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AN ANTICYCLONIC EDDY IN THE SARGASSO SEA

by V. A. Bubnov and K. N. Fedorov

During the final (27th) USSR POLYMODE cruise of R/V Akademik Kurchatov, an open ocean anticyclonic eddy was found and studied in detail. Two closely spaced XBT surveys were carried out in August-September 1978, and two hydrographic sections through the center of the eddy were made with a time separation of one month (Figure 1).

During the first XBT survey (24-25 August 1978) the eddy's center on the 15°C isothermal surface (Figure 2a) was found to be located at approximately 30°50'N, 68°10'W. The maximum depression of the 15°C isotherm was 687 m, and the diameter of the eddy measured within the 610 m depth of the same isotherm was equal to 85-90 n.m. (According to Taft/Baranov et al., POLYMODE News No. 47, the mean "historical" depth of the 15°C isotherm in this area $\bar{H}_{hist}$ is equal to 611 m.) The boundary of the anticyclone showed up equally well on the 17°C isothermal surface (Figure 2b), while a reversed image of the eddy disturbance was clearly seen in a mirror-symmetrical arrangement of the topography of the 20°C isothermal surface (Figure 2c) which represents the lower boundary of the seasonal thermocline. Within the limits of the positioning accuracy one may conclude that the axis of the eddy was almost vertical during the first XBT survey.

The second XBT survey, which was carried out almost one month later (20-23 September 1978), established a general shift of the eddy's center amounting to 100 n.m. towards west-south-west (to 245°) (Figure 2). The eddy preserved its circular form and characteristic dimensions. The maximum depression of the 15°C isotherm at this time was 696 m. Average speed with which the eddy shifted was 7.4 cm/sec or 3.4 n.m. per day over 29 days. This was clearly higher than the average eddy translation speed of 4.9±1.5 cm/sec as quoted by Taft/Baranov et al. (POLYMODE News No. 47).

According to Taft/Baranov, the ratio D/σ_{hist} of a maximum observed departure D of the 15°C isotherm from its long-term ("historical") average depth $\bar{H}_{hist}$, and of its "historical" rms departure, σ_{hist}, may be used to judge the degree of eddy activity in the Sargasso Sea. This ratio for the first XBT survey was equal to 2.05, and for the second XBT survey, 2.43, i.e., it too was substantially higher than the average one. This fact, together with the high shifting speed and the regular circular form of the eddy, may well indicate that the observed anticyclonic eddy was in its active phase.

The thermohaline structure of the anticyclone may be seen on the section given in Figure 3. These sections are based on standard hydrographic data obtained at stations occupied every 10 n.m. with sampling levels of 0, 10, 20, 30, 50, 75, 100, 125, 150, 200, 250, 300, 400, 500, 600, 700, 800, 900, 1000, 1100, 1200, 1300, 1400, 1500, 1750, and 2000 m.

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AN ANTICYCLONIC EDDY IN THE SARGASSO SEA (continued)

The maximum depth of isotherms and isohalines below 150-200 m at these sections (Figure 3) corresponds to the center of the eddy. The characteristic bend of these isolines in the central part of the eddy is observed through all the main thermocline and even deeper. In the seasonal thermocline, isotherms and isohalines have the opposite curvature as a result of which a peculiar lens of water is formed within the anticyclonic eddy with temperatures between 18° and 19°C and salinity of 36.5-36.6‰. Its thickness amounts to almost 400 m. This peculiar feature of an anticyclonic thermohaline structure in the presence of a well-developed seasonal thermocline has already been noted (see Leetmaa, POLYMODE News No. 10; Fedorov, Ginsburg, and Zatsepin, 1978). No doubt, this feature is related to the current velocity distribution in such eddies, which problem has not yet been sufficiently studied both theoretically and experimentally.

Not having any current measurements in the observed eddy at our disposal, we estimated current velocities through geostrophic computations based on our hydrographic data across the eddy. The observed density field was smoothed prior to the computations along each section with the aid of the expression $\bar{\rho} = \frac{\rho_{i-1} + 2\rho_i + \rho_{i+1}}{4}$, where i is the station number. Geostrophic velocities have been calculated from the 2000 m level.

According to our geostrophic computations (Figure 4), orbital velocities of water movement reach 25-30 cm/sec, maximum velocities in most cases being observed well below the water surface within the 100-300 m layer. This fact clearly relates to the change of curvature sign of isotherms and isohalines (therefore of isopycnals also) in the surface layer. This change of sign results in a greater horizontal pressure gradient within the lens of 18° water than in the surface layer.

Velocities in excess of 5 cm/sec fill the whole 100-1200 m thickness of the upper ocean which apparently corresponds to the limit of penetration of the observed eddy.

A comparison of velocity distributions in Figures 4a and 4b indicates that, beyond some relatively unimportant details, the general structure of geostrophic motion within the eddy preserved its characteristic features (e.g., the strongest horizontal velocity shear close to the center) over the one-month interval between our repeated observations. The second velocity section, however, showed a tendency of the eddy motion to become somewhat slower and shallower.

We note in conclusion that the observed eddy moved through the area of the Local Dynamics Experiment (LDE) of POLYMODE carried out by our US colleagues. Therefore, the information provided above on the results of observations of the eddy may prove to be useful in LDE data interpretation.

Reference

Fedorov, K. N., A. I. Ginsburg and A. G. Zatsepin (1978) On thermohaline disturbances related to eddies in the Sargasso Sea. *In*: Studies of Variability of Physical Processes in the Ocean, K.N. Fedorov, Editor. Moscow, Inst. Oceanol. Acad. Sci.

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OCEAN CIRCULATION AND EDDY FIELDS FROM SATELLITE ALTIMETRY

by Carl Wunsch and Mike Gaposchkin

A simple and clear lesson of POLYMODE is that mesoscale eddies are a global phenomenon having at least some characteristics which vary greatly with geography. Understanding a global phenomenon on very long time scales will obviously be an enormous undertaking by conventional means. One of the more interesting technical developments which has taken place during the past decade, and which may have a major impact on the post-POLYMODE study of eddies, is in the area of satellite altimetry.

