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Mesoscale dynamics in the Faroes Channels

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Mesoscale dynamics in the Faroes Channels

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1 INTRODUCTION

1.1 Background

The Faroe-Shetland Channel has been the subject of regular oceanographic study since 1893 (Turrell, 1995). Hydrocarbon exploration, fisheries research and the perceived role of the Channel in the basin-scale circulation of the North Atlantic (including the poleward transport of heat to the northern seas) have led to the region becoming one of the more intensively studied in the world. Physical processes occurring in the Channel thus have a wide impact and there is a requirement to further the understanding of the governing dynamics.

With the requirements of the offshore oil industry in mind, the UK Health and Safety Executive (HSE) sponsored the Centre for Applied Oceanography (CAO), at the University of Wales (Bangor), to undertake a two-year (part-time) investigation into the variability of flow in the Faroe-Shetland Channel and the southern end of the Faroe Bank Channel (Figure 1.1). Due to their association with elevated currents speeds, mesoscale features such as eddies are of particular interest, including how and where they may be formed. This report summarises the investigation, which utilised quasi-lagrangian surface drifters, *in-situ* hydrographic measurements and satellite observations. Observational data were collected during a two week cruise in the Faroese Channels aboard the FRV '*Scotia*', operated by the Marine Laboratory in Aberdeen.

This report draws upon and expands an interim report on the programme, submitted to the HSE in March 2000 (Williams and Sherwin, 2000).

1.2 Project Objectives

The objective of the project is to investigate, describe and if possible account for mesoscale variability, including eddies, in the Faroe-Shetland Channel. The main hypothesis is that the southern end of the Channel, in the vicinity of the Wyville Thomson Ridge is important to the formation of mesoscale eddies.

1.3 Geographical Setting

The Faroe-Shetland Channel or FSC (Figure 1.1) forms one of three openings in the Scotland-Greenland Ridge, which acts as a partial topographic barrier to the flow of water between the North Atlantic and the Greenland-Iceland-Norway (GIN) seas. The Channel forms a deep (>1500m) trough extending south from the Norwegian Sea. It is bounded on the western flank by the Faroes shelf or Plateau and to the east by the Scottish slope; its south-western extremity is defined by the Wyville Thomson Ridge (WTR), which has a sill depth of around 500m. The deepest bathymetry lies at the north-eastern end and shoals southwards to a sill of 1000m near 60.5°N, 5.0°W, before deepening again to the north of the WTR. The Faroe Bank Channel (FBC), which separates the Faroes Plateau and the Faroe Bank and has a sill depth of about 850m, is a continuation of the FSC towards the north-west. The two are known collectively as the Faroese Channels and the region where they meet is termed in this report the 'confluence region'.

1.4 Selected Previous Work

The importance of the Faroe-Shetland Channel in North Atlantic and GIN seas physical, chemical and biological oceanography is emphasised by the intensity of observation and measurement that has been conducted there for more than a century. Two standard hydrographic sections across the Channel (see Figure 1.1) have been regularly worked since 1903, more recently by the Scottish Office Marine Laboratory in Aberdeen, whose primary concerns are both climate and fisheries-related. High resolution CTD and nutrient data are collected at set stations along the sections. The recently completed Nordic WOCE (World Ocean Circulation Experiment) programme has yielded 6 years of more or less continuous current measurements from Acoustic Current Doppler Profiler (ADCP) moorings around the Faroe Islands (described in more detail in Chapter 3). In addition, interest from the oil industry, initially on the Scottish slope and more recently on the Faroes side of the Channel, has swelled the number of measurement data sets. Many of these are now beginning to find their way into the public domain.

The number of specific studies (reported in the public domain) of mesoscale variability in the Faroese Channels belies the frequency and extent of measuring activity in the region. Turrell *et al* (1999) analysed nearly 100 years of temperature and salinity, whilst Sherwin *et al* (1999) made direct measurements of a cold-core mesoscale eddy (diameter 50km) on the Shetland side of the FSC, and reported for the first time a large-scale deflection of the slope current across the Channel. Chambers (1999) employed Synthetic Aperture Radar (SAR) imagery to investigate mesoscale ocean features in the FSC. This study identified both warm and cold core eddies in the Channel and also cyclonic rotation of water associated with a large meander in the slope current that appeared to have no along-slope excursion. Smyth (1995) identified mesoscale cold-core eddies in records of currents measured from a drilling rig at the Foinaven field on the Scottish slope. An analysis by Otto and van Aken (1996) of three years of ARGOS drifter data in the north-east Atlantic, showed that the southern end of the FSC is important in terms of surface eddy kinetic energy. They also reported that one drifter was trapped in an eddy (diameter 30 km) at the northern end of the Channel for around two months. Hansen and Meincke (1979) observed eddy structures in moored instrument data on the Faroes side of the Channel, and on the same side Dooley and Meincke discussed a 30 km eddy evident in a satellite image. Using a 3-dimensional eddy-resolving model, Oey (1998) observed meanders in the slope current of the order 20 km, which propagated along-slope at speeds of around 0.14 ms $^{-1}$. The passage of the meanders appeared to induce vacillation in the slope current, where the main axis of the current shifts shoreward during the 'trough' phase of the meander and seaward in the 'crest' phase.

2 WATER MASS DISTRIBUTION

2.1 Surface Waters

In the following review of water mass distribution in the Faroese Channels, reference is made where applicable to both the literature and to a lesser extent results from the hydrographic surveys conducted during this project (including standard sections across the FSC undertaken three times per year). Full details of the surveys and the instrumentation used are presented in Chapter 3. Potential temperature is denoted as θ and salinity as S. Figure 2.1 gives schematic maps of the circulation around the Faroes Plateau, and Figure 2.2 annotated standard potential temperature sections across the FSC, based on data collected during this research programme.

Surface circulation in the FSC and FBC (to around 500m depth) is dominated by the influx of water from the Atlantic. Relatively warm, high salinity North Atlantic Water (NAW; $9.5 < \theta > 10.5$°C, $35.35 < S > 35.45$, Hansen *et al*, 1998) enters the FSC over the WTR and flows along the Scottish slope and outer shelf into the Norwegian Sea . NAW probably originates in the Rockall Trough (Sherwin *et al*, 1999) and is transported poleward by the Continental Slope Current (CSC), a narrow jet-like current that appears to adjust seasonally. In winter the current is broad and wedge-shaped in the near surface layers, in contrast to summer and autumn when it assumes a more core-like appearance beneath a fresher surface layer (Mowatt *et al*, 1997). Transport by the current is at a maximum in the summer and a minimum in winter, with an annual mean of $(7.65\pm2.58) \times 10^6$ m^3 s^{-1} (Mowatt *et al*, 1997). The current is associated in the FSC with a typical along slope speed over the 500m isobath of around 20 cm s^{-1} (Hansen *et al*, 1998). Although persistent, this Slope Current undergoes periodic relaxation and intensification that can result in significantly elevated current speeds (>1 ms^{-1}). Satellite imagery has revealed that the current is also prone to extensive lateral meandering which, as will be demonstrated in later sections of this report, can significantly affect the dynamics of the FSC.

The remainder of the surface layers in the Faroese Channels are occupied by a cooler, fresher water mass commonly referred to as Modified North Atlantic Water. The ultimate origin of MNAW is the North Atlantic Current (NAC) to the west of the

Rockall Plateau (Becker and Hansen, 1998), but there is some debate as to where the 'modification' occurs. Becker and Hansen (1998) found that the MNAW in the Faroes region is similar to the source NAC and hence perhaps the perceived 'modification' is possibly a result of comparison with NAW, which has entrained in it increased amounts of saline Mediterranean Overflow Water. From the source region, MNAW flows east over the Rockall Plateau into the FBC and the western FSC. A branch of MNAW also flows over the Iceland-Faroes Ridge and round the northern extremity of the Faroes Plateau, entering the FSC from the north-east. At depth this branch of MNAW mixes with intermediate water below, but in general characteristic θ-S values for MNAW (Hansen *et al*, 1998) are $7.0 < \theta > 8.5$°C, $35.15 < S > 35.30$. A comparison θ-S plot from stations in the deep southern FBC, the south-western end of the FSC and the north-east FSC (Figure 2.3) highlights the presence of MNAW in both channels. The locations of the stations, which were visited within days of each other, are given in Figure 1.1.

A persistent front located (variably) to the south of the southern tip of the Faroes Plateau (termed hereafter in this document the 'South Faroes Front' or SFF), marks the southernmost excursion of the MNAW in the FSC. This front is usually well-defined at depth (~300m), but there is often little signature of it at the surface (Hansen *et al*, 1998). Its existence suggests that most of the Faroe-Shetland Channel MNAW recirculates within the channel and flows north-east with the NAW into the Norwegian Sea.

Exchange of upper water between the Faroese Channels appears to be variable and principally due to the shedding by the front of eddies that transport water into the Faroe Bank Channel (Hansen and Jákupsstovu, 1992). This eddy activity is the subject of closer examination in later sections of this report, but drifter tracks (Otto and van Aken, 1996) suggest that there is also some exchange from the Faroe Bank Channel into the Faroe-Shetland Channel (also substantiated by drifters released in this study, see Chapters 5 and 6). A persistent anticyclonic circulation over the Faroes shelf/slope may form a complete revolution of the Faroes Plateau. In general, however, the circulation of the southern FBC is highly variable and flow patterns that may persist for weeks or months are inexplicably (to date) replaced by different circulation schemes. The associated time scales, frequency and seasonality of these changes are currently unknown.

2.2 Intermediate Waters

Whereas away from the Scottish shelf the surface layers of the FSC have similar temperature and salinity characteristics to those of the FBC, at depth there are clear differences. In the north-eastern FSC, θ-S analysis reveals mixing of MNAW with an intermediate layer that is found increasingly with depth, Arctic Intermediate/North Iceland Water (AI/NIW). With a characteristic temperature and salinity range of $2 < \theta > 5$°C, $34.93 < S > 35.03$ (Sherwin *et al*, 1999), this water originates north of the Iceland-Faroes Ridge through the mixing of Arctic and Atlantic waters during winter convection (Turrell *et al*, 1999). It lies in a band beneath the MNAW between 400 and 600m on the Faroes side of the FSC, and slightly shallower beneath the NAW on the Shetland slope.

A second, intermediate water mass, Norwegian Sea Arctic Intermediate Water (NSAIW; $-0.5 < \theta > 0.5$°C, $34.87 < S > 34.90$) is indicated on the θ-S diagram as a salinity minimum. It is probably created in the Arctic regions of the Iceland and Greenland Seas and subducted under the Arctic front before flowing south between the saline Atlantic waters above and Norwegian Sea Deep Water (NSDW) below (Sherwin *et al*, 1999). It lies beneath the AI/NIW at 600-800m on the Faroes side, and in a much thinner band on the Scottish slope where, occasionally, it does not impinge at all (Turrell *et al*, 1999), thereby allowing AI/NIW to mix directly with water nearer the bottom.

The FSC profiles in Figure 2.3 are somewhat atypical in that AI/NIW is usually characterised by an inflection at around 4°C (Turrell *et al*, 1999). In the figure this occurs at 2°C (in the northernmost profile) and is accompanied by an increase in salinity from around 34.87 to 34.90. The cause of this feature is not clear, although it would appear to be more a result of unusually low salinity associated with the lower end of the AI/NIW temperature range than any influence of NSAIW. It may be a artefact of interannual variability, but the salinity values above are at the lower end of the scale of those observed in the FSC (Turrell, 1995). There does, however, appear to be some evidence of mixing between the intermediate waters. The influence of the intermediate waters is less discernible in the south-western end of the FSC, suggesting not only mixing but also recirculation beneath the surface waters (Hansen *et al*, 1998). There is little exchange of intermediate water between the FSC and FBC (as indicated by the plotted profiles), and as mentioned previously it is believed that the transport of these waters from the FSC to the FBC is principally in the form of eddies or boluses.

2.3 Deep Waters

The deepest water apparent in Figure 2.2 has the characteristics of Faroe-Shetland Channel Bottom Water (FSCBW; typically θ < -0.5°C, 34.91 < S > 34.92). This water is a mixture of Norwegian Sea Deep Water (NSDW) and NSAIW and is evident in the θ-S profiles. The FBC is thus an exit route for deep water entering the FSC from the GIN seas. Deep water also flows out over the WTR, but the overflow across the Ridge is intermittent both spatially and temporally and perhaps 20% of that flowing through the FBC (Hansen *et al*, 1998). The composition of FSCBW is known to vary on a decadel scale, with a decrease in the proportion of NSDW in the FSCBW from 60% before 1985 to 40% in 1995 (Turrell *et al*, 1999). This decrease relates to a freshening of 0.01/decade.

3 STUDY DATA

3.1 Field Survey

3.1.1 Aims and Scope

The principal aim of the field measurement collection programme was to provide detailed knowledge of the physical regime in the confluence of the Faroese Channels, in order to quantitatively assess whether conditions in the region are conducive to the formation of eddies. The fieldwork had two main objectives:

1. To delineate the South Faroes Front through spatially intensive measurement of temperature and salinity by means of conductivity-temperature-depth (CTD), expendable bathythermograph (XBT), shipborne thermosalinograph measurements and an Acoustic Doppler Current Profiler (ADCP). In the event, no ADCP data were collected due to equipment malfunction after the vessel had sailed from Aberdeen. Figure 3.1 shows the vessel track and Figure 3.2 summarises the data collection locations.

2. To investigate the spatial structure of the flow in and near the frontal region. Near real-time analysis of the T-S data enabled an assessment of suitable locations for the deployment of eight satellite-tracked drifters across the front. It was intended that an eddy would be seeded with at least one drifter.

3.1.2 Measurement Programme

Data collection was carried out intermittently between 29th April and 10th May 1999 as a component of Cruise 0799S of *FRV 'Scotia'*, operated by the Scottish Office Marine Laboratory in Aberdeen. The cruise also included working the standard Fair Isle-Munken and Nolsa-Flugga hydrographic sections and was generally considered to be successful.

The data collection programme was carried out in three phases:

1. Frontal Survey: a CTD/XBT 'ladder' survey over the south-eastern end of the FBC. During this phase the thermosalinograph data were monitored to determine the location of the front.

2. Drifter Phase: deployment of the eight drifters, in locations that straddled the frontal region. CTD and XBT deployments were continued. Two drifters were retrieved after a week, the Minilog temperature sensors removed and the drifters re-deployed.

3. Eddy Survey: near real-time analysis of the satellite derived drifter positions in the southern end of the FSC showed two drifters displaying cyclonic rotational motion. A limited CTD/XBT survey was undertaken to investigate whether the motion was due to the presence of a cold-core eddy.

3.1.3 Meteorological Conditions

Figure 3.3 shows the synoptic isobaric maps for the period 29th April to 1st May ('Julian' days 119-121). On 29th conditions were initially fairly benign (winds 15-20 knots approximately, but a trough of low pressure moved through the region, giving winds of 40-45 knots for a time in the evening. This resulted in a suspension of survey activity from 1650 UTC on the 29th to 0220 UTC on the 30th April. Thereafter the calm conditions returned with light winds from the southeast, indeed until 7th May (day 127). A low pressure system approached the region from the southwest, bringing with it winds of 20-25 knots that backed into the northeast. These conditions persisted until the end of the data collection on 10th May.

3.1.4 Temperature and Salinity (T-S)

T-S data were collected by three means:

1. Seabird SBE 25 conductivity/temperature/depth sensor (CTD), that recorded vertical T-S profiles to a depth of around 1100m. Typical accuracies for the CTD are ± 0.002°C in temperature, ±0.005 psu in salinity and <0.5% depth in pressure

(Emery and Thomson, 1997). Post-cruise calibration and quality control of the CTD data were carried out by the Marine Laboratory (Aberdeen). Direct downcast conductivity measurements were calibrated from bottle samples prior to conversion to salinity using the linear equation:

$$Cond_{calibrated} = 1.00206 \text{ x } Cond_{measured} + 0.002142 \qquad (3.1)$$

2. XB-7 expendable bathythermographs (XBT), supplied by the Ministry of Defence, collected additional temperature profile information between the CTD stations. The XBT probe is attached by wire to a computer aboard the vessel, through which it transmits continuous temperature information. The depth is calculated using a standard fall rate equation:

$$\text{Depth} = \text{t x D1-t}^2 \text{ x D2} \qquad (3.2)$$

where t = time since splash detection
D1 and D2 are probe type-specific constants (6.4720 and 0.00216)

Any inaccuracies in the XBT data are likely to be due principally to fall velocity errors (Emery and Thomson, 1997), but a comparison of CTD and XBT data (Figure 3.4) for a station during the frontal survey shows they are in reasonable agreement with each other. The operational range of the XBT is around 800m depth.

