The Iceland–Faroe Front: A Synergistic Study of Hydrography and Altimetry

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(Manuscript received 19 July 1993, in final form 14 February 1994)

ABSTRACT

Two triangular hydrographic surveys were conducted in the Iceland–Faroe region with sampling corresponding to the ERS-1 3-day commissioning-phase orbit. The CTD, XBT, and ADCP data show a very active frontal region changing on the order of days. Likewise, the altimetric and infrared data also show a highly variable region. The CTD/XBT data and the altimetric data are synergistically combined in an implementation of the Bernoulli inverse method to calculate velocity fields. A priori information for the inverse are the ERS-1 altimeter height differences. The method produces a velocity field much different from that using a strict reference depth of 800 m for a level of no motion. Velocities of 15–20 cm s⁻¹ flow eastward along the shelf at a depth of 400–500 m. Farther north, a strong eddy feature is visible along the western section. Along the western section, the height differences calculated from the altimetry are similar to the height differences produced by the inverse method. The results of the research show that: 1) the Bernoulli inverse method produces a valid geostrophic velocity field, 2) altimeter data constrains the inverse method, and 3) the initial ERS-1 altimeter data are of good quality if handled carefully.

1. Introduction

Satellites, now more than ever, contribute a major portion of the data used to explore the oceans. With the influx of these data, new methods are required to relate them to the physics of the ocean. Hallock (1985) suggests, in his conclusion on the variability of the Iceland–Faroe front (IFF), that satellite data “would provide needed perspective on the survey. In particular, the availability of satellite altimeter data would provide a direct measure of the mesoscale fields.” This paper discusses the results of a cruise to the IFF in September 1991 (Srokosz 1991). The paper describes the hydrographic data sampled during the cruise and examines how three datasets (hydrographic, altimetric, and infrared) can be used synergistically (synergy is defined as combined action) to study an ocean feature and the limitations to the analysis when hydrography is not available. The hydrographic and altimetric data were combined quantitatively, while the infrared data were used qualitatively to examine the movement and location of the IFF and the associated water masses. The cruise was designed to contribute to the evaluation of the ERS-1 satellite instruments and to attempt to understand the relationship between the surface and subsurface view of the ocean. This section briefly describes previous research in the area and introductory details of the cruise. The process of combining the CTD and XBT data and the description of the circulation pattern and its variability as shown from the combined CTD/XBT dataset is in section 2. Section 3 describes, qualitatively, the limited amount of infrared data, followed by the description of the coincident altimetric data. Section 4 explains the synergism of the various datasets and the results obtained when using the datasets together with an implementation of the Bernoulli inverse method (Killworth 1986). The last section gives some conclusions.

Two CTD/XBT surveys, using the RRS Charles Darwin (cruise number CD62A) in September 1991, were conducted in the region bounded by 62°N and 64°N and 6.5°W and 3.5°W, along the two ERS-1 ground tracks, shown in Figs. 1a and 1b, one an ascending track and the other, a descending track. The oceanography of this region (the IFF region) has been examined by many researchers. The most comprehensive coverage of the surrounding larger area, the Greenland/Iceland/Norwegian Seas region, is the subject of a review paper by Hopkins (1991). Other papers of interest, more closely connected to this particular region, are Read and Pollard (1992), Meincke (1978), Hallock (1985), and Smart (1984). The papers discuss the water mass characteristics of the area and the temporal and spatial variability of the IFF. Briefly, two frontal structures converge in the area near the cruise surveys. The IFF passes through the middle of
Fig. 1. Region of the CD62a Cruise: (a) survey 1 and (b) survey 2. The large dots mark the CTD stations and the small dots are the locations of the XBT drops. The ERS-1 ascending and descending tracks are marked by dotted lines. The sections are also labeled. Section A is along the top, section B is the eastern section, and section C is the western section.
the survey area and a broader structure, the Norwegian Current front, is to the east. The area sampled was, most likely, to the west of this second front. The bathymetry, Fig. 2, shows that the surveys are also west of the Shetland–Faroe Channel, the location of overflows of deep water into the Atlantic from the polar seas.

The IFF varies on the order of days (Hopkins 1991). The circulation, from the various papers above, is eastward along the IFF. To the south of the front, the current flows anticyclonically around the islands, west of the northward flow through the Shetland–Faroe Channel that brings the warmer, saltier Atlantic surface water up into the Norwegian Sea. The Norwegian Current splits at the Voring Plateau (66°N, 2°E) with part of the current circulating around the Norwegian Sea. In the west, the East Icelandic Current, flowing east, south of the Jan Mayen Plateau (68°N, 10°W) joins the flow along the IFF. The importance of understanding these circulation patterns will become apparent during the discussion of the water masses found in the two surveys.

