Nearbed Flows and Sediment Movement on the Continental Slope

by

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Abstract

The steep continental slopes of the southern Celtic shelf have long been thought to be a major export region for the flux of sediment from the south west approaches to the deep ocean (47-49°N). Recent studies have suggested that the transfer of energy from the large barotropic $M_2$ tidal currents to internal tides, and higher frequency internal waves is locally enhanced, and provides a mechanism for the re-suspension and downslope transport of bed material on the upper slope region. This material is thought to be preferentially transported at the head of the many submarine canyons that exist along these ocean margins, where the barotropic $M_2$ tidal currents are locally amplified and internal wave energy focused.

A unique 23 day deployment of the benthic lander STABLE (Sediment Transport and Boundary Layer Equipment) in July 1990, was at a depth of 388m on La Chapelle Bank continental slope. The site was at the head of a canyon, and at a depth thought to be critical for the generation of internal tidal energy. It was also at the deep water end of a transect of two current meter moorings across the slope. These measurements have shown that during summer stratified conditions the barotropic and baroclinic tides are sufficiently energetic near to the bed to mobilise the sand/gravel sized sediment on the upper slopes and at the shelf break. Eulerian residual bottom currents and maximum tidal currents are orientated cross-slope and this has important implications for sediment transport. The shelf break is predicted to be a region of bed load parting with bedload transport shelfwards at the shelf break and downslope immediately oceanward. On the critical slope region peaks in suspended sediment concentration occur at times of locally enhanced maximum downslope flow (40cms$^1$) and maximum current shear. This suggests that sediment is being eroded from the bed locally and confirms a net flux of material downslope. Above the boundary layer suspended particulate matter will be transported with the net flow of water which is predominantly along-shelf and polewards.

A second 10.4 day deployment of STABLE (II) in January 1994 was at a depth of 879m on the Goban Spur. These observations demonstrated the variability of continental slope processes on the margins of the Celtic shelf. During the deployment, weaker $M_2$ tidal currents (maximum of 24cms$^1$) were orientated along-slope and there was no evidence to suggest that the fine cohesive sand/mud sized sediment was mobile. Current meter measurements show that any suspended material will be predominantly transported along-slope and poleward. This will be periodically reversed and the net flux will be equatorward.

The two studies have highlighted the temporal and alongslope variability of geological and hydrodynamical conditions near to the bed and highlight the difficulties in estimating shelf-ocean fluxes of material across the whole of the north-west European continental margin.
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Declarations

At no time during the registration for the degree of Doctor of Philosophy has the author been registered for any other University award.

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Relevant scientific seminars and conferences were regularly attended at which work was often presented; external institutions were visited for consultation purposes, and several papers prepared for publication.

Presentations and Conferences Attended


Faroe-Shetland Channel Centenary Symposium, SOAFD Marine Laboratory, Aberdeen, April 1993.

Two oral and one poster presentation of current work to Ocean & Fisheries Science Division Research Seminars, Institute of Marine Sciences, University of Plymouth, September 1992 to July 1995.
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NERC sponsored 3rd National Summer School in Geophysical and Environmental Fluid Dynamics, Department of Applied and Theoretical Physics, University of Cambridge, September 1993.

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Research Cruises

Research cruise, December 1993 onboard *RRS Charles Darwin* (NERC/PML/CD83). The main objectives of the cruise were to study shelf-ocean exchange processes and the North Atlantic gyre circulation. I was primarily responsible for the operation, data retrieval and preliminary analysis of an Undulating Oceanographic Recorder (UOR developed at PML).

Research cruise, April 1994 onboard *RV Cirolina* (MAFF). The main objectives of the cruise were to recover a bottom lander from the continental shelf off Norfolk, UK, which was primarily deployed to investigate the response of sediment to wave and current interaction.

Research cruise, October 1995 onboard *RRS Charles Darwin* (NERC/PML CD97). The main objectives of the cruise were to measure the geostrophic transport of the Azores current and to recover current meter moorings on the continental slope, deployed during the above CD83 cruise. I was again responsible for the UOR and was involved with post-processing CTD, XBT and SeaSoar data.

Signed: [Signature]

Date: 31/12/96
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Chapter 1

Introduction

1.1 Overview of research area

The shelf edge is perhaps the most complex of all dynamic balances within the world's oceans and shelf seas. The steep continental slopes serve to restrict the interaction of oceanic and shelf circulation processes, and until recently have been studied independently from one another. However a new trend in oceanographic research has emerged over the past decade in which the multi-disciplined, world oceanographic community is uniting in their studies of physical, biological, biogeochemical, and geological processes. The ultimate goals are to understand the mechanisms which transfer water masses and material across the shelf break, enabling us to develop global ocean-shelf physio-biogeochemical circulation models. An understanding of the dispersal and ultimate fate of man-made pollutants (or energy related materials) in the marine environment (see Olsen, 1995) has largely
been the impetus behind these multi-disciplinary studies, along with a recent demand to further our knowledge of the global atmosphere-ocean response to climate change. Recent examples off the northwest coast of America include the Shelf Edge Exchange Processes (SEEP) I & II programs in 1983/84 and 1988/89 respectively, and the current Ocean Margins Program (OMP) running from 1994-97. On the continental slopes of northwest Europe, recent examples are the Ocean Margin Exchange (OMEX I) program on the Goban Spur in 1992-95, and the current Land Ocean Interaction Study, Shelf Edge Study (LOIS SES) on the Hebrides slope, running from 1995-1998. OMEX II will begin in 1996, focusing on the Iberian slopes. The recent controversy over the environmental and ecological aspects to deep sea dumping (Gage & Gordon, 1995) has highlighted current limitations in the understanding of continental slope processes making the subject of this thesis timely in debate.

1.2 Specific aims

An important link to the processes outlined in section 1.1 are the mass budgets for sediment transfer across the continental margins (Nittrouer et al., 1988). The recently proposed Continental Margin Integrated Sedimentation Studies (CMISST) program along the northern California and New Jersey continental margins (Nittrouer & Kravitz, 1995) is an attempt to integrate a multi-disciplinary study of the continental margins, whilst remaining focused on fundamental geological questions. One of the goals outlined by Nittrouer & Kravitz (1995) is to determine the geological relevance of short-term physical processes that erode and transport sediment across the shelf-ocean boundary.

The focus of this present study is an investigation of the physical processes at the shelf-slope boundary which effect the sediment transfer across part of the northwest European continental margin. It is proposed in chapter 2 that one of the mechanisms

* In light of the recently proposed, and later abandoned proposal by Shell UK, to dispose of the Brent Spar oil platform in the Rockall Trough region in 1995.
responsible for the erosion of sediment along the upper slopes of the Celtic and Armorican shelves is the relatively short period motions of internal waves/tides. It is also proposed that the flux of sediment varies along-slope with a preferential cross-slope transport within submarine canyons.

There are very few investigations which have coherently and comprehensively measured the boundary layer hydrodynamics, suspended sediment fluxes and bedforms on the continental slope. The specific aim of this thesis is to investigate the shelf-slope hydrodynamical processes which are responsible for the transport of sediment from the shelf break and slope region of the Celtic Sea and into the deep ocean. Two separate field studies with contrasting geological and hydrodynamic environments also enabled a study of the along slope variability of these processes. This provides an insight into the practicalities of being able to accurately predict sediment fluxes for the whole of the northwest European continental margin.

1.3 Outline of thesis contents

The introduction and literature review are intended to give the reader a clear indication of the specific objectives of this current work. This is based on a clear understanding of the hydrodynamic processes which influence near-bed flows and sediment transport at the continental margins. Wherever possible the review has centered on the processes which are particularly relevant to the northwest European continental shelf edge and upper slope, in particular the La Chapelle Bank region of the Celtic Sea. The reader will also have an appreciation of the variability of these processes from one site to another. An experimental hypothesis is shown in schematic form at the end of chapter 2 and is based on the aforementioned review. Chapter 3 describes the site selection, instruments, and the limitations of the data collected at the main field study site. Chapter 4 describes the hydrodynamic field results and the methods used in the analysis of these data sets. Particular attention has been drawn to the accurate calibration of the near-bed turbulent current measurements on a critical upper slope region. This is because an accurate
interpretation of these results is fundamental to the hypothesis formed in chapter 2 and for the subsequent analysis described in chapter 5. Chapter 5 begins by describing the nature of the bed material. Estimates of the bottom stresses at the critical slope region are then described. These results are used to compare in situ sediment transport observations with predicted transports rates, and previous sediment transport estimates just shelfward of the shelf break. Chapter 6 is a brief description of a second opportunistic study of near-bed flows and sediment transport processes on the Celtic Sea continental slope. The contrasting hydrodynamical and geological nature of the Celtic Sea continental slope region is one of the conclusions discussed in chapter 7. This section draws together all the observations and analysis in a short and precise manner focusing primarily on the main La Chapelle Bank field study results.
Chapter 2

Processes at the shelf edge; A literature review

2.1 Introduction

The interaction of shelf edge topography, with stratification, generates a multitude of physical processes over the full spectrum of length and time scales (Huthnance, 1981), which can encourage the flux of water, material, and energy away from, and across the ocean-shelf boundary (Huthnance, 1995). The transport of sediment occurs when currents generate bottom stresses in excess of a critical value, which, for bottom sediment is a function of grain size, density, water content, depositional history and bioturbation (Miller et al., 1977, Gardner, 1989a). During transport events the seabed undergoes changes in roughness and sediment size which in turn modify the boundary layer hydrodynamics (Nittrouer & Kravitz, 1995). Consequently, the net direction and movement of sediment across this boundary is determined by a complex physical and geological environment (e.g. Karl et al., 1983). During the recent SEEP-II experiment for example, Biscaye et al. (1994) summarised the complexity of motions on different time and space scales which governed the observed distribution of fluxes of particulate matter at the shelf-slope boundary of the Middle Atlantic Bight (see also Biscaye & Anderton, 1994). In this instance the flux of material was both down-slope, out into the oceanic waters' and along-slope.
To accurately monitor sediment transport at the shelf edge, several instruments are required to monitor the near-bed activity over these variable time scales, and should ideally be measured for a period of months. The present field study involved a twenty three day deployment of the instrument tripod 'STABLE' (chapter 3) on the upper slopes of the Celtic Sea (described in chapters 4 & 5). The data collected enabled a regional study of the geological response to strong cross-shelf tidal currents in the presence of a background along-slope residual current. A second Celtic Sea deployment of the Mk II version of this tripod (described in chapter 6) enabled a comparative study of these processes and a unique insight into the variability of transport mechanisms along the continental margins of northwest Europe.

With the publication of several reviews of the physical oceanography at the shelf edge, e.g. Huthnance (1981), Pietrafesa (1983), Wroblewski & Hofmann (1989), and Huthnance (1995), the emphasis of this review is to describe the physical processes which are fundamental to the northwest European shelf edge, and/or are relevant to the experimental hypothesis formulated in section 2.5. Since little in situ sediment transport field data has been accumulated to date, many of the ocean-shelf motions which interact with the benthic boundary layer, and thus influence the sediment transport on the sloping region, can only be inferred.

2.2 Processes relevant to the NW European shelf-ocean boundary

The transfer mechanisms for the transportation of sediment across the northwest European shelf-ocean boundary are poorly understood. For example, are storm surges, which intermittently transport large amounts of sediment down-slope as turbidity currents, the main process? Evidence for this transport mechanism comes from the Celtic Sea, where during average or cold winters, Huthnance (1981) describes how shelf waters may cool to become as dense as the adjacent slope waters, and cascade over the shelf slope (as observed by Cooper & Vaux, 1949), facilitating a route for the down-slope transport of sediment. Are the barotropic tides themselves able to penetrate the bottom mixed layer and resuspend sediment at
the shelf break / slope? Are the baroclinic tides sufficiently energetic to erode/transport sediment from the upper slopes? The northward flowing slope current is expected to be the dominant residual transport mechanism for the fine 'marine snow' particulate matter, but how significant is the slope current near to the bed?

Physical oceanographic processes which are important along shelf break regions of northwest European shelf edge are discussed by Huthnance (1981). The area of interest begins with Armorican shelf slopes which border the deep waters of the Bay of Biscay from -45-47°N, the Celtic shelf slopes at -47-51°N, and finally the Irish, Malin and Hebrides slopes at the northern extent, culminating at the Wyville-Thomson Ridge at -60°N.

Fig 2.1 illustrates the topographic complexity of the shelf edge in this region (STABLE deployments are also shown for later reference). The 200m contour defines the shelf break, with the 4000m contour representative of the deep ocean. The shelf broadens in the Bay of Biscay across the Aquitaine and Armorican shelves compared to further south off the Spanish coast, and again further at the northern/southern end of the Armorican/Celtic slopes at 47°-49°N. The region has a meandering along slope topography and steep slopes which are incised by many canyons. Further north are the slopes of the Goban Spur and the Porcupine Bank (-49-54°N). The upper slopes are far gentler and form the eastern boundary of the North Atlantic Current at approximately 51°N (Krauss, 1986) where the flow separates, moving north into the Rockall Trough, and southeasterly towards the stagnant region of the Bay of Biscay (Pingree, 1993). The Malin and Hebrides slopes to the north (-54-60°N) have a more uniform along slope topography, and extend down into the Rockall Trough below.

The review will now focus on those processes which are both influential and resolvable in the context of the 1990 La Chapelle Bank field study on the upper slopes of the Celtic shelf edge. These processes are then put into context with the adjacent slopes of the northwest European shelf. We begin with the tidal currents, the most dominant process for the northwest shelf.
Fig 2.1 The northwest European continental margins. Depth contours are in meters. The locations of recent shelf edge/slope deployments of the benthic lander STABLE are also shown and are briefly described in Section 1.2.
2.2.1 Tidal Currents

The dynamics of the northwest European shelf are mostly dominated by the lunar semi-diurnal constituent, $M_2$ (period of 12h 25m). The oceanic tide can be approximated, to first order, as a progressive Kelvin Wave (see for example, 'Atmosphere-Ocean Dynamics' by A.E. Gill) which is predominantly travelling from the south in the northeast Atlantic. Part of this Kelvin wave turns northeastward into the Bay of Biscay (see Huntley, 1980) and then splits up into waves entering the English Channel and the Irish Sea (part of it also continues to travel north across the Irish shelf and Scottish shelf where it turns into the North Sea across the Fair Isle Channel on the West Shetland Shelf).

Continuity of momentum for the Kelvin wave as it propagates across the shelf break implies an amplification of the tidal current. For example, current meter observations on the upper slopes of the Bay of Biscay show an increase in the $M_2$ currents from 20 to 50 cms$^{-1}$ from the 500m contour to the shelf break at 200m (Pingree & New, 1991, i.e. tidal velocities $\propto h^{-1}$). A steep continental shelf margin can act as an effective boundary were tidal energy is reflected constructively. The shelf break thus acts as an effective outer boundary where tidal amplification can result from a co-oscillation with the landward boundary, if the distance between the boundaries approaches the tidal resonance wavelength (Thorne & Swift, 1985). The enhanced response of tidal forcing over the shelf break across much of the northwest European shelf occurs in response to the rapid change in topographic slope (for example, the across-slope gradient, $\alpha$, is equal to 0.051 on the upper slopes of La Chapelle Bank [Pingree & New, 1989]), and partly due to the large shelf width, $W_s$, where the Kelvin wave is near quarter wave tidal resonance across the shelf. Quarter wave resonance will occur if $W_s \sim \left(gh\right)^{1/2}/\pi/2\sigma$, which is equal to 350km for $\sigma=1.4\times10^{-4}$s$^{-1}$ and $h=100$m (from Huthnance 1995), a value comparable with the width of the northwest shelf and thus producing the observed large $M_2$ currents at the shelf break in this region.
It is important to note that a resonant tide implies the complete reflection of tidal energy. The semi-diurnal tide is a progressive wave, whose large tidal velocities at the shelf break can be partly accounted for, or described by, a near resonant characteristic length scale, \( W_s \), for this region. In reality there is a downstream dissipation of tidal energy, or flux of energy onto the shelf, and numerical models are required to study the spatial variability along the complex topography of the shelf break (Huthnance, 1995).

On the Celtic and northeast Armorican shelves, the combination of shelf width and steep slopes result in the largest fluxes of barotropic semi-diurnal tidal energy across the shelf break (~118 GW from 47\(^\circ\) to 49\(^\circ\)N estimated from a non-linear numerical modeling study by Pingree et al., 1982), and dominate the barotropic currents shelfward of the shelf break (and to a lesser extent, the Irish, Malin and Hebrides Shelves, where \( \alpha \) and \( W_s \) are typically an order of magnitude smaller; cf. ~34 GW from 49\(^\circ\) to 52\(^\circ\)N). The propagation of this large energy flux across the slope is due to the co-oscillation in the English channel which is mainly dissipated through bottom friction in the English Channel (Pingree et al., 1982). The main results of the numerical model by Pingree et al (1982) were:

(i) The total \( M_2 \) energy flux across the shelf break and onto the Celtic shelf (for the region shown in fig 2.2) is approximately 141 GW. The vertically integrated tidal velocities at the shelf break can typically be of the order of 40cms\(^{-1}\) (as shown in fig 2.2) and 60% of the tidal energy flux onto the shelf occurs in regions where this value is reached.

(ii) The tidal ellipses are rotary clockwise with semi-major axis of the \( M_2 \) tide aligned perpendicular to the shelf break (~20-30\(^\circ\)T).

(iii) In the region of maximum tidal streaming at the shelf break between 47-49\(^\circ\)N, the ratio of \( S_2:M_2 \) is at its largest (approximately 38%), which is 3% larger than that further to the north.

(iv) Spring semi-diurnal barotropic tides reach maximum values of 50cms\(^{-1}\) (semi major axis) in the Chapel Bank region of the Armorican shelf edge (Pingree et al., 1983). These may however be increased locally due to
shoaling effects over banks (e.g. Chapel Bank) or due to the local curvature of the slope and shelf break bathymetric contours.

Further to the south, in the Bay of Biscay on the west coast of France, the amplitudes of the semi-diurnal tidal currents decrease considerably to $O(20\text{cms}^{-1})$ for the semi-major axis (LeCann, 1990), as the influence of English Channel is lost. These currents are now a maximum at the shelf break, diminishing towards the French coast. The ellipticity of the tidal flow at the shelf break remains the same ($a/b \sim 0.6$) as that in the Chapel Bank region of the northwest Armorican shelf with the major component of the tide also aligned perpendicular to the shelf break.

Fig 2.2 Modelled vertically integrated $M_2$ tidal currents (cms$^{-1}$, semi-major axis) at the shelf break region of the Celtic and Armorican shelves (from Pingree et al., 1982).
In summary, the barotropic tidal current strength across the shelf break between $6^\circ$ & $7^\circ$W is the largest found anywhere along the northwest shelf edge and it has some of the steepest upper slope profiles. It is appropriate at this stage to introduce the concept of internal tides. These are forced baroclinic motions of $M_2$ period which result from the interaction of the large tidal currents with steep slopes and stratification. Internal waves and tides are the subject of discussion in sections 2.2.3 and 2.2.4. respectively and are:

(i) predicted from models in a global context (Baines, 1982),
(ii) predicted from localised models (e.g. Heathershaw et al., 1987),
(iii) observed in situ locally (e.g. New & Pingree, 1990a),
(iv) observed remotely (e.g. Pingree & New, 1995)

to dominate the baroclinic energy flux across the upper slope and shelf break regions of the Celtic and Armorican shelves. The baroclinic energy flux has important implications for the flux of material across the shelf-ocean boundary due to the enhancement of tidal energy dissipation through mixing in the ocean interior and through the interaction with the sea bed.

The implications for sediment transport are now briefly discussed in the context of the barotropic tides. Shelf edge processes heavily influence the nature and supply of sediment to the slope. This exchange is heavily dependent on the rates of delivery of sediment to the shelf edge region however (Nittrouer & Kravitz, 1995). On the northwest European shelf the barotropic tidal currents play an important role in this supply.

Non linear components of the equations of momentum and continuity generate higher tidal harmonics. The accurate description of the non-linear contributions are required in sediment transport studies, since sediment dynamics is a very non-linear process. It has been known for some time that sediment on the northwest shelf is transported towards the continental slopes. Stride (1973) proposed this hypothesis from a study of the asymmetry of the large sandwaves across the shelf, and Pingree & Griffiths (1979) later confirmed this by showing that the distribution of the mean and maximum bottom stresses obtained from the combined contributions of the $M_2$
and M₄ tides could accurately represent the oceanward sediment transport paths in the English Channel and Celtic Sea, i.e. towards the shelf edge. Heathershaw et al (1987) concluded that the tidal currents alone were strong enough to mobilise the sediments just shelfward of the Celtic Sea shelf break.

The barotropic tides thus play a major role in the transport of sediment and its associated pollutants to the shelf edge. The processes that transport sediment from the shelf edge to the slope, and subsequently into the ocean, are now addressed.

In the context of the barotropic tides on the Celtic and Armorican shelf edges, Stride (1963) suggested that sediment would be transported preferentially down the many submarine canyons that exist along the upper slopes. In a laboratory simulation of a forced tidal motion across a canyon, Baines (1982) showed that larger tidal velocities were concentrated near the head of the canyon and were substantially larger than those outside the canyon at the same depth. The large amplitudes in velocity were due to the three dimensional nature (or 'narrowing' effect) of the canyon geometry, effectively forced by a time varying pressure gradient on the shelf directly above the canyon. Shepard et al. (1979) show a number of examples of the large tidal velocities observed in submarine canyons. Hotchkiss & Wunch (1980) observed this phenomenon in Hudson Canyon and concluded that the amplification of the barotropic and baroclinic tidal currents in the canyon would enhance the down-slope transport of sediment locally. Garner (1989) presented clear evidence (from transmissometer and near-bed current observations in Baltimore Canyon) that submarine canyons do accelerate tidal flows sufficiently to cause resuspension at predominantly tidal frequencies.

2.2.2 Slope Current

"Currents along the continental slope can and do occur with non-zero time averages and velocities exceeding or opposing those to either side in the adjacent shelf and ocean" (from Huthnance, 1995).
Residual currents along the continental slope can be:

(i) wind driven,
(ii) tidally induced,
(iii) density driven, or
(iv) forced externally by pressure gradients set up by large scale variations of (i) and (iii), (from Pingree & LeCann, 1989).

There are several mechanisms by which these residual currents can be generated (reviewed by Huthnance, 1992 & 1995). This discussion will be limited to the poleward flow of slope water along the eastern boundary of the North Atlantic (but is also relevant the other eastern boundaries of the Pacific and Indian Oceans).

Observations of this poleward current, which is confined to, and steered by the sloping bathymetry at the ocean-shelf boundary are well documented, and stretch the entire length of the eastern boundary of the North Atlantic. They include Barton (1989) on the North African slopes, Frouin et al. (1990) on the Iberian slopes, Pingree & LeCann (1989 & 1990) on the Armorican and Celtic slopes, Thorpe (1987) on the Porcupine Bank, Booth & Ellett (1983), Dickson et al. (1986), and Huthnance (1986) on the Hebrides slope and finally Gould et al. (1985) across the Wyville-Thompson ridge.

It is now widely accepted that the slope current persists throughout water depth along the continental slopes of the northwest European shelf, due primarily to a poleward decline in dynamic height (e.g. Huthnance 1984 & 1986, Pingree & LeCann 1989). This mechanism, which is explained below, has been described as the "Joint Effect of Baroclinicity And Relief" (JEBAR) by Huthnance (1984).

Huthnance (1986) describes how the density of the upper waters' decrease towards the equator, and consequently 'stand high' nearer to the equator to achieve meridional barotropic equilibrium. Across the shelf break, the equatorial waters on
the shelf do not need to 'stand as high' to compensate and thus achieve this meridional equilibrium state. The result is an oceanic equilibrium response which produces a larger sea surface slope, or pressure gradient force, than on the shelf. The poleward slope current is thus "a forced baroclinic flow, geostrophically balancing the longshore density gradient" (Huthnance, 1992), which typically increases in magnitude (i.e. along-slope current speed), and transport, with distance northwards (see fig 2.3).
For example, on the Celtic & Armorican slopes a maximum slope current speed 10cms\(^{-1}\) and a maximum total transport of 3.5Sv are estimated in late summer by Pingree & LeCann (1990). The mean density-driven transport inshore of the 1000m contour is 1Sv (Pingree & LeCann, 1989) and also shows evidence of variability / seasonality. This is compared to maximum currents of >30cms\(^{-1}\) on the west Shetland slopes and maximum estimated transports of 7 Sv inshore of the 2000m contour (Gould et al., 1985).

The poleward decline in dynamic height required to produce the observed currents of O(10cms\(^{-1}\)) is of the O(10\(^{-7}\)) with a frictional decay time of 10\(^6\) s (Huthnance, 1984). Pingree & LeCann (1989) explain that the frictional decay time of 10\(^6\)s implies that the evolution of the slope current at any given location would be influenced by the slope up to 1000km to the south. This is therefore a large scale phenomenon which cannot reliably be estimated geostrophically across a shelf-ocean transect due to the difficulty of defining a level of no motion near to the bed on the upper slopes (Huthnance, 1986). Neither can it be easily observed meridionally with an estimated sea surface elevation of only 2cm (estimated by Pingree & LeCann, 1989) to produce the observed slope current of 5-10cms\(^{-1}\) along the Armorican slope.