3. Seabird SBE 21 thermosalinograph (TSR), that recorded sea temperature and salinity throughout the cruise at the depth of the water intake (about 3m) on a 2-minute time step. The sampling interval was changed for operational reasons during the cruise to five minutes. Post-cruise calibration of the TSR data was carried out by the Marine Laboratory (Aberdeen).

3.1.5 Velocity

Eight Far Horizon drifters (type FHD-A) were deployed in the Faroese Channels (see Figure 3.2 for deployment locations), six in the southern end of the FBC and two at the south-western end of the of the FSC. Each buoy was equipped with a battery powered Argos transmitter, enabling its position to be fixed by NOAA orbiting satellites. The French organisation 'Service Argos' currently utilises three polar-orbiting NOAA satellites, with more planned for the year 2000, giving more location updates and better spread of results throughout the day. The number of position fixes received each day is a function primarily of the buoy latitude (at 60°N there are about eighteen satellite passes, with an average rate of thirteen fixes per day). Each position fix is given an accuracy descriptor with the highest, '3', equating to a position circle of 150m radius, and the lowest, '0', to 1000m radius. For this project, only those fixes of classes 3 to 1 were accepted.

The units had been deployed in a previous survey, recovered and completely renovated. Each drifter comprised a cylindrical buoy (1m long, 12cm diameter) attached to a holey sock drogue (2m long, 66cm diameter) by 100m of 6.5mm diameter polyester line (Figure 3.5). The drogue depth was chosen such that the drifter movement reflected conditions below the mixed layer (i.e. geostrophic rather than inertial currents). A small pellet float was attached to each line 3m from the buoy, to give some additional buoyancy and to reduce snap-loading on the drogue and line due to surface waves. Each carried an onboard temperature sensor (accurate to ±0.15°C). Drifter positions were updated typically fifteen to twenty times per day, and temperatures at perhaps twice this frequency. Data were transmitted from the satellites to a receiving station in southern France, downloaded via the internet to CAO and then relayed to the ship. The temperature sensors on units 91, 92 and 96 returned no data.

Table 3.1 summarises the drifter deployment and monitoring phases of the project.

Table 3.1 - Drifter deployment monitoring and recovery details

Drifter	Start Details						Recovery Details							Last Tx				Cross 0°W	
	Date	Time z	Lat °	Long °	Sounding m	CTD		Date	Time z	Lat °	Long °	Sounding m	CTD	Date	Time z	Lat °	Long °	Date	Time z
12389	1-May	14:33	60.462	-6.848	800	94	NR							21-Jun	19:07	60.892	3.811	29-May	07:28
12390	1-May	17:12	60.042	-6.751	558	95		7-May	17:30	60.671	-3.742	641	167	30-Jun	22:12	62.391	1.166	4-Jun	07:56
							RD	8-May	00:41	61.125	-5.199	427	168						
12391	1-May	19:13	60.132	-6.358	1180	96	NR							30-Jun	23:57	62.138	3.026	30-May	23:47
12392	1-May	13:42	60.501	-6.727	415	93	IR	8-May	16:47	60.265	-5.931		XBT 81	17-May	20:08	60.159	-6.87		
12393	1-May	03:38	60.747	-6.999	377	88	NR											18-Jul	01:24
12394	1-May	04:53	60.784	-6.803	210	89	NR							21-Jul	08:32	61.466	-7.53		
12395	1-May	07:39	60.715	-7.115	679	91	NR											30-Jul	08:31
12396	1-May	12:13	60.532	-6.479	247	92	NR							4-Jul	20:54	61.354	-6.11		

Notes: NR = Not recovered
RD = Re-deployed
IR = Immediately re-deployed
CTD = Conductivity-Temperature-Depth
XBT = Expendable bathythermograph
-ve longitude = west

The drifters were deployed from the deck of '*Scotia*' following an initial test period of twelve hours, during which communication between the drifter, the satellite and CAO was verified. After release, the drifters were tracked either until they stopped transmitting or they crossed the Greenwich Meridian to the north of the Shetland Islands. Drifter 90 was recovered on the Shetland slope a week after release and re-deployed on the Faroes side of the Channel. Drifter 92 was recovered eight days after deployment and immediately re-deployed. It stopped transmitting seventeen days after initial release. Drifter 94, which never entered the FSC, continued transmitting for over eighty days. Drifter 89 was the first to transit the FSC and cross the Greenwich Meridian, in twenty-nine days.

Attached to each drifter line at the top of the drogue was a self-recording Venco Minilog temperature sensor (accurate to ±0.1°C), set to record every five minutes. Additional Minilogs were also deployed (1m below the pellet float) on the three drifters whose onboard thermistors were faulty. However, only three units provided information as most drifters were not recovered. The timing of recovery has unfortunately compromised the potential usefulness of these data.

3.1.6 Eulerian Current Observations

As part of the on-going World Ocean Circulation Experiment (WOCE), in the Nordic-WOCE (NWOCE) component five fixed current meter moorings have been maintained across the southern end of the FSC (see Figure 3.2 for their locations). An acoustic doppler current meter (ADCP) in upward-looking configuration continues to record current velocity profiles through the water column at each site, providing an invaluable high resolution, long term data set. The Marine Laboratory (Aberdeen) maintains the moorings and carries out quality checking and archiving of the currents. It transferred the currents for this study in ASCII format, following preliminary screening.

The ADCP uses the doppler principle to compute currents. After transmission of four acoustic pulses, the instrument 'listens' to the echoes returned from scatterers such as plankton in the water column. The change in frequency of the echoed signal enables the water velocity to be computed from the Doppler relationship:

$$F_d = \frac{2F_t.(V\cos\theta)}{c} \qquad (3.3)$$

where F_d is the doppler frequency shift, F_t is the transmitted frequency, c is the speed of sound in seawater, V the water velocity along the beam axis and θ the angle between the acoustic pulse beam and the vertical. The current velocity profile is calculated by range-gating the returning signal into 'bins' and directions are converted to a polar frame of reference by means of an internal compass in the ADCP. Table 3.2 summarises the main attributes of the current data used in this study.

Table 3.2 – ADCP Current Measurements

Instrument	Frequency (kHz)	Water depth (m)	Instrument depth (m)	Bin length (m)	Sampling frequency (mins)	Start time	Finish time
NWOCE-A	150	295	294	10	20	15/04/99 00:00	12/06/99 17:00
NWOCE-B	75	782	674	25	20	15/04/99 00:00	12/06/99 08:00
NWOCE-C	75	1076	662	25	20	15/04/99 00:00	12/06/99 03:00
NWOCE-D	75	794	771	25	20	15/04/99 00:00	16/09/99 10:00
NWOCE-E	150	446	429	10	15	15/04/99 00:00	16/09/99 07:15

3.2 Satellite Observations

3.2.1 Sea Surface Temperature

To enhance interpretation of the in-situ observations, images of sea surface temperature (SST) were obtained from the Remote Sensing Data Analysis Service (RSDAS) at Plymouth Marine Laboratory (UK). The SST data were acquired by satellite-borne Advanced Very High Resolution Radiometer (AVHRR) flown on aboard NOAA operational satellites in sun-synchronous orbits, with equator-crossing times of 0730 hours and 1400 hours. The AVHRR is a broad-band, multi-channel scanner, sensing in the visible, near-infrared, and thermal infrared portions of the electromagnetic spectrum. Because there are two satellites operational at any time it is possible to see the same point on the Earth's surface about four times each day, and more frequently at latitudes around the UK. Currently about nine passes are received and archived each day. As the sensors measure the temperature of the sea surface 'skin', diurnal surface heating (which in this study was found to be of the order +0.5°C, the approximate sensing accuracy) must be considered when quantitatively using images taken during daylight hours.

Following calibration and cloud-clearing (AVHRR imagery is susceptible to cloud cover, which inhibits sea-surface radiation from reaching the satellite sensor), full colour SST images are produced like that shown in Figure 3.6, which is a rare, almost cloud-free image from early morning on 19th May 1999. Several key features are evident in the image:

1. The ingress of warm NAW, which is organised into the jet-like slope current north of 60°N.

2. The southward intrusion of cooler MNAW over the Faroes Plateau, and the frontal region between it and the surrounding water.

3. An intense cold core eddy at about 60.2°N, 4.0°W.

4. Deep meanders in the slope current, extending to nearly the width of the FSC.

The persistent cloud cover in the region throughout the spring and summer of 1999 resulted in few clear AVHRR images (Figure 3.6 being the best). However, when using the images to investigate specific features even partial images are useful (if not presentable) and hence a chronology of some six such images of was obtained for use in this study.

In recent years the Remote Sensing Data Advisory Service (RSDAS) at the Plymouth Marine Laboratory (UK), has developed algorithms for detecting in AVHRR data sea surface temperature gradients associated with mesoscale structures. The technique produces scenes of thermal fronts collated from sequences of AVHRR images that individually due to cloud cover may be of little use. Although comprised of information collected over several days, the scenes are nonetheless extremely useful in cloudy areas such as the north-east Atlantic. Such a scene has been used in Chapter 7 of this report.

3.2.2 Sea Surface Height Anomaly

Variation in sea surface height (SSH) implies horizontal gradients of density, which in turn give rise to geostrophic currents. Horizontal variation in temperature will induce density gradients, and such changes are often associated with discrete features such as eddies and fronts. Knowledge of the gradient of SSH anomaly (i.e. departure from the long term mean, here of cycles from 1993 to 1996) across an area can therefore assist identification of these features.

Measurement of SSH is undertaken using satellite-borne radar altimeters. These permanently transmit signals at high frequency to Earth and receive the echo from the sea surface. The returning signal is analysed to derive a precise measurement of the round-trip time between the satellite and the sea surface. The time measurement, scaled by the speed of light (at which electromagnetic waves travel), yields a range measurement. Averaging the estimates over a second produces a very accurate measurement (±2 cms) of the satellite-to-ocean range. To transform satellite altimeter range measurements into accurate sea-surface elevation measurements, a variety of models, algorithms, and corrections needs to be adopted and applied. These include the determination of the satellite position within a consistent geodetic reference frame, the correction for various media and environmental effects, and the removal of tidal and (possibly) atmospheric loading effects. A full list of algorithms may be found at www.neptune.gsfc.nasa.gov.

For this project, limited SSH anomaly data from the Topex-Poseidon (T-P) and European Remote Sensing (ERS) satellites were downloaded from the Colorado Centre for Astrodynamics Research (CCAR) Internet archive. T-P is a combination of two sensors launched in 1992 to an altitude of 1335 km into a 9.97-day repeat cycle, i.e. it covers the same ground approximately every ten days. Such temporal resolution is useful for observing mesoscale variations, but it does result in a coarse cross-track spacing of over 300 km. The two ERS satellites (of which only ERS-2 is now operational) were launched in 1991 and then 1996 to an altitude of 780 km into a 35-day repeat orbit, giving a better spatial resolution of 8 km at 60º of latitude. An application of altimeter data in this study is described in Chapter 7.

4 DATA PROCESSING

4.1 Methods

4.1.1 Software

Digital data processing was carried out in the Mathworks Matlab v5.3 environment, with occasional formatting and program checking undertaken in Microsoft Excel. In addition to the many Matlab processing scripts written during the course of this study, use was also made of the SEAWATER (v1.2d) suite of routines (Morgan, 1994), developed by the Australian Commonwealth Scientific, Industrial and Research Organisation (CSIRO) to compute seawater properties (including potential temperature and density). All plots and non-contour presentations were produced in Matlab. The Matlab scripts were checked by performing the same functions either manually or in other software on a limited subset of the data.

Transforming oceanographic data onto regular horizontal or vertical grids for contouring purposes requires considerable care, especially if the data are irregularly spaced. Grid resolution is particularly important: too coarse a grid will result in important features at sub-grid scales being smoothed out of the data and too fine a grid may create misleading, artificial features such as closed quasi-circular structures. Mapping in this study was carried out in the surface-mapping package Surfer (Golden Software, v6), which offers an extensive selection of data gridding algorithms.

Through detailed experimentation, every effort was made in this study to ensure that all contour maps represent as faithfully as possible the actual conditions sampled. The preferred gridding method was Delaunay triangulation, which creates a network of triangles across the data domain connecting the original data points (Golden Software, 1997). Each triangle defines a plane over the grid nodes lying within the triangle, with the tilt and elevation of the triangle being determined by the original data points defining the triangle. All grid nodes within a triangle are defined by the triangular surface. The major advantages of this method are (i) that using original data points to define the triangles results in the original data being honoured closely, and (ii) no spurious gridded data is produced outside the spatial extents of the original data. This is especially important in the production of horizontal contour maps.

4.1.2 Identification of Periodicity

In their raw state, the drifter and current data encapsulate motion at frequencies not associated with mesoscale activity, including tidal, inertial (period approximately 14 hours) and high frequency noise not already removed by the interpolation process. Spectral analysis is a useful of way of identifying precisely where the 'super-mesoscale' motions exist in the frequency spectrum, and hence will assist in the prescribing of filters to remove the unwanted components.

Spectral analysis partitions the variance of a time series as a function of frequency, and contributions from the different frequency components are measured in terms of 'power spectral density' (PSD). The process involves the Fourier analysis of a time series (here drifter positions and Eulerian currents) to produce a crude spectrum or periodogram, which may then be modified to improve the spectral estimate. The basic premise of Fourier analysis (Emery and Thomson, 1998) is that any finite length time series *y(t)* can be reproduced using a linear summation of sines and cosines (the Fourier series):

$$y(t) = \frac{A_0}{2} + \sum_{p=1}^{\infty} \left[A_p \cos(\omega_p t) + B_p \sin(\omega_p t) \right] \tag{4.1}$$

where $A_0/2$ is the time series mean, A_p, B_p are constants (the Fourier coefficients) and in which

$$\omega_p = 2\pi f_p = 2\pi p f_1 = 2\pi p/T \tag{4.2}$$

is the frequency of the *p*th consituent in radians per unit time (f_p is the corresponding frequency in cycles per unit time), T is the record length and f_1 is the fundamental frequency $1/T$ (the lowest frequency resolvable by the data series). The collection of

Fourier co-efficients A_p, B_p constitute a periodogram, which defines the contribution of each periodic component ω_p to the total energy in the time series.

Considering a time series of finite duration to be a truncation of an infinitely long data series is analogous to viewing the latter through a narrow, rectangular 'window'. The response function of such a window causes energy to 'ripple' away from the central lobe of the function at each frequency within the time series into side-lobes, leading to leakage of spectral energy from the main frequency components of the time series into neighbouring frequency bands. This effect reduces and contaminates the energy in the main frequency bands (the Gibbs phenomenon, Emery and Thompson, 1998). Weak spectral responses can be masked by higher side-lobes from stronger spectral responses at other frequencies.

Rippling can be reduced by applying a function or 'window', which in the frequency domain smoothes the spectrum and is analogous to weighting the data in the time domain. Segmenting the data and applying the smoothing window to each segment also allows ensemble averaging of the spectral estimates in each frequency band (i.e. the periodogram), thereby increasing the statistical reliability (or degrees of freedom) of the PSD estimate. The disadvantage to this approach is a loss in resolution through 'smearing' of the spectral energy.

Spectral analysis will only resolve frequencies lying within the principal range

$$-f_N \leq f \leq f_0, \; f_0 \leq f \leq f_N$$

where f_0 is the fundamental frequency and f_N is the Nyquist frequency ($1/2\Delta t$). The presence of energy at frequencies greater than f_N means that the possibility of aliasing must be considered. This 'folding back' of energy about f_N into the principal range at frequencies beyond the Nyquist frequency can lead to false spectral peaks within the principal range. For example, it can be shown that observations at f = 0.03 cph (30 hours period) will be aliased with spectral contributions from 1.97 cph. Choosing a smaller sampling interval will reduce aliasing effects.