There have been two satellites launched with useable altimeters -- GEOS-3 and SEASAT-1 (e.g., Dunne, 1978). The intrinsic accuracy of the more recent SEASAT-1 altimeter appears to have been about 4 cm, i.e., relative to the spacecraft one can resolve changes in a nominal sea surface elevation of this amount with a spatial resolution of about 10 km. If the height of the satellite and the gravitational equipotential surface (the geoid) of the earth were known with equal or better accuracy, one could determine the absolute elevation of the sea surface relative to the geoid with 4 cm accuracy. The horizontal spatial derivative of the sea surface elevation would then clearly yield the absolute geostrophic velocity at the sea surface. Combined with hydrography, the old dream of oceanographers of determining the absolute field of flow in the ocean would be achieved.

There are a large number of difficulties. To be specific, we will focus on SEASAT-1. Figure 5 is a chart denoting a few sub-satellite tracks in the western North Atlantic in late 1978. The satellite was at an altitude of about 800 km. Figure 6 displays, along one of the tracks, the measured elevation of the sea surface. This was obtained in the following way. Let a be the altimeter-determined distance from spacecraft to sea surface; s is the distance of the spacecraft above a reference surface (the reference ellipsoid) obtained by laser tracking. Then Figure 6 displays $\zeta=a-s$, and thus represents the measured shape of the sea surface relative to the reference ellipsoid. One sees vertical fluctuations of some tens of meters. Most of this is due to fluctuations in the gravitational equipotential surfaces (in particular, at the southern end, one sees the Puerto Rico trench system, and near the northern end, the Bermuda high). On a global basis, the vertical changes in the geoid are of O(100 m). Because we anticipate deviations of the sea surface from the geoid due to geostrophic ocean currents of O(1 m), Figure 6 is an excellent estimate of the geoid. There are several caveats. Errors exist in the estimated spacecraft position but tend to be on very long spatial scales -- comparable to the diameter of the orbit. The mean value in Figure 6 cannot be trusted and there may be a slight tilt error. The tide introduces a time dependent fluctuation in the geoid; in this region it is probably O(.5 m), but is also on long scales. There are other errors too; the altimeter reflects from a random, rough, moving surface and there is a negative bias due to finite wave height. Atmospheric pressure loading will also introduce fluctuations into Figure 6, as will rain squalls and other phenomena.

One can correct Figure 6 for many of these problems, but in order to proceed to obtain an estimate of geostrophic surface velocity one must subtract from Figure 6 an independent estimate of the geoid. The best geoid available to us is one computed from shipboard gravity data by Marsh and Chang (1977) on a 5'x5' horizontal resolution. Unfortunately, it exists only over a limited area of the western North Atlantic.

Figure 7a displays the difference between Figure 6 and the Marsh/Chang geoid over the segment where the latter exists. A bulge remains at Bermuda where the geoid evidently does not fully resolve the spatial variability there (we have interpolated across this region in what follows). Figures 7b and 7c display the computation of surface geostrophic velocity from Figure 7a averaged respectively over 25 km and 100 km. The spectrum of surface velocity is displayed in Figure 8. (About 20% of the energy in Figure 8 is coherent with wave height, but we have not made a wave height correction to Figure 7.)

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In Figure 7, the Gulf Stream can be seen quite clearly, and in the velocity field, a westward flow on both flanks (most clearly in 7c) can be seen, as well as a series of fluctuations that one can infer to be eddies. The spectrum of surface velocity is remarkably white out to wave numbers of about 75 km (the rising power at shorter wavelengths is an artifact of the differentiation of a field whose spectrum is white because the system least-count error has been reached). Should one believe these results?

Robert Cheney (NASA Goddard Space Flight Center, private communication) has shown some remarkably good correlations between figures analogous to Figure 7a and independent measurements of the Gulf Stream and rings. In Figure 7, the position and speed of the Gulf Stream as well as the amplitudes of the westward flows are just about what one would anticipate. Even the eddy scales have a plausible look. On the other hand, if the geoid we subtracted were in error, there would be artificial structures in Figure 7 which would have no counterpart in any real ocean flows.

The problem of the geoid can be evaded (as many people have noted) by confining attention to the time dependent field only. In Figure 9, we display the counterpart of Figure 7, but from the same track 17 days later (the tracks are actually displaced by about 30 km, but this seems reasonably small compared to known features of the geoid over the abyssal plain and of ocean flow -- other tracks more nearly coincident give the same kinds of changes we perceive from Figures 7 to 9). There are many similarities and differences between the two track results. The changes that occur overall suggest that most of the features on the 100 to 500 km scale are in the water.

Because of a hardware failure, SEASAT-1 only lasted for 3 months; clearly with such a satellite in orbit for 3 to 5 years, one could easily determine the global distribution and frequency/wave number spectrum of the mesoscale eddy field insofar as it displays a surface geostrophic signal. This problem is now being pursued by many investigators using the limited SEASAT-1 data set and the more extensive (if less accurate) three year GEOS-3 data sets. A complete sorting out of the errors and the obtaining of quantitative estimates will take some time, but there appear to be no major obstacles. (Many of the details ignored here are discussed by Wunsch and Gaposchkin, 1979.)

In many ways, the more interesting and challenging idea is to use altimetric satellites to obtain the time mean (or instantaneous large scale) general circulation of the ocean. With a known geoid, an investigator working at sea doing classical hydrographic work, could, with an altimetric satellite, obtain the total geostrophic velocity field (possibly in real time) without using moorings, floats, or any other of the present encumbrances. In practice, the geoid itself will always have errors in it; one of the current difficulties is that the existing geoids are produced without error maps (in practice one really wants maps and error maps of the "deflection of the vertical" rather than of the geoid itself). A formulation for dealing with the errors in the estimated geoid, hydrography, and altimetry is described by Wunsch and Gaposchkin (1979). Essentially it is an extension of the geophysical inverse method described by Wunsch (1978), in which one simultaneously deduces the absolute velocity field, and corrections to the geoid and satellite orbit. The procedure depends upon careful specification of the errors in the fields, which is a subject we are actively pursuing.