The northward flow of the slope current implies that the slope water can be observed as a warmer and more saline tongue of water advected locally onto the slopes further to the north, as observed by Frouin et al., (1990) on the Iberian slope and Booth & Ellett (1983) and Hill & Mitchelson-Jacob (1993) along the Hebrides slope. The flow is therefore baroclinic across-slope and the width of the slope current (which is also topographically controlled) can be estimated from the Internal Rossby Radius of Deformation, defined by Huthnance (1980) as;

\[
R_L = \left[ \frac{gh_1h_2(p_2 - p_1)}{(h_1 + h_2)p_2f^2} \right]^{1/2} = 0 \text{ (10km)} \quad [2.1a]
\]

for a two layer fluid, where \(g\) is the acceleration due to gravity, \(h_1, \rho_1\) and \(h_2, \rho_2\) are the upper and lower layer depths and densities respectively and \(f\) is the Coriolis
parameter. For the continuously stratified case $R_I$ is defined by Frouin et al. (1990) as;

$$R_I = \left[ \frac{H N}{f} \right] = 0 \text{ (10km)}$$  \hspace{1cm} [2.1b]

for observations on the Iberian slope, where $H$ is the total depth and $N$ is the Brunt Väisälä or stability frequency defined by;

$$N = \left( -\frac{g}{\rho_0} \frac{\partial \rho}{\partial z} \right)^{1/2}$$  \hspace{1cm} [2.1c]

which is the frequency at which a displaced fluid will oscillate with period $2\pi / N$. Generally, the flow width is 20-50km, with the largest currents in 500-1000m water depth (Huthnance & Gould, 1989). These estimates are also consistent with observations from remotely sensed images by Frouin et al. (1990), Pingree & LeCann (1990) and Booth & Ellett (1983) and show that the slope current is indeed confined to the sloping region. This residual transport mechanism is therefore confined to the shelf-ocean boundary and has important implications for sediment fluxes at continental margins. These are discussed later in this section.

Although this poleward decline in dynamic height is now viewed the most significant mechanism for the driving force of the slope current, Huthnance (1989) emphasises that the currents are not steady, and estimates that transports may vary seasonally by 50%, in some cases reversing the flow completely. Frouin et al., (1990) discuss the flow off Portugal and show that during a winter (summer) period, poleward (equatorward) surface slope currents flow in response to the meridional component of the windstress which is essentially northward (southward). Booth & Ellett (1983) also observe flow reversals of the slope current on the Malin slope, with weaker flows equatorward in summer and stronger and poleward in winter. Pingree et al. (1996) observe seasonal flow reversals of ~17 days duration along the Goban Spur (see also STABLE II results in chapter 6). Thorpe (1987) observes reversals of 5-8 days duration along the deep slope waters of the Porcupine Bank and White (1994) observes reversals of 4-5 days duration on the Hebrides slope. Gould et al. (1985) also observe seasonal trends with maximum transports in the winter period north of the Wyville-Thompson Ridge.
The specific areas of interest in the formulation of the hypothesis outlined in section 2.4 are the upper slopes of the Celtic shelf and Armorican shelf, which have been extensively studied by Pingree & LeCann (1989 & 1990). The main results of these studies are listed below;

(i) The general trend is a persistent northwesterly flow between 47° & 48°N. Seasonal trends are apparent and reflect the seasonal trends in the windstress. Relatively steady maximum values of 10\text{cms}^{-1} occur in late summer and minimum values occur in winter, which contrast with the observations further to the north and south.

(ii) On the upper slopes (near the 500m contour), the strongest flows are near-bed, with a strong down-slope component e.g. 14\text{cms}^{-1} down-slope at 6°40'W. At mid-depths the flow has an up-slope component. These across-shelf residual currents on the upper slopes are discussed further in section 2.2.4.4. and are thought to be forced by the baroclinic response of the semi-diurnal tides across the upper slope.

(iii) Mean poleward currents are generally weak, reaching 7\text{cms}^{-1} at the 500m contour, increasing to 10\text{cms}^{-1} at the shelf break and reducing to 2\text{cms}^{-1} at 3-4km depth.

(iv) The prevailing wind for the Celtic Sea is west southwest, which is approximately in quadrature with the wind direction which would produce a maximum barotropic response. As a consequence of this, weak (<5\text{cms}^{-1}) wind driven residual currents are observed on the shelf compared to the slope. Strong tidal currents and enhanced bed friction also prevent the establishment of strong wind driven residual currents.

The residual current has significant implications for sediment transport and these are now discussed. The nature of the sea bed on the slopes reflect the combined effects of a variable number of the dynamical processes at the shelf edge (Huthnance 1981), and are difficult to separate. The meridional and seasonal variability of the slope current implies that sediment fluxes will also be variable. On the slopes of the
northwest European shelf there are few observations of the correlation between the slope current and sediment transport. Where the slope current has been described as particularly energetic, such as the Shetland slopes, Kenyon (1986) confirms its influence on the transport of sediment from the asymmetry of the sand waves found in the region. On the Celtic and Armorican slopes the residual currents are influenced mainly by the combined effects of the tides (sections 2.1 & 2.2.4.4) and the slope current (Pingree & LeCann 1989). With the diminishing effects of the baroclinic and barotropic tides on the slopes with increasing depth, the slope currents have the potential to transport the finer material, held in suspension at depth, for long periods of time. Current and future Lagrangian drifter studies on the Hebrides slope (as part of LOIS SES) may confirm this hypothesis since the net transport of suspended sediment generally follows the net transport of water. The along-slope transport of sediment may also become trapped within submarine canyons, where it could be preferentially transported oceanwards.

Near to the bed the residual slope current above the bottom boundary layer is not generally an indication of the net sediment transport as discussed above. Ekman veering within the bottom Ekman layer implies a down-slope (up-slope) component to the boundary layer current in the presence of a poleward (equatorward) residual currents. Thorpe (1987) observes a veering of 17.5° between 10 and 30 meters above the bed (mab) at 3314m depth on the Porcupine Bank in the presence of a mean poleward along-slope current. Houghton et al. (1994) observe a mean offshore current in the bottom 20m on the southern Middle Atlantic Bight, consistent with the effect of friction on a mean southerly along-slope current. This Ekman upwelling and downwelling close to the boundary can have significant effects on the boundary layer dynamics (Weatherly and Martin, 1978, Trowbridge and Lentz, 1991). Trowbridge and Lentz (1991) show that the up-slope (down-slope) bottom flux of water during equatorward (poleward) along-shelf flows enhances (reduces) the stratification, inhibiting (enhancing) the growth of a bottom mixed layer (bml). The height of the mixed layer, $D$ could then be correlated with the cross-shelf current speed. White (1994) showed that reversals of the slope current on the Hebrides slope correlated with changes in stratification within the friction layer. Lentz &
Trowbridge (1991) also showed this asymmetry across the northern California shelf slope and proposed that it may have profound implications for sediment transport processes on the slope, although no observations of this phenomenon relating to sediment transport investigations were known to the author.

Finally mention should be made of the implications for sediment transport arising from the formations of anticyclonic eddies. These have recently been observed, and are thought to develop, due to instabilities of the slope current along the slopes of northwest Europe (e.g. Pingree, 1994 in the southern Bay of Biscay and Pingree, 1995 along the Iberian slopes) in regions where rapid changes of along-slope topography occur. These eddies are potentially an important mechanism for the transportation of slope waters into the ocean. If the eddies impinge onto the slope they may also provide a mechanism for the cross-shelf transport of sediment from the slope region. This process has not been observed on the northwest European shelf, but Washburn et al. (1993) reported direct evidence of an anticyclonic eddy exporting sediment into the deep ocean off the northern California shelf.

2.2.3 Internal waves

"Near-critical frequency reflection of internal waves is a likely candidate not only for sediment movement but also for formation of deep mixed layers" (Eriksen, 1982)

In this section the general theory of internal waves is described and the potential influence of high frequency internal waves to sediment transport at the shelf edge is discussed. Internal waves of tidal frequency are then discussed in section 2.2.4.

The general theory of internal waves is approached similarly in all of the good reviews of the subject. The most comprehensive of these is by Phillips (1977), with others by Turner (1973), Huthnance (1981) and Gill (1982). A good overview of internal waves has been written by Munk (1981), in a Collection of Scientific Surveys in Honour of Henry Stommel, in which the conventional idea of internal
waves at a boundary between two fluids of different density is considered, followed by the more general case of a continuously stratified fluid.

This review will concentrate on the importance of internal waves at the continental shelf edge. The easily imaginable concept of a two layered ocean, in which waves propagate along a thermocline, is appropriate to the shelfward propagation of internal waves forced at the shelf break region of the northwest Europe (e.g. Sherwin, 1988). These first mode internal waves are phase locked with the barotropic semi-diurnal tide and are discussed in section 2.2.4 in the context of their ability to transport sediment from the shelf edge onto the upper slopes of the Bay of Biscay (Heathershaw et al., 1987). Within the deep ocean, continuous stratification allows internal wave energy to convert to higher modes, allowing rays of energy to penetrate the ocean at characteristic angles. The ray theory concept will be discussed, in which the dispersion relationship can be shown to describe some of the fundamental properties of internal waves in a rotating, stratified fluid, in particular their refractive, generative and reflective properties at the shelf edge.

Firstly, an important oceanographic parameter known as the Brunt-Väisälä, or buoyancy frequency is defined, which describes the static state of the ocean. This stability frequency can easily be derived from the linearised Boussinesq equations of momentum and continuity for an inviscid fluid, and is defined as:

\[ N = \left( \frac{g}{\rho_0} \frac{\partial \rho}{\partial z} \right)^{\frac{1}{2}} \quad (s^{-1}), \]  

where \( g \) = acceleration due to gravity (ms\(^{-2}\)), \( \rho_0 \) = a background density field (kgm\(^{-3}\)), \( \rho \) = the density (kgm\(^{-3}\)) at depth, \( z \) (m) measured from the surface downwards, and describes the smallest period wave that can exist in the ocean interior. The higher the degree of stratification (e.g. frontal zones such as the seasonal thermocline or shelf-slope fronts), the higher the frequency of internal wave that can exist. The corresponding period, \( 2\pi / N \), varies from a few minutes in regions of large vertical density gradients to several hours in the deep ocean. For stable static equilibrium \( N^2 > 0 \ \forall \ z \). The forcing for internal wave motion is supplied by the interaction of the barotropic flows with topography, and across the continental margins of
northwest Europe, this can occur at tidal and inertial frequencies (e.g. Huthnance, 1981). One would then expect a cascade of energy, through non-linear wave-wave interaction (to higher modes) and wave breaking, to the higher wavenumbers of the energy spectrum (Schott, 1977). This has important implications for near-bed dynamics, sediment resuspension and transport on the upper slopes of continental margins.

A relatively simple equation describing motion in a stratified fluid is described from linear theory by the Taylor-Goldstein equation (e.g. LeBlond & Mysak, 1978);

\[ \frac{\partial^2 \phi}{\partial z^2} + k^2 \left[ \frac{N^2 - \sigma^2}{\sigma^2 - f^2} \right] \phi = 0 \]  

with the linearised boundary condition \( \phi(0) = \phi(-h) = 0 \), where \( \phi(z) \) is a vertical displacement amplitude of a fluid particle which is assumed to take the form of a plane wave solution of the form \( \eta(z) = \phi(z) \exp[i(kx - \sigma t)] \) in which \( k = \) horizontal wave number vector \((k_x, k_y)\), \( x = \) position vector, \( N = \) Brunt-Väisälä frequency, \( \sigma = \) internal wave frequency and \( f = \) the Coriolis parameter. The equation forms an eigenvalue problem which has discrete solutions in the form of a family of modes, each with a specific vertical structure. Simple analytical solutions for simple forms of \( N(z) \) can be found, such as the two layer \( (n = 2) \) case, in which \( N(z) \) is assumed to be negligible everywhere except within a small depth range surrounding the seasonal thermocline, and for \( N = \) constant. The dispersion relationship for the simplest case \( (n=2) \) indicates that since the interfacial displacement \( \phi_1(z) \) (the first baroclinic mode) is acted upon by a reduced gravity \( g' = g \Delta \rho / \rho \), where \( \rho \) is the mean density and \( \Delta \rho \) is the density difference, the phase speed is reduced by a factor of \( 10^3 \) when compared to surface waves. On-shelf propagating internal tides for example, are therefore subject to advection by the barotropic tide and can in some cases be forced back in an off-shelf direction (Pingree & New, 1995).

Generally however numerical solutions are required to solve the Taylor-Goldstein equation for an \( n \)-layered ocean, in which there exist \( n-1 \) number of wavenumbers \( (k_1 \ [\text{the first baroclinic mode}] < k_2, \ldots) \), namely the eigenvalues, and corresponding eigenfunctions \( (\phi_1, \phi_2, \ldots) \) for a given internal wave frequency, \( \sigma \) (New 1987). This
is the subject of review in section 2.2.4.3 in which \( \sigma \) is of tidal frequency, and the generation of internal waves for realistic slopes and/or stratification is discussed.

For flows in a continuously stratified, rotating ocean in which \( f = \) constant locally, LeBlond & Mysak (1978) derive a solution for internal waves from the equation of motion in which the dispersion relationship reduces to:

\[
\frac{k_x^2 + k_y^2}{k_z^2} = \frac{\sigma^2 - f^2}{N^2 - \sigma^2}
\]  

[2.3]

for a simple plane wave solution. If we assume that \( N \) varies only in the z-direction and for simplicity also assume that \( k_y = 0 \) then the phase speed, \( c \), is in the \( k(k_x, k_z) \) direction, normal to the crests, which is inclined to the horizontal by:

\[
tan \theta = \frac{k_y}{k_x} = \frac{\left( \frac{N^2 - \sigma^2}{\sigma^2 - f^2} \right)^{1/2}}
\]  

[2.4]

This simple relationship is shown in fig 2.4 (in the x-z plane), which also illustrates that the group velocity, \( c_g \), is parallel to the crests and the wave packet moves sideways, i.e. the energy propagates normally to the phase, and in opposite vertical direction (\( c_c = 0 \)).

Figure 2.4 The simple relationship between group velocity (\( c_g \)) and the phase velocity direction, \( k \) (x,z). Full lines represent wave crests (c) and dashed lines troughs (t). Taken from Pond & Pickard (1986).

Equation 2.4 illustrates that for a given stratification, internal waves can exist between the limits \( \sigma = f \) (where we approach near inertial waves with \( \theta \) large) and
\( \sigma = N \) (where buoyancy waves exist with \( \theta \) small), resulting in horizontally and vertically propagating wave packets respectively. Variations in \( N \) and \( f \) can result in the refraction and reflection of wave energy. When \( f = \) constant, wave refraction will occur as \( N \) increases (a non-homogeneously stratified ocean) until they are reflected at a depth where \( N = \sigma \). Mowbray & Rarity (1967) and Turner (1973) have successfully demonstrated this relationship in the laboratory. Upon reflection, since the frequency of the internal wave is conserved, incident and reflected \( \theta \) must be symmetric with respect to a level surface rather than with respect to the reflecting surface (Munk, 1981). This has important implication for internal waves propagating from a point source on a sloping bottom (appropriate to internal wave generation on the continental slope) and for internal waves propagating into a sloping bottom, \( \alpha \), (appropriate to oceanic internal waves approaching the continental slope region) and are discussed next.

If we consider internal waves of frequency, \( \sigma \), in the (x-z)-plane propagating from a point source on the continental shelf edge where \( f = \) constant and the across-slope wavenumber \( k_y = 0 \), then an internal wave characteristic angle, \( \alpha \), can be defined by Eqn. 2.3 (Munk, 1981);

\[
\alpha = \tan \left( 90^\circ - \theta \right) = \frac{\sigma^2 - f^2}{N^2(z) - \sigma^2} \sqrt{2}
\]

When waves with a characteristic slope, \( \alpha \), propagate onto a wedge-like bottom with slope \( \alpha \), (such as the continental slope) three scenarios exist; (from Cacchione & Wunsch (1974), Huthnance (1989), Petrie (1975));

(i) \( \alpha > \alpha \) Energy of the travelling wave of frequency \( \sigma \) is reflected oceanward. The wave is reflected, and the topography is considered "steep bump".

(ii) \( \alpha < \alpha \) Energy of the travelling wave of frequency \( \sigma \) is reflected shelfward. The wave is transmitted, and the topography is considered "flat bump".

(iii) \( \alpha = \alpha \) One can expect an amplification of internal wave energy at frequency, \( \sigma \), producing an unstable bottom boundary layer
and a source of "beams" of internal wave energy that propagate both oceanward and seaward at the characteristic angle, c.

Along the northwest European shelf slopes internal waves do not generally undergo complete reflection at subtidal frequencies since \( c \gg \alpha \). Thus deep ocean and locally generated internal waves incident on much of the northwest European continental slope will be reflected shelfward and may become concentrated at the shelf break Huthnance (1981)*. This focusing of internal wave energy will be particularly enhanced in regions where steep upper slopes are incised by canyons.

Gordon & Marshal (1976) and Hotchkiss & Wunsch (1982), have observed high frequency internal waves to become trapped within the head of canyons. Near to the bed, Cacchione & Drake (1974) showed through a series of experiments, that this intensification can lead to strong near bed current shear, where the waves become unstable and break, cascading the energy into the turbulent frequency domain. Eriksen (1982) describes field results which suggest that sloping bottoms are indeed energy sinks for critical internal wave reflection.

The transition of internal wave energy to turbulence by critical internal wave reflection has also been studied using numerical models. The models are not discussed here in any detail but the key results are summarised as:

(i) Ivey & Nokes (1989) showed that at the critical frequency, a turbulent benthic boundary layer forms on a 30° bottom slope which varies in thickness over the wave cycle, the maximum thickness occurring during the up-slope phase of the cycle. This mixing period is then followed by a period of re-stratification.

(ii) Slinn and Riley (1996) showed that for slopes >20° a thermal front can develop near to the bed which produces strong vertical (relative to the bed) density gradients near to the bed. These strong gradients produce a region of strong down-slope flow. For slopes <10° the strong mixing occurred on the during the phase of the wave in which the wave set up a strong down-slope flow.

* On the northwest European shelf slopes the exception being the Porcupine Bank and Goban Spur.
(iii) Ivey et al. (1995) presented similar results showing that the periodic turbulent mixing event can occur over the frequency range $0.5 < (\sigma/\sigma_c) < 2.5$ and are not limited internal waves of critical period.

Locally increased near bottom current, bed shear and near-bed turbulence are thought to be strong enough to erode sediment on the sea bed at critical depths (e.g. Wunsch & Hendry, 1972, Baines, 1974 & Cacchione & Drake, 1986). Field observations of periodic boundary layer mixing events at tidal frequencies (as described by the numerical model results above) are described in section 2.2.4 and are thought to account for the formation of bottom and intermediate nepheloid layers (BNL's and INL's). These layers facilitate the offshore transport of fine particulate matter from many of the world's continental slopes. Observations of sediment erosion at sub-tidal frequencies are not widely reported. Karl et al. (1986) describe several canyons where they hypothesise that the intensification of the near-bed currents due to higher frequency internal waves could account for the large sand waves that are observed at each canyon head. The wavelengths of the internal waves are shown to be comparable to that of the sandwaves. Karl et al. (1986) estimated that local increases to the bed shear stresses would be sufficient to mobilise the bottom sediment, although no direct measurements were taken. Southard & Cacchione (1972) describe a series of experiments in which they simulate interfacial waves propagating on-shelf on a sloping bed. Although not to scale for the oceanic case the results showed that sediment did not move in suspension or as bedload until the point of wave breaking. At the point of breaking however the sediment is thrown into suspension just up-slope from the breaker point and then transported down-slope by the return current as bedload. From near bed in situ measurements of the water turbidity and from current meter observations on the Middle Atlantic Bight, Churchill et al. (1994) observe that high frequency internal waves are an important mechanism for the mobilisation of sediment at the shelf-slope boundary.

Direct turbulence measurements in critical regions of the slope are logistically very difficult to collect and until recently have not been possible. The 1990 STABLE deployment on the La Chapelle Bank (chapters 3-6) describe the first known in situ
observations of suspended and bedload sediment transport and near-bed turbulent shear at critical internal wave depths.

Finally, an important source of high frequency internal waves at continental margins are shelf-slope fronts (Flagg, 1988) since $N$ is locally increased. In regions where the front intersects the bottom, Flagg (1988) showed a direct correlation between shoreward propagating high frequency internal waves and sediment mobilisation. If the shelf-slope front is subject to strong cross-shelf advection by the barotropic tide, as would be the case (albeit hypothetically) for the Celtic Sea shelf edge, then the above hypothesis could provide an important source of turbulent energy for bottom resuspension and bottom mixing.

In summary, the intensification of high frequency internal waves near to the bed on the continental slopes of the northwest Europe could provide a mechanism for the resuspension of sediment on the slopes, which could then be advected with the residual current or the barotropic tide. The implication for internal waves of tidal frequency is now discussed.

2.2.4 Internal tides

This section is devoted to the study of internal waves of semi-diurnal period (namely 'internal' or baroclinic tides), on the northwest European shelf-ocean boundary. They have reduced length scale $0(10\text{km})$ when compared to the barotropic tide $0(1000\text{km})$ owing to the reduced density difference in the ocean interior, and are thus subjected to advection by the barotropic tide. In some instances the shelfward propagating internal tide can become arrested or even advected oceanward during off-shelf barotropic tidal flow on the Celtic shelf (Pingree et al., 1983).

Along much of the Celtic Sea shelf edge satellite images reveal a cold band of water during summer stratified conditions, which are associated with regions of increased nutrients and phytoplankton growth (Pingree & Mardell, 1981). These 'cool' regions
are associated with the production of large amplitude internal tides at the shelf break with crest to trough amplitudes of ~ 60m (Pingree et al., 1986) The internal tide is highly non-linear in form (section 2.2.4.2) at the shelf break due to the complex interaction of the internal tide with the barotropic tide and topography, and is thought to be responsible for enhanced mixing and the upwelling of nutrients (e.g. Briscoe, 1984, Sandstrom & Elliott; 1984, Holligan et al., 1985, New & Pingree, 1990a). Increased localised mixing causes a broadening of the thermocline across the shelf break during summer stratified conditions and could imply (via geostrophy) variations in the along-slope current magnitude (Pingree & Mardell, 1981).

The focus of this study however, are the implications for sediment transport both on the upper slope (New, 1987) and at the shelf edge (Heathershaw et al., 1987), which have until recently, only been inferred from numerical models in the Celtic & Armorican shelf-slope region. The author would recommend the review of internal tides by Huthnance (1989).

2.2.4.1 Internal tide generation

Baines (1982) identified the slopes of the northwest European shelf edge as a region for the worlds largest generation of internal tidal energy, resulting from the interaction of strong barotropic tidal currents with steep topography. This is greatest at the approaches to the English channel along the Celtic and Armorican shelf edges where Baines (1982) calculated an onshore energy flux of ~ 2 x 10^3 Wm^{-1} onto the shelf. This is compared to ~62Wm^{-1} for the Malin and southern Hebrides shelves, where the barotropic tidal currents are of order 20cms^{-1} (Cartwright, 1980).

Internal tidal oscillations may be interpreted as horizontally propagating modes in the seasonal thermocline, being mode 1 on the shelf (New, 1988, Sherwin, 1988) and mode 3 in the ocean (New, 1988). By contrast, in regions where the density gradient is considerably reduced, the internal tidal energy is made up of a large number of modes and the energy exhibits beam-like propagation.
In the idealised situation of a two layer density structure, which is representative of internal tidal forcing of the seasonal thermocline at the shelf break, the off-slope (on-slope) flowing barotropic tide produces a depression (elevation) of the interface. This interface propagates away from the generating region both oceanward and shelfward as a decaying progressive wave as the off-slope (on-slope) flow slackens (e.g. Pingree et al., 1986). Shelfward propagating internal waves can become highly non-linear (at spring tides) and propagate away from the shelf break with bore-like characteristics, which can then break up into a series of internal solitary waves or internal solitons. The largest amplitude internal solitons should theoretically travel at the front of the wave packet with those behind becoming progressively smaller (Gerkema 1996). The vertical displacement of the interface gives rise to baroclinic tidal currents which superimpose onto the barotropic current, to enhance or reduce it locally.

