—12.

17.5 cm/sec in direction N 69° E. At that time, the current at 50 meters can at most have been a few cm/sec. The second measurement was made at 19^h50^m, and gave 11.7 cm/sec in direction N 58° E. The drogue had then started to accelerate, but had probably not reached a speed comparable to that at 590 meters. The measurements thus seem to indicate that the remarkable current variations were confined to the Atlantic water and that a similar development did not exist in the deep water.

These variations took place during very quiet and constant weather conditions. In this case, and also in series 8 and 12, the acceleration (computed by means of velocity differences over 2—3 hours) was of the order of magnitude 10^{-3} (cm sec^{-2}) and thus comparable to the Coriolis acceleration by moderate velocities (10—20 cm/sec). In such cases, the geostrophic approximation will obviously not be valid. In most of the other series, however, the acceleration was much smaller, so that the magnitude of the acceleration term will not constitute any obstacle to the application of the geostrophic approximation.

As mentioned above, we tried in some of the later series to observe current crosses at more than one level simultaneously (series 9, Fig. 56; series 10, 11, and 12, Fig. 57). The differences in velocity between the three levels (4, 50 and 200 meters) were relatively slight, both in speed and direction. It seems that the Atlantic water down to at least 200 meters moved more or less as a single body, without any appreciable decrease of speed with depth. In one case (series 12), the current crosses at 50 and 200 meters moved even faster than the cross at 4 meters. In the two series (11 and 12, Fig. 57) where the currents were observed at all three levels, the closest similarity was between 50 and 200 meters. The motion at 4 meters seemed to be slightly more irregular than at the two other levels. In this connection it should be said that the distance between floats in the same series could be determined with greater accuracy than the absolute positions of each single float, so that the differences between the velocities of drogues at different levels in the same series can be considered real even if they are not more than 1—2 cm/sec.

In addition to the drift measurements, we also tried to make short hydrographic "sections" at right angles to the direction of drift in order to obtain comparable geostrophic current estimates. As a rule, the "sections" consisted of two stations, one on each side of the path of the float. The directions of the "sections", as well as the distances between the stations, are indicated in Figs. 54—57. Where such a "section" consisted of three stations, only the two outermost stations have been used for the dynamic computations. As far as practicable, we tried to make such a "section" for each of the drift series. Especially in the beginning, the stations were often made between two drift series, and in these cases the observed and the computed currents are not so easily comparable. The results of the geostrophic current computations are found in Table 8. The relation of the station pairs to the different drift series can be found by a comparison with Figs. 54—57. In the second column of Table 8, the distances between the stations are given. These distances are relatively accurate, as they are determined directly from the ship's log readings on the run between the stations, with the Loran readings as an additional control. Roughly, one can say that the error will rarely exceed 0.5 miles, corresponding to about 10% in the velocity. The geostrophic velocities relative to 500 db are given for the 0, 50, and 200 db levels. Letters in brackets indicate the side of the section toward which the computed current components were directed. It is seen from the table that the relative velocities (or, more precisely, their components normal to the section) are always in the direction of the observed currents. The only

Table 8. *Geostrophic velocity components from stations taken during the drift measurements (see text).*

Station pair	Naut. miles	Date (May)	Time	Relative velocity cm/s			Current (cm/s) at 500 db
				0—500	50—500	200—500	
76— 77	4	20	0815—0930	13 (W)	13 (W)	6 (W)	
79— 81	10	20	1913—2143	13 (N)	12 (N)	9 (N)	
82— 83	9	21	0645—0825	6 (E)	7 (E)	7 (E)	
85— 86	10	21	1407—1618	12 (E)	11 (E)	11 (E)	
88— 89	10	22	0403—0611	8 (E)	7 (E)	5 (E)	
91— 92	6	22	1228—1338	14 (E)	14 (E)	10 (E)	20
93— 94	6	22	1751—1911	18 (E)	15 (E)	10 (E)	13
95— 96	6	23	0013—0132	17 (E)	14 (E)	9 (E)	20
97— 98	6	23	0557—0723	12 (E)	12 (E)	10 (E)	12
99—100	5	23	1345—1527	17 (E)	15 (E)	11 (E)	20
101—102	6	23	2010—2139	21 (E)	19 (E)	16 (E)	11
103—104	6	24	0816—0935	22 (E)	21 (E)	18 (E)	
105—106	6	24	1646—1801	13 (E)	11 (E)	7 (E)	22

exception is the first station pair, 76—77. These stations were, however, taken between the first and the second drift, when the current changed from E to N. It is seen, furthermore, that the relative velocities are generally much smaller than the observed currents, a fact that again indicates that the 500 db level cannot be taken as a "zero-surface". In the last column of Table 8, we have entered the velocity component that must exist at 500 db in order that the "absolute" current at 50 db shall achieve a value corresponding to the observed velocities at that level. This could be done only for the cases when the "sections" had been taken within one of the drift periods. The values are seen to fall well within the range of velocities observed at 500 meters on previous occasions. In one case (series 9, Fig. 56) a bottom current measurement (at 560 meters) was made just prior to a section (stations 99—100). The bottom current was exactly in the same direction as the current in the upper layers, and the velocity 16.2 cm/sec must be said to agree reasonably well with the tentative value 20 cm/sec for the current at 500 meters given in Table 8 for the "section" 99—100.