At the present time we do not have significant hydrography data contemporaneous with the two extant altimetric satellites. Thus, an interesting question is whether the existing altimetry can be used to infer anything about the general circulation. There is a simple method. At least in the region where the Marsh/Chang geoid exists we can deduce the statistics of the surface geostrophic field (as in Figure 8). Prior hydrography, e.g., that displayed in Fuglister (1960), will not coincide with the particular surface velocities obtained in any particular sub-satellite track made 20 years later. On the other hand, it is not unreasonable to expect that it will remain statistically the same. In extending the circulation scheme of Wunsch (1978) to the entire Atlantic using inverse methods, we can, at least in the western North Atlantic, require that the resulting surface velocities have the same

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OCEAN CIRCULATION AND EDDY FIELDS FROM SATELLITE ALTIMETRY (continued)

statistics as those the satellites measured; this is particularly easy to do with the inverse techniques and is the basis of our current procedure.

This brief description of a complex system may give POLYMODE scientists some idea of the potentially revolutionary impact of satellite altimetry. Intense discussions are now taking place concerning possible future altimetric spacecraft and the requirements for independent improvements in the gravity field. If one combines altimetric measurements and improved gravity fields with acoustic tomography systems like the one described by Munk and Wunsch (1979), one can envision in the next 5 to 10 years a complete system for mapping entire ocean basins. This would yield the complete three dimensional density and velocity fields with mesoscale resolution in real time.

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*POLYMODE is derived from the names of the USSR POLYGON experiments and the Mid-Ocean Dynamics Experiment (MODE).

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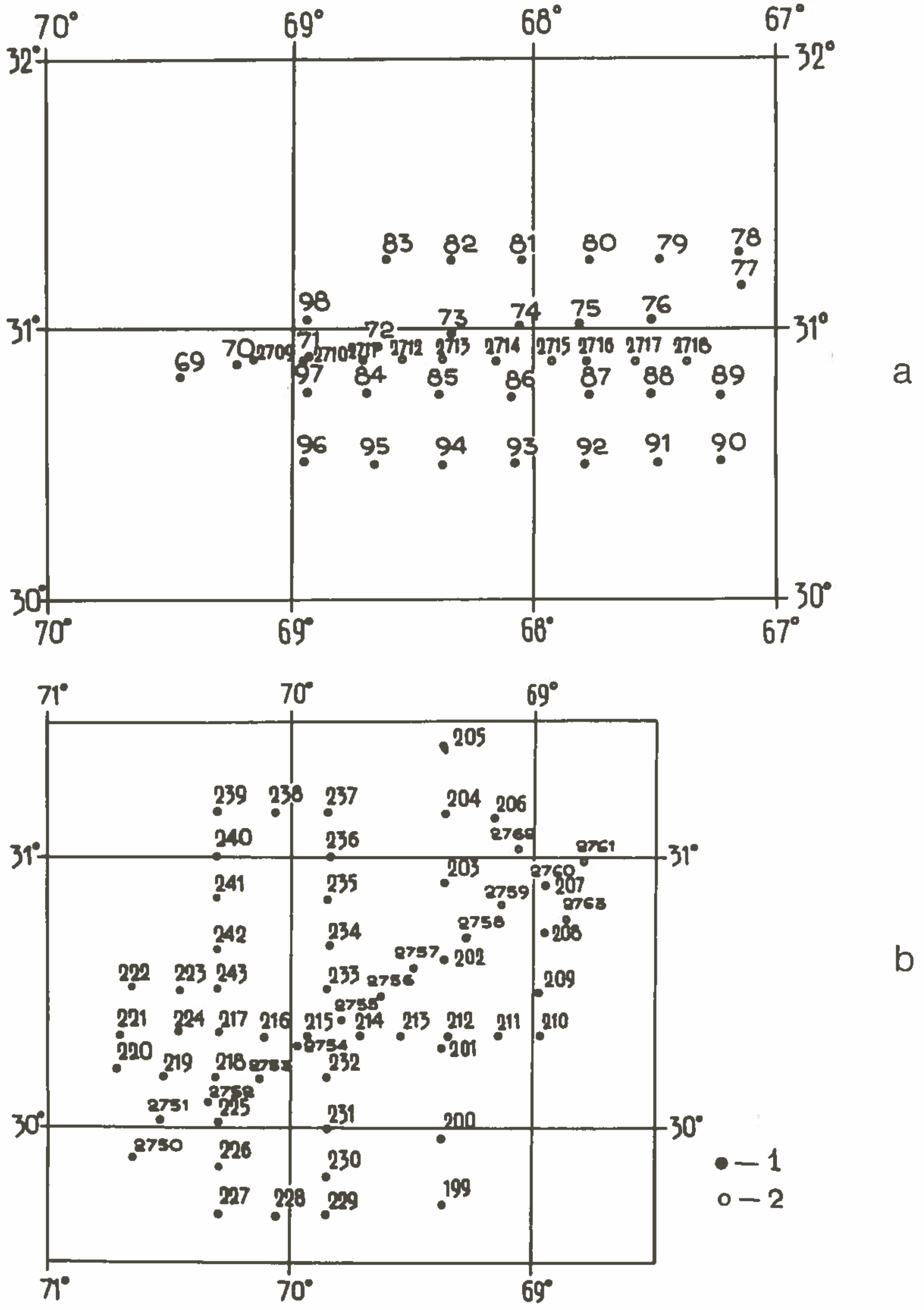
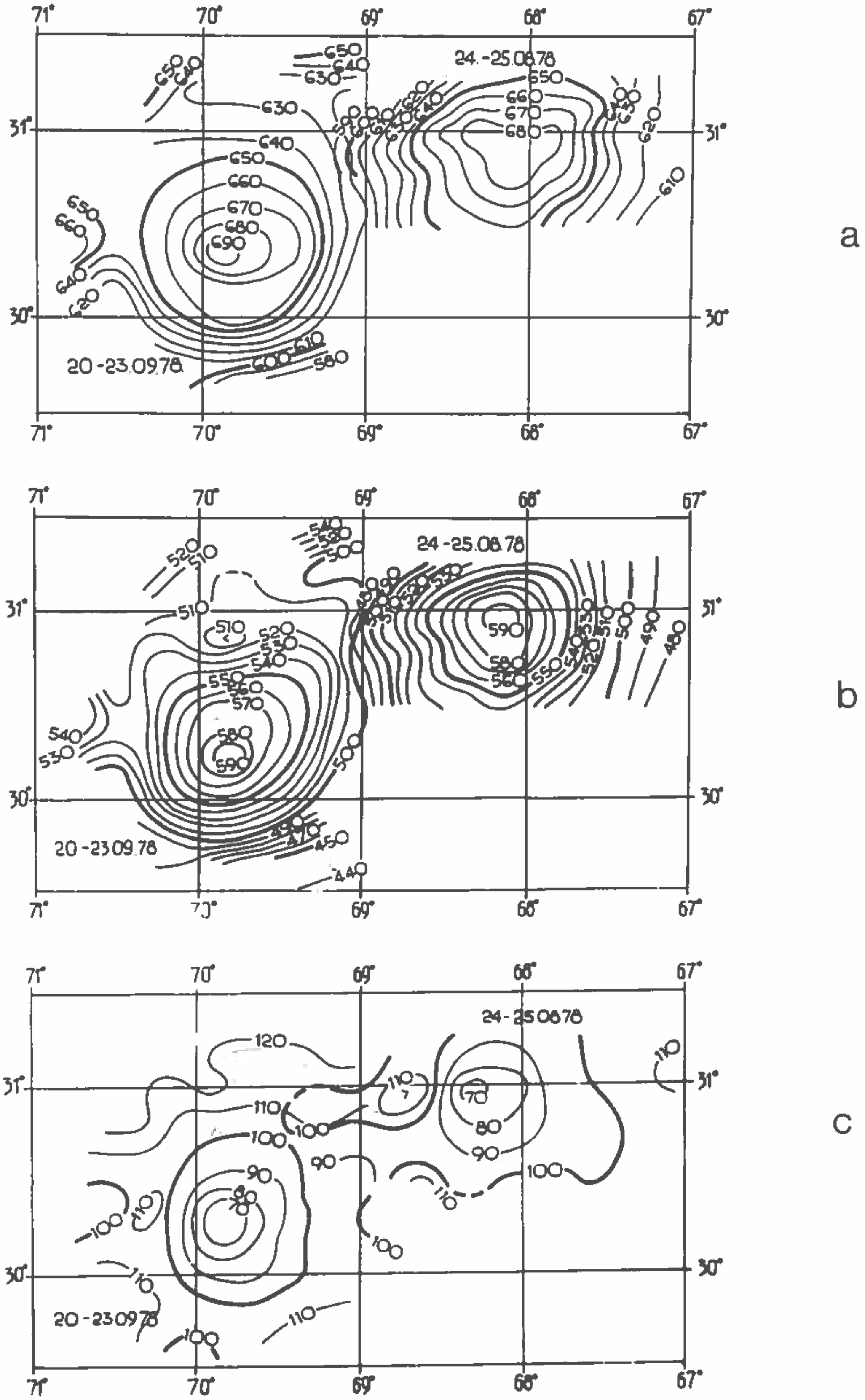