Within the deep ocean the density gradient is considerably reduced with depth and internal tides are expected to propagate as "beams" of internal tidal energy along characteristic ray paths (Pingree & New, 1989; hereafter referred to as PN89), which emanate from "critical" generating regions on the continental slope (section 2.2.3). Internal tide generation occurs when the bottom slope, $\alpha$, matches the slope of the internal tide characteristic, $c$, defined by Eqn. 2.5, where $\sigma$ is the semidiurnal tidal frequency. For constant $f$, the generation of internal tides is determined by the stability of the water column and the gradient of the topography. Along the northwest European shelf edge the expected location for internal tide generation of $M_2$ frequency varies considerably (Huthnance, 1986). Along the upper slopes of the Bay of Biscay ($=47^{1/2}_{0} \, N$), $\alpha = c (0.051)$ at 385m (PN89); along the gentle slopes of Goban Spur $\alpha \neq c$ (section 6.5.2); on the Porcupine Bank ($=50^{1/2}_{0} \, N$) the upper slopes are again much gentler and $\alpha = c (0.05 \, & \, 0.035)$ at 1850 & 2850m respectively (Thorpe & White, 1988); and along the Malin and Hebrides shelf slopes ($55-57^{0}_{0} \, N$) $\alpha = c (0.05)$ at the shelf break at 200m (Huthnance, 1986).
Baines (1974) has shown that at the critical depth the amplitudes of the waves should become infinitely large. This is of course an assumption of linear wave theory and does not account for wave breaking and other non-linear effects. It is expected however, that at the generation site (where \( c=\alpha \)), and along characteristic "ray" paths of angle \( c \), the vertical displacement of the isotherms should be greatest, and will be characterised by an intensification of the baroclinic velocity amplitude (Petrie, 1975) and regions of instability and mixing (New, 1988). Wunsch & Hendry (1972) suggest that the intensification of baroclinic energy at tidal frequency is quickly dissipated into higher frequencies, which near to the bed would be associated with boundary layer turbulence rather than internal waves. The slopes are therefore regions of high dissipation of tidal energy which obviously have important implications for sediment transport (2.2.4.4).

2.2.4.2 Observations on the northwest European shelf

The structure of the internal tide ~25km shelfward of the Celtic Sea shelf break is shown in fig 2.5, which was obtained from a thermistor chain mooring in position 47°41.8'N 6°18.2'W (Pingree & Mardell, 1985). A strong variability over a spring-neap tidal cycle is clearly apparent. At spring tides the crest to trough amplitude of the internal tide is 50-60m (in a water depth of 170m) and is highly asymmetric, with flattened crests and steep 'bore-like' or 'soliton-like' troughs (other examples include Sandstrom & Elliott, 1984 and Holloway, 1987). The time of occurrence corresponds to spring tide vertically averaged \( M_2 \approx 80 \text{cms}^{-1} \). By comparison, the internal tidal signal on the Malin shelf edge is less coherent with oscillations within the main thermocline of ~25m (Sherwin, 1988). In fig 2.5 there is also evidence that the energy associated with the internal tidal trough is diffusing into a packet of two internal soliton waves with a wavelength of 1-1.5 km. The troughs are formed during off-shelf tidal streaming and propagate on-shelf against the tidal current causing the wave to steepen. In comparison, at neap tides the asymmetry disappears and peak to trough amplitudes of 13-17m are observed.
Current fluctuations that have been attributed to the shelfward propagation of internal tides have been observed at the La Chapelle Bank shelf edge near to the observations of Pingree & Mardell (1985). Heathershaw (1985) described internal wave 'pulses', 8 km shelfward of the shelf break that are phase locked with the barotropic tide, occurring regularly about 1-2h before peak on-shelf flow. Referring to them as internal tide current surges (ITCS's), Heathershaw (1985) observed currents with amplitudes of 30-40 cms$^{-1}$ and of 30-50 min duration at spring tides, which superimposed onto spring barotropic tidal currents to give an observed flows exceeding 100 cms$^{-1}$. The measurements showed a phase change of π/2 above and below the thermocline, enhancing the flow above and reducing it below, suggestive of a first mode internal wave. The ITCS's were also correlated with rapid temperature increases of 2-3$^0$C, consistent with the on-shelf propagation of a an internal tide with a steepened trough. Heathershaw (1985) proposed that since the ITCS's easily penetrated the bottom mixed layer they could, depending on their direction of propagation and phase (in relation to the barotropic tide) directly influence sediment transport at the shelf edge. This is discussed further in section 2.2.4.4.
Oceanward of the shelf break, the ray concept is more applicable for studying the near-bed implications of internal tide generation. In fact, the logistics for the present study were based on a field survey (PN89) across the upper slopes of La Chapelle Bank aboard the RRS Challenger, in September 1987. An investigation of the slope water stability and local bathymetry (section 3.1) determined the expected depth for the generation of internal 'ray-type' waves of semi-diurnal tidal frequency, where $\alpha = c = 0.051$ at 385m. Using an array of thermistor moorings across the slope, PN89 convincingly demonstrated the existence of a 'beam' of internal tidal energy initially propagating downward, then reflected on the ocean bottom (Pingree & New, 1990b) 140km from the critical slope in the deep waters of the Bay of Biscay. Fig 2.6 summarises the internal wave propagation observed by PN89. Close to the region of critical slopes the amplitude of the internal tide increases towards the bottom with near bottom amplitudes of 150m. The trough of this isopycnal depression occurred $1\frac{1}{2}$h after maximum off-shelf tidal streaming on the upper slopes.

![Fig 2.6 A summary of the internal tidal oscillations oceanward of the shelf break in the Bay of Biscay. Vertical lines represent CTD station numbers and the amplitudes of the internal tidal oscillations. The depth of the maximum amplitude at each station is shown by the dot and the dashed line represents a ray path passing as closely through these points as possible and is compared to the results of a numerical model by New (1988, section 2.2.4.3). The real topography is shown on the upper slope region by the bold line and is critical at 385m (taken from PN89).]
The results of PN89 confirmed model results (e.g. Prinsenberg & Rattray, 1975, Baines, 1982 and New 1987 & 1988; see next section) and laboratory simulations (e.g. Mowbray & Raritry, 1967), which suggested that the transfer of tidal energy away from the critical slope region and into the deep ocean should exist in nature.

The near-bed implications for internal tide generation on the slopes are now discussed. Near to and at the critical point on the slope, an intensification of baroclinic tidal energy is associated with a strong asymmetry in the magnitude of the cross-shelf currents on the upper slopes of the Celtic and Armorican shelves (Pingree, 1988). Current meter measurements in position 47°28'N 6°38'W [mooring 081], 33m off the bottom in 548m water measured maximum down-slope Eulerian currents of 95cms⁻¹ (Pingree, 1988). The Eulerian residual current was 15cms⁻¹ at this depth and was also orientated down-slope (Pingree & LeCann, 1989). The depth averaged residual current of 7cms⁻¹ at 548m water depth was poleward and along-slope with a small up-slope component. Both current and temperature profiles showed an advance of phase with depth (Pingree, 1988), which is consistent with the downward propagation of internal tidal energy from the critical slope region. Pingree & LeCann (1989) suggested that the internal tides on the sloping boundary layer were responsible for the observed net off-shelf and down-slope Eulerian mean currents near to the bed.

Elsewhere, Butman (1988) also observed a strong asymmetry of the across-shelf near-bed current strengths on the upper slopes south of the New England shelf. Maximum down-slope currents exceeding 40cms⁻¹ were observed 7mab in 200m water (in some cases stronger than flows recorded at mid-depths) which compared with up-slope flows of ~10cms⁻¹, producing a net Eulerian near-bed down-slope flow. Again the mean Eulerian flow 50-200mab was along isobath (0[5cms⁻¹]) with a veering of ~30⁰ between the mid-depth and near bottom mean flows. These characteristics were typical of the entire upper slope and with other observations made at water depths of 250, 500, 1150m. The observations of this asymmetry were again attributed to strong current fluctuations, primarily of tidal frequency (Butman, 1988).
Within canyons, internal tidal energy can be focused along the canyon axis accelerating the tidal flows at the head of the canyon far more than the adjacent slopes (Baines, 1983). Wunsch & Hendry (1972) also observed intensified near-bed motions in Hudson Canyon, in regions where the bottom slope becomes critical for internal tides. Once again the Eulerian residual flow was along-slope, but near to the bottom is was directed down-slope. Gardner (1989a) also observed this phenomenon in Baltimore Canyon, where the net transport of water 5mab at 275m depth was consistently down canyon.

Observations of near-bed mixing events on the continental slope region which are phase locked with the semi-diurnal tide and associated with critical internal tide reflection (previous section) are now described. It is first necessary to define the gradient Richardson number as:

$$R_i = N^2 \left( \frac{du}{dz} \right)^2 = -g \frac{d \rho}{dz} / \rho_0 \left( \frac{du}{dz} \right)^2$$

[2.5]

which is a measure of the stability of the water column over the depth interval $dz$. Values $< \frac{1}{4}$ (implying instability and mixing) were calculated by Pingree (1988) for periods of 1 hour in the lower half of the water column at mooring 081 on the Celtic slope.

On the Hebrides slope and Porcupine Bank, Thorpe et al. (1990) and White (1994) assumed that the T-S relationship remained constant, so that density inversions could be inferred from a vertical array of thermistors in the bottom 100m. Periodic mixed layers were found on both slopes at depths near critical or resonant for $M_2$ internal tide generation. These were characterised by a strong asymmetry in the temperature signal found over a $M_2$ period, with rapidly rising isotherms (decreasing temperatures/up-slope advection) being associated with periodic bml's with heights 30-50m. This followed a subsequent and more gradual fall in isotherms (rising temperatures and re-stratification) near to the bed. These mixed 'patches' (White, 1991) were identified as regions of low Ri number ($<\frac{1}{4}$) and low beam attenuation. The vertical array of thermistors showed evidence of a downward phase propagation of temperature on the Hebrides slope which Thorpe et al. (1990)
attributed to an upward propagating internal tide. Near-bed current meter measurements did not show any appreciable increase in magnitude during these mixing periods.

The asymmetry in the temperature record was also observed by Gardner (1989b) in Baltimore Canyon. Where the slope was critical for internal tides Gardner (1989b) described the generation of a 'bore' of cold water which advected up the slope during on-shelf flow. The turbulent nature of this 'bore' (rapid cooling events, dropping 2-4°C in < 2min) resuspended large amounts of sediment, which remained in suspension and was advected down-slope during off-shelf flow.

The above discussion has identified the generation of internal tides on continental slopes as a potential mechanism for the erosion of sediment and its subsequent inferred off-shelf transport with the mean flow. The net movement of suspended sediment remaining in the water column moves with the net transport of water between isopycnal surfaces. The Eulerian mean down-slope flow does not necessarily infer a down-slope transport of water (although it does imply the subsequent down-slope transport of sediment as bedload). This is the subject of discussion in section 2.2.4.4.

2.2.4.3 Modeling of internal tides

There are many analytical and numerical models of the generation of internal tides near to the shelf edge. These models have been extensively reviewed by Huthnance (1989), and recently updated by Huthnance (1995). This section will be restricted to the modeling exercises which specifically focus on the spatial and temporal evolution of internal tides on the slopes of the Bay of Biscay, where a strong barotropic tide, coupled with steep topographic slopes and summer stratification, can be simulated to produce to the observed intensification of near-bed oscillatory currents of tidal period (section 2.2.4.2). Other comparisons with models in northwest Europe include the oceanward propagation of internal tidal energy from
the slopes of the Rockall Bank (OeWitt et al., 1986) and the shelfward propagation on the slopes of the Malin shelf (Sherwin & Taylor, 1990).

The different approaches to the modeling are described by Huthnance (1989) as:
(i) vertical structure modal models,
(ii) 2 and 3-layer nonlinear models,
(iii) calculations in terms of the internal wave characteristic (section 2.2.4.1) and
(iv) full numerical models (3-dimensional) of realistic topography and stratification.

Table 2.1 lists a variety of numerical models developed for the study of, and comparison with, observations of internal tides on the Celtic & Armorican shelf edges (section 2.2.4.2). For the layer models [1], [2] & [3], the non-linear deformation of the internal tide is studied. This is achieved by balancing the non-linearity of the wave form with a diffusive term. This allowed temporal (e.g. spring-neaps) and/or spatial variations to be investigated without the waveform steepening to the point of breaking. Since the hydrostatic approximation is used (i.e. the length scale of the internal tide is much larger than the depth of the ocean) the solitary form of the waves could not be studied.

Pingree et al. (1984) showed by running their two layer model with and without rotation that waves amplitude dispersion due to the earth’s rotation allowed waves with near permanent form to exist with deeply penetrating troughs and flattened crests. These internal waves were caused by the advection of the barotropic current and were in agreement with observations by Pingree & Mardell (1981). Heathershaw et al. (1987) essentially reproduced this model (but without rotation and the inclusion of a more rounded shelf break profile) to study the implications for sediment transport over the La Chapelle Bank shelf break. They suggested that ITCS's at the shelf break could cause a net sediment transport off-shelf as the trough of the internal tide tries to advance on-shelf against the ebbing tide. On the assumption that the sediments would predominantly travel as bedload in this region (Heathershaw & Codd, 1986, mean grain size of 500μm), Bagnold's (1966)
<table>
<thead>
<tr>
<th>Author</th>
<th>Location</th>
<th>Model</th>
<th>Topography</th>
<th>Stratification</th>
<th>Coastal Boundary</th>
<th>Limitations</th>
<th>Aims of study</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pingree et al (1983), Heathershaw et al (1987)</td>
<td>Celtic Sea shelf break [La Chapelle Bank]</td>
<td>2-D, long wave equations, non-linear, [+ rounded edge topography]</td>
<td>level shelf &amp; uniform slope, 2-layer</td>
<td>Radiative</td>
<td>Although non-linear, the model cannot reproduce the observed bore-like onshelf propagating IT's. No dispersion and the non-linear effects are balanced by diffusive attenuation. To reproduce some idealized responses of the thermocline due to the barotropic tide and to examine sediment transport at the shelf break.</td>
<td>To reproduce some idealized responses of the thermocline due to the barotropic tide and to examine sediment transport at the shelf break.</td>
<td></td>
</tr>
<tr>
<td>Willmott &amp; Edwards (1987)</td>
<td>Celtic Sea shelf slope</td>
<td>2-D, non-linear, long wave equations.</td>
<td>level shelf &amp; uniform slope.</td>
<td>3-layer</td>
<td>Radiative</td>
<td>No realistic shelf break profile. No dispersion &amp; non-linear effects are balanced by diffusive attenuation. As for [1], plus the third layer allows for a more accurate reproduction of the observed stratification.</td>
<td>As for [1], plus the third layer allows for a more accurate reproduction of the observed stratification.</td>
</tr>
<tr>
<td>Serpette &amp; Mazé (1989)</td>
<td>Bay of Biscay region (~43-49°N)</td>
<td>2½-D, non-linear, long wave equations</td>
<td>arbitrary</td>
<td>2-layer</td>
<td>Radiative: 100m depth required to: (i) study localized open forcing, (ii) give an improved comparison with observations</td>
<td>As above + smaller grid size required to: (i) study localized open forcing, (ii) give an improved comparison with observations</td>
<td>To predict geographical variations in the intensity of internal tide generation, propagation, and the advective effects of the barotropic tides.</td>
</tr>
<tr>
<td>New (1987, 1988)</td>
<td>Celtic Sea shelf slope</td>
<td>2-D, Modal model (after PR75°), linear.</td>
<td>arbitrary</td>
<td>Radiative</td>
<td>Internal wave steepening within the thermocline &amp; non-linear wave-wave interactions and wave breaking are not reproducible</td>
<td>Internal wave steepening within the thermocline &amp; non-linear wave-wave interactions and wave breaking are not reproducible</td>
<td>To reproduce regions of large oscillations at tidal frequency which emanate from a critical point on the shelf slope. Also to predict regions of internal tidal mixing.</td>
</tr>
</tbody>
</table>

*IT - Internal Tide  † - Location of model simulation for comparison with observations.  + - Prinsenberg and Rattray (1975)
sediment transport equation\(^1\) was incorporated into the model and the results (fig 2.7) showed that the shelf break would be a region of bed-load parting, providing a mechanism for transporting sediment from the shelf break onto the upper slopes. The results of this model are discussed later in chapter 5.

Fig 2.7  The tidally predicted averaged transport rates \(q_{sb}\) at different locations across the shelf break (a) shown in relation to various schematic shelf-edge topographies in (b). \(q_{sb}\) is calculated for \(h_1 = 60\,\text{m}\) and \(\delta p = 5 \times 10^{-4}\,\text{g cm}^{-3}\). Positive values of \(q_{sb}\) correspond to on-shelf sediment transport.

\(^1\) the sediment transport rate \(q_{sb} = (u^2 - u_{ex}^2) \cdot u\,(\text{g cm}^{-1}\,\text{s}^{-1})\) where \(u = \text{velocity} \, \text{mab using the relationship}\ u = \{u_0/[\ln (h_2/ez_0)] \cdot \ln [100/ez_0]\} \times [100/ez_0]\) by integrating a logarithmic velocity profile over the lower layer (Edwards et al., 1984).
The Willmott & Edward (1987) 3-layer model essentially increased the vertical resolution of the two layer model of Pingree et al. (1984). Serpette & Mazé (1989) described the spatial distribution of internal tides in the Bay of Biscay with their 2½-D layered model. This model showed that the generation of internal tides take place over the whole of the Bay of Biscay with maximum values in the Armorican Celtic shelf regions, especially along the shelf break between 6° & 7°W. This includes the La Chapelle Bank region and the observed ITCS's by Heathershaw et al. (1987). Recently, Pingree & New (1995) have demonstrated the potential of remote sensing in the study and comparison of the spatial and temporal characteristics of the internal tides in the Bay of Biscay.

Finally, New's (1988) linear, modal model [4] of internal tide generation in a continuously stratified ocean (but with a linear shelf slope) successfully predicted the propagation of energy from critical points on the Celtic slope as observed by PN89 (see fig 2.6 for the comparison with observations). During summer stratified conditions New (1988) predicted that modes 3 & 4 would dominate the seasonal thermocline oceanward of the shelf break, and that mode one would dominate on the shelf, with thermocline depressions emanating from the shelf break at the time of maximum ebb. Further, during summer stratified, spring tide conditions, numerically calculated values of Ri were found to be < ¼ near to the bed at the critical slope, and within the seasonal thermocline, in regions where the upward propagating rays cross. Pingree (1988) observed regions near to the critical slope where Ri < ¼, and New & Pingree (1990) observed regions within the thermocline where Ri < ¼. An intensification of the baroclinic tidal currents is also expected within the beam and at the generation point, producing maximum baroclinic currents of 50cms⁻¹, which could potentially produce combined currents near excess of 100cms⁻¹ near to the bed as was observed by Pingree & LeCann (1989).
2.2.4.4 Implications for sediment transport

The formation of bottom nepheloid layers on the northwest European continental slopes imply that the critical shear stress is periodically exceeded for a certain fraction of the bottom sediment. The increase in optical beam attenuation on the slopes of the Celtic shelf (Pingree, 1988) and Hebrides slope and Porcupine Bank (Thorpe & White, 1988, Thorpe et al., 1990 and White, 1991), and the subsequent formation of nepheloid layers are consistent with the observed intensification of baroclinic tidal motions described in section 2.2.4.2. Other observations of nepheloid layers are reported by Dickson & McCave (1986) on the Porcupine Bank on a region of the slope not thought to be critical for internal tides. In this instance a temporary increase in $N$ occurred during an upwelling event and temporally reduced the critical angle for internal $M_2$ tides, $\alpha$, to be equal to the local slope, $\alpha$.

Observations of a mean Eulerian down-slope current near to the bed (section 2.2.4.2) have been used to implicate internal tides in a net flux of fine grained suspended sediment down-slope. A net down-slope advection of slope water must imply that either the bottom flow becomes unstable and mixes with the water above, or the Lagrangian flow must separate from the bottom and enter the interior of the ocean along isopycnal surfaces. Houghton et al. (1994) showed that during particularly energetic down-slope flows some of the off-shelf bottom flow must intrude into the interior of the water column over the slope. This mechanism produces intermediate nepheloid layers and assumes that during off-shelf tidal streaming, some of the particulate matter will settle out to greater depths, unable to return on-shelf during on-shelf flow. This would provide a direct route to the deeper ocean. Both Shaw et al. (1994) and Houghton et al. (1994) observe turbid water departing from the bottom boundary layer and moving seawards along isopycnals in the presence of a mean Eulerian down-slope flow, albeit not thought to be due to internal tide intensification in this instance. It would seem that canyon heads are a primary location for the transport of significant amounts of sediment off the shelf break and upper slope region and into the deep ocean. If the resuspension of sediment is intense, then it is feasible that sediment could be transported directly oceanward in
the form of gravity currents. Field observations implicating internal tides in destabilising sediments at canyon heads and thus initiating turbidity flows have not been reported to date, but it does seem a feasible episodic transport mechanism in some of the more tidally energetic continental shelf edge regions.

In summary a schematic is presented below from Gardner (1989b, fig 7) which summarises the internal wave mechanism which causes periodic re-suspension and down-slope transport of sediment in Baltimore Canyon. This hypothesis would also seem equally applicable for the upper slopes of the Celtic shelf in the Bay of Biscay (fig 2.8).

"At regions on the slope where the slope is critical for internal tide generation, a bore of cold water is generated which advects rapidly up the slope entraining high concentrations of suspended sediment."

"The water then sloshes back down-slope until it arrives at a layer of equal density, where the suspended sediment layer spreads out along an isopycnal. The particles then settle out into deeper water."

Fig 2.8 Periodic resuspension of sediment on the continental slope due to focusing of internal tides (taken from Gardner, 1989b)

The unique feedback mechanism for this hypothesis is that the erosion of material on the upper slopes maintains or encourages the internal tide to remain critical and to continue to erode the canyon shelfward. It is interesting to speculate whether the numerous canyons found meandering along shelf-ocean boundaries, in regions that are tidally energetic, are not only a conduit of sediment to the deep ocean (Gardner, 1989a), but also a lifeline to its very existence? This would imply that critical slopes only occur in canyons and this is certainly not the case. It is true however that in
such critical regions the enhanced near-bed shear may still not be large enough to erode the sediment (e.g. Wunsch & Hendry, 1972). The critical threshold velocity (chapter 5) for bed erosion may only be exceeded in canyon heads where there exists a combined near bed intensification of the flow due to critical internal tide reflection, accelerated barotropic tidal currents and high frequency internal wave trapping.

2.3 Schematic hypothesis for the La Chapelle Bank shelf-slope field study

A schematic hypothesis is presented in fig 2.9 for the La Chapelle Bank shelf-slope field study. This picture essentially summarises this review chapter and has focused on the cross-slope hydrodynamics which are significant for sediment transport in regions of strong barotropic tidal currents, steep slopes and a meandering along-slope topography.

Observations of large amplitude internal tides have been described in the Celtic Sea region propagating both on and off-shelf. The implications to sediment transport shelfward of the shelf break have also been reviewed. Observations oceanward of the shelf break have largely been inferred from near bed mean logged (e.g. 10 minutes sampling period) current meter measurements and in situ optical turbidity probes. This review has highlighted the need for near bed turbulent current measurements and in situ measurements of suspended sediment and bedload transport on sloping regions that are critical for internal tide generation. This was the fundamental goal of the field program and has provided valuable information of the interactive internal tide-sediment transport mechanism. Finally the specific hypotheses of the field program are listed below (in numbers (i) - (iv)). The final hypothesis (iv) is included in light of a second opportunistic Celtic slope deployment of STABLE. It suggests that hypotheses (i)-(iii) are very site specific and that near-bed sediment transport fluxes are highly variable along the upper slopes of the Celtic shelf.
Fig 2.9 A summary of the hypotheses tested at the La Chapelle Bank field site.

(i) A concentration of turbulent current and suspended sediment measurements is required on the upper slope region of La Chapelle Bank which is critical for internal tide generation. It is hypothesised that sediment is being preferentially transported oceanward at this location.

(ii) The bed stresses at this critical location periodically exceed the critical threshold for mobilisation and resuspension of bed sediment. The critical slope is a site of active sediment erosion.

(iii) Strong interfacial depressions of the seasonal thermocline at the shelf break are coincident with a region of bed-load parting across the shelf slope/break.

(iv) In regions where the cross-slope tidal currents are much weaker the residual transport of fine material will be in the direction of the residual current. This is predominantly poleward and along slope at the northwest European continental margins.
1990 La Chapelle Bank shelf-slope field study: Site selection and preliminary measurements.

3.1 Location

To test the shelf-slope tidal processes hypothesis outlined in section 2.3, the field program concentrated on a transect across the upper slopes of the La Chapelle Bank continental slope, (fig 3.1), during the RRS Discovery Cruise 193 of July 1990. The general locality is an area of steep slopes where the intensity of barotropic and baroclinic tidal streaming are known to dominate the across-shelf hydrodynamics on the upper slopes (e.g. Pingree et al., 1986). A meandering along-slope topography is incised by many canyons which themselves generate a complexity of hydrodynamic processes (e.g. Baines, 1982, Huthnance, 1981, reviewed in chapter 2).
Pingree & New (1989, PN89) found that the critical depth for internal tide generation on the slopes of La Chapelle Bank coincided with a minimum in the stability profile of less than 1 cph (see figure 3.2a). This minimum was also found to extend throughout the tidally averaged section (see figure 3.2b), at a depth of 350m, impinging on the slopes at a depth of 385m. A seasonal study of the slope water stability found N to vary from 0.7 to 1.0 cph during March to September 1987. This does not have a large effect on the critical depth for the generation of the internal tidal energy, but does cause it to advect seasonally up and down the slope on a small scale, the magnitude of which is dependent upon the seasonal change in stratification.
The observations of PN89 on the upper slopes of La Chapelle Bank led to the deployment of a benthic lander at the critical depth of 385m for internal tide generation. The deployment was the first of its kind along the Northwest European continental shelf slope.
3.2 Equipment and measurements

The deployment site of the benthic lander 'STABLE' (see section 3.2.1) at the critical depth of ~385m, was located at the head of a canyon, at the deep water end of a transect of moorings across the upper slopes of La Chapelle Bank (fig 3.3). Mooring 124 (section 3.2.2) was located at a depth of 305m, and mooring 123 (also section 3.2.2) was located at 200m, approximately 4km shelfward of mooring 124. The 200m contour is defined as the shelf break. The transect of moorings across the upper slopes were to enable a study of the thermocline response to tidal forcing at the shelf break, and to correlate this to the near-bed response at the critical depth of 385m in the continuously stratified slope waters.

