GEOPOTENTIAL ANOMALY FIELDS
AND GEOSTROPHIC VELOCITY PROFILES
IN THE WESTERN ICELAND SEA.

by

Jacob W. Nielsen
Danish Technical University
ISVA, Bygn. 115
2800 Lyngby.

and

Erik Buch
Royal Danish Administration of Navigation and Hydrography
Overgaden o. Vandet 62 B
1023 København K.

Abstract.

As part of the international Greenland Sea Project (GSP) joint Danish-Icelandic hydrographic surveys have been performed each year in September since 1987 in the Western Iceland Sea.

Based on the calibrated and error-checked CTD data from 1987 and 1988 the horizontal distribution of the geopotential anomaly and vertical sections of the geostrophic velocity have been calculated.

Maximum long-shore geostrophic velocities observed were 0.20 m/sec at 71°N (the Greenland-Jan Mayen section) increasing to 0.40 m/sec in the Denmark Strait.
1. Introduction.

Since 1987 joint Danish-Icelandic hydrographic surveys have been performed in September in the Western Iceland Sea as part of the international Greenland Sea Project (GSP). The net of sections and stations are shown in Fig.1. On each station CTD casts were performed, supplemented with ordinary Nansen casts for temperature and salinity calibration as well as for analysis of oxygen and nutrient content.

In 1987 a Neil Brown Mark III CTD was used while on the following cruises a SEABIRD CTD was operated.

Scaling, error checking, calibration and vertical smoothing of data from 1987 and 1988 have been performed, see Kristmansson, Malmberg and Briem (1989, 1990). Based on these data the horizontal and vertical distribution of the specific volume anomaly, maps of the geopotential anomaly and profiles of the geostrophic velocity have been calculated, see Nielsen (1989) and Nielsen and Buch (1990).

2. Dynamical topography.

The geopotential anomaly $\phi$ of the layer from pressure $p$ down to reference pressure $p_{ref}$ is equal to the integral of the specific volume anomaly $\delta$ (the numerical method of integration is described in Nielsen (1989)).

$$\phi = \left[ \int_{p_{ref}}^{p} \delta * dp' \right] (J/kg)$$

where

$p[deci\text{bar}] = z[\text{metre}]$
Since isolines of $\phi$ are streamlines for the geostrophic velocity at the topography level relative to the reference level, mapping of $\phi$ gives the dynamical topography, where the geostrophic current speed $V$ is inversely proportional to isoline spacing $D$:

$$V \text{ [m/s]} = \text{const} \times D^{-1} \text{ [cm]} \quad (1)$$

The constant depends on the isoline difference, the map projection ratio and the local value of the Coriolis parameter.

In Figs. 2-3 maps of the dynamical surface topography relative to 200 m in 1987 and 1988 are shown and the geopotential anomaly of the 200 m layer relative to 500 m is shown in Figs. 4-5.

The two surface topography maps infer a similar current pattern in 1987 and in 1988, with longshore flow from north to south, trapped in a coastal region of about 150 km width. The most remarkable difference is a much wider flow at the northernmost section in 1987.

From Figs. 4-5, a considerable part of the longshore current is present at the 200 m level, although surface current speeds are about 3-5 times larger. Also at this level the current was less concentrated close to the coast in 1987 compared to 1988 in the northern part of the research area.

3. The dynamic method.

The dynamic method has been widely used to infer the velocity of large-scale, slowly varying ocean currents from density field observations. The relative velocity is computed using the geostrophic force balance (Coriolis force equals horizontal pressure gradient force), the Boussinesq approximation, and the assumption of static pressure distribution in the vertical. A detailed description can be found in e.g. Pond & Pickard (1982).

The equation for the geostrophic velocity profile normal to a
line between stations A and B reads:

\[ v(z) = v(z_{\text{ref}}) + \frac{(\alpha \Phi_A - \alpha \Phi_B)}{f \times L} \]

where \( f \) is the Coriolis parameter (\( \approx 1.36 \times 10^{-4} \text{ sec}^{-1} \)), \( L \) is the station separation (obtained from navigation), and \( z_{\text{ref}} \) is the reference level relative to which the velocity is computed.