Spectral analysis in this study has been carried out in the Matlab environment, which by default uses the windowed, average periodogram method, also known as Welch's method (Mathworks, 1996). The principal advantage of the method is that it reduces the variance of the periodogram, further achieved by overlapping the segments of input time series to increase the number of available. There is a trade-off between the fundamental frequency (which is lower for longer input segments) and variance reduction through the number of segments used in the analysis.

The application of spectral analysis to the drifter and ADCP observations is described in sections 4.2 and 4.3.

4.1.3 Signal Decomposition

Since mesoscale features with periods of days rather than hours are of primary interest here, it is useful to remove from the observations periodic events associated with non-mesoscale frequencies. Digital filters provide the means of removing all unwanted fluctuations from oceanographic observations, in this case the drifter observations (including temperatures) and the Eulerian current measurements.

A digital filter is an algebraic process by which an input sequential time series x_n is systematically converted into a sequential output y_n, where n = 0,1….,N-1. The filter employed in this study is a recursive filter, since it generates output by a feed-back loop specified by the second summation in term in the generalised form of the filter (Emery and Thomson, 1998):

$$y_n = \sum_{k=-M}^{M} h_k x_{n-k} + \sum_{j=-1}^{L} g_j y_{n-j}, \qquad n = 0,1......, N-1 \qquad (4.3)$$

where M, L are integers and h_k, g_j are nonzero weighting functions. In the course of applying this filter, all past output values contribute to all future output values. Unlike their nonrecursive counterparts, recursive filters do not lead to the loss of output data from the ends of the record i.e. $N_{in} = N_{out}$. The filter is applied in the time domain through convolution of the input signal with the weighting or impulse function $\{h_k\}$ by:

$$y_n = \sum_{k=0}^{M} h_k (x_{n-k} + x_{n+k}) \qquad n = 0,1......,N-1 \qquad (4.4)$$

where $h_k = h_{-k}$, a set of time invariant weights, and $k = -M, -M+1.....,M$. A function which describes the effectiveness of the filter in transmitting or blocking power within specific frequency bands is its transfer function, which in the frequency domain is given by:

$$H(\omega) = h_0 + 2\sum_{k=1}^{M} h_k \cos(\omega k \Delta t) \qquad (4.5)$$

where $\omega = 2\pi f$. In complex notation, this is has the form:

$$H(\omega) = |H(\omega)| e^{i\phi(\omega)} \qquad (4.6)$$

The amplitude $|H(\omega)|$ would be unity in an ideal filter for the frequency band of interest (the passband) and zero outside of it. The phase Φ should be zero for all ω so that the filter does not induce a shift in phase in the output signal. The Matlab software includes a specialist function (*'filtfilt.m'*) for preventing this, by passing the input forwards and then backward (after inverting the data order) through the same set of weights. This process corresponds to squaring the transfer function, so that the final function is $|H(\omega)|^2$ (Emery and Thomson, 1998).

As in spectral analysis, the range of frequencies that can be covered by a filter is defined at the high frequency end by the Nyquist frequency and at the low end by the fundamental frequency (see previous section for definitions). The cutoff frequency should lie sufficiently far from either of these limiting frequencies, a criterion shown in the next section to be fulfilled. The objective in processing the drifter and ADCP observations was to isolate low frequency events within the data. It was therefore necessary to set a cutoff frequency, ω_c, above which all frequency components would be removed from the data set. A lowpass filter was specified where (ideally)

$$\begin{aligned} |H(\omega)| &= 1 \text{ for } |\omega| \leq \omega_c \\ &= 0 \text{ for } \omega_c \leq \omega \qquad (4.7) \end{aligned}$$

The cutoff frequency is thus the transition between the passband and the stopband and ideally would be step-like, but in reality has finite width. This leads to another similarity with spectral analysis, that of the Gibbs phenomenon, which if the transfer function is too steep can produce 'ringing' or rippling in the output series through the leakage of energy into the filtered record. Frequencies close to ω_c will thus be distorted by these 'overshoot' ripples unless a smoothing or tapering function (i.e. a window) is applied to the input data. This was not done in the present case because the trade off is to increase the fundamental frequency (thereby possibly discarding the signature of mesoscale processes encapsulated within the data), and because frequencies close to ω_c are of lesser importance here. The degree of ringing can be controlled by the filter *order*, which is directly proportional to the attenuation rate of the transfer function $H(\omega)$ and hence to the amount of ringing in the output data.

The design of a suitable filter thus requires an amount of care and experimentation. The application of a specific filter to the drifter and ADCP data is described in the forthcoming sections.

4.2 Drifter Processing

4.2.1 Computation of Drifter Velocity

The downloaded drifter positions were plotted and examined for large discontinuities not readily associated with naturally occurring phenomena. The offending points were identified in the data files and removed. The resulting irregular time series of drifter positions (referenced in latitude and longitude) were then linearly interpolated on to a regular ½-hour time step (i.e. frequency $f = 2$ cph), chosen to remove high frequency noise from the data whilst minimising data distortion resulting from the interpolation. Appendix One contains plots of the raw drifter tracks and Figure 4.1 shows an 8-day section of raw drifter track from the FBC overlain by the same interpolated track.

The distance between successive pairs of positions was computed by the 'plane sailing' method used in marine navigation (suitable for distances of up to a few hundred miles):

1) Compute the zonal distance between the two longitudes, known as departure:

$$\text{Departure} = \cos(\text{mean latitude}) \times \text{difference in longitude} \qquad (4.8)$$

2) Compute the distance:

$$\text{Distance} = [(\text{difference in latitude})^2 + (\text{departure})^2]^{0.5} \qquad (4.9)$$

3) Orthogonal components of drifter speed were found for each successive pair of positions using the departure (the 'u' component) and the difference in latitude ('v' component) and the elapsed ½-hour time step. A mean position was assigned to each speed value for the purposes of plotting vectors.

4.2.2 Frequency Analysis of Drifter Data

The PSD associated with drifter movement was derived in the following sequence of operations:

1. Time series of orthogonal distance components were constructed, based on east/west and north/south distance of successive raw drifter observations from a pre-defined origin (60.5ºN, 006.0ºW). The mean was removed from each time series.

2. A single complex time series was formed from the orthogonal components ($y(t) = u(t)+iv(t)$).

3. The complex time series was divided into lengths of 512 (2^9) observations, each of which was linearly detrended and overlapped its neighbour by 50% of its length. If required the final section was zero-padded to length 512.

4. A Hamming window was applied to each segment, defined as:

$$w(n\Delta t) = 0.54 - 0.46\cos(2\pi n / N) \qquad (4.10)$$

where n = 0,1,….N – 1, Δt = sampling interval and N is the data length. Figure 4.2 shows the Hamming window for N = 512 data points, and for comparison the Hanning window of which the former is a derivative. Both are commonly used in oceanographic applications, but the Hamming window was chosen here as it has more efficient side-lobe attenuation (Emery and Thomson, 1998). The equivalent degrees of freedom (EDoF = DoF/2 for overlapping segments) have been computed for each spectrum from (Emery and Thomson, 1998) by:

$$\text{EDoF} = 2.5164(\text{N/M}) \qquad (4.11)$$

where N is the number of observations in the time series and M the half width of the window (256).

Using segments of 512 half-hourly observations gives a lowest resolvable frequency of (24/(512*0.5)) cycles per day, or a period of 11 days. Any apparent energy in the spectrum at frequencies below this is therefore to be ignored. To compensate for loss of spectral energy through the windowing process, the spectral estimates were multiplied by √[5/2] (Emery and Thomson, 1998).

This analysis was carried out on the movement of the drifters for the period during which they remained west of longitude 6ºW, and example spectra are shown in Figure 4.3 for drifters 89, 92 and 94. The spectra show energy up to the sampling frequency (2 cph) and since this is greater than f_N the possibility of aliasing must be considered. Choosing a smaller drifter position interpolation interval may have reduced aliasing effects (as would pre-filtering), but as most of the variability in the spectra is at frequencies higher than 0.03 cph, this was set as the low-pass frequency for filtering the drifter data. Most of the aliased energy will therefore be confined to near the upper end of the low-pass frequency range, minimising the impact on future analyses of mesoscale activity (which is associated with periods of days rather than hours).

4.2.3 Filtered drifter data

From the previous discussions the requirement is for a lowpass, recursive filter with a steep transition between the pass and stopbands, applied by a function which minimises phase distortion, with a cutoff frequency that is sufficiently far from the fundamental frequency (for a time series of 30 days, 0.001 cph) and the Nyquist frequency (1 cph). After review of the spectral analysis and considering the requirement to remove all energy of tidal and higher frequencies a cutoff frequency of 0.03 cph (period of 30 hours) was selected, to be applied by a Butterworth filter which is routinely used in oceanographic applications and meets the above criteria. Its transfer function can be expressed as (Emery and Thomson, 1998):

$$|H_L(\omega)|^2 = \frac{1}{1 + [\tan(\pi\omega/\omega_s)/\tan(\pi\omega_c/\omega_s)]^{2q}} \qquad (4.12)$$

where ω_c is the cutoff frequency, ω_s the sampling frequency and q the filter order.

The Butterworth filter as applied by the Matlab software has the advantage of a squared response (i.e. $|H(\omega)|^2$), which gives zero phase distortion, and its amplitude is attenuated by a factor of two at the cutoff frequency ω_c, for which $\omega/\omega_c = 1$ for all values of filter order q. The response of the filter for a range of q is shown in Figure 4.4. A 5th order low-pass filter was applied to each time series of latitude, longitude, orthogonal speed components and temperature from the on-board thermistors. Filtered scalar speeds were derived by re-combining the filtered orthogonal components. Figure 4.5 shows the effects of the filtering process on a particularly noisy drifter trajectory and all processed drifter tracks are contained in Appendix Two.

4.2.4 Verification of Drifter Integrity

A principal failing of drifter measurement is loss of drifter integrity through inadvertent detachment of the drogue from the surface buoy. If this occurs the drifter will follow more closely the direction of the wind and waves as well as that of the desired current components. The drifter data were subjected to three checks to ensure that this had not occurred:

1. Time series of raw drifter speeds were examined for sudden sustained increases.

2. The raw drifter speeds were linearly interpolated onto a 3-hour time step that coincided with a time series of wind vectors extracted from the Centre for Applied Oceanography forecast model, at 61.5°N, 005.0°W (central Faroe-Shetland Channel). This enabled direct comparison with wind vectors.

3. Time series of variance were constructed from the daily averaged drifter speeds. In the event of separation, the dampening action of the drogue and line would be lost and the variance in the drifter speed would increase.

Of all the drifters only unit 94 returned doubtful results. There is no evidence of drogue loss in the first two checks, but at around day 195 there is a sharp increase in velocity variance (Figure 4.6). Examination of the drifter trajectory showed that at around this time the drifter was very close inshore to the Faroe Islands. It was concluded therefore that this sudden increase in variance indicated drogue detachment. It was also concluded that all other drifter systems remained intact throughout the monitoring period.

4.2.5 Drifter Response to Wind

Errors can be introduced into the drifter data by forces acting on the drifter system that are not associated with the flow beneath the mixed layer (i.e. the currents the drifters were deployed to follow). These forces are (Krauss *et al*, 1989) (i) wind drag on the surface buoy, F_w, (ii) current drag on the submerged part of the buoy, F_s, (iii) non-linear wave forces (Stokes drift) on the surface buoy, F_{wa}, (iv) current drag on the tether between the buoy and drogue, F_t, and (v) on the drogue itself (if the other components inhibit the drogue from following the current), F_d. On average,

$$F_w + F_{wa} + F_s + F_t + F_d = 0. \qquad (4.13)$$

The forces are generally assumed to follow a drag law

$$F = \rho c_d A \;\; v \;\; v/2 \qquad (4.14)$$

where ρ is the density of the surrounding fluid (sea or air), c_d a drag coefficent, A the cross-sectional area (length x diameter) of the drifter system on which the force acts and v the relative velocity of the surrounding fluid. For optimal drifter performance, the drag area ratio $\sum_i c_{di}A_i/c_{dD}A_D$, where D represents the drogue and i the other individual drifter system components, should be as small as possible. For the drifter configuration in Section 3.1.5, this ratio is 1.1/1.8m^2 or 0.6, assuming a uniform value for c_d of 1.4. The drag ratio of area below and above the surface (assuming the surface buoy to be 2/3 submerged) is 2.9/0.05m^2 or ~52:1.

The drifters should be thus be regarded as quasi-lagrangian particles as they do not truly follow the paths of water parcels into which they were released. Following Sherwin (1990) and assuming (i) zero vertical shear in the surface layers occupied by the drifter system and zero wave-induced Stokes drift, (ii) a constant drag coefficient for air and water and (iii) the surface buoy is 2/3 submerged, equating forces above and below the surface gives:

$$|u| = \left[\frac{\rho_{air}}{\rho_{ater}} \cdot \frac{Area_{above\ surface}}{Area_{below\ surface}} \right]^{0.5} W \qquad (4.15)$$

where u is the slip magnitude, ρ_{air} is the density of air (1.2 kg m^{-3}), ρ_{water} is 1025 kg m^{-3} and W is the wind velocity (W>>u). This gives a slip magnitude of 5x10^{-3}W. In practice, Stokes drift can account for up to 50% of the combined wind and wave forces acting on the drifter buoy (Krauss *et al*, 1989).

From combined studies of TRISTAR and (6m) holey sock drifters centred at 15m depth, Niiler *et al* (1995) developed the theoretical slip model for winds up to 10 ms^{-1}:

$$U_s = (a/R).U_w+(b/R).\Delta U \qquad (4.16)$$

where U_s is the slip, R is the drag area ratio, Uw is the wind speed, ΔU is the velocity shear across the drogue length (determined empirically) and a, b are empirically determined constants (0.0534 and 5.57 respectively). A uniform C_d of 1.4 was assumed for the holey sock and the tether (and 1-1.2 for the TRISTAR drogue). The model accounted for up to 77% of the variance in the observations on which it was based and at 95% confidence there was little reported difference between the water-following abilities of the two drogue types. Applying the drag area ratio of 50 and a hypothetical $\Delta U = 0.5$ cm s^{-1} gives a slip value in the direction of a 9 m s^{-1} wind of 1.0 cm s^{-1}.

There are obvious differences between the drifter system deployed in the FSC and those used in the modelling experiments (e.g. buoy shape, tether length and most importantly drogue dimensions), but the models may at least give an indication of the order of slip magnitude that may be expected in the FSC data. However, in relation to other errors in the drifter data and in view of the extent of post-processing, the expected slip magnitude is not believed to be significant and hence has not been applied.

4.2.6 Additional Sources of Error

In addition to the drifter response to its environment and to drogue loss, there are other possible sources of error and data loss:

1. Errors in the velocity estimates associated with satellite reception and transmission of the positional data. It was noted that there were numerous duplicate positions and a high number of clearly erroneous positions or spikes in the raw drifter data.

2. Submerging of the surface buoy such that contact with the satellite is lost. In rough seas this becomes more of a problem, but in general the weather was comparatively settled during the monitoring period.

3. Drifter data processing – as discussed previously there are a number of areas that potentially could create inaccuracies in the drifter data. These include:

   - Careless manual de-spiking of the raw tracks could lead to the removal of real events or equally an impression of falsely high flow velocity.

   - Aliasing in the frequency domain of the drifter data may create false spectral energy peaks, leading to mis-interpretation of the spectra.

   - Whilst smoothing noisy data any interpolation and filtering process will alter the character of those data, potentially leading to mis-interpretation of actual events.

   - The filtering appears to have induced some sharp apparent deviations in drifter trajectory. This occurred where movement through the water slowed and the drifter oscillated about a position for a period of time. These are not false changes in heading, but the filter has removed the oscillations leaving just the resultant change. Two examples of this are evident in Appendix One in the raw tracks of drifter 89, as it approached the sub-mesoscale feature discussed in Chapter 7, and 93 as it abruptly changed direction in the mid-FSC. In both cases however, the drifters did eventually take the trajectory shown in the filtered data.

4.3 Processing of Eulerian Currents

The Nordic-WOCE ADCP measurements were visually inspected for bad and missing data and the resulting gaps (up to six hours) were linearly interpolated both in time and vertically through the water column. Although not ideal, this approach was considered suitable for the present purpose. A long break in the record resulted in a continuous, but truncated, data set from 6th May to 6th June 1999 being taken forward for further analysis.