Fig 3.3 Bathymetry (in metres) of the 1990 survey area on the upper slopes of the La Chapelle Bank continental shelf slope. Also shown are the deployment sites for STABLE and moorings 123 & 124. The three Shipek grab bottom sediment samples are labelled S1, S2, & S3, and the continuous CTD yo-yo station as CTD 140Y.
During the study three Shipek grab samples were taken in the local vicinity of the STABLE deployment. A continuous CTD yo-yo station (defined as CTD 140Y in fig 3.3) is also presented in the following discussion, to aid the interpretation of the mooring data.

3.2.1 Sediment Transport and Boundary Layer Equipment (STABLE)

The ‘pop-up’ version of this benthic lander (Plate 3.1) is one of two rigs that have been developed by the Proudman Oceanographic Laboratory (Humphery, 1987) to comprehensively measure tidal and wave induced currents, and the response of the sea bed in the re-suspension and advection of sediment. ‘STEEL’ STABLE is its shallow water equivalent, and is heavier and more robust, so that the large northwest European continental shelf tidal currents can be withstood. Several successful deployments of STEEL STABLE have recently been made which have provided a further insight to the sediment mobility in the southern North Sea (e.g. Soulsby & Humphery, 1989, Huntley et al., 1991, Huntley et al., 1993, Huntley et al., 1994 and Williams, 1995). There are numerous examples of bottom mounted structures being used in the investigation of near-bed hydrodynamics and sediment dynamics (reviewed by Green et al. (1992) and more recent examples include Green et al. (1995), Cacchione et al. (1995) and Voulgaris et al. (1995). All deployments published to date have either been in coastal or shelf sea environments. The deployment of ‘pop-up’ STABLE (here on referred to as just STABLE), is believed to be the first ‘deep-sea’ deployment of a benthic lander on the slope region of the world’s continental margins. The deployment was therefore as much of a learning process for the engineers at the Proudman Oceanographic Laboratory, who built the rig, as it was a scientific exercise. It is a tribute to their skill that, though some instrumental problems were encountered, the deployment yielded unique and very valuable data.

In order to study the boundary layer physics, and the subsequent transport of sediment on the slope, the following parameters have to be determined (from Humphery, 1987);
Plate 3.1 STABLE instrumentation: A, Two electromagnetic current meters; B, four vertically stacked Aanderaa rotor current meters; C, two Optical Backscatter Sensors (OBS's); D, Acoustic Backscatter Sensor (ABS); E, camera; F, flash unit; G, shadow bar; (pressure sensor not shown). (i) STABLE deployed at 0105 hrs on 3rd July, 1990 in 388 metres of water. (ii) STABLE successfully recovered at 0910 hrs on 25th July 1990.
i) Short-term, turbulent current
ii) Long term steady currents
iii) Pressure
iv) Suspended sediment flux
v) Nature of the bottom.

Table 3.1 describes the instrumentation fitted to STABLE in an attempt to adequately determine the near-bed processes outlined in points (i)-(v) above. A more comprehensive account of the design, self-logging, and instrumentation aspects of the rig can be found in Humphery (1987). Pre-deployment and post-deployment photographs of the STABLE rig are shown in Plates 3.1(i) and 3.1(ii) respectively, and the individual sensors that are described in table 3.1 are clearly labelled. For help in the interpretation of the data set, sensor heights are identified in Plate 3.2 in cm above the bed.

The only sensors not shown in plates 3.1 & 3.2 are the Digiquartz pressure transducer, and the compass and pitch and roll sensors. The pressure transducer is contained within the main body of the rig, 1.6m above the seabed. The compass and pitch and roll sensors are contained within the logger tube (see Humphery, 1987 for further details) at a height of 1.4m above the seabed.
### Table 3.1 Summary of STABLE instrumentation for the 1990 La Chapelle Bank deployment.

<table>
<thead>
<tr>
<th>Data collected</th>
<th>Instrumentation</th>
<th>Sampling frequency</th>
<th>Height, (z), above bed (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A)* Short term 'turbulent' current</td>
<td>2 electromagnetic current meters (EMCM's) mounted orthogonal to each other, measuring U,V and a comparable W component of the flow (see fig 3.4).</td>
<td>4Hz for 9 minutes every 2 hours.</td>
<td>z=39.3cm</td>
</tr>
<tr>
<td>(B) Tidal currents and current shear.</td>
<td>A vertical stack of 4 Savonius Aanderaa rotor (RTR) current metres, and direction vane.</td>
<td>Continuous 1 minute sampling, (number of counts per minute).</td>
<td>z=27, 45, 63 &amp; 81cm for RTR's 1 to 4 respectively.</td>
</tr>
<tr>
<td>Tidal elevation</td>
<td>Digiquartz pressure transducer.</td>
<td>Continuous 1 minute sampling</td>
<td>z=160cm</td>
</tr>
<tr>
<td>Rig orientation</td>
<td>Compass, pitch &amp; roll sensors</td>
<td>Continuous 1 minute sampling</td>
<td>z=140cm</td>
</tr>
<tr>
<td>Suspended sediment</td>
<td>(C) 2 Optical Backscatter Sensors (OBS's)</td>
<td>2 hourly bursts for 9 minutes, synchronised with the EMCM bursts</td>
<td>z=11.3 &amp; 41.8cm for OBS 1 &amp; 2 respectively.</td>
</tr>
<tr>
<td></td>
<td>(D) Acoustic Backscatter Sensor (ABS)</td>
<td>3 MHz transducers synchronised with EMCM burst data. Sampling at 0.5 Hz for 512 sec every 2 hours. System has a pulse repetition rate of 155Hz, pulse length of 155(\mu)s and beam width of 1.4°.</td>
<td>z=120cm looking vertically down in front of the EMCM's (10cm starboard offset). The distance of the ABS transducer beam forward of the line joining the centre of the EMCM coils was 25cm.</td>
</tr>
<tr>
<td>Sea-bed stability</td>
<td>(E) Camera &amp; (F) flash unit, &amp; (G) shadow bar.</td>
<td>Photograph taken every hour, synchronised with the beginning of each EMCM burst.</td>
<td>Photograph covered an area of approx. 42cm X 29cm.</td>
</tr>
</tbody>
</table>

* Letters in brackets are consistent with plates 3.1 & 3.2 for labelling individual sensors
Plate 3.2 Close-up view of the main sampling area of the STABLE rig. The same notation is used as in plate 4.1, with the height of each instrument above the bed in brackets. The ABS (not shown) looks vertically down to the sea-bed at a height of 120cm.
The orientation of the electromagnetic current meters (EMCM's) relative to true north are shown (not to scale) in figure 3.4. The compass sensor indicated that the EMCM spar bears 286° from magnetic north for the entire deployment. The deviation of magnetic north to true north was obtained from the Hydrographic Office, Taunton. The magnetic deviation, $\phi$, for position 47°30'N 06°30'W in July 1990 was approximately 007°W (annual rate of change of 7.6'E). The heads of the EMCM's were orientated so that the port EMCM head (No.6763) measured a positive horizontal component of the flow along 144°T, and the starboard EMCM head (No.6764) measured a positive horizontal component of the flow along 054°T. The starboard EMCM therefore measured a relatively undisturbed cross-shelf component of the flow, whilst the along-shelf, northward flow would go through the tripod before being read by either EMCM. The turbulent current measurements from one or both of the EMCM's would be relatively undisturbed by the tripod when the mean current flowed from the direction 144-360°T and 0-54°T. Finally, both port and starboard EMCM's measured a duplicate vertical component of the flow.

![Diagram](image)

**Fig 3.4** Orientation of the EMCM's on STABLE (not to scale). $\phi = 007°W$ is the deviation of magnetic north from true north. The EMCM spar bears 286° relative to magnetic north and vertical flow is +ve upwards.

This interference problem is also of concern for the Savonius rotor current meter stack, in so much as the results suggest a possible shielding problem for the rotor.
nearest to the main body of the rig (i.e. rotor 4). The data limitations to the mean flow measurements are discussed in section 5.2.2.1.

The pitch and roll sensors gave near constant readings of $2.8^\circ$ and $5.2^\circ$ respectively, providing an estimate for the bottom slope, $\alpha$, equal to 0.103 (discussed further in section 4.2.2). Thus the tripod was mounted securely on the seabed throughout the deployment.

Two differing suspended sediment sensors were fitted close to the EMCM heads in order to quantify the suspended sediment concentration in the near-bed region. The two Optical Backscatter Sensors (OBS's), have a high frequency response which makes them ideal for looking at the sediment response at turbulent length scales (Osborne et al., 1994). However, because the beam of infra-red light source has a short path length in water, a vertical array of sensors are required if one is to obtain a profile of suspended sediment. The 3MHz Acoustic Backscatter System (ABS) on the other hand, is not as greatly attenuated by absorption and can penetrate the water column by up to a meter or so. By range gating the return acoustic backscatter signal, a vertical profile of suspended sediment can be obtained from the concentration estimates at each range "bin". By mounting the ABS 1.2 meters above the bed looking vertically down in front of the EMCM's, an attempt was made to obtain vertical profiles of suspended sediment concentration from which sediment flux estimates could be calculated. Its 0.5 Hz sampling frequency was also synchronised with the other STABLE burst records (see table 3.1).

For both the OBS, and the single frequency ABS, the post calibration for absolute concentration is complex, because sensor output depends strongly on suspended particle size and provenance, as well as concentration. The two sensors are most commonly used in the nearshore zone where field calibrations are a matter of routine, ensuring as great a confidence in the results as possible. The OBS for example, is particularly sensitive to sediments with a large grain size (Osborne et al., 1994), which was the case for the region under study at the STABLE deployment site (section 3.2.2.2). Ambient sediment calibrations are essential in this instance.
Unless in situ near-bed water samples are taken from the field site, concentration estimates are not reliable. The OBS sensors are therefore invaluable in providing spot height inter-calibration checks for the ABS suspended sediment profile. Other studies have shown good agreement between the two (e.g. Osborne et al., 1994, found agreements to within ±10% for concentrations averaged over several minutes).

Finally, one of the most important features of the instrumentation mounted on the STABLE rig was the camera and flash unit. The unit was used to monitor the nature and stability of the sea bed. Observation of moving bedforms or bed sediment in sequences of photographs provides a mechanism for observing bedload transport (e.g. Huntley et al., 1991). More details of the camera system are presented in section 5.1.1.

3.2.2 Supporting Field Measurements

This section contains a summary of the across-shelf transect of current meter moorings and their position relative to the predicted upward propagating rays of internal tidal energy. A description of the nature of the sea bed on the upper slopes of La Chapelle Bank is also given.

3.2.2.1 Current meter moorings

Currents, temperature, and transmission were all derived from the three current meters located on each mooring (moorings 123 and 124). The upper and lower meters were Aanderaa RCM current meters (Table 3.2) with the mid-depth meters being the more rapid response S4 current meters. The lower current meters were both positioned 8m above the bed, of which CM8511 on mooring 124 also measured temperature and transmission, and CM7643 on mooring 123 measured temperature only. Temperature was also measured on CM421 on mooring 123, located 128m above the bed.
Table 3.2 Summary of Current Meter Moorings

<table>
<thead>
<tr>
<th>Mooring No.</th>
<th>Position</th>
<th>Depth (m)</th>
<th>Current meter No.</th>
<th>Distance above bed (m)</th>
<th>Start-stop date</th>
<th>Record length (days)</th>
<th>Other sensors</th>
</tr>
</thead>
<tbody>
<tr>
<td>123</td>
<td>47°31.1'N 200</td>
<td>CM421</td>
<td>128</td>
<td>02.07.90 - 23.07.90</td>
<td>20</td>
<td>Th -</td>
<td></td>
</tr>
<tr>
<td></td>
<td>06°32.0'W</td>
<td>S4</td>
<td>68</td>
<td>02.07.90 - 25.07.90</td>
<td>23</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>CM7643</td>
<td>8</td>
<td>02.07.90 - 25.07.90</td>
<td>23</td>
<td>Th -</td>
<td></td>
</tr>
<tr>
<td>124</td>
<td>47°28.9'N 305</td>
<td>CM7765</td>
<td>162</td>
<td>-</td>
<td>-</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>06°32.5'W</td>
<td>S4</td>
<td>85</td>
<td>02.07.90 - 25.07.90</td>
<td>23</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>CM8511</td>
<td>8</td>
<td>02.07.90 - 25.07.90</td>
<td>23</td>
<td>Th - &amp; Tr -</td>
<td></td>
</tr>
</tbody>
</table>

* Th - Thermistor, + Tr - transmissometer

The position of the current meters relative to the slope, and the critical depth for internal tide generation, (estimated from $\alpha=\phi$ [equation 2.5 in section 2.2.3]) are shown in fig 3.5a. Knowledge of the bottom slope is fundamental to understanding the propagation and generation of internal waves on the continental slope. The profile of the bottom topography along the mooring transect is also shown in fig 3.5a. The theoretical "ray" path for the main beam of the internal tidal energy is shown as it propagates away from the slope at the critical depth of 385m. The ray paths have been calculated using CTD data from Pingree & Morrison, (1973) & PN89, to calculate $N(z)$ profiles for $z>150$m (see fig 3.2a), and from the towed cycling CTD (Sea-Soar) data collected in September 1985 from Pingree et al. (1986) (see fig 3a in New, 1987), for $z<150$m. These profiles were used by New (1987 & 1988) to construct the numerical modal model of the topographic generation of internal tides described earlier in section 2.2.4.3.

The rays propagate away from the critical depth with a slope of around 0.05 ($N=1.7\times 10^{-3}$ rad s$^{-1}$, or 0.95 cph at 385m) but the upward propagating shelfward and oceanward rays become more horizontal as $N$ increases considerably through the seasonal thermocline, with a slope of around 0.004 ($N=1.3\times 10^{-2}$ rad s$^{-1}$, or 7.5 cph at 75m). This means that the rays take longer to pass through the seasonal thermocline (New, 1988) which may allow the interaction of the higher modes of the
Fig 3.5 (a) Across-shelf bathymetric profile of the mooring transect. The shelf break is defined by the origin of the x-axis, with the STABLE deployment at the depth shown to be critical for internal tidal generation of semi-diurnal period. Characteristic "ray" paths for the propagation of internal tidal energy are shown as a dashed line emanating from the critical point on the slope. These have been derived using data from Pingree & Morrison, (1973) and PN89.

(b) The critical period for internal wave generation as a function of distance across-shelf for the given bathymetric profile in (a) above. Calculations are again made using data from the two sources cited above. Note that internal waves of semi-diurnal period are critical at 6km offshelf corresponding to the STABLE deployment.
"ray" type wave with the forced motions of the seasonal thermocline to take place (New, 1987).

The author suggests that the higher period waves observed within the internal tidal troughs of the shelfward propagating wave by Pingree & Mardell (1985), may be due not only to the non-linearity of the wave as suggested by Pingree et al. (1983), but also by the interaction of the forced thermocline oscillation with the higher mode internal waves. Non-linearity, wave-wave interaction and viscous dissipation are all important factors which are not addressed in the internal tide model described by New (1988). This almost certainly accounts for the absence of observations in nature of upward propagating "rays", in regions where \( N \) can be an order of magnitude higher than the value given above for the seasonal thermocline.

The main objective of fig 3.5a is to estimate the regions of maximum currents which have been predicted to be along the theoretical ray paths by New (1987). Maximum currents are therefore expected at the generating site (i.e. the STABLE deployment site), and also at the depth of the S4 current meter on mooring 123, located 130m below the surface.

The bathymetry shown in fig 3.5a has then been used to estimate the critical period for internal wave generation as a function of distance across-shelf (following Wunsch & Hendry, 1972) in fig 3.5b. The accuracy of the calculation is very dependant on an accurate representation of the gradient of the slopes, and the information given in fig 3.5b should be used as a guide only. The critical \( M_2 \) period is at a depth of 385m, where upon the critical period continues to reduce with distance off-shelf to a minimum of approximately 5 hours at a depth of 740m, which corresponds to the steepest slope. The slope then becomes less steep and becomes critical at \( M_2 \) tidal period once again at approximately 810m.
3.2.2.2 Shipek Grab Samples

Details of the nature of the bed in Table 3.3 below are derived from a detailed analysis of the three Shipek grab samples taken in the local vicinity of the STABLE site. This type of bed release grab sampling technique is very effective for the full range of fine sand to gravel bed material (Buller & McManus, 1979) that was expected at this site (Heathershaw & Codd, 1986) The sediment analysis was performed by 'Sediment Analysis Services', Somerset, UK. The mean Phi value, deviation and skewness statistics are provided in Table 3.3, with the cumulative distribution of the three sediments shown graphically in fig 3.6 for inter-comparison.

Table 3.3 Statistical analysis of the Shipek grab samples using the phi system of units

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Position</th>
<th>Depth (m)</th>
<th>Mean(\phi)</th>
<th>Deviation(\phi)</th>
<th>Skewness(\phi)</th>
</tr>
</thead>
<tbody>
<tr>
<td>S1</td>
<td>47°27.4'N</td>
<td>344</td>
<td>0.20</td>
<td>1.90</td>
<td>0.03</td>
</tr>
<tr>
<td></td>
<td>06°29.9'W</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S2</td>
<td>47°29.0'N</td>
<td>317</td>
<td>0.22</td>
<td>1.29</td>
<td>-0.26</td>
</tr>
<tr>
<td></td>
<td>06°33.2'W</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>S3</td>
<td>47°28.3'N</td>
<td>377</td>
<td>0.75</td>
<td>1.25</td>
<td>-0.14</td>
</tr>
<tr>
<td></td>
<td>06°33.3'W</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* Statistical parameters from Inman (1952)

The statistical parameters in Table 3.3 have been derived from the cumulative distribution curve in fig 3.6 rather from multi-modal analysis, which is unsuitable for the open ended curves shown above. The open ended nature of the grain size curve is typical of most natural sediments. The moment mean grain size at all three sites is indicative of a coarse sand. Sample S3 was taken closest to the STABLE site (approx. 0.3N miles away), and at the greatest depth, with a mean grain size of 870\(\mu \text{m} \) (\(\phi = -\log_2(\text{size in mm})\)). The cumulative distribution of all three sediment samples indicates an approximate log-normal distribution (by the approximate straight line), which is representative of a single sediment type.
Fig 3.6 A cumulative weight frequency curve for samples S1-S3.

The sorting, or Phi Standard Deviation for each sediment is >1, indicative of a poorly sorted, or non uniform sample. The Skewness (a measure of the symmetry of the distribution about the mean) skews negatively for two of the samples (i.e. towards the coarser grains) and is close to zero for the third sample.

In summary, the samples are characterised as a poorly sorted gravelly sand, with the lack of fine material being representative of a reasonable flow rate within the boundary layer (see section 4). The coarser fraction of each sample was composed mainly of carbonate material which appeared to be broken down chemically rather than physically (J. Malcolm, Sediment Analysis Services; pers comm.). The biota were particularly rich in samples S1 and S2. Unfortunately the sediments were not analysed for carbonate content or biota. Each sample could also have been used in flume tests to estimate a critical threshold for initiation of sediment movement for
the larger carbonate material. Estimates of the critical threshold velocity for this bimodal sediment are discussed in section 5.2.3.1.

3.2.2.3 Remaining Data Sources

Finally, the remaining data sources are shown below;

Table 3.4  Remaining data sources

<table>
<thead>
<tr>
<th>Data source</th>
<th>Position</th>
<th>Depth (m)</th>
<th>Start / stop date</th>
<th>Record length (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>STABLE</td>
<td>47° 31.1'N 06° 32.0'W</td>
<td>388</td>
<td>03.07.90 / 23.07.90</td>
<td>23</td>
</tr>
<tr>
<td>CTD140Y</td>
<td>47° 29.3'N 06° 34.8'W</td>
<td>-</td>
<td>11.07.90 / 12.07.90</td>
<td>0.62</td>
</tr>
</tbody>
</table>

3.2.3 Data Limitations

STABLE was deployed at 0105 hrs on July 3rd 1990 (Table 3.4), in 388m of water on the western edge of La Chapelle Bank, and recovered at 0910 hrs on July 25th, 1990. Not all of the instrumentation fitted to the rig collected data successfully, and a description of the usable data is given below in Table 3.5.

Perhaps the potentially most useful data set not to be collected was the directional information for the rotor stack. The EMCM's did however provide an accurate description of the mean current velocity at two hourly intervals after utilising the rotor speed time series for the calibration of the EMCM offsets. This calibration is described in detail in section 4.2.2.
Table 3.5  Data collected from the 1990 STABLE deployment

<table>
<thead>
<tr>
<th>Sensor</th>
<th>Start / stop dates</th>
<th>Record length</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>EMCM bursts</td>
<td>0219hrs, 03.07.90 / 0819hrs, 10.07.90</td>
<td>88 bursts for 7 day neap/spring cycle.</td>
<td>The burst logger failed with 15 days of the deployment remaining</td>
</tr>
<tr>
<td>Optical Backscatter System (OBS)</td>
<td></td>
<td></td>
<td>No sensible data was collected during the deployment.</td>
</tr>
<tr>
<td>Mean logged data sets</td>
<td>0320hrs, 03.07.90 / 0323hrs, 11.07.90</td>
<td>Continuous 1 min sampling for 8 days</td>
<td>Rotor 2 (45cm above the bed), and the direction vane gave no readings throughout the deployment. Data recovered for rotors 1, 3, &amp; 4, pressure, compass, pitch and roll, sensors.</td>
</tr>
<tr>
<td>Acoustic Backscatter System (ABS)</td>
<td>0219hrs, 03.07.90 / 0619hrs, 24.07.90</td>
<td>256 backscatter records synchronised with the EMCM burst samples</td>
<td>The 512 second, 2 hourly backscatter returns arranged into 7 bins, giving sediment concentration profiles at 13, 20, 27, 40, 55, 79 &amp; 103cm above the bed.</td>
</tr>
<tr>
<td>Camera and flash unit</td>
<td>0219hrs, 03.07.90 / 2019hrs, 22.07.90</td>
<td>1 sea-bed photo hourly. (frame Nos. 7-474).</td>
<td>Frame Nos. 1-7 are pre- and during deployment photographs.</td>
</tr>
</tbody>
</table>

It should also be pointed out that the relative changes in concentration from the ABS are reliable, but the absolute values can not to be interpreted with any confidence, as sediment size is unknown. The backscatter signals are representative of small quantities of sediment in suspension. Concentration is in milligrams/litre, assuming sand sized material in suspension (based on the results of the sediment size analysis of the Shipek grab samples (Table 3.3)), but the exact sediment size distribution in suspension at the STABLE site remains uncertain. Bottle samples were not collected on the STABLE rig during the deployment.
The calibration of the Digiquartz pressure sensor has a discrepancy with the echo sounder concerning the deployment depth for the STABLE rig. The echo sounder gave a depth of 388m, whereas the mean pressure record suggests a 412.4m deployment. The detailed bathymetry shown in figure 3.3 also suggests a depth close to 388m for the given documented GPS position for deployment. The pressure sensor was calibrated pre deployment using a third order polynomial, with constant coefficients measured for a temperature of 9.2°C. Other temperature coefficients were estimated ranging from 0-21°C, but this was the closest calibration to the expected benthic temperature of 11-12°C at the deployment site. In the following discussion, no attempt has been made to change the offset of the mean pressure sensor value, to that measured by the echo-sounder.

There were also data limitations with the mooring data. Current meter data collected from Aanderaa CM7765 was extremely noisy and the results are not presented here. Finally, fig 3.7 shows that the direction data from the vane on Aanderaa CM8511, located 8mab on mooring 124 was also suspect due to compass sticking. Direction for this current meter has been obtained from the S4 current meter located 85m above the bed. This is discussed further in section 4.1.2.

![Fig 3.7 East verses north current velocity for CM8511 located 8mab on mooring 124, showing clear evidence of compass sticking.](image-url)
Chapter 4

La Chapelle Bank Field results – Dynamics

4.1 Mooring Stations 123 & 124: An introduction

Sedimentation processes are forced by a hierarchy of fluid motions with differing length and time scales. Since the mooring data sets are only of the order of 20 days duration, filtering the records to resolve lower than tidal frequency fluctuations did not, in this instance, resolve anything other than the residual current.

This chapter begins by describing the smaller wavenumber processes of the tides and residual currents. A general description of the time series data will be given first in section 4.1.1, followed by the results of tidal analyses of the current and temperature data in section 4.1.2. The residual 'slope' current and the observed variations over a neap-spring tidal period are also discussed in this latter section. The energy spectra for each current and temperature record are analysed in section 4.1.3. An investigation into a specific two day period during spring tide conditions will identify a specific baroclinic tidal 'event' in the bottom boundary layer at mooring 123 at the shelf break (section 4.1.4). The response of the bed to $M_2$ tidal forcing is considered in section 4.1.5 from near-bed temperature and transmissometer observations. Finally a brief discussion in section 4.1.6 will bring together previous sections of 4.1 by summarising the near bed responses of the upper slope and shelf
break region to the observed internal tidal motions. Section 4.2 will then proceed to correlate these results with the near-bed current measurements at the STABLE deployment site, located in 388m water depth. A considerable amount of time has been spent calibrating the electro-magnetic current meter (EMCM) data since it was initially thought to be of limited scientific value. The analysis of this data set is described in details in section 4.2.