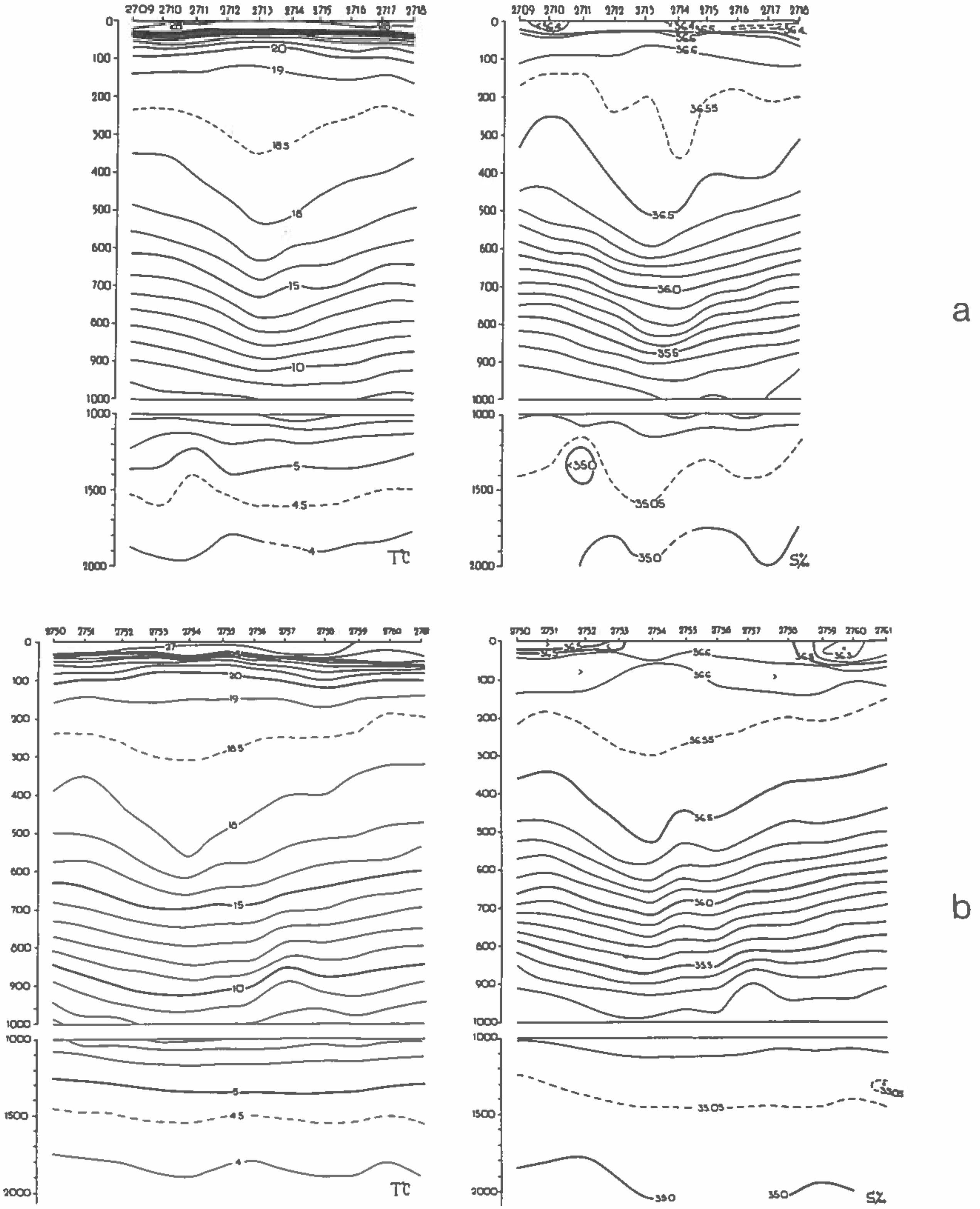
Positions of XBT soundings (●) and of hydrographic casts (o) taken during the survey of the anticyclonic eddy. a) XBT survey from 24-25 August 1978, and hydrographic section from 25-27 August 1978; b) XBT survey from 20-23 September 1978, and hydrographic section from 23-25 September 1978.