4. Velocity profiles and sections.

Vertical profiles of geostrophic velocity are calculated for each of the neighbouring pairs of stations along the transects, using the bottom of the shallowest station in the pair as reference level. Profiles corresponding to any other reference level are obtained simply by subtracting the velocity on the level in question from the entire profile. It is still not clear at which level one should assume zero (or known) velocity, and there is evidence (see Aagaard & Coachman 1968a,b) that there exists a barotropic internal circulation in the Greenland Sea, making the assumption of zero velocity invalid at any level. It therefore appears best only to discuss relative velocities.

The vertical distribution of geostrophic velocity at the section between Greenland and Jan Mayen and the Denmark Strait section in 1987 and 1988 are given in Figs. 6 and 7.

At the Greenland – Jan Mayen section southward motion is dominating both years with maximum velocities within 100 km from Greenland. In 1987 maximum velocities of 0.12 -0.14 m/s were observed at a distance of about 50 - 75 km from shore, while in 1988 the velocity maximum was observed closer to Greenland and attaining almost twice the velocity i.e. 0.26 m/s. Also the vertical extension of high velocity region was greater in 1988. Velocities above 0.10 m/s were only observed in the upper 10 meters in 1987 compared to 80 m in 1988. It can therefore be concluded that the southward transport of the East Greenland
Current over the Greenland shelf and continental slope was much higher in 1988 than in 1987, which also was the impression obtained from Figs.2 - 5.

East of the area of intense southward transport the current velocities are much smaller (less than 0.03 m/s) and more complex. Several examples of northward transport are observed. In 1987 these countercurrents were found at a distance of around 205, 280 and 395 km, while in 1988 they were found at 140, 285 and 400 km from the coast i.e. almost at the same locations both years. In 1987 maximum velocities were observed in the surface layer, this was not always the case in 1988.

In the Denmark Strait section 1988 did also reveal a greater southward transport compared to 1987, although not as outspoken as at the Greenland - Jan Mayen section. Maximum velocities both years around 0.40 m/s, and velocities above 0.10 m/s did reach depths of 100 m each year.

In this section it is remarkable to notice that close to Greenland an intense northward flow reveals itself, this flow was especially intense in 1987 occupying the innermost 70 km off Greenland and attaining velocities above 0.40 m/s. In 1988 the northward movement was narrower and less intense. Note, however, that the geostrophic current estimate is probably too high in this shallow region, as bottom friction is not accounted for.

The existence and the intensity of this countercurrent close to Greenland can be discussed. Northward transport of icebergs in this area has been observed in connection with environmental research carried out as part of the oil exploitation in Jameson Land. The presence of the above mentioned barotropic current component may overrule or at least weaken the northward geostrophic flow resulting in a net southward flow or a less intense northward transport.

At the two sections treated above eight current moorings were
established in September 1988, five at the Greenland - Jan Mayen section and three in the Denmark Strait section. The results from the first year of registration are being processed at the moment. When the results are available it will be possible to evaluate the importance of a barotropic current component in this area.

5. Summary.

Based on CTD data from two GSP cruises dynamic calculations have been performed, giving an impression of the distribution of the geostrophic flow, vertically and horizontally, in the research area in September 1987 and 1988.

Great variations in the intensity of the East Greenland Current in the Greenland - Jan Mayen section between the two years has been documented. The same variability, although less outspoken, was observed in the Denmark Strait.

In the Denmark Strait an intense countercurrent close to Greenland was registered both years, strongest in 1987. Its existence has earlier been visualized by ice drift, but its intensity may be weakened by the presence of a southward barotropic current component. The strength of a barotropic current component in this area can hopefully be evaluated by in situ current measurements carried out at several sites since September 1988 on the two sections treated in this paper.
References.


Fig. 1. Location of CTD-stations and sections carried out from R/V BJARNI SÆMUNDSON in 1987 and 1988.
Fig. 2. Geopotential anomaly of the 0 - 200 db layer, September 1987. Isoline units are J/kg. This is equivalent to a map of surface topography relative to 200 db, when isoline units are interpreted as decimeters.
Fig. 3. As Fig. 2, September 1988.
Fig. 4. Geopotential anomaly of the 200 - 500 db layer, September 1987. Isoline units are J/kg.
Fig. 5. As Fig. 4, September 1988.
Fig. 6. Vertical distribution of the geostrophic velocity component perpendicular to the Greenland - Jan Mayen section. Reference level is 200 m. Units are cm/s. Hatching indicates current towards the south.

Fig. 7. Vertical distribution of the geostrophic velocity component perpendicular to the Denmark Strati section. Reference level is 200 m. Units are cm/s. Hatching indicates current towards the south.