Concluding remarks. The investigations reported on in the present paper are of a varied nature. As mentioned in the Introduction, they were started because the results of the repeated occupation of a certain section across the Norwegian Atlantic Current (the Sognefjord section) indicated that a better knowledge of the structure of this current was necessary in order to understand the many irregular variations in the computed velocities and transports. The means to reach this end have been partly a more detailed mapping of the hydrography of an area through which this current flows, partly direct current measurements. Attempts have also been made to get an idea of the time variation of the fields of observed and derived quantities. Large and sometimes very rapid variations have been found in the hydrography of the investigated area. Furthermore, it has been confirmed that the current at the "reference surface" may at times be strong. These two circumstances may be responsible for much of the great variation found in the transport through the Sognefjord section as computed by the conventional methods. The first factor seems to indicate that, even if the Sognefjord section took only a few days to occupy, it is possible that the time variations were so large and rapid that the stations of the section could not be considered synoptic to a sufficient degree of approximation. With regard to the second factor, all the values of velocity and transport given in the first part of this work were relative values computed with reference to 1000 db in the open ocean and the bottom line on the slope. The values would be correct, or "absolute", only if the velocity on the "reference line" were everywhere zero. It was pointed out already then that there was no reason to believe that this was the case (S I, p. 18), and observations were presented showing considerable velocities close to the bottom. In the present paper, additional bottom current measurements have been presented, all of them confirming the impression that currents of considerable magnitude and great variability may exist on the slope close to the bottom even at relatively great depths. Currents of 10 cm/sec at 800 meters were not uncommon, and at 600 meters we might have 20 cm/sec or more. Even at 1000 meters, measurable currents were sometimes found, although less often. In this connection it should be mentioned that MOSBY (1953, 1955a) during the first trials of his bottom current meter in 1952 measured 18 cm/sec at 1030 meters and 10 cm/sec at 1110 meters on the slope in the Sognefjord section. The instrument had then not yet been fitted with a direction measuring device. The variability of the bottom currents makes one suspect that much of the apparent variation of the transport values might be related to such variations. It is becoming more and more evident, however, that a much larger amount of empirical evidence will be needed if we are to get anywhere near to a solution of this question. All we can do at present is to point out the possibility. Quite a different problem, which has not been discussed in the present paper, arises out of the results of the many bottom current measurements made in the real deep water. This has been shown to ascend to much shallower levels on the slope than in the deep basin of the southern Norwegian Sea (S I, p. 24). It is the question of the movement of the deep and bottom water in the Norwegian Sea. The measurements have shown that, in many cases, the cold slope water moves with a considerable velocity. A few examples taken from Fig. 44 (see also Table 5) will show this. At station 18, we have at 1000 meters measured 9 and 13 cm/sec in water of temperature — 0.94° C; at st. 21: 9 and 11 cm/sec at 880 m and — 0.95° C; at st. 35: 8 and 9 cm/sec at 830 m and — 0.92° C. These measurements were thus in real Norwegian Sea bottom water, only very slightly diluted. It is very hard to believe, however, that such velocities are representative of the general flow of the bottom water. More likely, they will have to be explained by a mechanism such as that proposed by ROSSBY (1938), to which we have previously alluded (p. 57). In fact, we have also in many cases observed no current in the cold slope water. Observations over a longer time and with improved techniques might give a reliable estimate of the average movement of the cold bottom water ascending on the slope, and thereby also give important clues to the problem of the circulation of the bottom water of the southern Norwegian Sea in general. In any case, our observations show that the conception of this circulation as a slow and regular drift probably does not hold.

In the present paper, we have on several occasions tried to compare observed currents with velocities computed from the geostrophic equation. These attempts have met with only moderate success. Most often only a qualitative agreement has been found, and in many cases no conclusions could be drawn because of incomplete observational data. The hydrographic conditions of the region we have worked in are rather complicated, so it has been difficult to collect comparable data for the two aspects to be compared. It has also been difficult to sort out different kinds of variations superimposed on the geostrophic current. Such variations will tend to be more of a nuisance in small-scale work than in larger surveys. On the whole, it seems that the experience from the experiments reported on in this paper tends to show that geostrophically computed velocities from small-scale surveys should be used with caution, especially as regards quantitative statements.

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