Figure 1 (Bubnov and Fedorov)



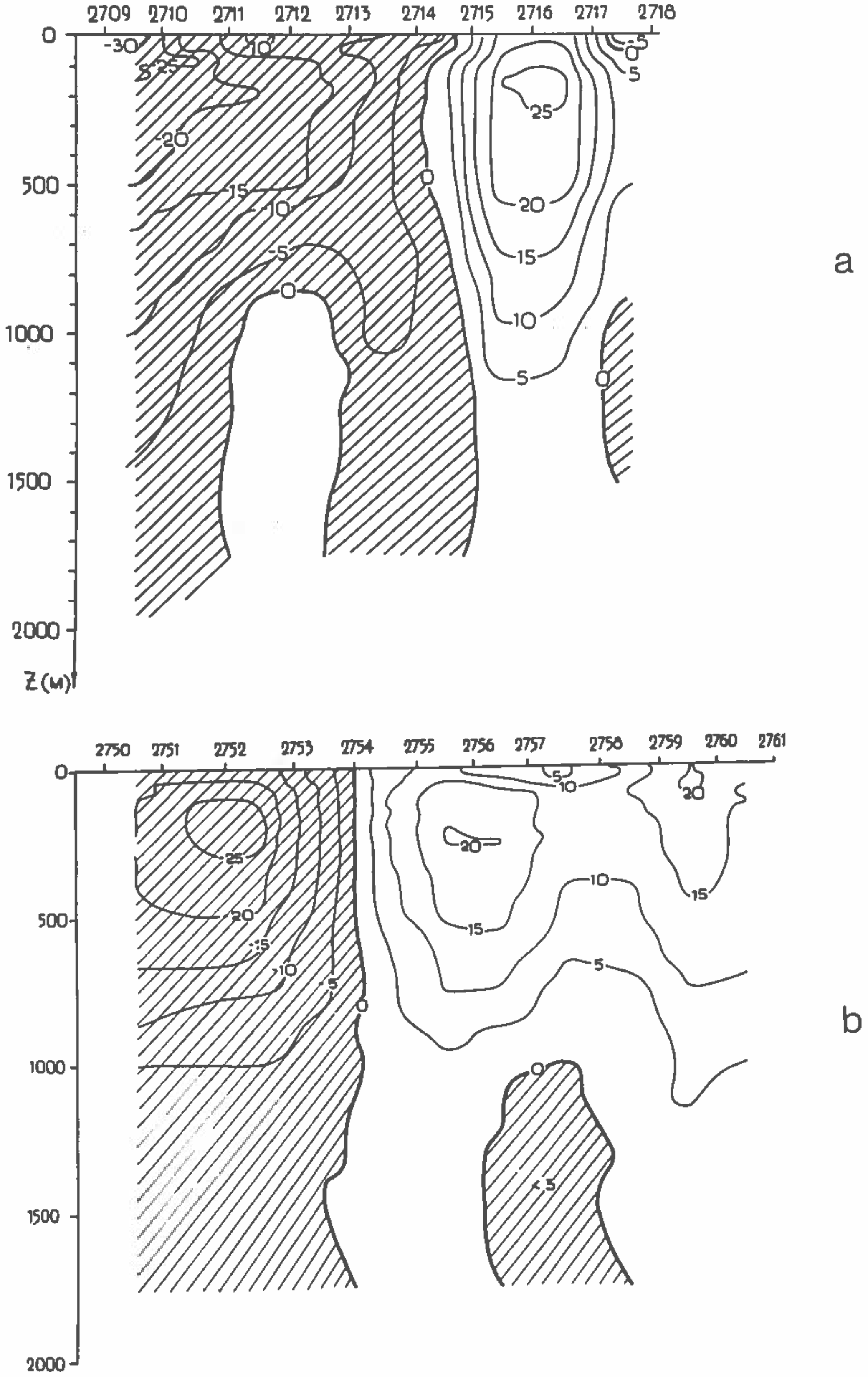
Topography (in m) of the isothermal surfaces (a) 15°C, (b) 17°C, and (c) 20°C in the area of the anticyclonic eddy as based on the XBT surveys. Note that the results of two cruises are plotted on each figure.

Figure 2 (Bubnov and Fedorov)