Spectral analysis of the currents was carried out using the Welch method and a Hamming window (see sections 4.1.2 and 4.2.2). The orthogonal current components were segmented into sections of 1024 (2^{10}) records, giving a fundamental frequency of 0.07 cycles per day (period of 14 days) for NWOCE_A, B, C and D (20 minute sampling interval) and 0.09 cycles per day (11 days period) for NWOCE-E (15 minute sampling interval). The segments were overlapped by 50% and windowed using a Hamming window, giving a total of four tapered segments per mooring. Statistically, this is lower than would normally be desirable, but consideration was also given to the lower frequency bands that needed resolving. The process resulted in some smearing of energy into the bands immediately adjacent to and below the fundamental frequency. The orthogonal spectral estimates at each frequency were added prior to plotting to give a total spectrum. Aliasing was not identified as a potential problem.

Low-pass filtering was undertaken by following the same methodology as for the drifters, as described in Section 4.1.3. Whilst a longer time series would increase confidence in these analyses, the month of currents are a useful corroboration of the drifter data.

5 THE FAROE BANK CHANNEL

5.1 Interpretation of Frontal Survey

The original principal objective of this study was to investigate the existence of mesoscale eddies in the FSC, and in particular to establish the role of the southern end of the Channel in the formation of such eddies. The existence of a persistent frontal region to the south of the Faroe Islands, and its propensity to shed eddies, have been reported previously by Becker and Hansen (1988). The front is variably located south of the southern tip of the Faroes Plateau and is better defined at depth than near the surface.

A detailed description of water mass distribution is given in Chapter 2, but in summary the front separates waters in the west (MNAW to some 500m and NSDW flowing out from the FSC below this), and waters in the east consisting of MNAW mixed with intermediate AI/NIW increasingly with depth. Pools or boluses of cold water are sometimes evident on the south-western extremity of the Faroes Plateau (e.g. Hansen *et al* 1998), and T-S analysis (Hansen and Jákupsstovu, 1992) shows a presence of intermediate water in such features, which in addition to being cooler are less saline than the surrounding waters. It is believed from this that they may originate in the front and that they are the principal means of upper water exchange between the Faroe Channels.

This section describes the results of the CTD and XBT survey, carried out in the south-eastern FBC on 28th-30th April 1999 with a view to establishing the frontal structure and identifying suitable drifter release locations. There is generally good agreement between the XBT temperatures (collected over nearly forty-eight hours) and those from the AVHRR image in Figure 3.6, which was taken nearly three weeks later and in the early morning.

Additionally, in response to observed cyclonic drifter motion a limited hydrographic survey was also undertaken at the south-western end of the FSC, as described in Chapter 7.

A CTD-derived potential density anomaly (σ_θ) section through the middle of the survey region, along the axis of the FBC, reveals a pronounced doming of the isopycnals between the 100 and 700db pressure levels that is most apparent in the σ_θ = 27.5 kg m^{-3} layer (panel 1, Figure 5.1). The depth of this layer (panel 2) varies across the survey region: in the FBC, the layer slopes down to the west from 50m on the upper Faroes slope to 500m in the deep FBC. Due south of the Faroes Plateau the layer depth increases from 50m in the north to around 400m in the FSC, a gradient of around 0.4° or 0.7% (c.f. the Iceland Faroes Front slope of >1-1.5%, Allen *et al*, 1994). From the configuration of the isopycnal layers in the section, the dome peaks appear to tilt through the water column at a gradient of ~1° or 1.7%. In the upper part of the water column (to 300db) the doming is coincident with what appears from the figures to be a tongue of relatively dense water extending south off the Faroes shelf.

The tongue of water has caused a marked discontinuity in the isopycnal layer depth that is also broadly spatially coincident with a similar feature in the AVHRR image of Figure 3.6. In the image, an apparently topographically-constrained protrusion of cool water (7.5-8ºC) extends south of the Faroe Islands into the deep FSC and FBC and forms part of a body of water inundating the entire Faroes shelf. As it is also evident in other AVHRR images taken at later dates (not shown here), this tongue of 'shelf' water is presumed to be a persistent feature although its extents vary between images.

The tongue of shelf water is also apparent in θ-S plots south of the Faroes shelf (Figure 5.2). From the profiles at stations 78 (FSC – see Figure 5.1 for location) and station 81 (FBC), the region was inundated by NAW with increased presence of intermediate AI/NIW in the FSC. However, the separated temperature and salinity profiles show that at the time of the survey the water at station 80 was significantly cooler and fresher than the other stations, and displayed the characteristic signature of MNAW. At the surface these differences were approximately 1°C and 0.09 salinity units, at 300db 2.5°C and 0.15 salinity units and at 500db more than 6°C and around 0.35 salinity units. Its position in relation to the presentations in Figure 5.1 suggests that at the time of the survey station 80 was located in the shelf water intrusion, which itself appears to be MNAW emanating from its resident location on the Faroes shelf.

XBT *in-situ* temperature contours at 5db, 100db and 300db across the survey area indicate that over the Faroes slope the influence of the shelf water is not confined to the surface layer (Figure 5.3). The temperature gradient increases with depth from 1.5°C (<8°C to >9°C) across the whole survey region at 5db to around 4ºC at 300db, with an associated minimum temperature at this depth of 5ºC. However, accurate demarcation of the shelf water below the surface is difficult. The discontinuity in depth contours of the σ_θ = 27.5 kg m^{-3} isopycnal layer suggests that it extends beyond CTD station 80, but CTD stations to the south and west are approximately 2°C warmer at 300db. At 550db, the θ profiles from station 80 and its immediate neighbour station 81 (on the western border of the survey region) converge at just over 1°C and are very similar below this depth. The data suggest that at the time of the survey, the shelf water was very approximately bounded in the surface layers by the 8.5°C isotherm, at 300db by the 7°C isotherm and at 550db it was detectable out to near the edge of the survey region.

In the FBC, the isotherms at 300db are broadly aligned with the bathymetry of the Faroes slope and this is reflected in XBT sections across the northern half of the survey region (Figure 5.4), by the tilting of the isotherms up the Faroes slope from deeper in the water column. The top 5db of data were removed prior to analysis, in order to remove apparent spurious surface readings from the XBT probes. The structure of the sections are interpreted in terms of the shelf water extending down the Faroes slope into the deep FBC. In the northern section to around 400db Atlantic water is fairly evenly distributed across the FBC, although the upward tilting of the 8°C isotherm at the western end suggests a slight cooling of water over the Faroe Bank. The main thermocline broadly occupies the 450db-550db layer across the Channel to the Faroes slope, where the isotherms slope up marking the presence of cooler water there. Across this section, the distribution of bottom water is biased towards the western flank.

In the middle section, the presence of cool shelf water positioned over the Faroes slope and spreading into the FBC induces a marked adjustment of the thermocline. The result is a confinement of bottom water to a deeper portion of the water column on the western side of the channel than in the northern section, whilst allowing cold water from near the bottom to sit higher up the Faroes slope. This interpretation of the section suggests that the shelf water is traceable for most of the section, whereas in the north it seems to be more confined to the Faroes slope. At the western flank, the isotherms slope upward, perhaps marking the actual limit of shelf water influence.

In the south section, the shelf water is again evident over the Faroes slope by the configuration of the isotherms, but in the western side of the section its effect is diminished if indeed the sloping isotherms do indicate the presence of shelf water. This theory is consistent with the station 80 and 81 θ-profiles (which lie on the south section) described in the previous paragraphs.

It is informative to portray the structure of the section in Figure 5.1 in terms of potential vorticity (q), in order to investigate the dynamics of the frontal region, since this parameter can indicate instability and the presence of eddies and frontal features. A requirement (but not necessarily an indicator) for baroclinic instability (i.e. the mechanism by which a perturbation to mean conditions may gain kinetic energy from available potential energy) is a change in sign of the horizontal potential vorticity gradient (Gill, 1982). Potential vorticity can be expressed in terms of static stability E (Wade *et al*, 1997) by

$$E = -\frac{N^2}{g} = \frac{q}{f} \qquad (5.1)$$

and hence

$$q = -f\frac{N^2}{g} \qquad (5.2)$$

where f is the Coriolis parameter (planetary vorticity), g the acceleration due to gravity (~9.81 m s^{-2}) and N^2 the Brunt-Väisälä or buoyancy frequency given by:

$$N^2 = \left(\frac{-g}{\rho}\right) \cdot \left(\frac{\partial \rho}{\partial z}\right) \qquad (5.3)$$

where ρ is the water density and z is the water or layer depth. Relative vorticity is thus considered negligible in relation to f in this case, a reasonable assumption in the FBC from measurements reported by Saunders (1990). Increased stratification is thus indicative of increased potential vorticity.

In panel 3 of Figure 5.1 the shelf water is evident in thickness of the σ_θ = 27.5-27.6 kg m^{-3} isopycnal layer, as a steep cleft in the contours that is coincident with the discontinuity in layer depth. At its maximum the thickness exceeds 160m and the minimum is less than 40m, indicating a gradient in potential vorticity across the shelf water protrusion. The main pycnocline is clearly shown along the section as a band of high q lying between the 450 and 650db levels (panel 4, Figure 5.1), with the highest values broadly coincident with the σ_θ = 27.8 kg m^{-3} layer. The highest potential vorticity (q = 900 x 10^{-12} $m^{-1}s^{-1}$) occurs at the northern end of the section as a result of the tight pinching of the isopycnals there, but the reason for this is unclear from the data. A secondary peak in q occurs on the eastern side of the shelf water protrusion, where again the isopycnals are closely spaced. Above and below the pycnocline q decreases quickly, although there are lobes of increased vorticity on both sides and beneath it the weak stratification has resulted in the interleaving of two low vorticity tongues with the main band.

Attempts to plot potentially useful presentations of q on isopycnal surfaces were unsuccessful due to the sloping topography and the corresponding number of available data points. Plotting potential vorticity along the cross-channel sections, to investigate (i) the region of reduced layer thickness in the west of the survey region and (ii) the maximum in q, were similarly unsuccessful.

In summary:

- The survey took place during a time when a tongue of cool, fresh water extended south off the Faroes Plateau and protruded into surrounding NAW. θ-S analysis shows that this water is of Atlantic origin (MNAW) and satellite imagery indicates that it is a persistent feature and is resident on the Faroes Plateau.

- The shelf water is characterised by temperatures typically at least 1.5°C cooler than the surrounding water.

- Although the shelf water appears to be mainly confined to the Faroes shelf/slope, the configuration of the isotherms from CTD stations on the western flank of the survey region seem to suggest that it is detectable at depth across the FBC.

- In a CTD section approximately coincident with the axis of the FBC, major potential vorticity anomalies appear to be confined to the pycnocline, with weaker stratification above and below (i.e. no anomalies associated with eddy structures were apparent). The highest anomaly is apparent at the northern end of the section.

- Despite the variation in potential vorticity gradient, there is no evidence in the presented data of instability or eddies shed from the frontal zone. Panel 3 of Figure 5.1 shows a circular structure at the western edge of the survey zone, but there is no supporting information in the temperature contours of Figure 5.3.

- The South Faroes Front itself appears to be a result of the presence of the cool shelf water, rather than any differences between the indigenous sub-surface waters of the FSC and FBC.

5.2 Geostrophic Circulation

The geostrophic relationship expresses a balance between the Coriolis force (CF), which is imparted by the rotating earth on a moving body, and a horizontal pressure gradient force (PGF) arising from lateral density differences in the ocean (through the hydrostatic relationship). The effects of friction and wind stress are therefore not accounted for. The balance is expressed in the geostrophic equation as

$$V\,2\Omega\sin\phi = -\frac{1}{\rho}\left(\frac{\partial p}{\partial x}\right) \qquad (5.4)$$

where V = speed of flow, Ω is the angular speed of rotation of the earth, ϕ is latitude, ρ is water density and ($\partial p/\partial x$) the horizontal pressure gradient. The geostrophic current exists due to the balance of the CF and PGF forces and acts perpendicular to them.

The pressure gradient force is a function of the isobaric surface slope and a parcel of water exposed to this gradient will move down it, whilst being deflected to the right (in the Northern Hemisphere) by Coriolis force. Where there are lateral changes in temperature and salinity (and hence density) the isobaric surfaces intersect the isopycnal surfaces, the two slope in opposite directions and conditions are described as baroclinic. Since the isobaric surface slopes vary with depth in baroclinic conditions, the magnitude of the geostrophic current will also vary with depth, tending to zero as the isobaric slopes tend towards the horizontal.

The dynamic method of computing geostrophic currents negates the need to compute the horizontal pressure gradient by referring the isobaric slope to a horizontal or geopotential surface, everywhere on which the value of geopotential (constant gravitational energy) is the same. The distance between two geopotential surfaces (Φ_1-Φ_2) is known as dynamic height or geopotential distance, and relates to changes in potential energy rather than simple distance (but in units of dynamic metres where 1 dyn m $\cong$ 1.02m depth). The geopotential distance is comprised of the standard geopotential distance ($\Delta\Phi_{std}$), which is a function of pressure only, and the geopotential anomaly ($\Delta\Phi$), which is function of temperature, salinity and pressure.

The form of the geostrophic equation arising from the dynamic method is

$$V_1 - V_2 = \frac{1}{L.2\Omega\sin\phi}\left[\Delta\Phi_B - \Delta\Phi_A\right] \tag{5.5}$$

where (V_1-V_2) is the vertical change (shear) in geostrophic current between levels 1 and 2, L is the horizontal distance between two stations A and B, $2\Omega\sin\phi$ is the Coriolis parameter (ϕ = latitude) and $\Delta\Phi$ is the geopotential anomaly between the two levels at A and B.

The results from equation (5.4) give only baroclinic current shear between the two levels and hence there is a potentially significant ageostrophic part of the total flow that is not represented (for example astronomical tides in the FBC are of the order 10 cm s^{-1}, Saunders, 1990). The usual practice of referring the shear to a reference level where the current velocity is either assumed to be zero or has been independently measured is difficult in the present context, due to the dynamic and topographic variability. Indeed, ADCP measurements in the FBC suggest that the no motion concept may not be applicable there (Saunders, 1990). An attempt was made in this study to refer the calculations to the drifter velocities, but the strong variability in their tracks makes this approach unfeasible. The alternative, adopted here, is to refer the geostrophic values to the seabed assuming the motion there to be zero. In regions of sloping topography this means that the reference level is more accurately the deepest common depth of each pair of stations, and hence in some cases the geostrophic currents are referred to a level slightly above the seabed, where the assumption of no motion may be false.

Minimising the spacing of the hydrographic stations between which the geostrophic currents are to be computed will reduce this potential error and increase the resolution of the results. Vertical density profiles from the three cross-channel sections (see Figure 5.4) were therefore constructed from XBT potential temperatures and CTD salinity linearly interpolated onto the depth increments of the closest XBT stations. This process therefore assumes no horizontal salinity gradient between the locations of XBT profiles merged with each CTD salinity profile and resulted in the truncation of deep profiles due to the 800m operating range of the XBT probes. The actual reference level used was therefore either the seabed or 800m, whichever was the shallower.

Despite their limitations, the geostrophic currents add a useful dimension to the previous sectional analyses and aid interpretation of the drifter tracks (as described in the next section). The sections (Figure 5.5) suggest strong horizontal shear across the FBC with the isotachs showing cores of opposing flow across the Channel. Over the Faroes slope, a weak north-westward flow (typically 10 cm s^{-1}) marks the eastern side of the survey zone. A strong central core of up to 70 cm s^{-1} is evident in all the sections. The geostrophic currents at 100db from the three sections have been vectorised and show schematically a weak north-west flow over the Faroes slope associated with the shelf water intrusion, perhaps forming part of the anticyclonic circulation around the Faroes Plateau. A strong south-eastward current is apparent in the deep channel, particularly in the middle section, which appears to be deflected offshore by the shelf water tongue. A counter-current flows north-west on the western side of the survey region, also appearing to be deflected further offshore and strongest in the middle section.