**Sampling strategy**

As stated above, two surveys were conducted around an inverted triangle, two sides of which coincided with ascending and descending ERS-1 ground tracks. The third side, to the north, was a west–east leg connecting the tracks. The apex of the triangle (see Fig. 1), the shallowest point, is the location of the crossing point

![Fig. 2. Bathymetry of the surveyed region.](image-url)
TABLE 1. Water mass characteristics.

<table>
<thead>
<tr>
<th></th>
<th>Temperature (°C)</th>
<th>Salinity (psu)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NAW</td>
<td>&gt;8.5</td>
<td>&gt;35.3</td>
</tr>
<tr>
<td>MNAW</td>
<td>&gt;8.5</td>
<td>&gt;35.1</td>
</tr>
<tr>
<td>IASW</td>
<td>&gt;8.0</td>
<td>&lt;35</td>
</tr>
<tr>
<td>Intermediate</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NNAW</td>
<td>3</td>
<td>34.95</td>
</tr>
<tr>
<td>NI/Al</td>
<td>1–4</td>
<td>34.6–34.9</td>
</tr>
<tr>
<td>AtlW</td>
<td>1.5</td>
<td>34.88</td>
</tr>
<tr>
<td>Deep</td>
<td></td>
<td></td>
</tr>
<tr>
<td>NSDW</td>
<td>−1.1</td>
<td>34.91</td>
</tr>
</tbody>
</table>

drographic data collected on the cruise were of three types: conductivity, temperature, and depth (CTD); expendable thermobathygraph (XBT); and acoustic Doppler current profiler (ADCP) measurements.

a. CTD/XBT data

The T–S relationship for the combined two surveys is shown in Fig. 3. The characteristic \( \sigma_0 \) density values overlay the plot. Four distinct water masses are shown. The high salinity surface water is MNAW (\( \sigma_0 \) of 27.4 kg m\(^{-3} \)). No pure NAW with a salinity greater than 35.3 was sampled. This suggests that our surveys were completely to the west of the Norwegian front. The distinctive intermediate salinity maximum water is NNAW (Read and Pollard 1992) with a \( \sigma_0 \) value between 27.8 and 27.9 kg m\(^{-3} \). The intermediate salinity minimum denotes the NI/Al. The bottom water is NSDW. The T–S diagrams for each of the CTD stations, Figs. 4a and 4b, show how the water mass characteristics change around the triangular survey and from survey 1 to survey 2. It is clear from this set of figures that the characteristics of the region change on order of days not weeks (Hopkins 1991). Only the stations in the northwest corner of the triangle and the three southern stations on the shelf remained consistent between the two time periods.

The surface waters are complicated. There are, at least, two distinct water types making up the surface waters in these surveys, one, a relatively high temperature, high salinity water mass (\( T > 8.5^\circ C \) and \( S > 35.2 \) psu) and the other, a slightly fresher water mass (\( T > 8^\circ C \) and \( S < 35 \) psu). The first is MNAW at approximately 100 m on stations 4, 6, 9, and 12–15 from the first survey and stations 16–19, 25, 26, and 28 of the second survey. At some of these stations, the second distinct surface water mass, IASW, overlays the MNAW (e.g., at station 9). Stations 2, 3, and 5–8 from the first survey and stations 21, 22, 23, 24, and 29 of the second survey have only the IASW in its surface waters. Pure NAW with a salinity > 35.3 psu is not seen at any station.

The most distinct water mass is that which Read and Pollard (1992) named the NNAW. This has a characteristic signature of around 3°C and 34.95 psu. The water mass, with its intermediate level salinity maximum, is visible in stations 2, 3, 4, 5, 8, and 13–15, from the first survey and in stations 21, 22, 23, and 28 from the second survey. Water with the same characteristics also appears in the T–S plots (Fig. 4) for station 12 (first survey), 17, and 18 (second survey) as a point on the mixing line between the surface water, MNAW, and the NSDW, but not as a distinct layer.

Another intermediate water mass is prominent in the two surveys, that of NI/Al, also known as Icelandic Current Winter Water (ICWW). This water mass lies over NNAW. These surveys show that it has a signature of 3.5°C, 34.88 psu, a salinity minimum, and is visible

2. Hydrographic data

Various water masses are described in the above papers and not always with the same terminology. The water masses referred to in this region are North Atlantic Water (NAW), Modified North Atlantic Water (MNAW), Iceland Arctic Surface Water (IASW), North Icelandic/Arctic Intermediate Water (NI/Al), Norwegian North Atlantic Water (NNAW), Atlantic Intermediate Water (AtlW), and Norwegian Sea Deep Water (NSDW). Table 1 lists the water mass names and their characteristics as used in this paper.
at several stations—2, 3, 4, 5, and 15 from the first survey and 18–24 and 28 from the second survey.