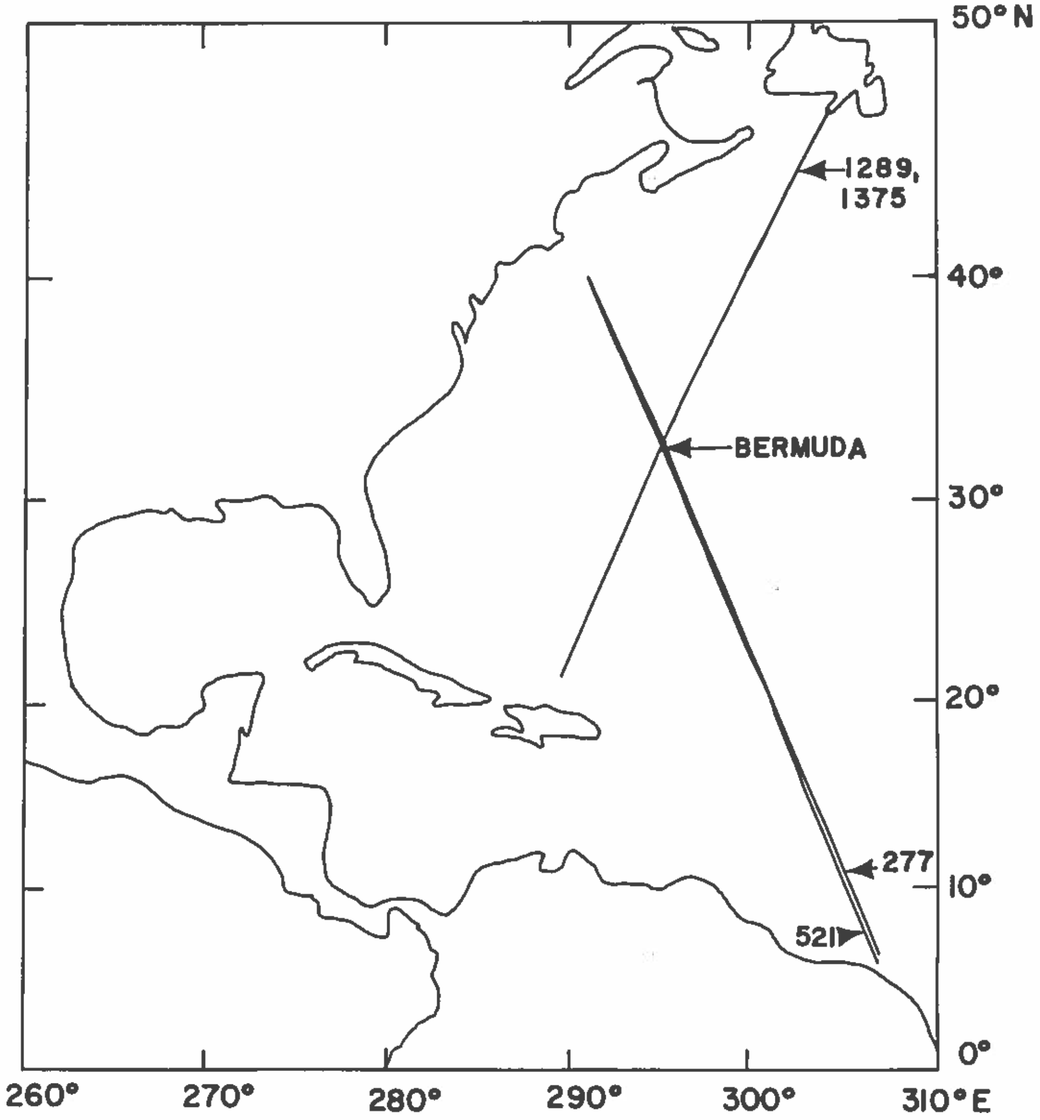
Temperature (°C) and salinity (°/₀₀) sections across the anticyclonic eddy: a) 25-27 August 1978; b) 23-25 September 1978.

Figure 3 (Bubnov and Fedorov)



Geostrophic velocity (cm/sec) sections across the anticyclonic eddy: a) 25-27 August 1978; b) 23-25 September 1978.

Figure 4 (Bubnov and Fedorov)



Sketch map showing 4 sub-orbit paths in the western North Atlantic in late 1978. The satellite was at an altitude of 800 km. Numbers are revolution identifiers. Revolutions 277 and 521 are displayed in Figures 6 and 7, and nearly pass overhead of Bermuda.

Figure 5 (Wunsch and Gaposchkin)