5.3 Drifter Circulation

5.3.1 Sub-mesoscale Motion

Drifter release commenced approximately five hours after completion of the frontal survey and spanned a further fifteen hours. Despite the close proximity of the release sites, the raw tracks in Appendix One show considerable variability and reveal a highly complex and variable sub-mesoscale circulatory regime, in anticyclonic and cyclonic senses, super-imposed on the larger scale flow. These rotations were of diameters ranging from 5-10 km up to 30 km. The local Rossby radius R_O, is around 9 km as given by:

$$R_O = \frac{\sqrt{g'h}}{f} \qquad (5.6)$$

where h is the upper layer depth (here 500m, chosen on the basis of CTD profiles at the time of drifter deployment, Figure 5.6) and f is the Coriolis parameter. The reduced gravity, g', is given by

$$g' = g.\left(\frac{\rho_{lower} - \rho_{upper}}{\rho_{lower}}\right) \qquad (5.7)$$

where g is the acceleration due to gravity (~9.81 m s^{-2}) and $\rho_{subscript}$ refers to the density of the upper and lower layers, here taken as 1027.65 and 1027.90 kg m^{-3} respectively. R_O is an important horizontal length scale in the analysis of baroclinic seas, as it describes the relative importance of rotational and buoyancy effects and is the scale at which disturbances to the mean conditions such as eddies grow. In the initial stages of a wave-like perturbation, the associated change in layer level is confined to a short distance and the pressure gradient is large. Gravitational forces are thus dominant. As the disturbance grows and the change in level is spread over a distance comparable to R_O the Coriolis acceleration is just as important as the pressure gradient and the growth of the disturbance is curtailed.

The sub-mesoscale variability is not evenly distributed throughout the study region. On the Scottish slope and the deep FSC, there is little evidence of drifter motion at these scales and the tracks appear much more stable. On the Faroes side of the FSC however, as in the FBC, the situation is reversed and there is a tendency towards both mesoscale and sub-mesoscale rotation. This difference is due principally to the presence of the persistent Slope Current on the Scottish side, as opposed to the weaker, apparently intermittent (see drifter 95, Appendix 1) south-west flow along the Faroes Slope and the complex dynamics of the southern FBC (see Chapter 6).

The actual causes of the sub-mesoscale variability have not been explicitly investigated here, but tides and wind-driven inertial effects are likely candidates. Anticyclonic barotropic tidal current ellipses in the FBC are reported by Hansen *et al* (1998) and it is believed that the region is subject to baroclinic tidal and internal wave activity. The strong semi-diurnal spectral peak in the drifter spectra (Figure 4.3) appears to support this. Filtering the drifter tracks removes nearly all the sub-mesoscale variability and thus clarifies the underlying residual motion of primary interest here. The distribution of the variability permits the drifter tracks to be split into two regimes for the purposes of describing them: (i) the post-deployment portion in the FBC and (ii) the remaining section in the FSC.

5.3.2 Post-Deployment Trajectories

Unless otherwise stated, all references to the drifter tracks and associated statistics relate to the filtered sets of positional and temperature observations.

The drifter trajectories over the first ten days of the deployment reveal a complex character to the region, with drifters deployed close together taking divergent paths from the outset (Figure 5.7). It is useful, therefore, to treat the period immediately after drifter release in isolation. However some coherence is also evident, in that those drifters deployed in the north and east of the region (93, 94 and 96) all remained in the FBC over the initial period, whereas the other five entered the FSC (and two, 89 and 92, returned to the FBC). The movement of many of the drifters appears consistent with the distribution of geostrophic currents in Figure 5.5.

Drifters 93, 94 and 96 were released into the frontal zone created by the cool shelf MNAW intrusion. The deployment location of the furthest north of the trio (94) was inundated with shelf MNAW, with (early morning) near surface θ-S values of 7.1°C and 35.21 (Figure 5.6). To the south at the release site of drifter 96 and to the west at 93, the effects of the shelf water were less evident near the surface, but further down the water column, the three sites shared similar characteristics.

Initially, drifter 96 moved north-west along and slightly across (into deeper water) the Faroes slope for five days at an average speed of around 10-20 cm s^{-1} (Figure 5.8). A subsequent abrupt turn to the south-west over the 700m isobath was accompanied by a rapid increase in speed to more than 50 cm s^{-1} and was consistent with the geostrophic velocity vectors in Figure 5.5. Two days later the drifter turned north-west, appearing to have entered the westernmost geostrophic core.

Drifters 93 and 94 were released spatially and temporally close together in 370m and 210m water respectively. Drifter 94 initially headed north along the Faroes slope at around 10 cm s^{-1} as expected from the geostrophic calculations. After a few hours it abruptly turned south-west and moved down the slope, increasing speed to more than 40 cm s^{-1}. The similarity of its track across the FBC similar to that of drifter 96, which followed two days after it, suggests a degree of persistence in the geostrophic circulation. The onboard thermistor recorded a gradual increase in temperature of around 1.5°C, as the drifter left the influence of the shelf water. The peak temperature coincided with the limit of the southerly excursion of the drifter: as it turned north-west and approached the Faroe Bank, the temperature dropped by 0.5°C.

After initially remaining close to its release site, drifter 93 also took a net south-westerly trajectory across the FBC, again increasing speed to nearly 30 cm s^{-1}. However, in contrast to the smooth, steady trajectories of the other two drifters in this group, its path is characterised by a series of jumps, in which the drifter abruptly turned to the north. The jumps coincided with small fluctuations in temperature (typically less then 0.5°C), and in speed (the speed approaching zero) as the drifter opposed and moved with the background flow. The data would suggest that the drifter may have encountered a cold-core eddy: certainly in the last three days of the period, it rotated cyclonically increasing speed from almost zero to 40 cm s^{-1} with a diameter of around 25 km. However, the path of drifter 96, straight across the area where the eddy may have been at the time that 93 was apparently encircling it, indicates that the feature must have been weak and possibly dissipating. As described in the next section, two other drifters also underwent mesoscale deflections in this area.

Deployed about 8 km west of drifter 93 and four hours later, drifter 95 left the release site to the south-east, apparently following the geostrophic currents located over the deep Faroes slope. However, after two days the drifter changed direction cyclonically and headed north across and up the Faroes slope, increasing in speed and water temperature as it did so (Figure 5.8). Sixty hours later (day 125), the drifter turned south again over the 400m isobath and moved away into the FSC, increasing in speed to 70 cm s^{-1} as it went. There is little evidence to explain the track of this drifter, although it is likely that its movement was influenced by the presence of strong temperature gradients associated with the shelf water intrusion (see, for example, the satellite image in Figure 3.6), which will have evolved over the ten days from frontal survey. It also indicates a transitory characteristic of the anticyclonic flow around the Faroes Plateau, in this region at least.

The remaining drifters were released further south and progressed more steadily into the FSC. Drifters 89 and 92 were released near 96, but moved away from the deployment zone to the east, rapidly increasing speed (Figure 5.9, unit 89, 60 cm s^{-1}), although for a short time 92 moved west until it reached the release site of 89. Both returned to the FBC after making a cyclonic rotation around the feature described in detail in Chapter 7. Drifters 90 and 91 were released in the confluence region of the

Faroese Channels, where it was expected that they would encounter poleward-flowing NAW and move steadily along the Scottish slope. In fact, the CTD profiles at the time of release (Figure 5.6) show that only drifter 90 entered characteristic NAW (this profile, from over the slope of the WTR was truncated at 550db). Initially, they both unexpectedly moved north for two days before entering the FSC, increasing speed as they progressed. This movement is probably a result of the strong temperature gradient that can exist between the NAW and cooler surface waters, as the satellite image in Figure 3.6 shows.

5.3.3 Longer Term Drifter Tracks

Figure 5.10 shows the drifter trajectories in the FBC from the end of day 130 until the drifters crossed the 6°W meridian on passage into the FSC. At the start of this period five drifters had left the FBC.

Drifters 89 and 92 returned after a period of five days and immediately impinged upon a cyclonic mesoscale feature with a diameter of ~30 km. Both also appeared to interact with the larger circulation prior to becoming fully entrained in it. The speed of both drifters decreased sharply (by 50%) during this interaction (Figure 5.11), with a corresponding slight fall in temperature recorded by unit 89 (not shown). The northward leg of this rotation, along the FBC, was characterised by an increase in speed to nearly 30 cm s^{-1}, but minimum speeds (10 cm s^{-1}) in the rotation occurred as they reached the northern limit. This deceleration corresponded with a rise in temperature of 0.5ºC. During the subsequent passage south the drifters again increased speed to more than 30 cm s^{-1}. Drifter 92 stopped transmitting at this time (day 137). The drifters were evidently entrained in the periphery of an apparently elongated, cold-core eddy-like structure (with a temperature anomaly at 8-8.5ºC of about -1ºC) that was oriented along the Channel and was associated with a cyclonic flow (either around the periphery or due to rotation). Examination of successive AVHRR images suggests that this feature is persistent but not permanent, and variable in precise location and dimensions, which may explain why drifters transiting the location at other times did not follow the same path. The deep water region of the southern FBC does appear to be influencing the eddy location.

The remaining drifter tracks in the FBC were characterised by rotational and meandering motions of varying scales and direction, and in the case of 94 and 96 appeared to be topographically constrained by the Faroe Bank. Their abrupt turn to the south around day 144, over the deep southern FBC may have been a result of the geostrophic circulation. Drifter 96 moved straight through the area where the mesoscale rotation of 89 and 92 took place and out of the FBC (after twenty-six days), and yet two days later drifter 94 appears to have encountered it and circulated northward again. Drifter 93 continued to exhibit a very unstable trajectory, and remained close to the Faroe Bank, finally leaving the FBC after a further 50 days. Drifter 94 remained in the Faroe Bank Channel until tracking was terminated, having moved west to beyond the Faroe Bank before abruptly returning to the FBC by a nearly reciprocal track.

6 THE FAROE-SHETLAND CHANNEL

6.1 Lagrangian Observations

Unless otherwise stated, all references to the drifter tracks and associated statistics relate to the filtered sets of positional and temperature observations.

While the drifter tracks reveal many features of the upper layer circulation, on the basis of these data alone it is difficult to isolate the underlying processes. However, combining the drifter information with other sources of data such as satellite imagery gives greater insight to the dynamics of the region, especially the apparent presence of eddy structures. The surface circulation of the FSC and the differences between it and the FBC, can to a large extent be summarised by Figure 6.1, which shows the AHVRR sea surface temperature (SST) image from early morning on the 19th May (day 139). Superimposed on the image are the drifter tracks for days 131 to 140 (red), 141 to 150 (green) and 180 to beyond day 200 in white. A mesoscale loop undertaken by drifter 91 around 10 days before the image capture is shown in purple. Although the AVHRR image represents a moment in time during the life of the drifters, it contains enough information to hypothesise on the causes of the drifter behaviour and hence the upper layer dynamics of the FSC.

The image shows warm NAW flowing along the Scottish slope (>9.5°C) as the Slope Current and cool MNAW (~8.0°C) inundating the Faroes side of the Channel. In between lies water of around 9.0°C, a mix of the two. As the drifters exited the FBC, the character of their trajectories significantly changed, becoming more stable and exhibiting deep mesoscale meanders and cyclonic rotations. There appears to be a favoured route into the FSC, with all the drifters that made the transition (over a period of more than 50 days) doing so via the deep water at the southern end of the Channel. From the figure the route appears to be determined by the distribution of NAW as it leaves the northern slope of the WTR (and hence the lateral extent meander M1 in Figure 6.1) and the spread of transit times and similarity in trajectories perhaps implies a quasi-steady upper layer exchange of water from the FBC into the FSC. Otto and van Aken (1996) show the tracks of Argos drifters deployed in the

North Atlantic between 1990 and 1993, which also followed this path from near the Faroe Bank into the FSC.

The drifter trajectories at the south-western end of the FSC are governed by a deep meander in the warm NAW (M2), which caused them to undergo a cyclonic deviation (the nature and formation of this and other meanders are investigated in the next chapter of this report). Some drifters escaped this cell and progressed further up the Channel, but two were steered into simultaneous complete revolutions. One drifter (90) made two such rotations, the first of approximately 35 km and the second of roughly 50 km diameter from which it escaped. Figure 6.2 shows that the maximum speeds in the cell (up to ~60 cm s^{-1}) occurred deeper into the trough (i.e. in deeper water nearer the centre of the Channel), and were followed in each case by a sharp reduction to approximately 25 cm s^{-1}. The temperature rose fairly steadily during this period, reaching a peak (9.3ºC) when the drifter was furthest up the Faroes slope during the second, broader rotation. This increase of 1.3ºC over a nine-day period was subsequently partially eroded as the drifter moved off the slope into the FSC. From the AVHRR image, which was taken as the drifter was on the Faroes slope completing the first rotation, it appears that after moving off the shelf during the second revolution the drifter interacted with the cold shelf water intrusion from the Faroes Plateau.

The region of the FSC upstream of M2 appears to be a zone of little net flow, where the south-westerly flow of surface waters along the FSC is buffered by the shelf water intrusion, forcing re-circulation within the Channel. However, the integrity of the cell is variable and whilst there is obvious correlation between features further along the FSC and the drifter tracks from day 180 onwards, the cyclonic cell appears to have eroded during this phase. Whether this means that the re-circulation of MNAW breaks down during such periods is not clear. If so, perhaps the variability of the FBC is influenced in part by the subsequent increased exchange of surface waters between the FSC and FBC.

Throughout the FSC, the drifters are subject to detachment from the persistent along-slope flow and to subsequent cross-channel excursions of up to 90 km in length. The temporal spread of track segments in Figure 6.1 suggests that this is a persistent feature of the surface circulation in the Channel. Those segments straddling the image acquisition (day 139) show cross-Channel movement that appears to be closely related to deep meanders in the Scottish Slope Current. Northward crossings occur on the upstream side of the meanders and return excursions on the downstream side. The track segment from day 180 and beyond shows similar character, although the apparent 'phase shift' between the crossing points and the meanders perhaps suggests propagation of the meanders along the FSC.

The segment marked A-B in Figure 6.1 represents a part of the track of drifter 95 that was significantly affected by the Slope Current meanders. The associated speed vectors (Figure 6.3) show an acceleration along the downstream side of the meander M2 to a peak in excess of 50 cm s^{-1} that was maintained along the upstream limb of a the smaller meander M3 (and accompanied here by a temperature rise of 0.75°C). The subsequent northward channel crossing induced speeds of almost 30 cm s^{-1}, at the northern limit of which the drifter turned south-west through point B in, according to the onboard sensor (Figure 6.3), water of nearly 10°C. The image in Figure 6.1 shows much cooler water in this area and hence the distribution of surface waters must have changed during the interval between image acquisition and the drifter observations.

A measure of the stability of the meander system is given by the drifter tracks in the vicinity of the meander M4. Here, two drifters (89 and 90) that were significantly influenced by the meander M2 thereafter moved swiftly along the Scottish slope, passing M3 and then M4 just five days after another drifter (95) had crossed the Channel from there. The cause of this behaviour is not immediately clear, but as will be shown in the next chapter there is considerable variability associated with the development and lateral extent of the meanders, and perhaps the NAW was more confined to the Scottish Slope at this time.

On the Faroes side of the FSC, the circulation is complex. The lateral extent of the cross channel excursions is dependent not only on the development of the NAW meanders but also the distribution of MNAW. The crossing towards the Faroes closest to the north-eastern end of the FSC, undertaken just before the AVHRR image in Figure 6.1 was acquired, was sharply curtailed as the drifter (91) encountered MNAW flowing anticyclonically round the Faroes slope. The drifter was abruptly deflected to the north-east for three days before becoming entrained in the flow across the end of the FSC. More than 60 days later, the later stages of the track of drifter 95 is strikingly similar (shown in white in Figure 6.1). However, as the raw trajectory plot in Appendix One shows, prior to this time the drifter remained on the Faroes slope for a many weeks, moving erratically along the isobaths and over the shelf. From these tracks and the water mass analysis of Chapter 2, it appears that there is a northward flow on the Scottish side of the interface between the MNAW and surface water over the deep FSC, whilst on the Faroes side the reverse prevails.

Drifters 91 (shown in purple in Figure 6.1) and 95 (in white) displayed distinctive loops in regions where, historically, eddies have been observed. Drifter 91 increased speed to around 60 cm s^{-1} on the northward part of the loop slowed as it transited the upper limb of the loop and accelerated again as it rejoined the main along-slope flow (Figure 6.4). There was little evident variation in temperature (not shown) associated with the rotation. It is possible from the AVHRR image, which was taken some ten days after the period in question, that the rotation was caused by its interaction with the meanders in the Slope Current, rather than a discrete eddy structure such as that visible due north of it.