4.1.1 Time series

The time series plots of the current, temperature and transmissometer data recovered from moorings 124 & 123 are shown in figures 4.1-4.5. Data recovery does not include CM7765 (162m above the bed on mooring 124) for this discussion (see data limitations in section 3.2.3). All of the time series in figs 4.1-4.5 are clearly tidally modulated. The cross shelf direction for all six current meters on moorings 123 & 124 is defined topographically from fig 3.3 as 020°N, with positive long shelf current velocities resolved along 110°.

4.1.2.1 Mooring 124

Time series plots of current, temperature and transmission for CM8511, located 8mab are shown in fig 4.1. There is a strong asymmetry in the cross shelf velocity form at spring tides. This asymmetry favours down-slope flow with a mean current of 4.5cms⁻¹ and a maximum recorded down-slope flow of 50.9cms⁻¹ occurring during Julian day 203 (table 4.1). This compares to a maximum instantaneous up-slope current of 33.6cms⁻¹ for the same tidal cycle. This asymmetry is not observed at neap tide which has currents of the order of 25cms⁻¹ in both directions (fig.4.3). A mean poleward slope current is also found in the along-slope time series data, the magnitude of which increases during spring tide.
Table 4.1  Current meter statistics for mooring 124 (Julian days 184-204 inclusive)

<table>
<thead>
<tr>
<th></th>
<th>CM8511 (8m above the bed)</th>
<th>S4 (85m above the bed)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cross-slope (cm/sec)</td>
<td>Along-slope (cm/sec)</td>
</tr>
<tr>
<td>Mean</td>
<td>-4.5</td>
<td>-4.8</td>
</tr>
<tr>
<td>SD'</td>
<td>20.7</td>
<td>12.5</td>
</tr>
<tr>
<td>Max*</td>
<td>33.6</td>
<td>22.7</td>
</tr>
<tr>
<td>Min*</td>
<td>-50.9</td>
<td>-46.8</td>
</tr>
</tbody>
</table>

Table 4.2  Current meter statistics for mooring 123 (Julian days 184-204 inclusive)

<table>
<thead>
<tr>
<th></th>
<th>CM7643 (8m above the bed)</th>
<th>S4 (68m above the bed)</th>
<th>CM421 (128m above the bed)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Cross-slope (cm/sec)</td>
<td>Along-slope (cm/sec)</td>
<td>Temperature (°C)</td>
</tr>
<tr>
<td>Mean</td>
<td>2.8</td>
<td>-3.2</td>
<td>11.8</td>
</tr>
<tr>
<td>SD'</td>
<td>29.7</td>
<td>20.5</td>
<td>0.2</td>
</tr>
<tr>
<td>Max*</td>
<td>77.1</td>
<td>43.7</td>
<td>12.3</td>
</tr>
<tr>
<td>Min*</td>
<td>-60.8</td>
<td>-52.8</td>
<td>11.3</td>
</tr>
</tbody>
</table>

* Max - maximum instantaneous onshelf flow, Min - maximum instantaneous offshore flow, SD - Standard deviation
Fig 4.1 Time series of cross shelf (+ve onshelf) and along shelf (-ve poleward) current, temperature and transmission at current meter station 8511.
The temperature time series is also of predominantly tidal period. A mean temperature of 11.4°C (shown plotted in fig 4.1) has maximum and minimum deviations of 0.8°C and 0.7°C respectively, but there is no long term advection of slope water up or down-slope during the deployment period. The most noticeable feature of this record is the rapid cooling event which occurs near maximum on-shelf $M_2$ tidal flow, when the bottom water is warmest. This is observed as a saw-toothed time series profile, and is more prominent at spring tide. The temperature time series is essentially responding to the cross-slope advection of slope water per tidal period, which is up-slope during on-shelf tidal streaming and down-slope during off-shelf tidal streaming. However, there does appear to be a phase lag between on-shelf tidal streaming and the response of the isotherms near to the bed, since the up-slope advection of slope water (slope water cooling) does not occur until maximum on-shelf flow. In conclusion, the data set is unable to resolve whether there is a net down-slope transport of slope water. The temperature observations suggest that the Eulerian residual down-slope transport of slope water shown in the cross-slope velocity profile does not infer a net down-slope transport of slope water (since there is no net warming or cooling throughout the entire sampling period), although it is possible there could be a down-slope transport together with changing (diffusive) water properties.

The transmissometer time series data is unique for this slope region (fig 4.1). The semi-diurnal, spring-neaps cycle is again clearly evident, even though the bit resolution of the signal is poor. The transmission is a measure of the optical attenuation of the water near to the bed and although not calibrated, is a measure of the turbidity of the slope water. Its distinctive semi-diurnal periodicity is discussed in section 4.1.5 in relation to the up-slope advection of a tidal front, or internal tide, which is shown in the near-bed temperature time series by its characteristic saw-tooth form at spring tide.

Time series plots of current for the S4 CM, located 85mab are shown in fig 4.2. The time series does not show the same asymmetry in the cross shelf direction as at 8 mab. A maximum down-slope current of $76 \text{ cms}^{-1}$ is recorded during Julian day 203.
Fig 4.2 Time series of cross shelf (+ve onshelf) and along shelf (-ve poleward) current at the S4 current meter station on mooring 124.
which is uncharacteristically 11% larger than the corresponding on-shelf current (compared to 34% larger for CM8511) and is measured in the presence of a small on-shelf residual current of 0.36 cms\(^{-1}\). In this instance the residual current is due mainly to the north-westerly flowing slope current with a flow speed of approximately 6 cms\(^{-1}\). The magnitude of this current is consistent with residual current measurements by Pingree & LeCann (1989) which were also on the La Chapelle Bank shelf slope.

4.1.2.2 Mooring 123

The asymmetry of the cross-shelf currents 8 mab at mooring 123 (200m depth, CM7643) is reversed when compared to the measurements further down-slope (fig 4.3), with maximum currents occurring during the on-shelf phase of the \(M_2\) tide (table 4.2). An instantaneous on-shelf current of 77 cm/sec is measured during Julian day 203 in the presence of an on-shelf residual current of 2.8 cm/s. Large down-slope current surges are also observed in cross-shelf time plot during the 'flood' phase of the \(M_2\) spring tide (the flood phase being defined as the period of flow between maximum off-shelf and maximum on-shelf tidal streaming). These current surges are most noticeable during Julian days 191-193 and 202-204 inclusive and are discussed further in section 4.1.4.

The S4 CM, located 68 mab (fig 4.4) and CM421, located within the seasonal thermocline, 128 mab (fig 4.5) are both consistent with CM7643, showing on-shelf tidal current asymmetry (table 4.2) and on-shelf residual flows. Increased variances in the current and temperature records on mooring 123 are also apparent, more so as the seasonal thermocline is approached. Maximum instantaneous on-shelf currents exceed 100 cm/s\(^{-1}\) at 68 mab and are more than 10 cm/s\(^{-1}\) greater than those recorded within the seasonal thermocline.

Within the seasonal thermocline a mean temperature of 12.8\(^\circ\)C is recorded (fig 4.5) and has maximum and minimum deviations of 2.7\(^\circ\)C and 0.8\(^\circ\)C respectively. The large isotherm depressions, represented by the positive temperature excursions, are
Fig 4.3 Time series of cross shelf (+ve onshelf) and along shelf (-ve poleward) current and temperature at current meter station 7643.
Fig 4.4 Time series of cross shelf (+ve onshelf) and along shelf (-ve poleward) current at the S4 current meter station on mooring 123.
Fig 4.5 Time series of cross shelf (+ve onshelf) and along shelf (-ve poleward) current and temperature at current meter station 421.
phase locked with the semi-diurnal tide. The temperature maxima occur later than
the off-shelf maximum flow, nearer to the maximum displacement which is 1/4 of a
M$_2$ tidal cycle later. The temperature maxima are also out of phase with the 'saw-
toothed' temperature profile near to the bed (fig 4.3, described earlier for CM7643).
Finally, it is worth noting from tables 4.1 & 4.2 that the minimum recorded
temperature of 12.0°C in the seasonal thermocline (72m depth at the shelf break) is
lower than the maximum recorded temperatures 8mab (at the same location) and
4km further off-shelf (also 8mab in 305m of water). These statistics suggest a large
vertical and lateral advection of the isotherms per tidal cycle during spring tide
conditions and is indicative of a large baroclinic response both near to the bed on the
upper slope and within the seasonal thermocline at the shelf break.

4.1.2 Tidal and residual current analysis

The tidal dynamics are investigated by harmonically analysing the current meter
time series data sets using the PML (Plymouth Marine Laboratory) computer
program TIRA (Tidal Institute Recursive Analysis). TIRA was originally developed
at the former IOS (Institute of Oceanographic Sciences) and uses a least squares
regression technique to produce details of the amplitude and phase of each tidal
constituent. A summary of the TIRA results are shown in fig 4.3 which show the
semi-diurnal tidal constituents (M$_2$ and S$_2$), three tidal harmonics (M$_4$, MS$_4$, M$_o$) and
the mean flow speed and direction.

The mean orientation of the major axis of the semi-diurnal M$_2$ tide is 037° for
mooring 123, whereas the alignment for the S$_4$ current meter (S$_4$ CM) on mooring
124 is rotated more anti-clockwise (028°), possibly being steered more by the local
bathymetry near to the bed at the head of the canyon. This orientation does not vary
significantly for the S$_2$ tide except nearer to the bed where maximum flows veer
cyclonically by up to 016° when compared to M$_2$ tide. One option when
investigating shelf edge current systems would be to resolve the coordinate system
for each current meter along the direction of the major axis of the M$_2$ tide (e.g. Shaw
Table 4.3  Current meter analysis

<table>
<thead>
<tr>
<th>Mooring No.</th>
<th>Depth, z (m)</th>
<th>Current meter</th>
<th>Depth, D (mab)</th>
<th>z/D</th>
<th>Mean flow Julian days 184-204 incl.</th>
<th>TIRA analysis [ M1 : S2 : M4 : MS6 : M8 ] *</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>speed (cms')</td>
<td>direction (°T)</td>
</tr>
</tbody>
</table>

| 124         | 305          | CM7765        | 162            | 0.47| -            | -              | -             | -             | -             |
|             |              | CM8511        | 8              | 0.97| 6.7 (17.5)   | 240 (256)      | -             | -             | -             |

* Tidal constituents with periods of 12.42 hours for M1, 12.00 hours for S2, 6.21 hours for M4, 6.10 hours for MS6 and 4.1 hours for M8.

* Direction is not used for the CM8511, [although Pingree & New (1995) have included this data set and calculated large downslope mean currents of 17.5cm/sec, *shown in brackets]. In the subsequent time series plots direction for this current meter is taken from the S4 current meter above.

Table 4.4  TIRA versus model results for current meter data

<table>
<thead>
<tr>
<th>Mooring No.</th>
<th>Depth (m)</th>
<th>Current meter</th>
<th>TIRA</th>
<th>A comparison of the M1 TIRA results and depth integrated model as described in Pingree et al (1982).</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>S2: M1 ratio (%)</td>
<td>a (cm/s')</td>
<td>b/a (%)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>TIRA</td>
<td>Model</td>
</tr>
<tr>
<td>123</td>
<td>200</td>
<td>42</td>
<td>48</td>
<td>51</td>
</tr>
<tr>
<td></td>
<td></td>
<td>68</td>
<td>45</td>
<td>56</td>
</tr>
<tr>
<td></td>
<td></td>
<td>8</td>
<td>33</td>
<td>42</td>
</tr>
<tr>
<td>124</td>
<td>305</td>
<td>85</td>
<td>43</td>
<td>58</td>
</tr>
</tbody>
</table>
et al., 1994). However, since the mooring and STABLE transect was aligned at approximately 020° (fig 3.3), it was decided not to rotate the coordinate system in this manner for the mooring data set.

Cyclonic veering of the major axis of the semi-diurnal tide is found on mooring 123 as you approach the sea bed with a maximum M2 current amplitude of 56 cms⁻¹ found 68 mab. The depth of maximum M2 tidal flow is aligned with the theoretical internal tide ray path shown in fig 3.5a which is also predicted by New (1988) to be the depth for maximum semi-diurnal tidal currents (barotropic + baroclinic) in this Celtic Sea slope region. The results from table 4.3 also show this to be true for both the S2 tide and the M4 tidal harmonic. A phase difference of 42° between M2 and S2 tides is observed 85 mab at mooring 124 which is consistent with the phase lag of spring tides after full or new moon and with observations by Pingree & New (1990). The cyclonic veering of both the M2 and S2 tides are in the same sense as that which is due to Ekman veering near to the bed.

Unfortunately the baroclinic and barotropic contributions from the M2 tide cannot be realistically separated with the data set in hand, especially in light of the limitations of the data and the sparse separation of each current meter. The barotropic-baroclinic separation is also complicated by the slope which couples these M2 tidal components. An estimate of the baroclinicity of the M2 tidal flow for this field study is therefore made by comparing the TIRA results with that of a numerical model. This was possible using the modified northwest European shelf tidal model of Pingree & Griffiths (1978) since it included the continental slope region. The numerical model is non-linear, extends down the slope and onto the abyssal plains, and is derived from spherical polar coordinates. The results of the model are presented in Pingree et al., (1982) and Pingree et al., (1984) and were described earlier in section 2.2.1 (see fig 2.2). The results of the TIRA analysis for the M2 tidal currents are compared to model results of the vertically integrated tidal currents in table 4.4 and are very consistent with one another, which perhaps gives confidence in the model as much as the TIRA results. A modelled depth integrated M2 tidal current of 53 cms⁻¹ is calculated for the shelf break (200m contour). This value is
exceeded by 3cms\(^{-1}\) within the expected region of internal wave intensification (68mab) and is diminished by 5cms\(^{-1}\) within the seasonal thermocline (128mab). Tidal flows are reduced near bed due to frictional stresses and are 11cms\(^{-1}\) less than the vertically averaged model results for the shelf break. Also included in table 4.4 are the \(S_2:M_2\) ratios for each time series which compares to maximum values of 38% for this region using the numerical model (Pingree et al., 1983, Pingree & New, 1990). These vary significantly between current meter stations at the shelf break but have a mean value of 40%, which is very similar to the vertically integrated numerical model value.

The residual currents for both moorings have a characteristic north-westerly component of flow (table 4.3) which is consistent with the well known poleward slope current that exists along the margins of the northwest European continental shelf edge (e.g. Pingree & LeCann, 1989, see also section 2.2). At the shelf break the residual flow has a large up-slope component at all depths. The largest residual current is measured within the thermocline at 72m depth with a speed of 10.8cms\(^{-1}\) and also has the largest on-shelf component of flow. This is in agreement with Pingree & LeCann (1989) who show that the residual flow is strongest at the shelf break (a mean flow of the order of 10cms\(^{-1}\) in late summer) and directed on-shelf. Pingree & LeCann (1989) also show that as the 500m contour is approached, the mid depth residual flow is directed on-shelf, but the near-bed flows are strongest and directed down-slope. We observe this phenomenon on the slope region at 305m water depth at mooring 124 One possible reason for this observation is discussed in section 4.3 and is attributed to the baroclinic \(M_2\) tide.

Progressive vector plots for the current meter data on moorings 123 & 124 are shown in fig 4.6. The east and north components of flow have been plotted for the 20 day period when all five current meters were operative. Due to compass sticking on the Aanderaa CM8511 located 8mab on mooring 124 (see fig 3.7), the direction data has been taken from the S4 current meter, located 60m above. It is clearly evident that in this case the residual flow near the bed at 305m depth is directed strongly off shelf. An attempt has been made to validate this result by comparing the
Fig 4.6  Progressive vector diagrams for instruments on mooring 123 (black), 124 (green) for Julian days (JD) 184-204 inclusive, and for the STABLE site (red) for Julian days 184-192 inclusive. Distances above the bed are shown in brackets and the insert from fig 3.3 is used to compare residual flow direction to the local bathymetry. See text for further details.
progressive displacement plot for CM7643 (8m above the bed) using its own directional data set (plot (i) for CM7643) with that using the directional data set from the S4 current meter, located 60m above (plot (ii) for CM7643). It would seem from plots (i) & (ii) that cyclonic Ekman veering in the bottom boundary layer could explain some of the difference observed. If the same argument were true for CM8511 on mooring 124, then the observed cyclonic veering associated with using real and reliable direction data could lead to an even larger down-slope component of the residual flow. If the original 'suspect' direction data set is used the magnitude and direction increase further in a cyclonic sense (Pingree & New, 1995).

It is therefore fairly certain that the flow 8mab at mooring 124 is directed down-slope and is larger in magnitude than the flow 68mab. The residual flow calculated from the Electro-Magnetic Current Meters (EMCM's) on STABLE, located 39cm above the bed at 388m depth, are also shown plotted in fig 4.6. The EMCM's residual current for Julian days 184-192 inclusive is larger than any of those measured further up-slope at mooring stations 123 and 124 for the same period. The EMCM residual flow is very sensitive to offset errors however, which are notoriously difficult to calibrate for. It is shown in section 4.2.2 that the results obtained are a reliable estimate of the near-bed currents at the STABLE site and are discussed further in section 4.2.3.

To summarise fig 4.6, the flow is clearly shown to be dominated by the M2 tide and a residual current, both of which seem to vary both in magnitude and direction throughout the record. The amplitude and orientation of the semi-diurnal tide can be approximated by studying the ellipticity of each vector plot. A strong residual flow is observed where the ellipses are not fully closed. There also appears on first inspection, to be a spring-neap variation in the residual flow. This is shown by the variation of the residual flow direction over successive tidal cycles compared to the amplitude of the elliptical orbit (a measure of the tidal excursion) for all current meter stations.

This phase relationship between the residual flow and spring-neap cycle is investigated further in fig 4.7, which is a repeat of fig 4.6 but with the M2 tide
Fig 4.7  Progressive vector diagrams (as for fig 4.6) for instruments on moorings 123 and 124, shown relative to the cross slope coordinate axes. In this instance the semi-diurnal tide has been removed using a low pass filter and circles are marked every 12.5 hours. The residual flow speed / direction for each mooring instrument and the EMCM's on STABLE, are shown both in the brackets and in vector form, by a straight line. The same scale is used for each progressive vector plot (in km's) and residual flow vector plot (in cm/s).
effectively removed using a Butterworth low pass spectral filter, which is a standard Matlab spectral analysis tool (The Math Works, Inc. [1994]: Signal Processing Toolbox User's Guide, for use with MATLAB). A circle is marked every 12.5 hours and the variation in distance between successive circles effectively shows a rate of change of the residual current over the 20 day neap-spring-neap-spring period. There existed the possibility that the strength of the residual current was increased during spring tides compared to neaps and the vector diagrams certainly show a variation of this kind. The hypothesis was tested by comparing the maximum cross-slope tidal excursion of slope water for each tidal cycle, as measured from the peak to trough amplitude of the temperature time series record at selected current meter stations, with the variation of flow speed and direction of the residual current over the 20 day period. Tidal forcing encourages a correlation, but near to the bed friction (greater at springs) inhibits it. The results of this hypothesis proved inconclusive and are not included here.

Two scenarios emerge from figs 4.6 and 4.7 which could have a significant effect on transport of the finer sediment by the residual current.

(i) Although the slope current is primarily driven by the pole ward decline of dynamic height (i.e. density driven), there is clear evidence of vertical (i.e. baroclinic) modulations to the observed westerly slope current and temporal variations over the spring-neaps cycle.

(ii) The near-bed flows can be significantly different in speed and/or direction to the mean residual current. The cross-slope flux of material could far exceed estimates made using the latter current information.

(iii) If the near-bed flows exceed the critical threshold for sediment movement on the upper slope/shelf break, a region of bed-load parting is likely since the maximum observed flows 8mab are on-shelf (77cms\(^{-1}\)) at the shelf break and off-shelf (51cms\(^{-1}\)) at mooring 124. Sediment transport processes are discussed fully in chapter 5.

The baroclinic M\(_2\) tides are thought to be responsible for the large cross-shelf instantaneous and residual currents on the upper slopes of the Celtic Sea (e.g.
Pingree, 1988, New, 1987, Pingree & LeCann, 1989). There are however other mechanisms which could account for the observations shown in fig 4.6. Pingree & LeCann (1989) suggest that local upwelling and downwelling will effect the residual flow (it is suggested however that upwelling and downwelling at tidal period is internal tide motion), but a more likely alternative in this instance may be the channelling of the tidally induced near-bed flows at the head of the canyon at mooring 124 (Pingree & LeCann, 1989). The apparent variation in the residual flow over the neap-spring cycle would suggest that the baroclinic tides are a significant process in this instance. This is not on its own conclusive, since a much longer data set may in fact implicate an entirely different process (at a longer wave period) for the observed residual flows shown in figs 4.6 & 4.7 (longer period motions are not resolvable with this data set).

The hypothesis that strong down-slope, near-bed residual and maximum currents exist on this upper slope region due to an enhancement in internal tidal motion, does become more plausible when coupled with similar observations by Pingree (1988) and Pingree & LeCann (1989), which are also in the immediate vicinity of the STABLE site. Evidence for enhanced near-bed baroclinic tidal motion will be presented in later sections for the shelf break (section 4.1.4) and the upper slope region (section 4.1.5) and later inferred from the STABLE results in section 4.2.

4.1.3 Energy spectra

Spectral analysis of time series data such as the current and temperature data collected from moorings 124 and 123, enable an investigation of the contributions to the flow from the different frequencies. The resolution of the frequency spectrum is dependent on the reciprocal of the total length of the record used for the spectral estimate, which can vary according the degree of smoothing or non-over-lapping segments that are required for the spectrum. In this instance $1/T$ is equal to 0.14 cpd for each current meter and temperature data set. The highest frequency, shortest period of motion that can be resolved is known as the cut-off or Nyquist frequency,
and is defined as \( \frac{1}{2} \Delta t \) (\( \Delta t = 10 \) minutes) and is equal to 72 cpd, i.e. several tens of minutes. These motions are usually intermittent internal wave fluctuations. The smallest period wave motion that can exist in nature is defined by the Brunt Väisälä frequency, \( N \) (refer to section 2.2.3).

Figure 4.8 shows the kinetic energy spectra for the cross-slope current on (a) mooring 124 and (b) mooring 123. Fig 4.9 is the same as fig 4.8 but for the long-slope current. The spectra were obtained using the standard 'Matlab' spectrum function (The Math Works, Inc. [1994]: Signal Processing Toolbox User's Guide, for use with Matlab) with \( 2^{10} \) sample pieces and a 50% overlapping Hanning data window, thus providing 8.22 degrees of freedom (dof). The number of dof is computed from Nuttall (1971) as \( \nu = 1.91 N_p - 1.33 \), where \( P = 3 \) (the number of non-overlapping samples), and \( N_p = 2P - 1 \). The 95% confidence intervals for the 8.22 dof are obtained from Fisher & Yates (1963) and are also shown plotted and reveal several 'real' peaks in the spectra. The reader should refer to Jenkins and Watts (1968) for further details on the level of confidence that can be attached to the measured values of the distribution of spectral density.

All spectra show the semi-diurnal frequency band (shown labelled by \( M_2 \)) to contain by far the largest fraction of the total horizontal current variance for all current meters. For the cross-slope components of current flow, an average of 94% of the kinetic energy is accounted for within the dominant semi-diurnal tidal peak. This reduces to a mean value of 87% for the long slope current, the deficit being accounted for by a 50% increase in energy above semi-diurnal tidal frequencies (mean value of 10% of the total energy) compared to the cross-slope current. Also labelled on fig 4.8a are the smaller but significant peaks for the first and second tidal harmonics (shown as \( M_4 \) and \( M_6 \) respectively). At lower than \( M_2 \) tidal frequencies, the inertial period (I) and the diurnal (\( K_1 \)) tidal peaks are also labelled. The inertial period is due to the earth's rotation, and is defined as \( I = \frac{2\pi}{f} = 1.47 \) cpd, where \( f = \) Coriolis parameter \( (2\omega \sin \theta) = 1.07 \times 10^{-4} \) s\(^{-1} \), \( \omega = \) angular velocity of the earth, and \( \theta = \) latitude. The diurnal (\( K_1 \)) peak is clearly apparent for all instruments except notably CM7643 on mooring 123.
Fig 4.8 Cross shelf velocity power spectra for current meters on (a) mooring 124 and (b) mooring 123. See text for further details.