When describing data from a cruise in June and July of 1977, Meincke (1978) noted that his NI/AI water had characteristics (≥3°C and ≤34.78 psu) that were definitely fresher than the historical definition of the water mass and concurred with the freshening trend shown by Dooley et al. (1984). However, the surveys presented here (e.g., stations 3 and 4) show that the characteristics are close to the historical definition. Similarly, Read and Pollard (1992) show the historical characteristics for the water mass (1990 data) in the same region, although they do not discuss it in their paper. And using 1980 data, Hallock (1985) also uses 34.88 psu to label his NI/AI water.

The analysis of the hydrography first required that the CTD and XBT data be combined into one dataset so that in situ sampling would be similar to ERS-I sampling. This meant that salinity values were computed for the XBT temperature data located between CTD stations. A bicubic spline fit was found for each CTD station's T–S relationship. Adjacent CTD station spline fits were then used for interpolating the XBT salinity values and the distances the XBT drop was from adjacent CTD stations. Of the 38 XBTs from the first survey, 28 have been used for the final analysis. Of the 19 XBT drops conducted during the second survey, 9 have been used in combination with the CTD data. Lagged correlation analysis (conducted to determine if the correct XBT drop rate had been used) of the five coincident CTD/XBT data shows no consistent pattern in the depth of the highest lagged correlation value between the CTD temperature-pressure data and the XBT depth data (Singer 1990). No conclusive statement can be made about the error in the drop rate on the XBT probes and no correction has been made to the manufacturer's specified drop rate. The XBT and CTD datasets have been combined to form a set of T–S values between 50 and 800 m. The XBT data were too noisy to attempt a salinity fit for the top 50 m.

The two surveys around the triangle have different spatial sampling due to the number of XBTs that were dropped between the CTD stations: three between CTD stations for the first survey and one between CTD stations for the second. Obviously, the first survey will resolve smaller spatial features. But are features between 10 and 20 km influencing the hydrographic structure extensively in this region? As a test, the first survey, with a 10-km sampling, was subsampled to re-
Fig. 4. Temperature-salinity diagrams for (a) survey 1 and (b) survey 2.
The number in the bottom right corner of each T–S diagram is the CTD station number.
duce the number of points in the first survey to approximately the number and spacing of the second survey (only one XBT between CTD stations). The magnitude of the rms difference was calculated between the points removed from the first survey to produce the subsampled survey and points at these locations interpolated from the subsampled survey. The rms differences were calculated for both temperature and salinity on the isopycnal surfaces described above. Temperature rms differences range from 0.47°C to 0.07°C, with the highest difference in the 27.46 kg m⁻³ layer near the ocean surface, whereas the deeper layers have rms differences below 0.2°C. The near-surface layer also is biased 0.23°C lower in the subsampled survey than in the full survey. The other layers are not biased in the temperature field. Salinity rms differences range from 0.08 to less than 0.01 PSU with all the differences less than 0.02 PSU except the top layer. Therefore, it is reasonable to compare the spatial variability of the two original surveys, even with the different sampling.

Density contours (Figs. 5 and 6) along each section show the spatial distribution of these water masses between 50 and 800 m. Figures 5a–c show the contours of density (σ₀) from the combined dataset of the first survey and Figs. 6a–c show the same for the second.

FIG. 5. Section density (σ₀) contours for survey 1: (a) section A, (b) section B, and (c) section C. The approximate shelf location in (b) and (c) is defined by the line in the southern region of the plot.

FIG. 6. As in Fig. 5 but for survey 2.

Section A is the section along the top of the triangle, section B is the eastern section, and section C is the western section as diagrammed in Fig. 1a. The CTD (C) and XBT (X) locations are identified along the top of each of the Figs. 5 and 6. The added XBT data extends the information about the structure in each section.

The first survey, section A (Fig. 5a), on the north side of the survey triangle, shows a fairly uniform density structure across the section, with a small increase in lighter water at 5.75°W and east of 5°W. The layer of NNAW (between 27.8 and 27.9 kg m⁻³) is about 150 m thick across the section. The short second survey (Fig. 6a) shows relatively little change in the structure.

Section B (Figs. 5b and 6b), the eastern section, shows the IFF location by the steepening of the isopycnals between 62.7°N and 63°N. To the north of 63°N, the slope remains fairly flat with only a small eddylike feature north of 63.3°N in Fig. 5b. The structure of the second survey is similar, but with no visible eddylike feature north of 63.3°N, due to the lack of data at the north end of this section. The NNAW (between 27.8 and 27.9 kg m⁻³) water mass is north of 63.7°N in the first survey, at a shallower depth of between 150 and 300 m. No data are available on the north end in the second survey due to the rough weather encountered on the cruise.
The IFF is clearly visible in section C. As in the eastern section (B), the isopycnals slope upward toward the north between 62.7°N and 63.2°N in Fig. 5c (first survey) and again at the northern end of the survey starting near 63.7°N. The horizontal change in the density of the water is an example of the meandering nature of the front or evidence of a cyclonic eddy. Similarly, in the second survey (Fig. 6c) section C shows upward sloping isopycnals beginning at 63.3°N, but the southern frontal structure with the steep slope in the first survey is not as evident in the second survey. The frontal slope is much flatter and does not extend to the surface as it does in the first survey. The water mass identified as NNAW (between 27.8 and 27.9 kg m$^{-3}$), in both surveys, increases in thickness northward (marked by hatched area) with the thickest section of about 150 m between 63.3°N and 63.7°N, centered at about 400 m in depth.