Other than the size of the loop there is little evidence to support interaction of drifter 95 with an eddy, indeed as the drifter entered the loop the temperature (not shown) rose slightly (<0.5°C) and the speed decreased slightly. The drifter accelerated until the it reached northern limit of the loop, slowed and then increased again to around 40 cm s^{-1} as it moved around the western limb of the rotation. Once clear of the loop, the drifter accelerated to over 60 cm s^{-1}. A poor, partial AVHRR image (not shown) taken at the same time suggests a tongue of warm (11-12ºC) water on the extending towards the Faroes side of the northern Channel, and hence on this occasion the looping motion may have resulted from the drifter moving across the interface between this and the cooler water over the Faroes slope. However, the onboard sensor did not capture the expected variation in temperature associated with this.

6.2 Statistical Description of the Upper Layer Circulation

Analysis of the drifter tracks undoubtedly gives great insight into the upper layer circulation, but to investigate energetics and important general trends over large areas the processed drifter data have been synthesised into boxes, each ½º in latitude by 1º in longitude. Due to the low number of drifters deployed, the box size was based not only on resolution but also on numbers of available drifter observations. The statistical analyses based upon the zonation must therefore be viewed in conjunction with some measure of the population in each box. The statistics of each box are given in Appendix Three. Figure 6.5 gives the box residence time in days, based on the total number of drifter satellite positions falling in each box (i.e. the number of drifter days), normalised by the number of drifters entering it. The residence time is thus a reflection of both the propensity of the drifters to populate a given box and also the drifter velocity (the higher the velocity the fewer the observations).

Vectors of mean drifter velocity for each box are shown in Figure 6.6. Also shown in the Figure is a directional stability ratio, *D*, computed from

$$D = \frac{\text{Vector mean speed}}{\text{Scalar mean speed}} \qquad (6.1)$$

A ratio of $D = 1$ indicates that the drifter moved in a straight line for the duration of the observing period and low ratios are an indicator of strong directional variability. *D* is thus a measure of the confidence that can be placed in the vectors for interpreting circulatory features.

While the process of zoning smoothes out much of the detail embodied in the drifter tracks, the figures do reveal several interesting features of the residual circulation. The Slope Current bringing NAW into the FSC is clearly evident from the high mean speeds (~70 cm s^{-1}), low residence times (0.2 days) and high *D* (>0.90) along the Scottish slope. Residence times generally increase with depth towards the central FSC, as distance away from the main axis of the Slope Current increases, and corresponding mean speeds are seen to decrease. At either end of the Channel the vectors give an impression of an anticyclonic circulation cell, the cross-Channel component of which is relatively strong and at the northern end is associated with maximum D (1.00, speed ~0.45 cm s^{-1}). The longest residence time (13.6 days) occurs to the west of the Faroes Bank, where a drifter was apparently stalled. On the Faroes slope to the south-east of the islands the residence time of 12.8 days is due mainly to the presence of one drifter for more than a month. The low net flow and inherent variability in the drifter release area is highlighted by the high residence times (around ten days on the eastern slope of the Faroe Bank), low mean speeds (4 cm s^{-1}) and low D (<0.2), although the circulation is more persistent around the northern extremity of the Bank (D = 0.46). The vectors suggest a weak background anticyclonic circulation around the Faroe Bank (also observed by Otto and van Aken, 1996, Hansen *et al*, 1998 and Saunders, 1990) and a cyclonic flow around the patch of deep water in the central southern Faroe Bank Channel.

Figure 6.7 shows eddy kinetic energy (EKE) derived from the zoned drifter speeds using the relationship

$$\mathrm{EKE} = 0.5[u'^2(t) + v'^2(t)] \qquad (6.2)$$

where $u'(t) = [u(t) - \bar{u}]$ and the overbar denotes the mean.

Peak EKE (700 $cm^2\ s^{-2}$) occurs at the southern end of the FSC, coincident with meander M2 in the NAW (Figure 6.1). Another secondary peak (600 $cm^2\ s^{-2}$) is located close to the position of meander M3 at the northern end of the FSC. Values are generally low over the Faroes slope. To the west of the confluence region, EKE decreases poleward through the Faroe Bank Channel from a peak of 400 $cm^2\ s^{-2}$ over the deepest section of the Channel, where there is very low directional stability.

7 SLOPE CURRENT MEANDERS

7.1 Background

The preceding chapters of this report have highlighted the existence of a series of four deep (>40 km) meanders in the Scottish Slope Current, with an impression of a fifth at the north-eastern end of the FSC. The existence of these features hitherto has only been inferred (e.g Sherwin *et al* 2000, Oey, 1998), but the interaction of the drifters and the corresponding statistics indicate that these features play a key role in the upper layer dynamics of the FSC.

Meanders are commonly observed in strong jet-like boundary currents, for example the Florida Current (Chew and Bushnell, 1987), the California Current (Reinecker and Mooers, 1989), the Norwegian Current (Ikeda *et al*, 1989), to name a few. The Iceland Faroes Front shares common attributes (such as meander length scales) with the Slope Current in the FSC (Allen *et al* 1994). Their generation is complex and quite site specific and has been the subject of numerous laboratory studies (e.g. Killworth *et al*, 1984). In the FSC, the topography over which the incoming NAW flows is irregular, the water mass distribution appears highly variable and persistent strong winds are a common occurrence there. These are all factors that may contribute to the generation of mesoscale meanders like those in the AVHRR image, and this chapter examines their nature, hypothesises on their generation and shows how they can promote the formation mesoscale eddies.

7.2 Physical Dimensions

In later sections of this chapter the developmental variability of the meanders is explored, but in Figure 6.1 (from 19th May 1999) the first of the meanders (M1) is situated about 70 km from the WTR; the wavelengths of the subsequent features is also around 70 km. The lateral excursion from the 500m isobath (the main axis of the Slope Current) is more variable: 65 km in the case of the first two meanders (M1 and M2), 50 km for the third (M3), around 80 km for the fourth (M4) and 25 km for the fifth (M5). In M2 there is evidence of a strong rotary circulation from the ring of warmer water (~+0.5ºC) extending out almost to the tip of the meander.

7.3 Eulerian Observations

The southeastern half of the NWOCE ADCP section lies in meander M2 (Figure 7.1, shown with the AVHRR image from 19th May, day 139), and hence it may be expected that the month of currents will reveal some pertinent features of the meander. Spectral estimates of unfiltered ADCP currents (Figure 7.1) for the period 6th May to 6th June 1999 show high levels of long period spectral energy, highest in the middle FSC at mooring C, but also evident at the other moorings across the FSC. The spectra represent the sum of the orthogonal spectral estimates; see Chapter 4 for a description of the techniques used. The low frequency variability represented in the spectra (around 10 days period) appears to be mainly confined to the upper 400m of the water column at the deep water sites, giving an impression of a baroclinic nature. Note that a longer current time series is needed to maximise confidence in low frequency spectral estimates.

Current vectors for four near-uniform depths at sites A, B, C and D are given in Figure 7.2, based upon the 30 hour filtered currents averaged over twenty-four hours. There is good directional coherence through the upper layers (whole water column at A) at each site and little vertical shear, but the time series at the three upper levels are strongly rotational suggesting the passage through the site of transient features. At B and C there is an impression of a ten day periodicity to the rotations, which is broadly consistent with the spectra. Stations A and B lie within the mesoscale revolution undertaken by drifter 90, as described in Section 6.1. The drifter was closest to A on days 131 and 140 and B on day 136-137 and the drifter speeds in Figure 6.2 are in good agreement with the corresponding ADCP currents, to within ~10 cm s^{-1}.

The passage of meander M2 through NWOCE-C is evident in the cyclonic rotation of the upper layer vectors between days 132 and 142. The leading edge of the meander passed through the site in the first three days of this period (vectors oriented to the south-east), during which time the current speed increased to around 40 cm s^{-1}. From the north-west orientation of the vectors, the trailing edge appears to have arrived on day 141. Close examination of the vectors at NWOCE-C and D during this period also seems to confirm the anticyclonic circulation within the meander itself, which has the appearance of a warm-core mesoscale eddy. Assuming that the meander took ten days to pass through NWOCE-C, an approximate phase speed for the feature is 6 cm s^{-1} (on the basis of a 50 km length scale).

Vertically, the spectra and current vectors suggest that the meandering flow is confined to the upper 400m. It is interesting to note that after day 143, the trough following M2 was associated with a period of strong current at B, C and D, reaching nearly 80 cm s^{-1} at C and directed generally across the FSC to the north or northwest. At all three stations, this persisted through the water column into the deep layers and was followed on day 150 by a reversal in direction. At B and C this was confined to the deep layers, but at NWOCE-D it was evident through the water column and strongest nearest the bed. This reversal is indicative of a pulse of bottom water moving south, but if that is the case it is coupled in some way to the Slope Current reversal in the upper layers at station D.

7.4 Satellite Observations

7.4.1 Sea Surface Temperatures

The importance of the meanders to the dynamics and variability of the FSC is demonstrated by the preceding discussion. To further investigate their inherent variability Figure 7.3 shows a chronological sequence of AVHRR images taken several weeks apart, between April and August 1999. Panel 2 of the figure is the same AVHRR image as in Figures 6.1 and 7.1. All the images except the first, on April 23rd, are taken in the early morning. Seasonal warming is evident, especially in Panel 4 (4th August) and although the temperature scales are not consistent between images, the differences in scaling are uniform across the whole temperature range thereby preserving the water mass distribution throughout the sequence.

Although the images are dominated by the presence of the warm NAW, there is considerable variation in the development of the meanders. In panel 4 (4th August), the Slope Current forms a coherent ribbon-like jet along the Scottish slope. There is little meandering and the interface between the NAW and cooler MNAW tends to undulate. Conversely, in panel 3 (17th June) the meanders extend across the FSC and onto the Faroes Slope. The presence of meanders M2, M3 and M4 (day 139, 19th May) is still apparent and broadly consistent with the 19th May positions, but over the Scottish slope they have coalesced and at the their extremity the flow is considerably more chaotic, with extensive filaments present over the Faroes Slope. Panel 1 also shows extensive lateral development, especially of M2 and M3. M1 is obscured by cloud, but is probably inundated with NAW. Cold-core eddies are clearly visible in M2 and (a larger structure) in M4, the formation of which is discussed in Chapter 8.

7.4.2 Sea Surface Height Measurements

Provided the features are sufficiently intense to produce SSH anomalies that are within the sensor detection range and are of sufficient size to be resolved by successive satellite tracks, information from altimeters provides a successful means of monitoring mesoscale variability. Taking the analysis a step further it is possible to compute geostrophic currents (V_g) associated with the variability directly from the SSH anomaly data (Wade and Heyward, 2001 and Samuel *et al*, 1994), since the change in sea-surface height between two stations is directly proportional to the geostrophic flow between them:

$$V_g = \frac{g \tan i}{f} = \frac{g \Delta H}{fL} \qquad (7.1)$$

where g is the acceleration due to gravity, f is the Coriolis parameter, i the angle between the sea surface and a level surface (geoid) and L the horizontal distance between the stations. The resulting geostrophic currents are thus comprised of barotropic and baroclinic components.

On 20th May 1999 T-P flew a track that ran along the Scottish Slope, and using the above relationship surface geostrophic currents associated with the Slope Current meanders were computed and overlaid on the frontal mosaic scene for 18th to 20th

May of temperature gradients based on AVHRR data[1] (Figure 7.4). Despite the time difference, there is a clear correlation between the meanders and the geostrophic currents, which peak at nearly 40 cm s^{-1} in M2. This value is in good agreement with the ADCP vectors at NWOCE-C (Figure 7.2).

7.5 'Eddy' Survey

7.5.1 Rationale

Near real time monitoring of the drifters showed that two (89 and 92) were released into a south-easterly flow that took them towards the confluence of the Faroese Channels (Figure 6.1). After three days they were observed to follow a cyclonic rotational trajectory for a similar period, after which they moved back towards the FBC. Based on information relayed to the *'Scotia'*, a fine resolution CTD and XBT survey pattern was devised to determine whether the motion was attributable to eddy activity. Figure 3.2 shows the measurement area, which lies at the southern end of the Faroe-Shetland Channel in the eastern part of the frontal zone. The main survey line lay approximately co-incident with the axis of the FSC for around 53 km, and in addition two orthogonal legs each around 38 km long were surveyed. CTD casts were made approximately every 10 km, with additional XBT launches in between.

7.5.2 Drifter Behaviour

From day 124, as the drifters entered the survey region they accelerated as they progressed along the southern limb of the cyclonic rotation (Figure 7.5) to between 20 and 35 cm s^{-1}. As they turned first north and then west, they slowed to perhaps half this maximum, before increasing speed as they left the area. The temperature sensor on drifter 89 indicated no more 0.5°C variation throughout the 5-day period.

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[1] Chapter 3 has details of the techniques used to produce the image. The principal features associated with strong thermal contrasts (i.e., the shelf water intrusion, the Slope Current meanders and a cold-core eddy in the central FSC) are clearly defined in the figure.

7.5.3 Spatial Structure

Depth slices of potential temperature across the area reveal a depth-dependent coherent structure, the signature of which is less apparent in the upper layers (Figure 7.6). The steep temperature gradient gives an impression of a sinusoidal boundary zone, aligned approximately south-west/northeast and separating warm water (>9.0ºC near surface) to the southeast from cooler water nearer the Faroes Plateau (~8ºC). The horizontal meander amplitude (~8 km) is deepest at 300db, where the wavelength is of the order 12 km (c.f the local Rossby radius of 10 km) and the temperature gradient is 3.8ºC<θ>6.4ºC. At 500db the boundary appears to have flattened out.

An interesting feature of the structure is the horizontal variation in meander position at the different levels through the water column, giving the appearance of a phase shift with the configuration at 300db apparently nearly 180º out of phase with the layers above and below. A vertical phase shift between layers is expected in a baroclinically unstable wave in channel flow (Ikeda *et al*, 1989), but such instability is difficult to prove without evidence of the phase shift of the associated horizontal wave modes. If the phase shift is real, some contribution of baroclinic instability is indicated, but further data from further afield are required to confirm this.

Geostrophic currents relative to the seabed or 800m, whichever is the shallower through the two cross-channel sections (Figure 7.7), show in the western section a coherent cyclonic circulation with westerly currents to the north of the and the opposite embedded in and to the south of it. There is also evidence of a weak counterflow further to the south. The strongest currents (>60 cm s^{-1}) lie at the southern end of the section. The cyclonic structure is present in the eastern section, but due to the wider spacing of the isotherms is less well-defined.

θ-S analysis of the CTD stations along the western section (Figure 7.8) shows that the station at the southern end of the section (CTD 173) was in NAW (surface θ-S 9.3°C, 35.38), whilst to the north the other stations lay in cooler fresher MNAW. Although the spatial distribution of sampling points at the extremities of the survey area is such that the horizontal temperature contours maybe misleading, at 300db Station 170 lies in what appears to be a coherent closed structure and the θ-S profiles show cooler, fresher water at this location and level. It is possible therefore that the pattern of geostrophic currents is due to the presence of a sub-mesoscale eddy structure embedded in the boundary zone, which in turn would seem to partition a NAW meander from MNAW on the Faroes side of the FSC. The relationship between the meanders and eddies is discussed in Section 8.4.

8 MEANDER FORMATION AND PROPAGATION

8.1 Theoretical Aspects

Typically, meanders in a current are an indication of dynamic instability. Such instabilities may be baroclinic, where the source energy for a disturbance is available potential energy and hence the process is reliant on vertical velocity shear. Alternatively the causative process may be barotropic instability, wherein there is no vertical shear and the source energy is associated with horizontal variations in velocity (and hence is kinetic in nature). Finally, the instability may be a mixture of the two. Griffiths and Linden (1981) performed laboratory experiments to generate buoyancy-driven flows in a rotating container. The currents developed wave-like disturbances that grew in amplitude and eventually formed eddies. Their experiments show that the disturbances are primarily barotropic when ($h/H < 0.1$), where h is the depth of the current and H the total depth. When the fractional depth of the current is large the disturbances are predominantly baroclinic. Huthnance (1986) suggests that south of the WTR energy transfer is predominantly barotropic, but tests for baroclinicity (described in the next section) indicate that is not the case in the FSC. From the current spectra in Figure 7.1 the above condition for baroclinic transfer appears to be satisfied.