Fig 4.9 Long shelf velocity power spectra for current meters on (a) mooring 124 and (b) mooring 123. See text for further details.
For the purpose of this study, the tidally averaged Brunt Väisälä frequency will be taken from Pingree *et al.* (1986) for the upper 150m and below this level from Pingree & New (1989) and Pingree & Morrison (1973), in the same manner as New (1988), for each current meter station on mooring 123 & 124 (shown in fig 3.2a). The spatial variation of N is shown by the shaded regions in fig 4.9 for each mooring site, with upper and lower values of N shown colour coded to represent upper and lower current meter stations. N is shown to progressively increase with decreasing depth below the surface, from 1.2 cph (29 cpd) at 297m depth (mooring 124) to 1.4 cph (33 cpd) at 132m depth (mooring 123), but to then significantly increase to 4.6 cph (110 cpd) within the seasonal thermocline at 72m depth (mooring 123). As the seasonal thermocline is approached the sampling rate of 10 minutes for CM421 (providing a spectral resolution of 20 minutes) is too long to detect the highest frequency internal wave motions which can theoretically exist within the seasonal thermocline (approximately 13 minutes). It should be emphasised however that N can vary both spatially and temporally in tidally energetic environments. For example, near to the bed, the variability of N over a tidal cycle may be attributed to varying degrees of bottom mixed layer thickness (examples for the northwest European continental slope include Thorpe *et al.* (1990) and White (1994)), or to a varying baroclinic response of the isopycnals which impinge on the slope (see later in section 4.1.5, also Thorpe, 1987), both of which will vary the stratification locally. In the bottom mixed layer internal waves should not by definition exist at all, but their motion may still be attenuated down from above this layer and be observed in the current meter data near to the bed. This is as might be expected for the current meters located 8mab at both mooring sites in this study. Such processes are actually observed in the inter-burst EMCM data and are discussed later in section 4.2.4.

The spectral plots shown in figs 4.8b and 4.9b show an increased contribution to the current variance within the internal wave and turbulent frequency domain nearer to the surface. Unfortunately, spectra which don’t diminish towards the Nyquist frequency could suffer from aliasing ('white noise' within a spectrum, or aliasing, can both cause mis-interpretations of spectral plots shown in 4.9, although this does
not appear to be a problem with the present data set). Section 4.1.3 does show that much more high frequency noise exists at 128mab and 68mab on mooring 123 in comparison with the measurements 8mab and at all stations at mooring 124.

The temperature spectra for sensors on mooring 124 & 123 are shown in figs 4.10a & b respectively. They agree well with the interpretation of the current spectra in figs 4.8 & 4.9 and also shows little cascading of energy at the high frequency end of the spectrum within the seasonal thermocline (fig 4.10b, CM421).

Cross spectral analysis of pairs of time series data sets enabled a correlation study of signals at frequencies where peaks were shown to be statistically significant in figs 4.8, 4.9 & 4.10. The degree of correlation over the frequency domain is represented here by the coherence (or more correctly the squared coherency spectrum) between the two signals $x(t)$ and $y(t)$. The coherence is a real value between 0 and 1 and is calculated using standard Matlab functions as:

$$c^{2}_{x,y} (\omega) = \frac{|\text{CSD}(\omega)|^2}{\text{PSD}_{x}(\omega) \times \text{PSD}_{y}(\omega)}$$  \[4.1\]
where CSD = Cross Spectral Density of \((x, y(t))\) and PSD = Power Spectral Density of \(x(t)\) or \(y(t)\), as previously shown in figs 4.8-4.10. The phase spectrum is also calculated to show how much the frequency component in one series leads or lags the component at the same frequency in the other time series, at frequencies where there is a statistically high degree of coherence. In what follows the coherence and phase (in degrees) is shown in three separate comparisons in the same manner as Thorpe (1987). The estimated 95% coherence value is calculated using \(2^9\) sample pieces (providing 19.68 dof) using a simple formula which calculates an upper limit on zero coherence. That is, it identifies whether the observed coherence is significantly different from a zero coherence and is defined as (see Thompson, 1979);

\[
c^2 = 1 - \alpha^{2\nu-2}, \quad \text{where } \alpha = 0.05 \text{ for 95\% level,}
\]

\[
= 0.29 \quad \text{and } \nu = \text{number of dof}
\]

Confidence levels on phase are not well defined and are therefore not included here. Jenkins & Watts (1968) do however suggest a method for calculating the 95% confidence intervals.

The coherence and phase for signals from the CM8511 station 8mab on mooring 124 are shown in fig 4.11. The estimated 95% coherence value for the true coherence of zero is shown by the dotted line. There is a high coherence between the signals at tidal frequencies for all three signal pairs, those being (a) cross-slope current and temperature, (b) long-slope current and temperature, and (c) cross-slope current and long-slope current. The correlation with temperature is strongest at \(M_2\) period where the cross-slope current leads the temperature by \(90^\circ\), whilst the long slope current is out of phase with temperature by \(180^\circ\). The cross-slope current leads the long-slope current by \(90^\circ\) for \(M_2\) frequency and the first and second tidal harmonics. These results confirm the elliptic, rotary clockwise nature of the tidal current on the slope. The temperature correlation’s are consistent with the up-slope advection of the isotherms during on-shelf \(M_2\) tidal streaming (a persistent cooling, with maximum rate of cooling occurring during maximum on-shelf flow) and the down-slope advection of isotherms during off-shelf \(M_2\) tidal streaming (a persistent warming, with maximum rate of warming during maximum off-shelf flow).
Fig 4.11 Phase and coherence verses frequency for (a) cross-slope current against temperature, (b) long-slope current against temperature, and (c) cross-slope current against long-slope current at CM8511 at 8mab on mooring 124. The 95% Confidence limit on zero coherence is shown by the dotted line for 19.7 degrees of freedom.

Fig 4.12 Phase and coherence verses frequency between (a) cross-slope currents, (b) long-slope currents, and (c) temperature at 128mab and 8mab on mooring 123.
The coherence and phase for time series of the same parameters from 128mab and 8mab at the shelf break (mooring 123) are shown in fig 4.12. The highest coherence at $M_2$ period is again observed with zero phase lag for the cross-slope current. The first and second harmonics are approximately $180^0$ out of phase with the $M_2$ current. Both the along-slope current and temperature at 128mab lag by approximately $14^0$ (or 29 minutes) compared to near the sea bed. The same phase lag is also found when comparing the measurements 8mab on moorings 124 and 123 (fig 4.13), albeit from different depths, with all the signals at the deeper instrument lagging by tens of minutes. In the case of the temperature signals 8mab, the $M_2$ tidal frequency on mooring 124 lags the temperature signal at mooring 123 by approximately $30^0$, or one hour. This very significant observation is discussed in section 4.1.5 in which the near-bed dynamics are studied further.

Finally it is notable that the cross-spectral analyses presented here do not suggest any significant coherence between pairs of signals below semi-diurnal tidal period, except that is, between the two temperature time series and the long-slope current time series pairs, located 8 mab at each mooring (fig 4.13). In this instance, at approximately 0.6 cpd, the temperature 8 mab on mooring 124 lags the temperature 8 mab on mooring 123 by $100^0$ whilst the long-slope current at mooring 124 lags that at mooring 123 by approximately $10^0$.

4.1.4 Internal tide current surges at the shelf break.

Internal waves cannot be resolved using a single current meter record, and if the vertical separation between current meters on a mooring is large, then the wave-number (and hence frequency) resolution for the internal wave reduces. Having temperature sensors at two of the current meter stations at mooring 123 has helped to identify the higher frequency internal wave fluctuations more easily, in particular, the on-shelf propagation of a first mode internal tide. This is because internal waves deform isothermal surfaces as they propagate, and can be identified by their co-oscillation with the velocity perturbations (McGrail & Carnes, 1983).
Fig 4.13 Phase and coherence verses frequency between (a) cross-slope currents, (b) long-slope currents, and (c) temperature at 8mab on mooring 124 and 8mab on mooring 123.
Internal wave current surges (IWCS's) are clearly observed at the shelf beak during spring tides to be phase locked with the semi-diurnal barotropic tide and are generally absent during neap tide conditions. Fig 4.14 shows the results of the current and temperature measurements taken at the shelf break (200m water depth) for three successive tidal cycles at spring tide. Events (A) to (C) identify well defined pulses in each current meter record and in the temperature record 128mab. The pulses occur approximately 3½ hours before maximum on shelf flow in the $M_2$ tidal cycle, and have their largest amplitude at spring tide. The events are characteristic of shelfward propagating, first mode internal waves and are essentially a travelling depression of the seasonal thermocline. Above this depression (or trough) the current 128mab increases the cross-slope flow from near zero up to as much as 60cms$^{-1}$ on shelf (event C) during the flood tide. Well below the thermocline, near to the bed, the current 8mab is increased off-shelf for up to 50 minutes duration, with a maximum off-shelf current amplitude of 35cms$^{-1}$. The IWCS's also have a long-slope component and are estimated from fig 4.14 to be propagating along 019$^\circ$T. The temperature record 128mab shows an associated temperature pulse of 2.5°C, with a rapid increase from 13-15.5°C followed by a rapid decrease over the 50 minute period. The 180 degree phase change in current direction above and below the thermocline is indicative of a first mode wave.

The IWCS's also seem to be preceded by a packet of up to four much shorter period waves of solitary form, particularly evident prior to event C. The cross and long-slope currents 128mab show the solitary waves to be also propagating on shelf. The associated current pulses observed 128 mab do not penetrate down to 8 mab and therefore have no influence on boundary layer processes. They are first observed during maximum off-shelf tidal flow and progressively increase in amplitude as the off-shelf flow slackens. These observations provide further evidence that internal soliton packets are generated at the shelf break through the interaction of shelf edge bathymetry, the barotropic tide and seasonal stratification (Sandstrom & Elliot, 1984, New & Pingree, 1991). The internal soliton packets are observed to propagate along the seasonal thermocline prior to the main internal wave current pulse which penetrates the sea bed. This is contrary to modeling studies such as Gerkema
Fig 4.14 Direction, cross-slope current, long-slope current and temperature at different heights above the bed on mooring 123 at the shelf break. The direction records from each record is offset by 90 degrees, the current records are offset by 50 cms \(^{-1}\), and there is no offset for the two temperature records. See text for further details.
(1994), but is in agreement with observations by Heathershaw et al., (1987) which were also in the La Chapelle Bank region of the Celtic Sea, approximately 8km shelfward of the present data set.

As can be seen in figure 4.14, once these internal solitons have passed, the IWCS easily penetrates down to the bottom mixed layer and consequently will (phase dependent) influence near-bed processes both at the shelf break and further on shelf (Heathershaw, 1985, Heathershaw et al., 1987). It would seem plausible that the internal solitary wave packet and the IWCS are all manifestations of forced oscillations within the seasonal thermocline which are phase locked with the semi-diurnal barotropic tide. They are formed during off-shelf tidal streaming as the thermocline is depressed at the shelf break and are released immediately the off-shelf tidal streaming begins to slacken. The process being described is therefore a non-linear response to what originated as a first mode oscillation of the seasonal thermocline. This immediately breaks up into a series of internal solitary wave-like features as shown in fig 4.14 as the off-shelf tide slackens. An IWCS which is grouped with a packet of internal solitary waves and which is phased locked with the semi-diurnal tide is described as an internal tide current surge (ITCS, Heathershaw et al., 1987). These ITCS's are also observed most noticeably in the temperature record within the seasonal thermocline. The temperature records at 128mab and 8mab are plotted without any vertical offset and so the increasing temperature difference observed during off-shelf tidal streaming indicates the vertical extent of the depression of the seasonal thermocline at spring tide. During on-shelf tidal streaming for example, the temperature difference between 8 & 128mab can be less than 0.5°C. As the barotropic tide decelerates off-shelf and approaches cross-shelf slack water, the temperature difference can exceed 2.5°C as the baroclinic tide propagates on-shelf, past mooring 123.

To illustrate this ITCS phenomenon further, contour plots of temperature and density have been constructed for the data collected from the CTD yo-yo station CTD140Y (fig 4.15). The CTD was continuously profiled for a complete $M_2$ tidal cycle in approximately 295m of water, down to a maximum of 270m and then up to within
Fig 4.15  Contour plots of (a) temperature and (b) $\sigma_1$, for the CTD yoyo station CTD140 at approximately 295m depth. The station was located approximately 6.5km longslopes and poleward of mooring 124 (refer to fig 3.3). See text for further details.
40m of the surface. The position (fig 3.3) was approximately 6.5km due WNW and along-slope of mooring 124, which itself was located in 305m of water. Also marked on fig 4.15a are the times of maximum off-shelf tidal flow (Max off), maximum on shelf tidal flow (Max on) and times of cross-slope slack water (slack), as recorded 8mab on mooring 124. The structure of the internal tide in relation to the barotropic tide is clearly illustrated, and it would seem plausible therefore that the position of the CTD station and therefore also mooring 124 is near the origin of the internal tide. This is because the contour plots shown in fig 4.15 agree with earlier modeling studies of Pingree et al. (1984) and Heathershaw et al. (1987) where the internal tidal trough is shown to become fully developed after maximum off-shelf tidal streaming (event C in fig 4.15), which will then advance on-shelf against a decelerating off-shelf tidal current (the trough will split in two and one depression will also propagate oceanwards). One expected consequence of the interaction of the barotropic tide with the internal tidal trough (ITT) is a highly non-linear form of the baroclinic tide at this generation site, with steep and deeply penetrating troughs (Pingree et al., 1984).

The event marked as 'C' in fig 4.15b is found to be consistent with the ITCS event that is observed at the shelf break (mooring 123), approximately 4km on shelf from the CTD yo-yo station. This internal tidal trough is observed near slack cross-slope tidal flow (fig 4.14), approximately 3½ hours before maximum on shelf tidal flow at mooring 123 (and is also labelled event 'C'). This is concluded since it is the only time in the tidal cycle where the 14.5°C isotherm is depressed below 68m. The internal tidal trough shown in fig 4.15a is out of phase with event C in fig 4.14 and leads by approximately 2.1 hours. Further on-shelf the phase lag increases still further. Heathershaw et al. (1987) show that 8km landward of the shelf break the ITCS event now occurs during on shelf tidal streaming approximately 1 hour before peak on shelf flow.

Based on the above argument we can assume therefore that the internal tidal trough becomes fully developed somewhere oceanward of the shelf break (200m contour) and the upper shelf slope at ~300m. This is consistent with calculations by Pingree
& New (1996) who estimate the origin for the internal tidal trough to be near the 250m contour for this time of year. Their estimate is based on the assumption that the origin of the internal tidal trough is fixed in space and lunar time, so that variations in the arrival time of the ITTs for seasonally varying stratification can lead (after a displacement correction for the barotropic tide) to an interpolated estimate of the ITT origin.

We cannot say with any certainty therefore that the ITT which is observed from temperature and density profiles at the CTD yo-yo station in fig 4.15, is precisely at, or oceanward of the origin. This is because the associated ITCS is not attenuated down to the shallowest current meter on mooring 124 (S4 CM at 220m depth). One reason for this is because in deeper water the density below the thermocline increases gradually with depth, and away from the thermocline the structure of the internal tide changes. The two layer situation is no longer valid and the baroclinic tidal energy is converted to higher modes (Pingree et al., 1986). It is not therefore possible to calculate the travel time (hence phase speed) of the ITCS based on the two observations at the CTD yo-yo station and mooring 123. If however we were to assume that the ITT does in fact propagate away from the slope region at approximately 300m depth which is then observed 2.1 hours later propagating onshelf past mooring 123, we can, by neglecting the barotropic tide, estimate a phase speed for the ITCS as being 53cms\(^{-1}\). If we compensate for a mean off-shelf tidal advection of \(-25\text{cms}^{-1}\) for the \(M_2\) barotropic tide, assuming the internal wave is travelling parallel to the mooring transect then the phase speed for event 'C' increases to approximately 80cms\(^{-1}\). Since this observation can be approximated as a two layer fluid problem, the estimate can be compared with predictions using linear internal wave theory (Turner, 1973). The phase speed, \(c\), for infinitesimal waves is defined as:

\[
c^2 = g\frac{\Delta \rho}{k} \left( \rho_1 \coth h_1 + \rho_2 \coth kh_2 \right)^{-1}
\]  

[4.2]

For long waves \(kh_1\) and \(kh_2\) both tend to zero so that \(c\) can be approximated to:

\[
c^2 = \frac{g \Delta \rho}{\rho_2} \frac{h_1 h_2}{(h_1 + h_2)}
\]  

[4.3]
However, one tidal cycle from the CTD yo-yo station is insufficient data to estimate $\Delta \rho$ reliably and so our estimate is compared to previous observations in the same region. Phase speeds of approximately $50 \text{ cms}^{-1}$ are reported for the La Chapelle Bank shelf break region using this approximation (Pingree & Mardell, 1981 and Pingree et al., 1983). Phase speeds of $\sim 70 \text{ cms}^{-1}$ for a typical July density structure (Pingree & Mardell, 1985) and $\sim 65 \text{ cms}^{-1}$ (Pingree & New, 1995) are more reliable estimates. These latter two phase speeds have been calculated from observations which have benefited from increased spatial coverage of the IT 'event'. The apparent increase in the phase speed compared to that using Eqn. 4.3 is due to non-linear effects and locally increased stratification (Pingree & New, 1995).

Finally, the peak to trough amplitude of these ITCS's are also similar in magnitude to the current surges described by Heathershaw et al. (1987) in the bottom mixed layer (20-30cm/sec), in water depths of 160-170m, slightly northwest of this site (latitude $47^038'N$). As in this case, the baroclinic ITCS's are not superimposed onto maximum off-shelf barotropic tidal currents, and so they do not enhance the magnitude of the maximum off-shelf currents at this site. They may very well contribute to the observed Eulerian residual currents at the shelf break / upper slope region however, and this is discussed further in section 4.1.6.

4.1.5 Near-bed temperature observations of the up-slope advection of an internal tide/front on the upper slope.

This short section discusses the rapid periodic decreases in temperature from near-bed thermistors (e.g. $0.7^0C$ decrease in 1 hour) which are mainly prominent at spring tides (as first mentioned in section 4.1.2.1). In conjunction with similar observations at other field sites along the slopes of northwest Europe, and from further afield, it is inferred that these cooling events are due to a near-bed baroclinic tidal response of the isopycnals impinging on the slope region.
The response of the boundary layer to isopycnal forcing by the barotropic tide on a large portion of the north-west European continental slope has been shown in section 2.2.4.2 to be significantly different to the response observed away from the sloping bottom. For example, at the shelf break, cross-spectral analysis of the temperature time series data at 8mab and 128mab (section 4.1.3) shows a 29 minute phase difference at M₂ tidal frequency (fig 4.12c). Peak values of temperature within the seasonal thermocline (128mab) are shown in fig 4.14 to occur approximately 2.8 hours before the peak values near to the bed (8mab, event [c]) for typical spring tide conditions. This compares to the current meter tidal analysis results (table 4.1) which show that M₂ currents 8mab lag those in the seasonal thermocline by only 6 minutes.

Critical semi-diurnal internal tide reflection on these sloping regions creates a variable and periodic signature in any temperature time series record near to the bed where the slope inclination, α, matches the angle of the internal wave characteristic, c (see chapter 2, equation 2.5 in section 2.2.3). To briefly recap, the internal waves are predicted through linear wave theory to become infinitely large, providing conditions favourable for large current shear, low Richardson number (section 2.2.4.3) and consequently energy dissipation through internal wave breaking and mixing (Eriksen, 1985, Thorpe, 1992).

The general pattern of the temperature structure near to the bed (which also characterises the density structure for much of the north-west European continental margins) where the bottom slope is close to critical for the semi-diurnal tide is one of a rapid increase in density associated with sharply rising isotherms during the up-slope phase of the tidal cycle (e.g. Thorpe et al., 1990 and White, 1994 on the Hebrides slope, and Thorpe 1987 and Thorpe et al. 1990 on the Porcupine Bank). These sharply rising isotherms are then followed by a more gradual rise in temperature during the down-slope phase of the tidal cycle. The rapid decreases in temperature are thought to be due to the up-slope propagating internal tide (section 2.2.4.2) and has been described as a 'bore' of cold water by Gardner (1989b) and as a 'thermal front' by Thorpe (1992). The implications to sediment transport (section 2.2.4.2) are discussed further in chapter 5.
The temperature time series 8mab at mooring 124 (305m water depth) and mooring 123 (200m water depth) are compared in fig 4.16a. The record covers a spring tide, five day period. Also shown in fig 4.16 are the transmission time series results (uncalibrated voltage) from the transmissometer at mooring 124 (fig 4.16b) and the cross shelf velocity time series for both sites (fig 4.16c). Events [D]-[I] show periods of rapid cooling (or rising isotherms) which are phase locked with the semi-diurnal tide. The dots on the temperature plots are at the ten minute sampling interval of each thermistor and the events are proceeded by a far more gradual rise in temperature during off-shelf tidal streaming. At mooring 124, event [D] occurs during maximum on-shelf tidal flow and shows rapid near-bed cooling, with the temperature decreasing by 0.7°C in one hour (0.45°C in the initial 20 minutes). This represents the full tidal excursion of the near-bed isotherms (isopycnals) past mooring 124 during this period. The temperature then remains approximately constant until the onset of on-shelf tidal streaming, when the isotherms begin to advect up-slope once again. This front is then observed to advect past mooring 123 approximately 100 minutes later. This represents a rate of advance of the front of approximately 65cms⁻¹ between moorings and occurs at the time of maximum 'peaked' on-shelf current as measured 8mab at mooring 123. This rate of advance is consistent throughout the record and is consistent with results from a least-squares fitting to the complete temperature time series records. This analysis indicates a 2 hour advance of phase of temperature at 12.42hr period at mooring 124 compared to mooring 123 8mab. In some instances however (e.g. event [F]) the frontal feature has dissipated completely during its tidal excursion from mooring 124 to 123.

Events [F], [G] and [I] also show short periods when the slope water is warmer 8mab than at the shelf break. This is at times of maximum down-slope tidal excursion of the isotherms at mooring 124. This could infer periods of instability within the boundary layer when enhanced vertical mixing might take place on the upper slope region, just prior to the cooling events described previously. Or this may infer that the formation of the ITT during down-slope tidal streaming (as described in the previous section) is at or near the 300m depth contour and which actually penetrates the bottom mixed layer at this depth. The formation the ITT would enhance the down-slope flows near to the bed locally.
Fig 4.16 Time series plots of (a) temperature, (b) transmission and (c) cross-shelf velocity, measured 8 mab on mooring 123 (red) and mooring 124 (blue). See text for further details.
Of particular interest is the corresponding transmission signal 8mab at mooring 124 which shows an increase in transmission associated with the up-slope advection of the frontal feature. This increase in transmission implies an advection of clearer water up-slope and is followed by the down-slope advection of more turbid water during the down-slope phase of the M2 tide. A least-squares fit to this data set shows a 180 degree phase difference compared to the temperature at this location consistent with high transmission during the up-slope phase of the semi-diurnal tide. It is also interesting to compare the magnitude of the cross-shelf currents at the two near-bed sites. Over successive tidal cycles the down-slope currents on the upper slope site are equal to or exceed the down-slope currents at the shelf break. Continuity requires the barotropic currents at the shallower site to be stronger in magnitude and so there would appear to be a strong baroclinic component at M2 frequency on the upper slope which infer maximum and mean flows down-slope.

4.1.6 Discussion

The dynamics of the continental slope region are highly complex. To understand the sediment dynamics at the La Chapelle Bank shelf break / upper slope region one requires a knowledge of the near-bed processes. The study of boundary layer flows on sloping bottoms is a scientific discipline for both physical oceanographers and modellers in its own right. So too are the hydrodynamical processes that occur above the boundary layer which have complexities beyond our present understanding. Many shelf sea and oceanic circulation processes are now well understood and easily reproduced in hydrodynamic models. At present the same cannot be said to be true for models which incorporate the continental margins. Although the complexity of these numerical techniques in regions of rapidly varying topographic depth are a significant contributory factor in such modelling studies, the main underlying barrier is the lack of knowledge of the processes themselves. This understanding is the focal point to the success in being able to predict shelf-ocean fluxes of material, whether it be water masses or polluted terrigenous material. It is also suggested that boundary mixing is the biggest contributor to overall oceanic mixing.
Faced with such complexities, it is not really possible to explain how the full spectrum of hydrodynamical processes communicate with the sea-bed (via the interaction of boundary layer and sedimentary dynamics). Sediment transport, boundary layer flows, boundary layer mixing, baroclinic and barotropic hydrodynamical processes, all above sloping bottoms, form multi-disciplinary studies.

The deployment of STABLE in July 1990 on the upper slopes of La Chapelle Bank was the first ever deployment of a benthic lander of this kind on the northwest European continental margin. It was the first time that in-situ sediment transport measurements were taken to test predictions that had previously been inferred by physical oceanographers and modellers alike. The cross-slope transect of STABLE and the current meter moorings were set-up specifically to investigate the sedimentary response of the sea bed to tidal forcing in the upper slope / shelf break region, at time of strong seasonal stratification. Sampling varied from 10 minutes continuously for 23 days to 4 times a second at nine minute discrete intervals. Its purpose was to investigate if the turbulent response of the boundary layer and the measured sediment transport fluxes were consistent with one another and if so, could they be inferred from what is observed and is already known from the hydrodynamics. This section has attempted to give the reader an insight into the complexity of the hydrodynamical processes involved before proceeding to describe the results within the bottom 1m at the critical STABLE depth. There are clear limitations with the present mooring data set, but still some very important conclusions can be drawn from these results which are also consistent with other studies in this region.