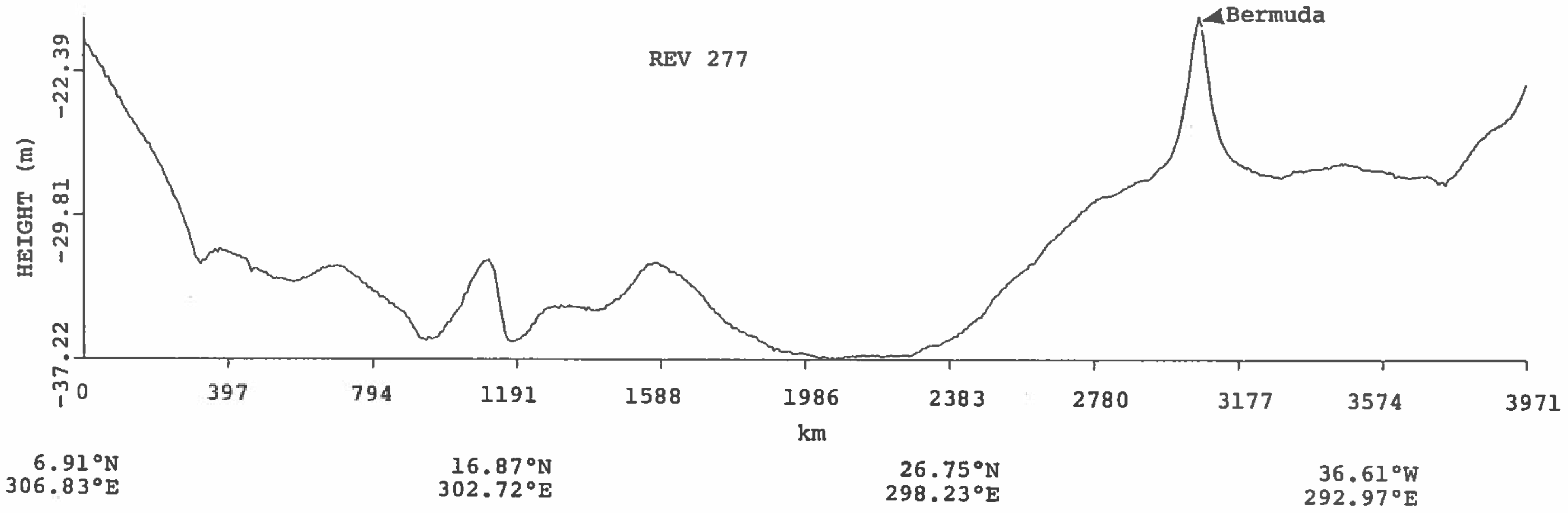
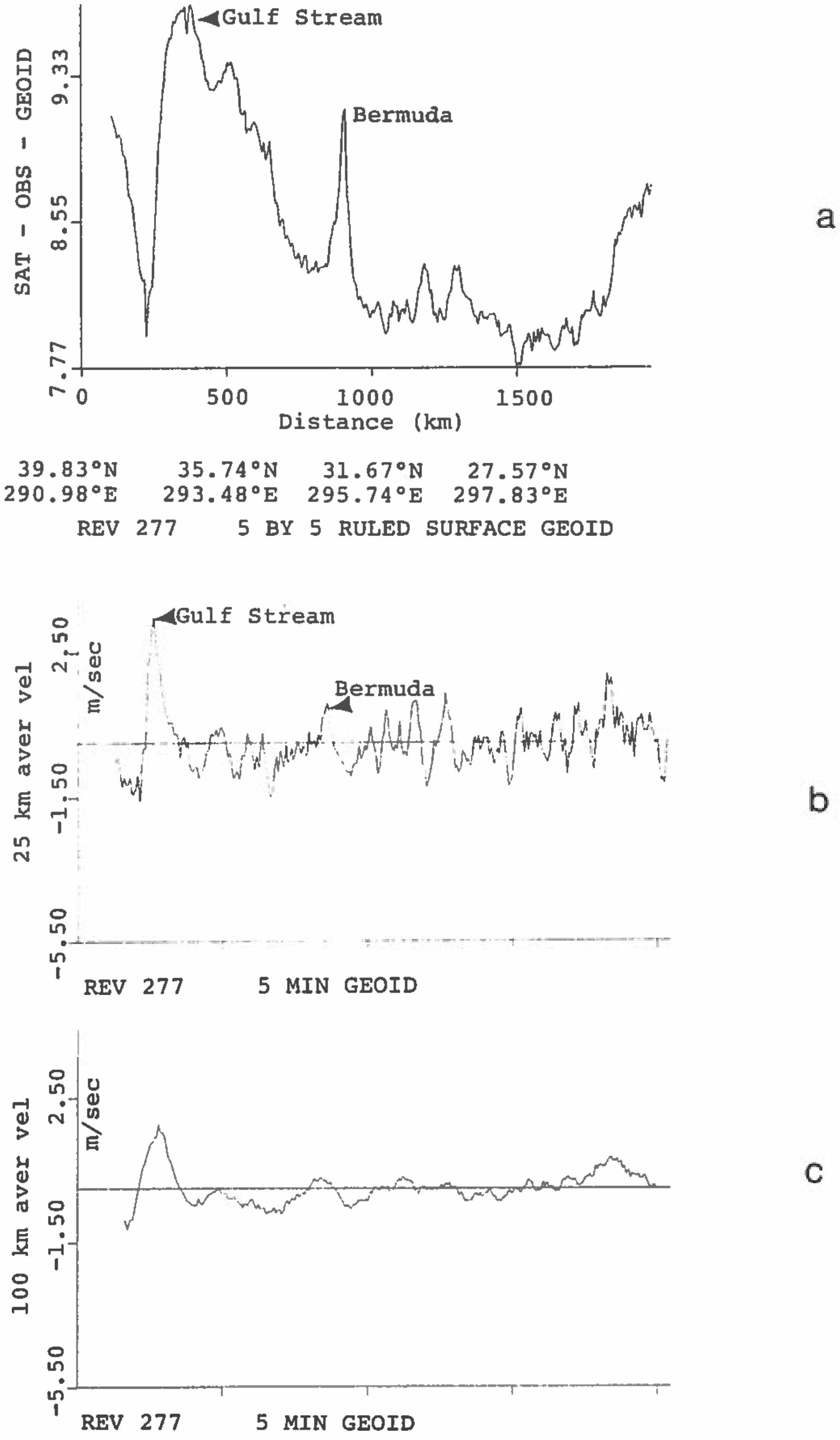


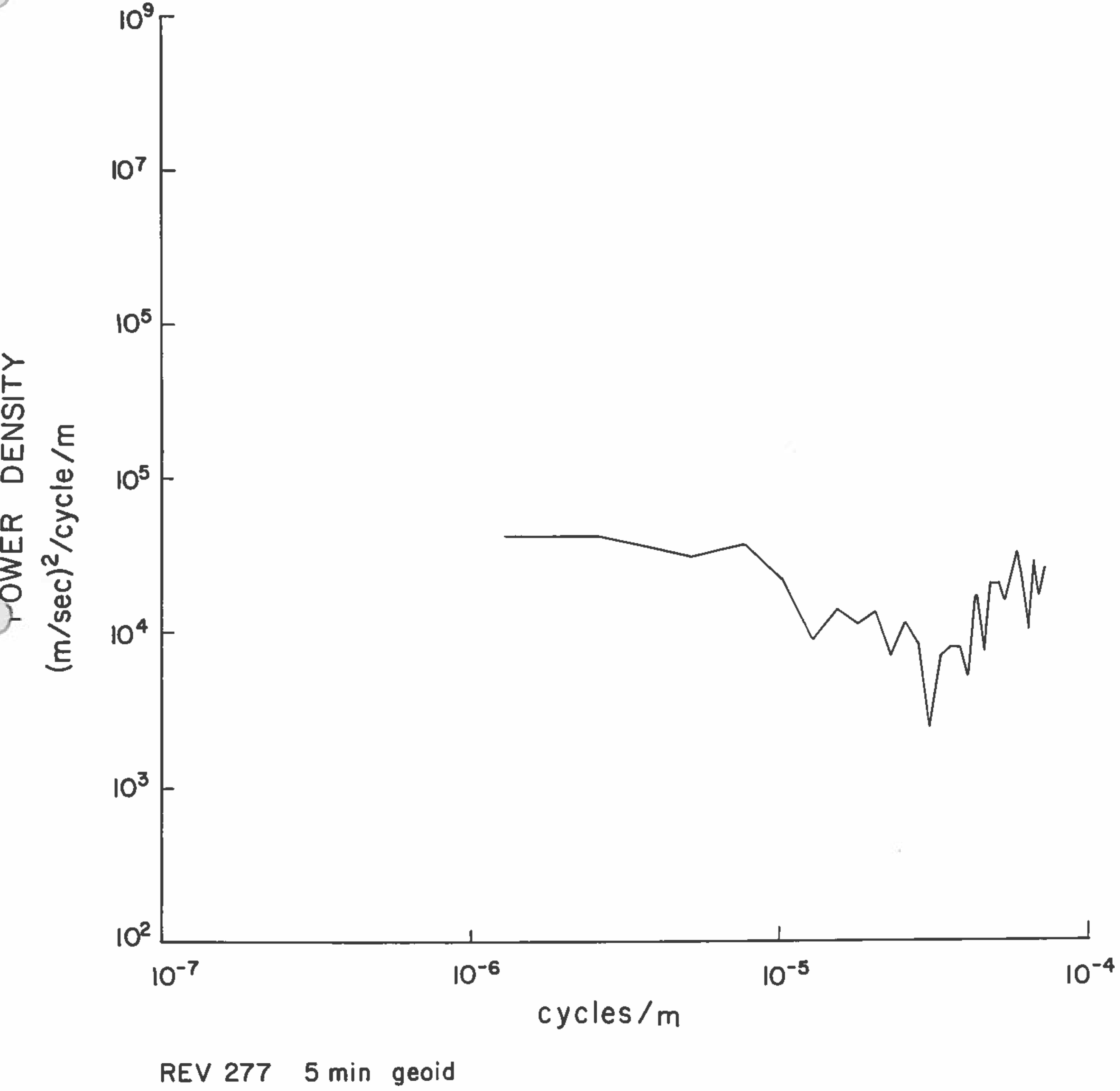
Illustration of the measured height of sea surface elevation relative to reference ellipsoid on revolution 277.