The NNAW 27.86 kg m$^{-3}$ isopycnal is not visible as a distinct layer in all the T–S plots. This distinct intermediate-layer salinity maximum can only be found north of the front. These surveys agree with the findings of Read and Pollard (1992) that, generally, the water mass is at deeper levels to the south than to the north ($\sim 450$ and $\sim 250$ m, respectively), sinking as it moves southward. Figure 6c shows this clearly: near 63.8°N it is at a depth of 240 m and at 63.3°N the water mass is at 350 m. The T–S diagrams of stations 17 and 18 (Fig. 4b) in the second survey clearly show the characteristic water on a mixing line between MNAW and the deep water. At these stations, the water mass is no longer a stable layer and has mixed with the layers above and below it. From the density plot of section C (Fig. 5c) both stations are south of the major frontal location, located north of 63.5°N. Thus, the data suggest that mixing of the intermediate layers are occurring only south (or the Atlantic side) of the front. The NNAW and NI/Al layers are distinct features corresponding to water that is north of the IFF. These layers are not observed south of the front because they have mixed with the overlying MNAW.

Figures 7 (survey 1) and 8 (survey 2) show plots of the geostrophic flow referenced to 800 m. Section 4 discusses these plots. The level of no motion is 800 m because this is the approximate depth to which most of the XBT drops were made.

b. ADCP data

ADCP data were collected on and off station. Only the on-station data are being used for this study due

FIG. 7. Geostrophic flow referenced from 800 m for survey 1: (a) section A, (b) section B, and (c) section C. The approximate shelf location in (b) and (c) is defined by the line in the southern region of the plot. Units are centimeters per second.

FIG. 8. As in Fig. 7 but for survey 2.
to the Global Position System (GPS) signal having been degraded by the U.S. Government, thus requiring the GPS data to be averaged. The ADCP data are used to evaluate the accuracy of the computed geostrophic velocity fields. The ADCP data were calibrated by the method of Pollard and Read (1989). Details of the preliminary processing of the data are in Crisp (1993, CD62A ADCP data report in preparation). For this study, the Flather tide [local tide model for the region (Flather 1981)] has been removed from the data. Using the model, the mean and standard deviation of the tidal signal at the CTD station points has been computed over a period of one year. The mean of the tidal velocity magnitude over the whole region is 3 cm s$^{-1}$. Only the two shallowest stations, 11 and 12, have a mean over 5 cm s$^{-1}$ (6 and 9 cm s$^{-1}$). With the magnitudes of the current speed reaching 60 cm s$^{-1}$ along the front, the tide has little effect except, perhaps, at the two shallow stations. The ADCP data, plotted in Figs. 9 (survey 1) and 10 (survey 2) were averaged over the time period on station. The plots show the contours of the velocity field normal to the sections A, B, and C. The data are plotted in this manner for the later comparison to the velocity fields calculated with an inverse method from the in situ data in section 4. The sign convention for the vectors normal to the sections are section A, north is positive and sections B and C, positive is the easterly direction (northeast for section C and southeast for section B).

**c. Frontal signature**

Both surveys show the front extending across the surveyed area. From the density sections shown in Figs. 5 and 6 (along each section), the front is much different in location at depth than at the surface. At depth, the frontal signature extends farther to the south than it does at the surface and is defined by the shelf edge as the density sections show (see also Smart 1984). The densely packed isopycnal layers intersect the shelf edge within 50 m of each other in the two surveys. The difference in the two surveys is the slope of the front, flattening from the first to the second survey along section C (the first survey has a frontal slope of about twice that of the second survey) and with the first survey having an eddy in the middle of the section. Section B shows the upward sloping isopycnals differing only slightly in steepness from survey 1 to survey 2. The stable intermediate layers, NIAI and NNAW, are north of the front.
3. Satellite data

a. Infrared data

One clear Advanced Very High Resolution Radar (AVHRR) image is available, taken on 5 September, a week before the first survey was sampled. Although ERS-1 carries an infrared instrument (the Along-Track Scanning Radiometer), either the region was cloudy or the satellite did not collect an image on the days coincident with the surveyed sections. Figure 11 shows the available AVHRR image with the cruise track superimposed on it. The IFF can be seen in the image with the western track crossing it (the front being defined by the sharp gradient between 9.5°C and 8.5°C).