(i) All current meter measurements show the residual flow to have an along-slope current component consistent in magnitude and direction with observations by Pingree & LeCann (1989 & 90) for the same region. There is also clear evidence of baroclinic vertical modulations to this mean residual flow and temporal variations over the neap-spring cycle. The vertically averaged residual current of 7.4cms\(^{-1}\) at the 200m contour was shown to increase in magnitude (10.8cms\(^{-1}\)) and direction on-shelf near the surface compared to mid-depths. At 305m depth the near-bed residual currents were shown to
be maximum (at least \(6.7\text{cms}^{-1}\)) and with a down-slope component of flow. This compared to a residual current 68mab of \(6.1\text{cms}^{-1}\). The near-bed residual current increased still further to \(11.1\text{cms}^{-1}\) at 388m depth (at the STABLE site), and Pingree (1988) have calculated bottom currents of \(15\text{cms}^{-1}\) orientated down-slope at 548m depth.

(ii) The dominant flux of energy across the upper slope and shelf break is shown to be of semi-diurnal tidal period with the other main contribution shown in the tidal and spectral analyses to come from the first and second tidal harmonics. Comparable increases in the total horizontal kinetic energy at semi-diurnal tidal frequency (barotropic and baroclinic) are observed near to the bed at 305m depth and at 68mab in 200m water depth, both of which coincide with the up-slope propagating theoretical beam of internal tidal energy (New, 1988) and regions of enhanced tidal currents.

(iii) Clear evidence of high frequency internal wave motion at the shelf break (defined as the 200m contour) within the seasonal thermocline. Internal wave current surges with maximum peak to trough amplitudes \(60\text{cms}^{-1}\) (on-shelf) are observed within the seasonal thermocline during the flood phase of the semi-diurnal tide, approximately \(3\frac{3}{2}\) hours before maximum on-shelf tidal streaming. The observed current surge penetrates down to 8mab, is 180 degrees out of phase with current measured above the depressed seasonal thermocline, and results in maximum off-shelf current surges of \(35\text{cms}^{-1}\) in the bottom layer. These IWCS's are preceded by a group internal solitary waves which also propagate on-shelf within the seasonal thermocline. The source region for these waves, which are phase locked with the semi-diurnal tide, are from between the 200-300m contours (more probably between 250-300m). They are shown from the CTD yoyo station to originate as a forced depression of the seasonal thermocline during off-shelf barotropic tidal streaming which then become non-linear, breaking up into solitary-like waves as they propagate away from the source region. In this instance we can term these events internal tide current surges (ITCS's). ITCS's are noticeably absent during neap tides.
Pingree et al. (1984) infer from modelling studies that immediately shelfward of the generating region the maximum bottom currents and residual currents will be weakly directed on-shelf, compared to stronger maximum and residual currents at the generation region which are directed down-slope. Heathershaw et al. (1987) use a similar model to examine the effect of these current surges on sediment transport and infer a region of bed-load parting, with large down-slope transport rates at the internal tidal trough (ITT) source and a weak on-shelf transport immediately on-shelf of the source region. The present results are consistent with these studies.

(iv) The asymmetric form of the temperature time series records 8mab at 305m and 200m depth over a semi-diurnal tidal cycle is a characteristic which seems confined to the near-bed sloping region. The pattern of generally steepening isotherms during the up-slope phase of the tide is evidence of an enhancement of baroclinic tidal energy near to the bed at a critical region on the slope. This has been shown in the laboratory (e.g. Thorpe 1987a, Ivey & Nokes 1989, Slinn & Riley 1996) and in the field (e.g. Thorpe 1987b, 1992). These fronts are observed as up-slope surging isotherms and are most prominent at spring tides. A mean rate of advance of this front up-slope is estimated as being approximately 65cms⁻¹, which exceeds the maximum currents measured at 305m depth, and is equal to the maximum observed currents directed on-shelf at 200m depth. Advection of the slope water due to the barotropic tide alone cannot therefore account for this phase speed and infers baroclinic enhancement. These flows are therefore likely to cause the boundary layer to vary in thickness at any fixed point near to the bed over a tidal cycle. Butman (1988) suggests that during the more turbulent up-slope phase of the tidal cycle the vertical mixing will increase the size of the bottom mixed layer. To have a net zero cross-slope flux of water (there are no trends in the temperature time series for the complete 23 day deployment) the flow will move more slowly up-slope. The opposite will be true for down-slope flows when the boundary layer re-stabilises itself and is confined to a narrow region near to the bed. The advection of this front 8mab at 305m water depth results in an immediate decrease in the turbidity of much cooler water up-slope (discussed further in section 4.3). The turbidity then increases during the down-slope phase of the semi-diurnal tidal cycle and infers a net down-slope turbidity transport close to the bed.
This section has shown that in all probability the observed Eulerian residual down-slope currents on the upper slope and cross-slope divergence of this Eulerian residual at the shelf break is tidally induced and would therefore be characteristic of much of the Celtic Sea continental margins south of 48°N. This tidal inducement is likely to be enhanced near canyons (such as the STABLE deployment site; refer to section 2.2.1) where large down-slope near-bed residual and maximum currents have been previously reported (Pingree, 1988). The STABLE results are now studied in detail in the following section.

4.2 STABLE results: an introduction

The hydrodynamic results from the STABLE deployment at 388m depth are discussed fully in this section. On initial inspection the data set was found to have many limitations, but a great deal of time has been spent on being able to confidently and accurately describe the mean (tidal) and burst (turbulent) environments at this site (the results of the residual current at this site already being described in section 4.1). Calibration procedures and results are presented in this section.

4.2.1 Mean flow data sets

Time series results from the mean logged rotor stack and pressure sensor are shown in fig 4.17. All 11524 data points at 1 minute sampling intervals are plotted for the 8 day period. The pressure record clearly shows the duration of the deployment to cover a neap-spring period with a maximum spring tidal range of 3.5m and a mean depth of 412.3m. Maximum flow speeds of the order of 40cms⁻¹ are clearly observed at spring tides within the bottom 1m of the boundary layer at the STABLE deployment site. There is also a large observed shear in the flow speed between rotors 1 and 3, with maximum flows reduced to 20cms⁻¹ for spring tide conditions at rotor 1, a distance of 27cm above the bed. The higher frequency contributions to the flow speed periodically enhance the instantaneous, maximum flow speeds such as is observed during Julian days 185, 187 and 189, and clearly marked by the arrows on
Fig 4.17 Time series of the mean logged rotor stack and pressure sensor. Rotor 2 dataset is unavailable.
The latter observation 'rides' on top of a maximum mean tidal flow speed of 40cmsg⁻¹ to produce a combined flow speed approaching 50cmsg⁻¹. These bursts are consistent in all three flow speed time series plots.

It is unfortunate that the direction information is not also available for the rotor stack. It is not therefore possible to resolve flow speed into velocity components, but it has been clearly shown in section 4.1 and in the pressure record in fig 4.17, that the flow in this upper slope region is dominated by the semi-diurnal tide, with maximum flows observed cross-slope. As a result, in the absence of a strong residual cross-slope current one would expect to observe approximately four peaks in the flow speed record each day in this region. With the exception of perhaps Julian day 187-188 in fig 4.17, there is a strong asymmetry in the semi-diurnal tidal signal with large maximum flow speed peaks alternating between a second much smaller flow speed peak, a contrast which is particularly apparent at spring tide. This would seem to suggest a significant cross-slope mean current at the STABLE site.

The similarity of the time series of rotors 1, 3 and 4 is confirmed in fig 4.18 which shows scatter plots of rotor pairs and a very high correlation in each plot. A line of best fit with zero offset is also plotted through the data points. An R² value greater than 0.979 is found for each rotor pair (r-squared evaluations are discussed further in section 4.2.2) although the value is smallest when rotor 3 is included as one of the variables. There is no observed shear associated with the variation in flow speed between rotors 3 and 4; in fact results suggest a slightly larger mean flow response by the flow meter 63cm above the bed (rotor 3) to that 81cm above the bed (rotor 4). This observation is discussed later in section 5.2.2.1 and implicates flow shielding (from the main body of the rig) or a reduced inertial response of the rotors (perhaps due to particles of sediment clogging up the rotor mechanism) for this observed result. The reduced response of rotor 2 at the beginning of the Goban Spur deployment of STABLE II (chapter 6) was shown, upon recovery and further investigation, to be due to the intrusion of sediment particles within the workings of the rotor. It is important to note at this stage that either explanation would only cause a change in the instrument gain and consequently each rotor is consistent in its
Fig 4.18  Scatter plots of pairs of flow speed records from Rotors 1, 2, & 3. Lines of best fit through zero are also shown.
profile with each other rotor. This is essentially the hypothesis used for the offset correction of the electro-magnetic current meters (EMCM's) in the following section, which is of vital importance for two reasons;
(i) to accurately measure the near-bed current velocity at the critical slope region
(ii) to estimate shear stresses, and hence infer sediment transport rates from the mean and turbulent current measurements (chapter 5).

It is commonly recognised that the sensitivity values for each EM head remain constant from laboratory calibration to field deployment, but that this is not the case for the zero-flow offset readings, which can vary between successive deployments (John Humphery, pers comm). The rotors therefore serve a secondary function, i.e. to correct for possible EMCM offsets variations. This is now discussed in detail in section 4.2.2.

4.2.2 EMCM calibrations and comparison with rotors

The orthogonal pair of EMCM's (here-after referred to as the port and starboard EMCM's) measured U, V and duplicate values of the W component of the flow. The calibration of the current meters was performed at the University of Liverpool, post deployment, by my colleagues at the Proudman Oceanographic Laboratory. The results of this calibration are shown plotted in fig 4.19. During calibration the port and starboard flow meters remained fixed to the EMCM spar (fig. 3.4) as in the field deployment, and the horizontal calibrations were carried out by towing the spar so that each EM head was inclined at 45 degrees to the measured water flow direction. The resulting calibration gains are scaled by a factor of $\sqrt{2}$ to compensate for this inclination (see table 4.5). The calibration details in table 4.5 were applied to all 88 EMCM's burst data sets and the mean value of each burst plotted in fig 4.20. The starboard meter was aligned at 054°T and is clearly more aligned to the direction of maximum flow than the port meter. The energy contained within the mean burst current meter time series is clearly dominated by the semi-diurnal tidal frequency in
Fig 4.19a  EM current meter calibrations for horizontal flow

Fig 4.19b  EM current meter calibrations for vertical flow
Fig 4.20 Flow meter calibrations to the E-M current meter burst dataset
the horizontal, with an observed neap-spring cycle and maximum spring tide velocities of 50cms$^{-1}$ in the horizontal starboard direction (-ve flow).

Table 4.5 Post deployment EMCM calibrations

<table>
<thead>
<tr>
<th>Head No.</th>
<th>Position</th>
<th>Velocity calibration (cms$^{-1}$)</th>
<th>R$^2$ value</th>
</tr>
</thead>
<tbody>
<tr>
<td>6763</td>
<td>vertical port</td>
<td>$Vel = [-0.3384 \times counts] - 12.6060$</td>
<td>0.9948</td>
</tr>
<tr>
<td>6763</td>
<td>horizontal port</td>
<td>$Vel = [+0.4682 \times counts \times \frac{1}{\sqrt{2}}] + 0.4999$</td>
<td>0.9996</td>
</tr>
<tr>
<td>6764</td>
<td>vertical starboard</td>
<td>$Vel = [-0.3429 \times counts] - 6.2511$</td>
<td>0.9990</td>
</tr>
<tr>
<td>6764</td>
<td>horizontal starboard</td>
<td>$Vel = [-0.4869 \times counts \times \frac{1}{\sqrt{2}}] - 9.6409$</td>
<td>0.9993</td>
</tr>
</tbody>
</table>

This is indicative of flow along 234°T which has a down-slope component of flow.

On initial investigation these results seem highly dubious for three reasons;

(i) There seems to be an extraordinarily large asymmetry in the horizontal starboard velocity time series with a weak return up-slope component of flow along 54°T (+ve flow). This brings into question the validity of the horizontal starboard offset calibration in table 4.5.

(ii) The vertical flow in both EMCM's also suggest offset discrepancies since there exist large downward residual flows in both port and starboard time series. If the flow meters are truly aligned in the vertical relative to the sea bed, then the mean vertical flow for each burst should be zero.

(iii) In the vertical starboard time series the semi-diurnal tidal frequency is clearly seen with an amplitude of the order of 4cms$^{-1}$, indicating that one, or possibly both sensors are slightly mis-aligned. There is also evidence to suggest a long term drift in the vertical offsets, which may or may not also be present in the horizontal.

The EM offsets therefore need to be calibrated by comparing the EM current speed time series with that of a selected rotor speed time series.

The power spectral densities for each EMCM time series are shown plotted in fig 4.21. Although the distribution of energy away from the semi-diurnal frequency is
Fig 4.21  Power spectral density (PSD) plots for the E-M current meter burst data sets.
not statistically significant, it is interesting to note that the low frequency 'offset drift' observed in the vertical EM current time series in fig 4.20 is also present in the PSD plots, with an approximate 3 day period. The same pattern is also observed for the horizontal PSD plots with a similar period.

The statistically significant peak in the horizontal current spectra is, as expected, in the semi-diurnal frequency band, with an order of magnitude difference in the PSD for port and starboard meters. The largest band is also at semi-diurnal frequency for the vertical starboard meter, but is notably absent from the vertical port meter. This implies, and has since been confirmed (John Humphery, *per comm*), that the port EM head is truly aligned in the vertical but that the starboard EM head is not. It is assumed therefore that the EMCM pair are truly mounted orthogonally on the EM spar but that the starboard EM head is not correctly fixed in the U, W plane relative to the EM spar.

A further implication of the lack of $M_2$ tidal signal in the vertical port time series is that it can now be confidently concluded that the EM spar and the STABLE rig itself has landed in a 'stable' position on the sea bed, with all three feet lying flat on the bed. The feet have not been individually scoured, buried, or sunk throughout the 8 days for which the EMCM's were operable. It can therefore be assumed with confidence that the near constant pitch and roll readings of $2.8^\circ$ and $5.2^\circ$ respectively measured throughout the STABLE deployment represent the true vertical orientation of the STABLE rig, i.e. the bathymetric slope locally. The pitch and roll values of $2.8^\circ$ and $5.2^\circ$ respectively imply a slope angle of $5.9^\circ$ locally (a gradient, $\alpha = 0.103$) in the direction $191^\circ$. This slope direction is offset by just 9 degrees when compared to the mooring transect along $020^\circ$. The gradient compares favourably with that assumed by Pingree & New (1988, $\alpha = 0.051$) for this region and with the profile obtained from the bathymetric survey shown in fig 3.5a.

The ratio of the variance of the semi-diurnal peak in the vertical and horizontal signals can now be used to estimate the degree of mis-alignment, $\phi$, of the U,W plane for the starboard head, using;
\[ \phi = \tan^{-1}\sqrt{\frac{PSD_{M2 \text{ Vertical}}}{PSD_{M2 \text{ Horizontal}}}} = \tan^{-1}\sqrt{\frac{37}{5694}} = 4.67^0 \] \[ \text{[4.4]} \]

The rotation of the orthogonal coordinate axes \( \Omega uv \) to the new axes \( \Omega u'w' \) through angle \( \phi \) is expressed as the product of the rotation matrix and the components of the original coordinate system, \( u \) and \( w \),

\[
\begin{bmatrix}
  u' \\
  w'
\end{bmatrix} = \begin{bmatrix}
  \cos \phi & \sin \phi \\
  -\sin \phi & \cos \phi
\end{bmatrix}\begin{bmatrix}
  u \\
  w
\end{bmatrix}. \text{ [4.5]}
\]

By applying this rotation to the vertical and horizontal starboard time series, the \( M_2 \) PSD peak in the vertical is shown in fig 4.22 to be eliminated.

The task of correcting for the EMCM offsets became quite a time consuming exercise but has definitely proved worth while. After a prolonged study of the time series plots with varying offsets, there was found to be a one hour phase lag in the EMCM real time data acquisition compared to that for the rotors. Once this time lag was corrected for and the following offset calibration was implemented, the EMCM results became highly coherent with those from the rotors.

For similar shelf deployments using 'steel' STABLE for the study of turbulent and wave induced velocities near to the sea bed, it has been sufficient to assume that the true residual flow is zero (e.g. Soulsby & Humphery, 1989), so that the record of all mean burst current velocity data is de-meaned to leave just the tidal component. In this instance, in the absence of a reliable direction data set on any of the slope near-bed flow meters, it was imperative to also account for the residual flow in a region where residual currents make significant contributions to the near-bed slope current (e.g. Pingree, 1988).

When an independent rotary current meter system is logging simultaneously with the EMCM's (as is becoming the norm for many wave and current instrumented benthic landers), the 'usual' procedure for calibrating the EMCM offsets would be to force the EMCM time series to zero flow at times when the rotor flow speed was zero. Fig 4.23 is a time series plot of the continuous flow speed of rotor 3 averaged over 9 minute periods and identifies those discrete 9 minute mean values which are
Fig 4.22 Power spectral density (PSD) plots for the starboard E-M current meter burst dataset after rotation of axes.

Fig 4.23 Rotor 3 flow speed during the period in which the E-M current meters were also operative.
coincident with the 2 hourly EMCM burst data samples. The loss of resolution from
the EMCM's is clear, but so too is the small number of instances when near zero
flow speed should be measured. The smallest coincident rotor flow speed of 3cms⁻¹
occurs during Julian day 185 and in principle the EMCM's resultant flow speed
could be forced equal to this by a variety of offset combinations. The STABLE
rotors have stall speeds of 2-3cms⁻¹ (largest for the bottom rotor) and this further
affects the use of zero rotor speed to calibrate the EMCM's. Another cause for
concern was the inertial response of the rotors described in section 4.2.1 and their
possible variable response at low flow speeds. Using the Steel STABLE rig at a site
where the near-bed log layer (refer to chapter 5) would be expected in all the
observed flow conditions, Williams (1995) observed that when the flow speeds were
less than 10cms⁻¹, the velocity profile was significantly different from the expected
logarithmic law. The reason for this was inconclusive, although deviations in rotor
response was rejected as an explanation in this instance. In the present case the
EMCM offset calibrations performed using this method were unsuccessful. Forcing
the resultant flow speed to near zero for near zero port and starboard EMCM
velocities, gave a poor correlation between the EMCM and rotor flow speed time
series over all.

It was finally found that the best method for estimating EMCM offsets was to find
the offsets of the U,V components of the EMCM flow speed time series which best
correlated with the rotor 3 flow speed time series, using regression analysis. This
was decided in view of the high degree of correlation between individual rotor flow
speed records. The original lab tested offset calibrations were removed from the port
and starboard mean horizontal time series and r-squared values of rotor 3 time series
versus the EM flow speed time series were calculated for all offset pairs between ±
20cms⁻¹ at 0.25cms⁻¹ intervals. The r-squared value is simply the square of the
correlation coefficient, in this case the 'Pearson Product Momentum Correlation
Coefficient', defined as;

\[ r = \frac{\left[ \sum (x - \bar{x})(y - \bar{y}) / n \right]}{\left[ \sqrt{\sum (x - \bar{x})^2 / n} \right] \left[ \sqrt{\sum (y - \bar{y})^2 / n} \right]} \]  [4.6]
where \(x, y\) are the two flow speed variables, \(n\) is the number of samples and the over bar denotes the mean value. Using standard Matlab functions;

\[
    r = \frac{\text{Covariance of } x \text{ and } y}{\left(\text{Standard deviation of } x\right) \times \left(\text{Standard deviation of } y\right)} \tag{4.7}
\]

Initially the optimum correlation was an \(r\)-squared value of 0.62 before the timing error was detected.

Fig 4.24 shows the results of this analysis for all \(r\)-squared values greater than 0.9. The plot shows that the port and starboard offset values converge to an optimal \(r\)-squared value exceeding 0.97. A contour plot of fig 4.24 is shown in fig 4.25, where the optimal \(r\)-squared value of 0.9728 is shown by the cross, with port and starboard offsets of \(-2.75\text{cms}^{-1}\) and \(-2.25\text{cms}^{-1}\) respectively. Also plotted on fig 4.25 are contours of the residual flow speed (in red) and the residual direction (in blue) resulting from the different offsets. With long slope defined as 290\(^{0}\)T, the results conclusively show that the mean flow is predominantly along-slope, in the correct sense with regard to the known poleward slope current, but also with a down-slope component of flow.

The comparison of the EMCM and rotor 3 current speed time series are shown in fig 4.26, with EM current speed verses rotor 3 current speed shown in fig 4.27. The comparison of the time series profiles is generally excellent, but there is an increase in the maximum flow speed of the EMCM's of the order of 6\(\text{cms}^{-1}\) at spring tide compared to rotor 3. This discrepancy remains unresolved since;

(i) The maximum flow speed observed from rotors 3 and 4 remains uncertain due to the possible shielding problems of the main body of the rig itself.

(ii) The long period offset variation observed in fig 4.20 for the vertical time series plots has a standard deviation of 2.4 \(\text{cm s}^{-1}\) (for the vertical port time series plot). If this offset drift also appears in the horizontal time series plots (as is suggested from the PSD plots in fig 4.21) then this could plausibly explain some of the difference observed.
Fig 4.24 Regression analysis identifying port and starboard offset pairs which have an R-squared value ≥ 0.9 when compared to Rotor 3.

Fig 4.25 A contour plot of the regression analysis shown in fig 4.24 above (black), whilst also showing the variation of residual flow speed (red) in cm/s$^{-1}$, and residual flow direction (blue) in °T for the port and starboard offset domain. See text for explanation of green contours.
Fig 4.26 A comparison of the E-M and Rotor 3 current speeds using the optimal offsets for the port and starboard flow meters.

Fig 4.27 E-M verses Rotor 3 flow speed for fig 4.26 above, with line of best fit plotted in red.
The latter point is investigated further in fig 4.28 which shows the comparison of the spectral density plots of horizontal flow speed for the EMCM's and rotor 3. The low frequency oscillation is also present in the rotor 3 flow speed record. The correlation between the two flow speeds across the frequency domain is shown in fig 4.29.

A very high correlation is found at this low frequency as is also found for the whole semi- and quarter diurnal frequency bands. Also, the increase in energy of the EMCM's compared to the rotors is found throughout the frequency domain in fig 4.28 and would therefore seem real at long period. As far as the 'apparent' long period motion in the vertical mean EMCM time series are concerned, the same conclusion cannot be reached. It is hard to think of a physical motion which has a vertical as well as horizontal oscillation, especially at low frequencies. This observation must be due to variable vertical offsets in successive EMCM data bursts.

The green contours in fig 4.25 show a narrow band extending through the port and starboard offset domain of the EMCM's where the difference between the PSD of the EMCM's and rotor 3 flow speed records is less than 20 (cm/s)²/day at M₂ tidal frequency. The plot shows that the r-squared value would be reduced if the offsets were chosen such that the total energy across the frequency domain was approximately equal for both flow speed time series data. To conclude, on the basis of the spectrum comparisons and the regression slope in fig 4.27, the calibration slopes of the EMCM's and the rotors don't exactly agree, though it can't be determined where the error lies. The EMCM's are actually positioned 23.7cm below rotor 3 and hence if both sensors are truly calibrated then the slope should be in the opposite sense to that observed in fig 4.27 for an assumed logarithmic vertical velocity gradient (section 5.2.1.1).

Nevertheless, this fairly exhaustive section has provided an important and accurate description of the tidal current velocity from the mean burst data set for the STABLE site. It has also been conclusively shown that the residual current is along-slope (poleward) with a down-slope component of flow for all probable offset pairs. This section also shows that the offset values do indeed vary between calibration and
Fig 4.28 A comparison of the Power Spectral Densities (PSD) of the E-M's and Rotor 3 times series.

Fig 4.29 The coherence function for fig 4.28 above.
deployment and shows the vital secondary role the rotors have played in this deployment. It was not possible to conclude whether or not the offsets varied throughout the deployment for each successive 9 minute burst sample period, but the vertical time series profiles from the port and starboard EMCM's suggested this may be the case. The final calibration details of the port and starboard current velocities in cm s\(^{-1}\) are shown in matrix form in table 4.7 (these results are used in later sections).

Table 4.7 Final EMCM calibration file (cm s\(^{-1}\))

<table>
<thead>
<tr>
<th>EMCM calibration</th>
<th>Rotation vector</th>
<th>EMCM sensitivity</th>
<th>inferred offset</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hor Port</td>
<td>1 0 0 0</td>
<td>(0.4682 \times \text{counts} \times 1/\sqrt{2})</td>
<td>-2.75</td>
</tr>
<tr>
<td>Vert Port</td>
<td>0 1 0 0</td>
<td>(-0.3384 \times \text{counts})</td>
<td>-6.17</td>
</tr>
<tr>
<td>Hor Star</td>
<td>0 0 \cos \phi</td>
<td>(-0.4869 \times \text{counts} \times 1/\sqrt{2})</td>
<td>+2.25</td>
</tr>
<tr>
<td>Vert Star</td>
<td>0 0 -\sin \phi</td>
<td>(-0.3429 \times \text{count})</td>
<td>+1.46</td>
</tr>
</tbody>
</table>

The vertical port and starboard offsets are set such that the mean of the mean burst time series is zero. Standard deviations of 2.40 cm s\(^{-1}\) and 1.80 cm s\(^{-1}\) for the port and starboard vertical time series records about this zero mean, are a measure of the accuracy, or perhaps the stability of these calibrations over the deployment period. Finally, table 4.8 compares the inferred and calibration offset values, to emphasise that calibration offsets don't seem to relate even qualitatively to field offsets. However, these differences are not unexpected because in situ offsets are generally smaller than lab calibrated values (D.A. Huntley, pers comm.).