The presence of available potential energy in a system is not sufficient to ensure instability and a necessary condition for instability to occur (Gill, 1982) is that the potential vorticity gradient must not have the same sign everywhere and should have positive and negative values. In the case of the FSC, potential vorticity (eqn. 5.2) derived from the Nolsa-Flugga and Fair Isle-Munken sections provides a basis for determining whether this condition is satisfied, and the resulting sections in Figure 8.1 appear to confirm that this is the case.

8.2 Application of an Instability Model

Idealised mathematical models predict wavelengths and phase speeds of instabilities, the results of which may then be compared to observations, and hence they can indicate which processes prevail. Such models often employ simplified stratification by assuming a two layer water structure and either uniform or no bathymetry.

Killworth *et al* (1984) developed a two layer model, the configuration of which is shown in Figure 8.2 and consists of an upper layer that vanishes at some point in the y axis direction, forming a surface front. The bottom layer was effectively stagnant. The governing equations are summarised below, as succinctly elucidated by Allen *et al* (1994), who applied the model to the Iceland-Faroes Front. Assuming uniform potential vorticity in the upper layer, the wavelength of the fastest growing mode is

$$\lambda_f = \frac{2\pi R_o}{\varepsilon_{\max}} = 5.5 R_o (r-1)^{1/4} \tag{8.1}$$

with corresponding dimensional growth rate (frequency)

$$(f\varepsilon c_r)_{\max} = 0.13 f (r-1)^{-3/4} \tag{8.2}$$

where R_o is the Rossby radius (defined previously in eqn. 5.6),

$$R_o = (g'H)^{1/2} f^{-1} \tag{8.3}$$

H is the thickness of the upper layer, f is the Coriolis parameter, g' is the reduced gravity, rH is the total water depth (hence r is the ratio of total water depth to upper layer depth) and c is the meander phase speed with imaginary and real parts c_i and c_r.

In addition, the phase speed may be found from

$$c = fR_o \left[0.13(r-1)^{-3/4} \right] \tag{8.4}$$

Table 8.1 summarises the input parameter values for the FSC. Density values and layer depth were chosen on the basis of the potential temperature section in Figure 2.2

in Chapter 2 and the corresponding potential density section (not shown). The 27.5 kg m^{-3} was selected to represent the boundary between the upper and lower layers.

Table 8.1 – Input parameters to Killworth *et al* (1984) baroclinic instability model

Input		Results	
Parameter	**Value**	**Parameter**	**Value**
H	350m	Wavelength	52.3km
rH (total depth)	900m	Period	6.2 days
r	2.57	Phase speed	10.1cm s^{-1}
f	0.000128rad s^{-1}		
Upper layer density	27.65kg m^{-3}		
Lower layer density	28.00kg m^{-3}		
g'	0.00334m s^{-2}		
R_o	8.5km		

The wavelength scale compares favourably with the observations, although the period is shorter than expected on the basis of the ADCP and AVHRR data. The model also gives a reasonable comparison of phase speed with that derived from the observations. Clearly, from the AVHRR images there is a range of depths that could be used in the model and so it is worthwhile to test its sensitivity to various input depths and layer thickness. As the section in Figure 2.2 shows, the isotherms (and hence isopycnals) slope down sharply over the Scottish Slope and so the boundary between the upper and lower layers is kept constant at 27.5 kg m^{-3} during the test.

This simple sensitivity analysis (Figure 8.3) highlights the relationship in Eqn 8.1, that the wavelength will increase with larger r, although for the range of depths and the density distribution in the FSC the fitted (cubic) curve tends towards asymptotic with higher r. Note the very small sample used. In the case of the FSC, increasing r implies a shift of the main axis of flow offshore. Model studies (Oey, 1998) have shown that this occurs during the passage of meanders in the FSC and appears to be confirmed by the AVHRR images in Figure 7.3, and hence it may be inferred that in the absence of other forcing mechanisms the wavelength will increase as the meanders deepen.

In the computations of Killworth *et al* (1984) the scale of c (the meander phase speed) is dependent not only on r, but also on the mean flow. In the FSC therefore, where the mean flow does vary due to the interaction with other near and far field dynamics, phase speeds are unlikely to compare well with modelled estimates. Furthermore, Allen *et al* (1994) suggest on the basis of further model computations that the effect of topography with the opposite sign to the isopycnal slope (as in the case of the front between the Slope Current and the MNAW) is to reduce the wavelength from that predicted above.

In computations of baroclinically unstable waves using a one-dimensional continuously stratified model, Johns (1987) determined that mode dispersion curves were highly sensitive to uncertainties in the vertical vorticity profile, especially changes in sign. By assuming a uniform potential vorticity distribution in the upper layer, the Killworth (1984) model reduces the potential for barotropic modes of instability and hence it is important to establish the contribution of the barotropic component. Killworth *et al* (1984) analytically derived an expression to describe the total perturbation energy E integrated in the y (zonal) direction:

$$\frac{d}{dt}\int_{-\infty}^{0} E\,dy = \frac{\varepsilon c_i}{2}\left\{\int_{-\infty}^{0}\bar{h}\left(\frac{\partial \bar{u}}{\partial y}\right)^2 dy + \int_{-\infty}^{0}\overline{u^2}dy\right\} \qquad (8.5)$$

where the first term represents the barotropic contribution and the second the baroclinic component, the overbar denotes the mean state and h is the upper layer depth at distance y. From Allen *et al* (1994), by re-arranging Eqn. 8.5 it is hence possible to determine the relative contributions of the horizontal and vertical shears through:

$$\frac{\int_{-\infty}^{0}\bar{h}\left(\frac{\partial \bar{u}}{\partial y}\right)^2 dy}{\int_{-\infty}^{0} g'\left(\frac{\partial \bar{h}}{\partial y}\right)^2 dy} \qquad (8.6)$$

Assuming geostrophic balance

$$g'\left(\frac{\partial \bar{h}}{\partial y}\right) = f\bar{u} \qquad (8.7)$$

the energy partition is of the order

$$\frac{g'H}{f^2L^2} = \frac{R_0^{\ 2}}{L^2} \qquad (8.8)$$

Putting L = 30 km (based on AVHRR images), the barotropic energy contribution is thus approximately 13%.

From their rotating tank and theoretical experiments, Griffiths and Linden (1981) found that the instability was likely to be predominantly baroclinic if the current width is much greater than the Rossby radius, a condition that is satisfied in the FSC assuming a width of 30 km (from AVHRR images). In their model the onset of instability occurs if

$$6 < \frac{f\lambda}{(g'h_1)^{1/2}} < 30$$

where λ is the wavelength of the (dominant) disturbance and the other variables take the values in Table 8.1. Using a wavelength of 60 km (from the AVHRR images) gives a value of 7 and hence this condition is also satisfied. Griffiths and Linden (1981) derived an expression for phase speed

$$c_r = \frac{(\gamma U_1 + U_2)}{(1+\gamma)} \qquad (8.9)$$

where γ = ratio of the upper layer current and lower layer depths from Table 8.1, U_1 and U_2 are the along stream velocities of the upper and lower layers respectively, set from the Nordic WOCE data in Figure 7.2 to 40 cm s^{-1} and 20 cm s^{-1}. This configuration gives a phase speed of 12 cm s^{-1}, which is good agreement with both the observations and the estimate from the model by Killworth *et al* (1984).

There is a plethora of other models in the literature, most of which do not include some important physical attribute of the FSC, such as a true representation of bathymetry or stratification. Whilst the models use a simplistic approach to what is undoubtedly a highly complex regime, they do however give an insight into the some of the governing dynamics. On the basis of available evidence, it seems that the Slope Current is (primarily) baroclinically unstable to a perturbation (possibly initiated by passage over the WTR) and the configuration of the FSC enables the disturbance to grow at a wavelength of around 50-60 km. Whether as the models suggest there is some critical width the Slope Current must reach, or some critical value of vertical shear, before the meanders develop requires further investigation. From the AVHRR images in Figure 7.3 and others not shown, when the Slope Current is confined to the upper Scottish slope there is little lateral development. However, it is not immediately clear if the reduced width of the current is inhibiting the meander development or is itself a result of the absence of some other mechanism essential to that development.

8.3 The Role of the Wyville Thomson Ridge

The topography of the WTR is such that the Ridge acts as a sill to the FSC, and as such plays a prominent role in the meander dynamics. However, in the absence of measurements from close to the WTR, recourse is made to AVHRR images and the literature to hypothesise on what the role may be, although finding an absolute solution to the problem will require measured data from the vicinity of the WTR.

The images in Figure 7.3 (and others not included) show warm NAW flowing over the WTR and intensifying to form a coherent ribbon along the Scottish slope, characterised by a 'street' of regular meanders that vary temporally in lateral extent. Observations reported by Huthnance (1986) show the Slope Current south of the WTR following the bathymetric contours and speeding up to maintain transport between converged contours. As the Current flows over the WTR the resulting acceleration is accompanied by an anticyclonic deflection, evident in the AVHRR image of 19th May. Consideration of the conservation of potential vorticity (*PV*), given by $(f+\zeta)/d$ (where f is the Coriolis parameter, ζ the relative vorticity and d the isopycnal spacing) means that as the flow moves off the northern slope of the WTR into deeper water, f and d increase requiring an adjustment of ζ. In this case the adjustment is negative. This interaction between the topography and the mean flow over it (topographic steering), and the subsequent re-distribution of vorticity may be sufficient to generate the meanders.

Results described in Oey (1998) from a primitive equation model of the FSC also show the development of a meander just downstream of the WTR, which in the experiments is always accompanied by a change in sign in the cross-slope potential vorticity (*PV*) gradient. *PV* was seen to peak on the inshore side of the meander, close to the 500m isobath and nearly coincident with the Slope Current axis. Such positioning is consistent from the AVHRR images with the intensification of the slope current over the Scottish slope downstream of the WTR, which apparently triggered the modelled meanders. It is interesting to note that the model predicts a meander scale of about 20 km and periods of 5-10 days, the latter at least agreeing with computations made in the previous section.

Close examination of useable AVHRR images shows that despite the elapsed time between the images, the lateral extent of the meanders and especially the estimates of phase speed along-Channel meander positioning varies little. Drifter tracks from Sherwin *et al* (1999) appeared to follow a deflected track consistent with an encounter with meander M4 – these drifters were deployed in 1996. The calculations in Section 8.2 predict a wavelength of 50-60 km, shown by the AVHRR images to be a reliable estimate, but they also show that the first meander (M1) is consistently formed about one wavelength from the WTR, which acts a sill at the end of the FSC. Assuming that this is not coincidence, the behaviour of the meanders is consistent with some form of standing wave with initial forcing being provided by the Ridge.

An initial impression is of a lee-wave scenario, with a series of regularly spaced waves downstream of a topographic obstacle, as sometimes observed downwind of terrestrial mountain ranges. The response of a flow over elevated topography depends on the width of the topography and on this basis Gill (1982) identifies five flow regimes over a bell-shaped ridge. Of interest here is the case of $L>>U/|f|$, where L is the width of the ridge (~20 km from Figure 1.1), U is a uniform mean flow (50 cm s^{-1}) and f the Coriolis parameter (0.00012 rad s^{-1}), giving $U/f = 4.2$ km. In this case the solutions are evanescent and so waves are not produced, suggesting that lee waves are not responsible for the quasi-stationary appearance of the meanders.

8.4 Eddy Generation

In panels 1 and 3 of Figure 7.3 well-developed Slope Current meanders extend across the FSC to the Faroes slope. At their extremities meanders M2 and M4 in panel 1 and M3 and M4 in panel 3 have developed into filaments that appear to curl round against the expected direction of propagation towards the WTR. Within these structures pools of cold water have been cut off from the main body of MNAW, forming discrete cold-core, cyclonic eddies that have temperatures some 1º-1.5ºC colder than the surrounding NAW. The 'eddy' survey described in Section 7.5 took place very close to the location of the eddy formed by meander M2 in Panel 1.

This 'backward breaking wave' phenomenon has been observed in nature (e.g. Johannessen *et al*, 1989, Mertz *et al*, 1988 and Oey, 1988) and described in model studies (Griffiths and Linden, 1981, Oey 1998). Griffiths and Linden (1981) simulated a boundary current by continuous release of a buoyant fluid into an anticyclonically-rotating tank. As wave-like disturbances grew on the upstream side of the current, cyclonic vorticity accumulated behind the crests eventually producing closed zones of cyclonic motion, while the flow remained anticyclonic in the wave crests. Each wave thus comprised an asymmetric dipole. Figure 8.4 shows the well-developed instabilities produced by these experiments. The growth of the disturbances was observed to have a finite limit, suggested by the authors to be due to friction effects. The final current was diffuse with large spatial variations in density and velocity (c.f panels 1 and 3 of Figure 7.3).

There are strong similarities between the AVHRR images and results of the experiments, which may thus explain a major means of eddy generation in the FSC. The meanders do have the breaking wave appearance, but only when mature, and they do appear to be generating eddies at this stage. Once fully developed these eddies presumably propagate along the FSC with the mean flow and hence they are responsible for entraining cool MNAW and AI/NIW water in the warmer Slope Current.

The most detailed observations of mesoscale eddies in the FSC are from Sherwin *et al* (1999), who investigated such an eddy at the position of M2 in Figure 7.3, using *in-situ* currents and temperature measurements, satellite imagery and vessel mounted ADCP and thermosalinograph data. This cold core eddy had a diameter of 50 km and appeared to be attached to the 800m isobath. It propagated northeastward at about 8 cm s^{-1} and was associated with surface current speed of 50 cm s^{-1}. Its passage was evident at 300m depth, giving some indication of the vertical scale these features in the Channel. The observations also showed that warm Slope Current water was swept cyclonically round the north-eastern edge of the eddy.

Contrary to the conclusion by Sherwin *et al* (1999) that the eddy was not related to instability of the front along the deep water side of the Slope Current, when considered in conjunction with the findings of the present study the evidence actually suggests that this in fact was the case. The accompanying satellite image (not shown here) reveals the (fragmented) Slope Current meanders to be in a dissipation phase: there is a remnant of M4 close to it's position in Figure 7.3 and a suggestion of M2, although this is partially obscured by cloud. The consistency of the dimensions and phase speed of the eddy with those of the meanders, its location relative to M2 and the entrainment of NAW round the downstream limb of the eddy indicate that it was an artefact of this meander.

9 DISCUSSION

9.1 Circulation and Exchange in the Faroese Channels

9.1.1 The South Faroes Front

The original principal objective of this study was to investigate the existence of mesoscale eddies in the FSC, and in particular to establish the role of the southern end of the FSC in the formation of such eddies. The premise for this was the existence of the front that lies to the south of the Faroes Islands, and which separates the surface waters of the FSC and FBC. The drifters have revealed a wealth of information about the physical regime in the Faroese channels, which varies significantly throughout the region. Broadly speaking, the circulation in the FSC occurs on a scale of many days and several tens of kilometres. In the FBC, however, both the spatial and temporal frames tend to be smaller.

The detailed hydrographic survey of the south-western end of the FBC showed that the 'front' is a tongue of MNAW, perhaps further modified over the Faroes Plateau and extending south off the shelf with steep temperature gradients between it and the surrounding water, especially at depth. It marks the confluence of: (i) NAW/MNAW from the FBC; (ii) NAW/MNAW from the FSC, variably mixed with intermediate water; (iii) MNAW that has been resident on the Faroes shelf. From AVHRR images, the positioning of the tongue is variable and appears to be strongly topographically influenced by the shelf break. Some images show an inundation in the FBC by warm NAW and hence the surface temperature gradient across the western flank is stronger (up to 3ºC over a few kilometres) than that surveyed in 1999. This is possibly a seasonal effect, and further investigation is required to establish conditions in the winter months. The southerly excursion can impinge sufficiently far into the FSC to influence the path of the incoming NAW over the WTR, and hence is possibly important in the generation of Slope Current meanders. The presence of the shelf mode is indicative of a lack of exchange between the Faroes shelf and deeper surrounding waters, although the water is traceable at depth across the FBC.