b. Altimetric data

The data from the above research cruise were, as stated in the introduction, specifically sampled with the ERS-1 ground tracks in mind. This section describes the ERS-1 altimetric data and the associated processing of the altimeter data so they can be used in conjunction with the in situ dataset. ERS-1 was launched in July 1991 and the cruise was designed to collect in situ data for the evaluation of some of the capabilities of ERS-1: height, wind speed and direction, and wave height. The cruise coincided with the satellite’s 3-day commissioning phase cycle. That is, every 3 days the ground tracks of the satellite orbit were repeated. This orbit existed from approximately mid-August to mid-December 1991, giving four months of repeat cycle data. (Note: The orbit is not the same as the 3-day ice phase orbit, which followed until April of 1992, preceding the 35-day repeat orbit.) Because ERS-1 was in its initial commissioning phase, there were inevitable losses of data due to various adjustments being made to the satellite. Unfortunately one of the critical passes, with respect to this research, coinciding with section C in survey 1 was not collected. This loss is discussed in section 4 where the datasets are combined.

c. Processing

There were several ERS-1 altimeter datasets available at the time this study was done: 1) the European Space Agency (ESA) fast delivery data; 2) the offline reprocessed “geophysical data record” set available from ESA [the official, validated dataset or ocean product record (OPR) dataset]; and 3) an intermediate geophysical data record (IGDR) set assembled by the

Fig. 11. AVHRR image of region for 5 September 1991 showing sea surface temperature with overlaid ERS-1 track coincident with surveyed sections B and C. Units are in tenths of degree Celsius.
group at the National Oceanic and Atmospheric Administration (NOAA) (R. Cheney 1992, personal communication). Each dataset is a little different. The third dataset (IGDR) is the first dataset (ESA fast delivery) plus an orbit, tide, and a few NOAA-defined atmospheric corrections added in. This dataset does not exist for the time period of this cruise. The OPR dataset is the ESA ocean product, with the sea surface height values resulting from ESA reprocessing the altimeter radar signal on the ground and adding ESA computed tide values, a precise orbit, and atmospheric corrections. The OPR dataset, available through ESA, is used for this study. The noise level is between 4 and 5 cm, comparable with the computed noise level of the fast delivery data (R. Francis 1993, personal communication).

The altimeter height data were processed following the standard steps as described in Tokmakian and Challenor (1993). (Various corrections were made to the "raw" OPR height value and the removal of the orbit error.) The following corrections were applied to the altimeter height field from the OPR tapes: orbital, dry tropospheric, inverse barometer, solid earth tide, and ocean tide. The orbit and dry tropospheric correction were used as supplied on the OPR tape; the dry tropospheric correction is an ESA modeled correction. The inverse barometer correction was determined from the dry tropospheric correction in the standard manner. The solid earth tide was calculated from the model used for the Geosat GDR data. (There was an error in the ESA ERS-1 OPR solid earth calculation on the initial OPR tape release.) The last correction applied was the tidal correction. For this, as with the hydrographic data, the local Flather tidal model was used (Thomas and Woodworth 1990; Flather 1981). A tilt and bias method removed the residual orbit error. ESA (B. Greco 1992, personal communication) advised against using the wet tropospheric correction from the initial OPR tapes.

d. Variability/mesoscale features

It is straightforward to examine the variability in time of the sea surface heights produced from altimetric data. Figures 12a (section B) and 12b (section C) show the altimeter height anomalies (or residuals) along track after the removal of a 4-month mean signal and the application of a 42-km low-pass filter along track to remove the white noise (the value having been determined from a frequency spectrum of the data). The IFF can be located in the plots of alongtrack residual heights by finding the locations where the absolute value of the residual height (sea surface height — mean sea surface height) is relatively large and the gradients
of the heights are relatively large. The approximate location of the front, denoted by the arrows on Figs. 12a and 12b, is seen as it moves north and south between several of the repeat tracks. The height residuals are on the order of 10–15 cm at their greatest, well above the noise level of the ERS-1 altimeter.

The monthly alongtrack variability can also be examined as shown in Figs. 13a (eastern track) and 13b (western track). The monthly variabilities (all the track residual heights for one month averaged), referenced to the mean from the complete set of tracks for the four month period, are shown with their associated deviations. Both August and December appear to have much more variability than the other three months, but this is due to the smaller number of repeat tracks (only a half-month of data are available for each) contributing to the monthly mean. The variability along track for September of 1991 shows that it is very close to the mean of the four months of data, suggesting that the front is moving on a timescale less than a month. When the front is relatively stationary over a period of a month, the alongtrack monthly variability should show a strong frontal signal as it does in the month of October (at about 63°N in Fig. 13a), but in September there is no strong frontal signal.