Figure 6 (Wunsch and Gaposchkin)



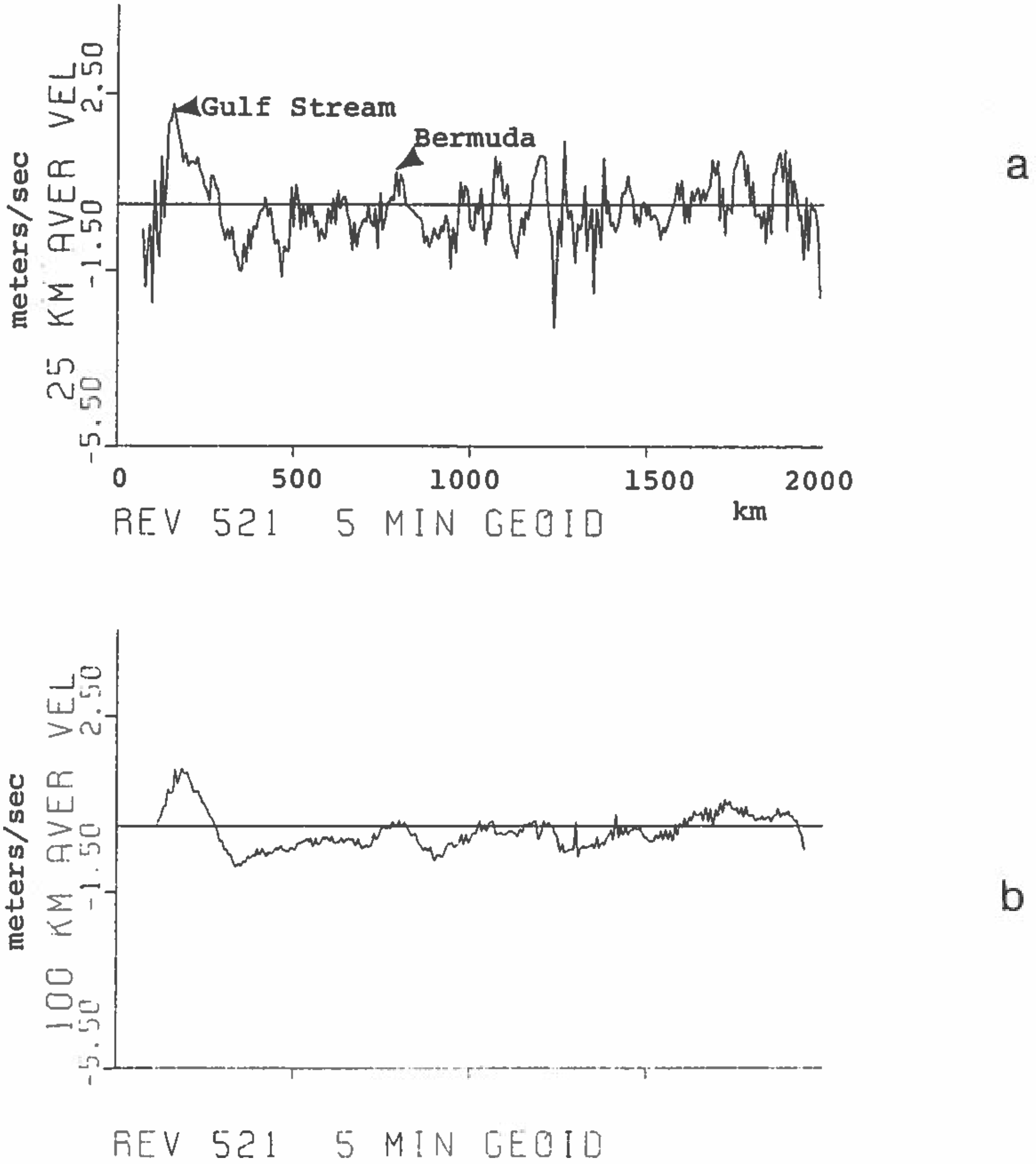
(a) The difference between Figure 6 and the Marsh/Chang geoid. Notice that north is now on the left. Surface geostrophic velocity computed from 7a after: (b) 25 km running average; (c) 100 km running average.

Figure 7 (Wunsch and Gaposchkin)



The power density spectrum of geostrophic velocity computed from Figure 7a. No averaging was done.

Figure 8 (Wunsch and Gaposchkin)



The counterpart of Figures 7b and 7c, except from revolution 521, from the same track 17 days later than revolution 277.

Figure 9 (Wunsch and Gaposchkin)