From the trajectories of drifters passing between the FBC and FSC (and back again in some cases), the role of the shelf water intrusion is primarily one of steering the exchange of upper layer water. The drifter tracks suggest that the intrusion forces the (partial) recirculation of MNAW in the FSC, and hence when it is well-developed off the southern tip of the Faroes Plateau it may inhibit the inter-channel exchange of water that from the tracks does occur, both to and from the FBC. A repeat of the frontal survey may reveal a different regime, and hence the results described in this report represent something of a snapshot. An important question relates to the time scales of variability in the FBC, which is less-well studied than the FSC.

On the FBC side of the shelf water intrusion, there is a tendency towards sub-mesoscale motion of the order of the local Rossby radius, possibly attributable to barotropic and baroclinic tides and inertial effects. The circulation is complex and variable, and there was little net movement through the region with many of the drifters remaining in the FBC for several weeks before progressing into the FSC. This is indicative perhaps of the intermittence of the surface water exchange between the channels. Geostrophic currents are an important circulatory component in the vicinity of the shelf water intrusion and had a strong influence on the drifter trajectories.

There is little evidence in the data to suggest that eddies are being generated by the front, as reported by Hansen *et al*, (1998). Indeed the boluses of cool water they describe on the Faroes slope may in fact have been attributable to shelf water although this cannot be substantiated from the present study. Despite strong vertical density gradients, baroclinic instability cannot be demonstrated through the application of a model such as that described in Chapter 8, as predicted wavelengths are inconsistent (larger) with the length scales of the shelf water intrusion.

The potential fate of any eddy forming along the surveyed western front of the shelf water intrusion is unclear. The tendency for some drifters to move back into the FBC from the FSC means that any eddies formed in the frontal area could theoretically propagate in either direction. It is likely that such structures transiting the FBC would not travel far along the channel and would become trapped or dissipate through frictional effects from the complex bathymetry. Any entering the FSC would become entrained in the NAW and move along the Scottish slope. However, there is no evidence to suggest that either occurred during the field surveys for this study, nor in any AVHRR images examined. In any case, the flow patterns appear too chaotic and variable to support the development of an unstable disturbance as observed in the FSC. On this basis, therefore, it is concluded that mesoscale eddies observed in the FSC are not generated by the South Faroes Front.

9.1.2 The Influence of the Faroe-Shetland Channel

In the FSC, the circulation is dominated by deep wave-like meanders in the Slope Current meanders. The application of a baroclinic instability model has shown that the meanders have a wavelength of around 50-60 km with a similar amplitude, a period of 6-10 days and a phase speed of about 10 cm s^{-1}. Confidence in these values is increased by their reasonable agreement with peaks in spectral energy estimates from low frequency, local ADCP current components and with meander dimensions in AVHRR images. The meanders are correlated with 'hotspots' of eddy kinetic energy (EKE), which occur at intervals of around 100 km from the WTR, the highest of which (600 cm^2 s^{-2}) is closely associated with meander M2.

The trajectories of the drifters in the FSC showed them to be drawn into the frontal region between the NAW of the Slope Current and the MNAW, where their movement was closely correlated with the meanders. Currents produced by strong thermal gradients associated with the meanders resulted in the drifters travelling up to 60-70 km in a day (~75 cm s^{-1}), which was faster than those observed at NWOCE-E in the Slope Current (50 cm s^{-1}, not shown). Similar magnitudes were observed by Sherwin *et al* (1999), who tracked drifters progressing around what has been identified here as meander M4.

AVHRR images show the meanders to have a developmental lifecycle and when well developed they extend across the FSC onto the Faroes slope, thereby transferring warm NAW throughout the Channel. The variable extent of the meanders thus influences the re-circulation of MNAW in the Channel and in combination with the spawning of cold-core eddies changes the ambient balance of water masses in the upper and intermediate layers. Increased NAW in the FSC has implications for the transport of heat and hence the meanders may have a role to play in climatic terms (discussed in Section 9.2).

Through the use of simplified models, baroclinic instability in the Slope Current has been shown to be the probable cause of meander generation. However, although the WTR is believed to provide the initial perturbation this has not been proven here and remains an important unresolved issue. Changes in the flow structure are forced over the WTR resulting in re-distribution of potential vorticity, but it is of interest to note that in Figure 7.3 both instances of increased lateral meander development occurred when the Rockall Trough south of the WTR is inundated with NAW. Increased transport of NAW crossing the WTR will pump momentum into the system in the FSC, but the implication is that the conditions south of the Ridge may influence the development of the FSC meanders. Of course, two images are insufficient evidence on which to base anything other than a working theory, but detailed measurements either side of the WTR would provide a suitable basis from which to investigate any link and would help to establish the reasons for the variation in NAW transport approaching the Ridge.

An apparent contradiction exists between basic calculations of meander propagation speed and the AVHRR images, which seem to show that the meanders develop laterally but do not move along the FSC (i.e. they are a form of standing wave). In addition, the deflection of drifter tracks in May 1996 (Sherwin *et al*, 1999) coincides very closely with meander M4, both in terms of along-slope positioning and lateral extent. In their investigation of continental shelf waves (see Section 9.1.3), Gordon and Huthnance (1987) show that resonance (i.e. the amplification of a standing wave) occurs in the FSC at a frequency of around 23 hours, due to energy being reflected by the WTR in the south and the Norwegian Trench to the north. Further research may show that such reflection at lower frequencies (an order of magnitude lower) influences the meander characteristics.

Mesoscale eddies appear to be a common occurrence, especially on the Scottish side of the Channel where the meanders develop sufficiently to spawn these structures. However, not all eddies present in the FSC are generated there, and there is strong evidence to support eddy propagation from the north into the Channel (e.g. Hansen *et al*, 1998). Generation may be at the north-eastern tip of the Faroes Plateau, or as a result of cool, low salinity water tongue extending into the Channel from this direction. The Iceland-Faroe Front is also a well-known breeding ground for cold-core eddies (Allen *et al*, 1994), of dimensions similar to those observed in the FSC and also generated by the baroclinically unstable frontal meanders. It seems possible that FSC may receive some of these structures as well, which would travel south west along the Faroes side of the FSC. The assertion by Sherwin *et al* (1999) that eddies may be formed at the southern end of the Channel seems, on the evidence presented here, less likely.

9.1.3 Wind-forced Mesoscale Variability

Wind-induced variability is of particular concern over the Scottish continental slope which is subjected to frequent storms, producing large currents and transport variability that account for a major portion of low-frequency variance (Oey, 1998a).

Mesoscale variability in the form of low frequency wave structures is known to exist over the slopes of the FSC, particularly on the Scottish side. These features, known as continental shelf waves (CSW), are forced primarily by longshore wind stress (a frequent occurrence in the FSC). The waves decay exponentially toward the ocean and travel cyclonically (i.e have the coast to the right in the northern hemisphere) at sub-inertial frequencies, in an infinite sequence of modes that correspond to increasingly complex structures (Huthnance, 1995).

CSW propagation along the slope is maintained by the conservation of potential vorticity. A water mass at rest will acquire planetary vorticity, but if it is displaced up or down the slope changes in depth and hence relative vorticity will cause it to return to its starting position. The relative displacement of adjacent parcels results in a longshore propagation. The trapping of the waves against the slope in this way has resulted in their alternative name of coastal trapped waves. In unstratified conditions the barotropic current magnitude associated with CSW's varies spatially approximately as $h^{-0.5}$ (Huthnance, 1981), where h is the water depth.

Most observations of CSW's are of the barotropic mode 1 form (the fastest travelling), which consists of a longshore current that is confined to the shelf and typically uniform across it (Huthnance, 1981). This mode appears to be the natural response to the wind stress. The scale of the response is limited by the across-shelf scale of the topography and the along shelf scale of the most severe winds within storms; on the Scottish shelf in the FSC both are of the order of 100 km.

Direct observations on the Scottish shelf/slope (Gordon and Huthanance, 1987) show that there two, barotropic, response types to winter storms:

i) High frequency oscillatory anticyclonic currents caused by short-duration (half day), localised wind forcing, rotating with a period of about a day. Model studies show that the response frequency is close to the resonant frequency of the slope/shelf area between the WTR and the Norwegian Trench, both of which tend to amplify the response through energy reflection ;

ii) Quasi-steady along-slope currents that flow as long as the wind blows. The wind forcing for this response is typically 2 or more days in duration, and the resulting currents are parallel to the local bathymetry.

Both responses are barotropic over the continental shelf, due to wind-induced mixing through the water column. Both are reported to be lowest mode continental shelf waves, with wavelengths exceeding 100 km and response amplitudes of 20-40 cm s^{-1}. These features thus potentially make a major contribution to current variability in the region.

Smyth (1995) identified spectral peaks at near diurnal and 50 hour periods in three months of winter ADCP and current meter data, from close to the 500m isobath at the southern end of the FSC. These features were tentatively attributed to mode 1 barotropic CSW's. The month of currents at NWOCE-E on the Scottish continental shelf shows only very weak energy at 25 hours, probably attributable to barotropic tide, but also a slightly enhanced signal at 48 hours (Figure 9.1). At NWOCE-A on the Faroes shelf, the energy at these periods is much stronger and better resolved with narrower bandwidth. The spectra may be displaying the signature of barotropic

CSW's, but in both cases the spectral density associated with this periodicity is about an order of magnitude less than the semidiurnal tide (shown in the figure as a very high energy peak at 1 to 2 cpd). As such, therefore, it is of less significance in terms of extreme currents. In a less quiescent period than May 1999, when the ADCP data were collected, this may not be the case.

9.2 Comment and Future Work

The initial impetus for this study came from the offshore oil industry, which for design, safety and operational reasons needs to fully understand the potentially severe environment of the deep water Faroe-Shetland Channel. Considerable labour has been applied through numerical modelling to developing a means of quantifying the physical conditions in the FSC, but without *in-situ* observation (including satellite measurements) there is no way of verifying of the model output. Direct investigation is therefore a crucial component of the on-going effort to define the oceanography of the FSC.

Tracking quasi-lagrangian drifters over several weeks permits detailed investigation of dynamic processes occurring over a wide range of spatial and temporal scales. Here, the drifters have mapped the regional structure of the circulation at their drogue depth, whilst zonation of the tracks has enabled the development of a suite of statistics based on the drifter movement, that can be used to characterise and compare the flow in different parts of the region of interest. Eight drifters were deployed during the work, but this represents a minimum to obtain meaningful data from this type of instrument. Trajectories from numerous other investigations cover the Faroese Channels (most notably perhaps the Shelf Edge Study; see Burrows and Thorpe, 1999) and these should be incorporated in future studies to verify and expand on the results and theories presented in this report.

The extensive variability in the circulation of the FSC is the net result of several mesoscale processes, each individually associated with potentially elevated current speeds and collectively with the capability of disrupting offshore operations. Mesoscale eddies and an intense north-eastward flow along the Scottish continental slope are well-documented phenomena, but this study has also shown that deep meanders in the Slope Current are associated with steep temperature gradients and high velocity currents. Seasonal changes in the Slope Current have previously been documented, and future work should seek to establish a link between this and the development of the meanders. The scales of variability associated with circulation around the Faroe Islands are also important, as changes in the fluxes of MNAW and intermediate water will alter the balance of water masses in the FSC and may influence the development of unstable disturbances. These tasks will necessarily involve the investigation of conditions well beyond the Faroese Channels, and are perhaps ideally suited to numerical modelling.

The circulation of the North Atlantic and deep water convection are important factors in modulating climatic conditions in Europe. As one of three openings in the Scotland-Greenland Ridge, the FSC is a major conduit for the poleward transport of heat and the equatorward removal of water of Arctic origin. The surface currents in the FSC are estimated to transport about half the total northward heat and water across the Greenland-Scotland Ridge (Sherwin *et al*, in press), but the variation in quantities of NAW in the Channel will modify the transport. Quantification of the variability of the meanders, including frequency, seasonality and life expectancy is therefore important to the estimating the role of the FSC in climatic terms, and in accurate numerical climate modelling.

The Faroese Channels are host to commercially important fisheries, including that of blue whiting, which migrates through the FSC after spawning. However, on occasion the fish use the FBC, the comparatively narrow shape of which results in a better yield. Advance knowledge of which route the fish are taking will therefore clearly be advantageous to the fishing fleet. Hansen and Jákupsstovu (1992) suggest that this species migrates passively, i.e. it drifts with the flow, and also that it congregates in frontal regions, including to the west of the South Faroes Front (although this may be the result of some biological or dynamic phenomenon). As described in Chapter 2, the surface MNAW in the FSC mixes with intermediate water beneath and when this mix is stronger (i.e. more AI/NIW is flowing south west along the FSC) there is likely to be a higher flow rate to the south west, hindering the passage of the fish into the FSC and prompting them into using the FBC. If the meanders do influence the conditions in the FSC as hypothesised above, greater awareness of their generation and life cycle will therefore aid the fishing industry. In addition, further investigation is required to establish whether the temperature gradients associated with the shelf water tongue south of the Faroe Islands (shown here to significantly influence inter-Channel exchange) are maintained as the surrounding water undergoes seasonal cooling.

In-situ observations are invaluable, but costly and difficult to obtain and so the major thrust of future studies will require the extensive use of satellite data, which encapsulates many processes in a readily available format. Altimeter data has been shown here to accurately reflect the presence of the meanders, and by using ever-increasing databases of regular observations and AVHRR images it would be possible to develop a climatology of meanders at a relatively low cost. In tandem with long term measurement sets such as the hydrographic sections maintained by the Marine Laboratory in Aberdeen, the climatology will document past trends of meander development, including statistics on formation, size, life expectancy, intensity and propagation, and hence provide insight into future conditions, perhaps through linkages with the North Atlantic Oscillation and far field conditions or events. The work would also give real insight into the dynamics governing the formation of the meanders.

10 CONCLUSIONS

Through the analysis of *in-situ* observations from eight surface drifters and fixed station survey data, in combination with satellite measurements, this study has revealed new information about the dynamics of the Faroese Channels, particularly the Faroe-Shetland Channel.

The measurements have for the first time characterised the nature of baroclinically unstable Slope Current meanders in the FSC, and shown that they play a fundamental role in the circulation of the upper layers. With a wavelength of 50 km and a period of 1 to 2 weeks, their developmental cycle is such that by transferring NAW right across the FSC onto the Faroes slope, possibly on a time scale of 1 to 2 weeks, they also directly affect the exchange of water between the FBC and the FSC and the recirculation of upper and intermediate waters within the FSC. In addition, the meanders generate mesoscale eddies along the deep water side of the Scottish Slope Current, which then propagate north-east along the Scottish slope. In addition to the popular view that eddies are responsible for elevated current velocity measured at fixed locations in the FSC, the meanders themselves are also associated with strong currents (~75 cm s^{-1}) that flow around their extremities. Interaction between the meanders and the surface waters on the Faroes side of the FSC results in cyclonic mesoscale cells between the meanders.

The drifters showed that the meanders impose on the surface variability of the FSC time and length scales much greater than those in evidence in the FBC, which broadly is sub-mesoscale in nature with little net through flow. On the upper Faroes slope and shelf, the circulation is highly variable and encapsulates a range of length scales. The bilateral exchange of surface waters between the Channels, as highlighted by the drifter trajectories, is regulated not only by the meanders, but probably also by the intrusion of cool Faroes shelf water into the Faroese Channels confluence region. This intrusion constitutes a frontal zone that is better defined at 300m than near the surface and is a persistent feature of variable horizontal area. The intrusion markedly influences the circulation in the southern FBC, through the maintenance of density driven currents that are evident across the width of the FBC. In contrast to the initial hypothesis of this study, there was no evidence that the frontal zone sheds eddies, although historical reports suggest that this may be the case. These, however, have not been observed to propagate into the FSC and the scales of any such structures are unlikely to be compatible with the mesoscale features observed there.

To advance climate change research, further work is required to investigate in more detail the time scales and periodicity of the meanders, the role of the Wyville Thomson Ridge in their formation and their place in the wider North Atlantic circulation. The importance of satellite measurements in this work has been established here, and the opportunity exists to develop a comprehensive meander and eddy climatology through the judicious use of altimeter data and AVHRR imagery. A 'tool' of this nature would also assist with numerical model verification and with the development of a prediction capability for mesoscale variability and associated strong currents.

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