4. Synergism and inverses

Described above are two independent and different datasets showing different views (both in time and space) of the same area of the ocean: a hydrographic view and a satellite view. Each set has its advantages and disadvantages, of course, but can more information be obtained when the two sets are used together: specifically, can the deep flow be determined more accurately if the surface height signatures from the altimeter are used in conjunction with the hydrographic data (giving an estimate of the barotropic flow)? The assumption of geostrophy is applied even though the region is a very active frontal region.

The geostrophic method using a uniform level of no motion (referred to as the "traditional" method later in the paper) for calculating currents, unfortunately, gives only a relative estimate of the velocity field. And altimetry, still, despite all the advances, does not yet provide us with an absolute measure of the sea surface height. The mean flow field cannot be separated from the geoid (at least not in the unclassified world), because both are unchanging with time. Because of the uniqueness of the datasets that are presented here, coincident hydrography and altimetry repeated over a short period of time, the joint use of the sets might

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**Fig. 13.** Monthly variability, 4-month mean removed.
provide the needed information to determine absolute velocities. The technique used to determine these velocities was the Bernoulli inverse method (Killworth 1986).

Three types of inverse methods are applied to oceanographic data: the box method (Wunsch 1978), the beta-spiral (Schott and Stommel 1978) method, and the Bernoulli method. This paper is not intended to be a comparison of inverse methods and, therefore, only the most appropriate method was applied to the datasets. Killworth and Bigg (1988) gives a very complete comparison of the three methods. The three methods were compared using a common dataset for several ocean regions at various resolutions. Four different types of ocean regimes were examined: 1) strong baroclinic signal, rapid flow, and medium eddy activity; 2) weak flow and low energy; 3) homogeneous; and 4) close to the equator. The oceanic region in this present paper fits most closely to the first category. Killworth and Bigg state that the Bernoulli inverse method was "best" for this type of region, especially at fine resolution. Their finest resolution was one-thousand of a degree, while our data were sampled at one-fifth (first survey) and one-half degree (second survey). For completeness, the beta spiral method works well for regions near the equator and away from regions of complicated dynamics, and the box method, designed for large-scale flow, gives the "best" results when the sampling grid is coarse.

a. Bernoulli inverse method and altimetry

The Bernoulli function is

\[ B = P + \rho g z, \tag{1} \]

where \( g \) is gravity, \( z \) depth, \( \rho \) density, \( P \) pressure (decibars); and \( B \) at station \( i \) is

\[ B_i = B_i + g \int_D z \rho_1 dz, \tag{2} \]

with \( B_i \) an unknown constant. The inverse method finds the set of \( B_i \) by assuming that density and potential vorticity \( (q) \) are conserved on a streamline and, thus, the Bernoulli function is conserved. If \( \rho_1(z_1) = \rho_1(z_3) \) and \( q_1(z_1) = q_1(z_3) \) (where \( 1 \) and \( 3 \) are station numbers), then

\[ B_1 + g \int_D z \rho_1 dz = B_3 + g \int_D z \rho_3 dz, \tag{3} \]

where \( D \) is the reference depth (surface in this implementation). A search through the data by station pairs finds all the locations where the density and potential vorticity match between stations. These equations are then used in a least squares fashion [Eq. (4)] to solve for all the unknown values of \( B_i \), from which geostrophic velocities can then be computed, with \( k \) representing the \( k \)th station pair match (e.g., station 1 and 3 may have one or more depths at which \( \rho \) and \( q \) match):

\[
\sum \left( [B_i(k) + g \int_D z(\rho_1(k)) dz] - [B_j(k) + g \int_D z(\rho_2(k)) dz] \right)^2 = \text{error}^2, \tag{4}
\]

summed over \( k, j = 1, 2, \cdots, M \), and \( M \) is the number of pairs.

A solution set for the \( B_i \) can be found for each of the two surveys separately—that is, a set \( B_1 \) for the surface pressures of survey 1 and a set \( B_2 \) for the surface pressures for survey 2.

This unique dataset provides a set of a priori information from the altimeter data. From repeat passes of the altimeter, a sea surface height difference can be calculated at the locations of the in situ measurements.

The altimeter measures a height, \( H \), at location \( i \) at time \( \tau \) and defined as \( H_i(\tau) = \langle \eta_i(\tau) \rangle + H_i(\text{geoid}) \). Both the mean, \( \langle \eta_i \rangle \), and the geoid, \( H_i(\text{geoid}) \), are constant in time, and cancel when height differences, \( \Delta \eta_i \), are computed between repeat tracks (in units of meters):

\[
\begin{align*}
\eta_i'(\tau = 1) + \langle \eta_i \rangle + H_i(\text{geoid}) & = \Delta \eta_i, \\
- [\eta_i'(\tau = 2) + \langle \eta_i \rangle + H_i(\text{geoid})] & = \Delta \eta_i, \\
\end{align*}
\]

where 1 and 2 denote the survey number. If, also, \( \Delta \eta_i = B_i(k) - B_j(k) \), the altimeter data can be used as a priori information in the inverse method. Equation (4) is now rewritten to include the a priori altimeter data as

\[
\sum \left( [B_i(k) + g \int_D z(\rho_1(k)) dz] - [B_j(k) + g \int_D z(\rho_2(k)) dz] + [B_1(k) - B_2(k) - \Delta \eta_i] \right)^2 = \text{error}^2, \tag{5}
\]

where \( k = 1, 2, 3, \cdots, M \). The altimeter data were only used when a CTD or XBT station in survey 1 coincided with a station in survey 2. The solutions are found using a weighted least squares; the stations with the altimeter data weighted more than the other stations to force the solution toward these a priori data values. If this is not done, the altimeter data only provides a few more equations in an overdetermined set of equations. Survey 1 contains a total of 38 stations and survey 2 has 21 stations. Four stations matched between survey 1 and 2 for section B, the eastern section and seven stations matched between the surveys for the western section, C. Of course, no altimeter data are available along
pared below to ADCP data and velocities computed in the traditional sense, using 800 m as a level of no motion.

Figures 15 and 16 show the total velocity field along the three sections of the triangle for surveys 1 and 2 computed with the Bernoulli method and including the altimeter data as a constraint. The inverse-produced velocities along section A (Figs. 15a and 16a) show very little flow north or southward. Only near 6°W in Fig. 15a does the flow reach 10 cm s⁻¹ southward. This compares favorably with the flow shown on Fig. 9a, the plot of the normal component to the section of the ADCP velocity vector. The ADCP data also show a flow greater than 10 cm s⁻¹ centered on 5°W, which is only very weakly (less than 5 cm s⁻¹) represented on the plot of the inverse velocities. The traditional geostrophic velocities (Figs. 7a and 8a) show very little flow, all less than 2 cm s⁻¹.

Section B, the eastern section, shows a westward flowing undercurrent between 5 and 20 cm s⁻¹ near the shelf edge at 62.3°N in Figs. 15b and 16b, the plots of the inverse velocities. There is also an eastward surface current (most likely flowing along the front) with a maximum of 20 cm s⁻¹ (at 62.8°N, Fig. 14b). The velocities near the surface on the ADCP plot (Fig. 9b) are also between 5 and 20 cm s⁻¹ and on the traditional section A, the base of the triangular region. Figures 14a and 14b show the alongtrack height differences for the eastern and western sections. The dots on the figures mark the values that have been extracted from the altimeter data to use as prior information in the Bernoulli inverse. The altimeter height differences were interpolated to the in situ data locations from the filtered, corrected altimeter height data. The two repeat tracks used for the computed differences for the eastern track, B, were sampled on 13 and 22 September, coinciding with the days when the in situ data were sampled. The two tracks used for the western section, C, were sampled on 16 and 19 September. An altimeter repeat track should have been sampled the day when the in situ data were collected on 13 September but was not; altimeter data from 16 September are used instead.

b. Results

The Bernoulli method was run twice, one without altimeter data and one with altimeter data. The velocity fields, from the run with altimetry included, are com-

FIG. 14. Alongtrack height differences for (a) track B, days 13 and 22 Sep 1991 and (b) track C, days 16 and 19 Sep 1991. Used in the inverse method as constraints.

FIG. 15. As in Fig. 7 except for inverse method velocities: survey 1.
and ADCP fields show lower velocity water above and below 240 m.

To determine how well the inverse method reproduces the geostrophic flow, it is more enlightening to examine the dynamic height field difference between survey 1 and survey 2 for the altimeter data, the traditional method and the two inverse runs (with and without the inclusion of altimeter data). This eliminates the ageostrophic component of velocity that is included in the ADCP data. Figure 17b, section C, shows the plots of the difference in dynamic height for the traditional calculation, for two inverse runs, and for the altimeter data. Ideally, the dynamic height from the inverse method should match what the altimeter sees, if the altimeter is "truth." As expected, use of the altimeter data in the inverse method constrains the results, not only at points where the altimeter data have been used (the points on the graph where the inverse method and the altimeter lines intersect) but also at other locations such as 63.7°N. Since results from the traditional calculation differ significantly from both runs of the inverse method, the choice of where to place the reference level is a problem. An educated guess

Figure 16. As in Fig. 8 except for inverse method velocities: survey 2.

Figure 17. Differences in the dynamic height field, survey 1 minus survey 2, for altimeter height data (dash–dot), geostrophic referenced to 800 m (solid), inverse method with altimeter data (short dash) and without altimeter (long dash). (a) Section or track B and (b) section or track C.
could be made as to where a “good” level is for each station, but it is wiser to use an inverse method, such as the Bernoulli method which applies some conservation laws, to make the estimate of the level of no motion.

Unfortunately, the comparison for section B is not as good (Fig. 17a) as it is for section C. Neither the dynamic height field computed in the traditional manner nor the inverse dynamic height field reflect what appears to be large changes in the surface height field observed by the altimeter and vice versa. In section 2, it was suggested that the different spatial sampling between the first and second survey should not matter and it does not for leg C. However, in the section B comparison, near 63.3°N, the hydrography indicates a large change in the surface height, whereas the altimetry shows little change. The reason for this is not the spatial sampling, but the temporal sampling of the hydrographic station at approximately 63.3°N. It was sampled two days after the stations on either side of it due to rough weather. There could also be some error in the corrections applied to the altimeter data, mainly the tidal correction, though the tidal model is very good. Other corrections, inverse barometric pressure and ionospheric corrections, change on a spatial scale larger than our sampled region and would not cause the difference as seen between the hydrographic data and the altimetric data. Rain did occur off and on during the cruise, but it is thought that it covered the whole area; thus, the wet tropospheric correction should be the same along each section.

The inverse velocity fields can also be in error if the initial prescribed depths and the associated densities used in the inverse are not chosen correctly from the original hydrographic set. In this case, 20 depths were chosen, which were 51.0, 55.5, 70.5, 82.5, 91.5, 102.0, 130.5, 150.0, 171.0, 190.5, 210.0, 240.0, 291.0, 340.5, 390.0, 450.0, 550.5, 651.0, 750.0, and 801.0 m. If a different set of prescribed depths are chosen (as was done in an earlier test where more layers between 50 and 180 m were defined), the bottom flow in the southern region is even stronger, 30 cm s⁻¹, instead of 15 cm s⁻¹. Therefore, the depths should be chosen so that any linear interpolation of the densities between the depths chosen reflects the same gradients that were in the original in situ data.

5. Conclusions

The circulation in this small region is known to be very complicated. It is near the point where the front changes from being a west–east front to a north–south front. The tip of the triangle in the south almost exactly corresponds to the tip of a tongue of cold water as seen in Fig. 11, an AVHRR image taken on 5 September 1991. From this image, many meanders can be seen, showing how difficult it is to describe the circulation in this area. The two surveys have, most likely, cut across the front or its meanders.

The hydrographic variability of the region is on several timescales. At the CTD stations near the frontal region, the variability is on timescales of days, but farther away from the frontal region, such as along the top of the triangle, the variability is on much longer timescales. The hydrography did not vary over the two weeks of sampling in this northern section. The western
section has cut across either a meander or an eddy. This eddy may cross into the eastern section (Figs. 5a,b). This is supported by examining the $T$–$S$ plots (Fig. 4), which show differences in the water masses encountered along each section. Only south of the front and at station 6 is there any MNAW along section B, the eastern section.

What about the deep flows, are they reasonable? Figure 18 shows a schematic of what could be happening along the IFF. As the eddy in the near-surface layer moves westward from the first (Fig. 18a) survey to the time 6 days later of the second survey (Fig. 18b), the water in the intermediate layer below is also moving westward. The $T$–$S$ diagrams (station 14 and 17) suggest that a layer of NNAW is developing at about 63.3°N, coming from the west (Fig. 18c). The velocity magnitudes are reasonable as others (e.g., Dorey 1975) have also recorded relatively large velocities at these depths.

Can altimetry tell us something about the subsurface flow in a complicated frontal region? Unfortunately, because the region is so highly variable, only with more data, both in situ and satellite, sampled at different times of the year over many years will we be able to infer the subsurface flow from satellite altimetry alone. Models may be a good tool to also look at these relationships.

To summarize, this research found that

1) even at depths and in a frontal region, the Bernoulli inverse method has reproduced a valid geostrophic velocity field, especially from the evidence in Fig. 17b.

2) the altimeter height data, even if only the temporal change is used, can be used to effectively constrain the inverse method.

3) the ERS-1 OPR data is of good quality shown by the qualitative similarities between the hydrographic results and the altimetric results.

Acknowledgments. I would like to thank Meric Srokosz for conducting the cruise and the members of the scientific staff and RVS staff for helping with the collection of the data and I thank the RSS Charles Darwin captain and crew for a well run ship. Thanks goes also to Roger Proctor and Roger Flather for the tide code, Phil Woodworth for various codes for altimeter corrections, Nick Crisp for processing the ADCP data, and to Paul Sherliker and Peter Killworth for the Bernoulli code. Jane Read, Dave Ellett, and Raymond Pollard are thanked for teaching me a little about GIN Sea water masses and to some of the above and the referees for their comments on a previous draft. ESA provided the ERS-1 altimeter data under the ERS-1 Announcement of Opportunity.

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