Table 4.8 Inferred and calibrated EMCM offset values.

<table>
<thead>
<tr>
<th>EMCM calibration</th>
<th>Inferred (cm s(^{-1}))</th>
<th>Calibrated (cm s(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hor Port</td>
<td>-2.75</td>
<td>0.4999</td>
</tr>
<tr>
<td>Vert Port</td>
<td>-6.17</td>
<td>-12.606</td>
</tr>
<tr>
<td>Hor Star</td>
<td>-2.25</td>
<td>-9.641</td>
</tr>
<tr>
<td>Vert Star</td>
<td>1.46</td>
<td>-6.251</td>
</tr>
</tbody>
</table>
4.2.3 EMCM tidal and residual flows

Using the calibration details given in table 4.7, a residual current of 11.2 cms\(^{-1}\) in direction 261°T was measured 39cm above the bed at the STABLE site by the EMCM's. The progressive displacement plot for the EMCM results was shown earlier in fig 4.4 in section 4.1.1. This is the largest residual current to be observed across the mooring transect and infers a poleward along-slope component of flow of 9.84cms\(^{-1}\) along 290°T and a down-slope component of 5.35cms\(^{-1}\) along 200°T.

The east versus north components of flow have been plotted in fig 4.30 and show the characteristic elliptical motion of the flow very close to the bed. A least squares fit has been added to this short data set for the M2 tidal constituent (by comparing the observed tidal amplitude with a predicted tidal amplitude via the sum of the squared differences). The results show the orientation of the major axis of the M2 tidal ellipse to be aligned along 041°T, which is 12° clockwise of that calculated 68mab at mooring 124 (which itself will be veered more cyclonically nearer to the bed). The amplitude of the M2 tidal current along the major axis, 'a', is equal to 25cms\(^{-1}\), with the ratio of the major to minor axes, 'b/a', equal to 16%. Compared to the mooring results described earlier in section 4.1, the flow is far more rectilinear near to the bed at the STABLE site. Clockwise polarisation is often maximal near the shelf break (J.M. Huthnance, pers comm.).

The current speeds are resolved along their major and minor M2 tidal axes in fig 4.31 such that the M2 tidal flow is minimised along the minor axis, and compared to the flow speed. Maximum flow speeds approaching 40cms\(^{-1}\) are shown to correlate with the maximum down-slope component of flow along the major axis of the M2 tidal ellipse. The second much smaller flow speed maxima which alternate between successive flow speed peaks is shown to correlate with the maximum up-slope component of flow along the major axis of the M2 tidal ellipse. The cross-slope asymmetry in the tidal flow predicted from the rotors in section 4.2.1 is therefore confirmed and infers that both the residual current and the maximum flow speeds
Fig 4.30 East (+ve) verses north (+ve) velocity for E-M flow meters in black. A semi-diurnal least-squares fit to the orthogonal components of the flow is plotted in red.

Fig 4.31 Time series plots of the E-M flow meter velocities resolved relative to (a) the minor axis and (b) the major axis of the semi-diurnal tide, and (c) the flow speed. See text for further details.
have down-slope components of flow. This is discussed further in section 4.3 and has important sediment transport implications (discussed in chapter 5).

4.2.4 Inter-burst EMCM flows

The rapidly sampled burst data sets are now examined to consider the turbulent response of the boundary layer to the tidally dominant mean flows described in section 4.2.3 at the STABLE site. Time series plots of standard deviation for each burst are shown for the horizontal port (HP) and starboard (HS) and vertical port (VP) and starboard (VS) components of flow in fig 4.32. The variance within each burst is again shown to be tidally modulated. The magnitude of the variance is shown to be similar for all but one EMCM component, the exception being the VP record which has standard deviation values twice as large as the other three EMCM components. For future shear stress estimates which utilise the turbulence data (Chapter 5), the $u'$, $v'$ and $w'$ turbulent components of flow will be taken from the HS, HP and VS data sets. There is clearly a calibration error with the VP meter since the turbulence should be almost isotropic and therefore approximately equal in magnitude for each plot in fig 4.32.

The standard deviation for each burst in the HS EMCM flow component is shown plotted in fig 4.33 in red and compared to the current velocity resolved along the major axis of the semi-diurnal tide ($040^\circ_T$). The general increase in background variance at the beginning of the record (JD 184-185) and at the end of the record (JD 188-191) is of some concern. Shear stress measurements in chapter 5 reveal that the bursts were contaminated with noise during these periods and are described further in section 5.2.2.2. Maximum current flows along $220^\circ_T$ of up to 40cms$^{-1}$ are shown to coincide with peaks in the turbulent intensity of the near bed flow throughout the neap-spring cycle. This implies that the turbulence intensity generally increases during periods of strong down-slope flow which one might intuitively expect to be the case. A smaller maximum variance peak is generally associated with on-shelf flow but is not a consistent trend. Similar inconsistencies
Fig 4.32 Time series plots of standard deviation for each EMCM inter-burst dataset.

Fig 4.33 A comparison of the horizontal EMCM current velocity resolved along the semidiurnal major axis (040°T) and the horizontal starboard flow variance (standard deviation, (cm/s)).
correlate minimum flow speed with minimum variance. These results have important implication for sediment transport and are described in chapter 5.

Individual EMCM inter-burst time series plots are shown in figs 4.34-4.37. The sampling time and phase in the tidal cycle is shown marked in red on the mean bursts time series plots in fig 4.31. Fig 4.34 is a typical burst showing the U & u' (HS) V & v' (HP) and W & w' (VS) mean and turbulent components of flow over the 9 minute sampling period. Fig 4.35 shows an example of a period of tidally accelerating flow in the HP and HS time series plots which is consistent with the on-shelf flooding phase of M₂ tidal cycle in fig 4.31. It is re-assuring not to see the same trend in the vertical burst component. Fig 4.36 shows an clear example of internal wave activity within the bottom 1m at 388m depth on the upper slope. With a period of a few minutes, the event is recorded during maximum on-shelf tidal streaming and is associated with a rapid up-slope advection of isotherms 4km up-slope at mooring 124. There is no general pattern to these observations and so they cannot be attributed to any specific phase of the M₂ tidal cycle. Once again it is re-assuring to see the fluctuation absent from the vertical burst component and increases our confidence in these results. Finally, fig 4.37 shows an example of a typical problematic data set. A common characteristic of the VP burst time series was for the turbulent intensity to be large at the beginning of the record and to taper to smaller values at the end of the record. This trend persisted for a twelve hour sampling period during Julian day 189.

4.3 Discussion

The near-bed hydrodynamics at the STABLE site show a region of enhanced down-slope tidal flows and corresponding periods of increased turbulence. High frequency internal wave fluctuations are also observed in the horizontal inter-burst records. The example shown in fig 4.36 occurs during maximum on-shelf tidal streaming. Pingree & New (1989) estimate the tidally averaged stability frequency at this critical depth of 388m to be 0.95 cpd, which is also a minimum value for the upper 1000m. This theoretically creates a large 'spectral gap' between the transfer of this
Fig 4.34 A typical burst EMCM time series, sampled at 4 Hz for 9 minutes every two hours. These records were collected at 1619hrs on Julian day 189 (08th July, 1990; marked as a red dot in the mean burst time series plot in fig 4.31 in position [G]).

Fig 4.35 As for fig 4.34 above but collected at 2019hrs on Julian day 184 (03rd July, 1990; position [D] in fig 4.31) showing a period of accelerating tidal flow.
Fig 4.36 Near-bed internal wave activity is clearly evident from the horizontal time series collected at 0019hrs on Julian day 188 (07th July, 1990; position [E] in fig 4.31).

Fig 4.37 A problematic dataset of which this is an extreme example. The larger variance at the beginning of vertical starboard record is characteristic of several other bursts in the vertical component. These time series were collected at 0419hrs on Julian day 189 (08th July, 1990; position [F] in fig 4.31).
baroclinic internal wave into higher frequency internal waves and finally near bed turbulence. There must be a transfer of energy from the tide to turbulence via the 'spectral gap' and so these horizontal current motions are assumed to be real.

Recent modelling study by Slinn & Riley (1996) showed that there is strong periodic diapycnal mixing within the boundary layer associated with internal wave breakdown into turbulence above sloping bottoms at and near critical angle reflections. The study showed the formation of a distinctive front which moves upslope with the phase speed of the oncoming wave and can resemble a bore-like feature for steep slopes (section 4.1.5). This flow is quasi-periodic with strong mixing and small scale dissipation followed by a quieter period of re-stratification. At the STABLE site the observations suggest that increased bed shear turbulence is associated with the stronger maximum down-slope phase of the (barotropic + baroclinic) $M_2$ tide. These results are in agreement with Slinn & Riley (1996).
Chapter 5

La Chapelle Bank field results – Sediment transport observations and predictions

One of the main goals of this field study was to obtain direct evidence of sediment transport rates in the upper shelf/slope region of La Chapelle Bank by making continuous *in situ* measurements of suspended and bedload transport. This has been achieved at the STABLE site using the ABS sensor and camera system, but also inferred from the near-bed transmissometer results at mooring 124. This chapter discusses these results, compares them with predicted bedload transport rates, to previous sediment transport studies (Heathershaw & Codd, 1985 & 1986) and to an internal tide model, the latter suggesting that the shelf break is a region of bed-load parting (Heathershaw *et al.* 1987).

5.1 *In situ* sediment transport observations at the STABLE site.

Pingree (1988) suggested that the upper slopes of La Chapelle Bank are regions of active erosion of bottom sediments. Pingree (1988) infers this erosional state from
near-bed optical beam attenuation profiles, which show increased levels of turbidity near to the bed during spring tides. This increase in turbidity is associated with increased bottom currents on slopes which are critical for internal tide generation. Little is known about the nature of the bed and so the magnitude of the stresses required to erode the slope material is mostly unknown. Sediment transport predictions to date have therefore not been possible.

Previous work undertaken in this region (Heathershaw & Codd, 1985 & 1986, Heathershaw et al., 1987) fell short of direct near-bed current measurements at the shelf break (< 170m depth). The in situ measurements of sediment transport and near-bed turbulence which are now described are therefore thought to be unique to this shelf break / upper slope region.

5.1.1 Sea-bed photographs

In addition to the Shipek grab samples described in section 3.2.2.1, a synoptic view of the current-bed interaction was investigated at the STABLE site using the on-board camera and flash system. Photographs were taken of the sea bed at two hourly intervals and were synchronised with the beginning of each EMCM burst. Plate 5.1 is typically representative of the photographs taken and shows (i) the first sea bed shot and (ii) the last sea bed shot of the ~19.75 day deployment. The results are consistent with those from the Shipek grab samples, i.e. that the bed is composed of a badly sorted gravely sand. The larger material is clearly visible in both photographs and is made up of mainly shell fragments. There is no noticeable movement of some of the larger material (some examples are shown numbered in plates 5.1(i) & (ii)) but there are also examples of shell fragments which are visible in plate (i), such as directly above the shadow of the compass, but absent from plate (ii). The movement of this larger material may have been caused by the benthic biology, examples of which are shown on plate 5.2. The series of photographs taken are representative of a well consolidated gravelly bed and are indicative of an energetic environment which is able to winnow out and erode the finer sandy material.
Plate 5.1  STABLE sea-bed photography. (i) Frame No 7; First sea-bed shot (0219hrs on 3rd July, 1990).  (ii) Frame 474; Last sea-bed shot (2019 hrs on 22nd July, 1990). Numbers identify individual pebbles which are clearly visible on both shots.
Plate 5.2  STABLE sea-bed photography. (i) Frame No 115; A crab approximately 22cm across the carapace.  
(ii) Frame No 126; A fish approximately 30cm long.
The STABLE photographs and the Shipek grab sample results are also consistent with the bed composition shelfward of the shelf break on La Chapelle Bank. Heathershaw (1986) describes the results of 24 sediment samples taken using the Shipek grab at depths ranging from 160-200m. The mean grain size of each sample varied from 339-807\(\mu m\) with an overall mean of 506\(\mu m\). The carbonate content varied from 20-80\%. Unfortunately, carbonate analysis of the present Shipek grab samples was not undertaken but the grain size results are generally consistent with the results from Heathershaw (1986). The mean grain sizes for each sediment are 870, 860 and 595\(\mu m\) which have an overall mean of 775\(\mu m\). Of the 24 samples taken by Heathershaw (1986) only one sample had a mean exceeding 700\(\mu m\) (807\(\mu m\)) and so these results suggest a slight increase in flow rates on the critical slope compared to further on-shelf. It is significant in itself that the nature of the bed should suggest comparative flow rates on the upper slopes at 385m to those near the shelf break (160-200m).

Transport of the mean grain size sediment described above will primarily be as bedload and saltation along the bed. The finer material will be re-suspended at peak flow rates but will quickly fall to the bed once the friction velocity, \(u_\star\), reduces below the critical threshold (see following section). The 84th percentile of the three Shipek grab samples had a mean grain size of 250-350\(\mu m\). Sediment of this size has a 10cm settling time of the order of seconds at \(20^\circ C\) (Buller & McManus, 1979). It is therefore fair to predict that the sediment will only be transported at times of peak flow.

5.1.2 ABS results

A total of 256 acoustic backscatter records were collected from the 3MHz Acoustic Backscatter System (ABS), mounted approximately 1.2m above the bed looking vertically down. The backscatter returns were time gated into 8 bins and sampled synchronously with the EMCM bursts at 0.5Hz for 512 seconds every 2 hours. A total of 8 bins were arranged with a gap of 15cm from the transducer to the first bin, followed by successive bin widths of 2 \(\times\) 22.5 cm, 2 \(\times\) 15cm and 3 \(\times\) 7.5cm. The
calibration of the raw ABS data was performed by P.J. Hardcastle and P.D. Thorne at POL and the preliminary results of the ABS post-processing are reported by Hardcastle & Humphrey (1991). The inter-burst calibrated ABS data has not been available for analysis but Hardcastle & Humphrey (1991) conclude from a preliminary investigation of the data set that there was no evidence of any ejection of material from the sea-bed into suspension. Some records do show evidence of bottom movement which Hardcastle & Humphrey (1991) suggest was due to biological activity near to the bed.

The mean burst ABS data set has been kindly made available by POL for further interpretation. A time series contour plot of the results are shown in fig 5.1 below. Concentration is in milligrams/litre, assuming sand sized material in suspension.

![Fig 5.1 Suspended sediment concentration derived from acoustic backscatter.](image)

Because suspended sediment size is unknown only relative changes in concentration should be interpreted with any degree of certainty. The transport of suspended sediment is clearly tidally modulated with neap-spring and individual M2 tidal cycles clearly evident. As one would expect, concentrations are highest nearest to the bed which suggests that the sediment source is from locally eroded bed material.
general lack of mobility of much of the observed bed material in plate 5.1 supports the hypothesis that the material detected by the ABS was of a fine nature and may therefore have been below the threshold for detection at low concentrations. The results presented in fig 5.1 do however suggest a continuous background concentration throughout the deployment.

Figure 5.2c shows the product of the mean burst cross-shore velocity (from the EMCM's at 39.3cm above the bed) and the mean burst suspended sediment concentration (from the ABS sensor at 40cm above the bed). Although absolute suspended sediment concentrations (ssc) is not confidently known the relative magnitude of the cross-slope flux of sediment can still be determined.

![Fig 5.2 Time series plots of (a) cross-slope velocity, (b) ABS suspended sediment concentration and (c) cross-shelf suspended sediment flux approximately 40cm above the bed at the STABLE site. Negative values represent down slope fluxes.](image)

The dashed lines have been put in fig 5.2 to show typical correlation's of the ABS ssc and EMCM cross-slope velocity and identify times of maximum up-slope, down-
slope and slack cross-slope current. The ABS modulations are clearly of $M_2$ period and show increased concentrations at times of maximum current and not maximum advective excursion. This would suggest that the sediment is being resuspended locally. Fig 5.2c shows the largest fluxes occur at spring tide during off-shelf tidal streaming. The smaller up-slope flux of sediment during the up-slope phase of the tide implies a net transport of sediment down slope at the STABLE site.

It is slightly confusing that the results described to date are not in complete agreement with one another. The Shipek grab samples collected near to the STABLE site (section 3.2.2.2) suggest that the bed is of a sand-gravel composition. The sea-bed photographs are generally in agreement with the sediment analysis results and do not show any visible evidence of sediment in suspension. There is however evidence of bedload movement. The ABS burst analysis (Hardcastle & Humphery, 1991) does not show any resuspension events, but the mean ABS and EMCM time series suggests that peaks in ssc correlate with maximum flow speeds.

One of the fundamental questions to be addressed from section 5.1 is whether sediment is being eroded further up-slope and then being advected past STABLE or whether the near-bed turbulence is energetic enough to re-suspend sediment locally. This uncertainty is investigated further in section 5.2.

5.2 Sediment transport predictions

The measurement of the near-bed turbulent flows from the EMCM's have been used to estimate the near-bed stresses (section 5.2.2) to determine if they are large enough to erode the bed material at the STABLE site (5.2.3).

For a sand-gravel bed the sediment will predominantly move as bedload if the critical threshold velocity is exceeded. One would expect that if the $M_2$ tidal currents are preferentially winnowing out the fine material at the bed, then it will also be possible to correlate predicted bedload transport with in situ measurements.
of suspended sediment. The procedure for predicting initial motion and bedload transport of sediment of a known mean diameter is summarised by Sternberg (1972) and is shown in the schematic below (fig 5.3).

1. Calculate the mean velocity 1 m off the bed.
2. Estimate the boundary shear stress, $\tau_0$, from the velocity measurements.
3. Is the critical shear stress, $\tau_{cr}$, exceeded to cause grain movement?
4. Estimate the mass transport as bedload (for sandy/gravelly beds).

Fig 5.3 Procedure for predicting initial grain movement and transport as bedload (after Sternberg, 1972)

The procedure involves being able to accurately predict the shear stresses within the boundary layer, of which there are several currently available methods. Techniques include the velocity profile method, the Reynolds stress or eddy correlation method, the quadratic stress law method, the inertial dissipation method and the turbulent kinetic energy method. If the measured critical stress exceeds the threshold for sediment movement then the mass transport of bedload can also be predicted, again using several theoretical approaches. Bed-load transport equations are briefly described in section 5.2.2. The structure of the boundary layer is described in detail by Wimbush & Munk (1970), Weatherly & Martin (1978) and Soulsby (1983).

5.2.1 Boundary layer bed stresses

Three methods are discussed in the following sections, namely the logarithmic profile method (section 5.2.1.2), the Reynolds stress method (section 5.2.1.3) and the turbulent dissipation rate method (section 5.2.1.4). All three methods are briefly
reviewed by Huntley (1988) and Dewey & Crawford (1988) and rely on velocity measurements being taken within the boundary layer.

5.2.1.1 Fundamental assumptions for calculating bed stresses.

When water flows over sediment lying on the sea bed the flow experiences retardation due to frictional forces at the bed. The frictional force causes shear, \( \tau_0 \), between successive layers such that at \( z=0 \), \( U=0 \) and at \( z=\delta \) (at the top of the boundary layer), \( U= \) the mean stream velocity. For steady flows of an unstratified fluid over a boundary, the vertical distribution of the current is known to vary logarithmically with depth close to the bed. The velocity gradient responsible for maintaining this stress, \( \tau_0 \), is related to a friction velocity, \( u_* \), defined by:

\[
\tau_0 = \rho u_*^2
\]

by using the dimensional analysis argument that:

\[
\frac{dU}{dz} = \frac{u_*}{\kappa z}
\]

where \( z \) is the distance from the bed and \( \kappa \) is Von Karman's constant = 0.4. Upon integration Eqn. 5.2 becomes (e.g. Tennekes, 1973);

\[
U(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right)
\]

for hydrodynamically rough flows (which are discussed later in this chapter and assumes no viscous sub-layer), where \( z_0 \) is a roughness length (which is related to the geometry of the boundary), and \( U \) is the mean stream velocity at height \( \delta \) above the bed. In physical terms, the logarithmic layer is described as the range of heights for which the height, \( z \), is simultaneously too great for the details of the geometry of the bed to affect the flow, and too small for the flow to be influenced directly by the free-stream velocity (Soulsby, 1983). The logarithmic layer thickness is therefore determined by the free stream velocity and the geometry of the bed and typically varies between 1m in the deep ocean and 10m on the continental shelf (Wimbush & Munk, 1970). For plane beds with a uniform grain size, \( z_0 = D/30 \); where \( D \) is the grain diameter. The measured value of \( z_0 \) will deviate from this relationship.
considerably if bedforms are present or if the sediment is poorly sorted (see later in section 5.2.2).

It should be noted that there are limitations in the assumptions made in the derivation of Eqn 5.3 which can be violated in the field. These assumptions are (i) that the mean current $U$ is steady and (ii) that the fluid is unstratified. Examples where the assumptions do not hold true are (i) in shelf sea regions during periods of rapidly accelerating or decelerating tidal currents and (ii) when the boundary layer becomes stratified due to variations in temperature and/or density, or variations in the effective density due to suspended sediment load.

Above the logarithmic layer there exists an outer Ekman layer which can extend tens of meters above the bed. This layer defines the outer limit of the bottom boundary layer (bbl). For ‘turbulent’ shelf sea environments the thickness of the bbl from the sea bed to the top of the Ekman layer, $\delta$, is often estimated from the velocity shear, $u_*$, as (e.g. Wimbush & Munk, 1970);

$$\delta = 0.4 \frac{u_*}{f} \quad [5.4]$$

For oceanic boundary layers, Weatherly & Martin (1978) have suggested that eqn. 5.4 may not generally apply for a bbl formed in a stably stratified ocean. Solely on dimensional grounds, Weatherly & Martin (1978) give an expression for the thickness of the bbl as;

$$\delta = \frac{A \cdot u_*}{f(1 + N^2 / f^2)^{\frac{1}{2}}} \quad [5.5a]$$

where $A =$ constant and $N =$ Brunt Väisälä frequency (which reduces to $h = A \cdot u_*/f$ for $N<<f$ and $h = A \cdot u_*/(f \cdot N)^{\frac{1}{2}}$ for $N>>f$ ), and their model results suggest that $A=1.3$ such that;

$$\delta = \frac{1.3 \cdot u_*}{f(1 + N^2 / f^2)^{\frac{1}{2}}} \quad [5.5b]$$
The numerical model results are also consistent with observations by Weatherly & Martin (1978) and therefore suggest that the bbl thickness can be appreciably different to that defined by eqn. 5.4 for bbl's formed in stably stratified oceans.

Each of the methods used for calculating shear stresses rely on current measurements being made within the logarithmic part of the boundary layer and each have their own physical limitations. These methods are now considered in more detail.

5.2.1.2 Velocity profile method

Equation 5.3 shows that near to the boundary the mean velocity is expected to vary logarithmically with height. By measuring the mean velocity at several points above the bed a best line fit to $U$ versus $\ln z$ can be used to estimate $u^*$ and $z_0$ directly. This is the easiest and the most widely used method in the field. In some instances an assumed $z_0$ and one current meter measurement within the expected logarithmic layer is all that is used to estimate the shear in the boundary layer. There are limitations to this method which make shear stresses very difficult to predict accurately, some of which are described below (from Grant et al., 1984, Huntley, 1988 and Dewey & Crawford, 1988);

(i) The mean current must be measured at several heights
(ii) The measurements must be to a high degree of accuracy since the logarithmic layer predicts only small shear at typical current meter heights (between 30 & 200cm for example).
(iii) The exact heights of the current meters above the bed must be known.
(iv) Estimates at low flow speeds are difficult due to the stall speeds and resolution of the current meters.

The four rotor current meters on STABLE and STABLE II (described later in chapter 6) have successfully collected data on two previous continental shelf deployments and are described by Soulsby & Humphery (1989) and Williams (1995) respectively.
Soulsby & Humphery (1980) found considerable scatter in the accuracy of the results but Williams (1995) calculated shear stresses with linear regressions in the logarithmic profile with $R^2$ values $> 0.95$ (rejected if not), but usually $> 0.98$. Grant et al. (1984) showed that an $R^2$ value $> 0.993$ was required from velocity measurements at four heights above the bed, in order to maintain uncertainties in $u_*$ of $\pm 25\%$.

5.2.1.3 Reynolds stress method

The boundary layer stress is transmitted to the fluid above by Reynolds stresses $\rho u'w'$ which arise due to a correlation between the horizontal and vertical turbulent velocities (Turner, 1981). If it is assumed that the turbulent stresses within the bottom part of the logarithmic layer are representative of the stresses at the bed (i.e. constant stress) then turbulent current measurements can be used to infer bed stresses directly. The Reynolds stress or eddy correlation method (e.g. Soulsby, 1983) uses the high frequency turbulent fluctuations to estimate $\tau_0$ from the time averaged products of the horizontal and vertical velocity turbulent fluctuations, i.e. $-\rho u'w'$ and $-\rho v'w'$. These Reynolds stresses represent turbulent fluxes of momentum, equal to $\tau_{xz}$ and $\tau_{yz}$ across the $x$-$y$ plane. The bed shear stress is found using:

$$\tau_0 = \rho \left( u'^2 + v'^2 \right)^{1/2}$$  \hspace{0.5cm} [5.6]

where the time averaged product is typically over the order of 10 minutes duration (Soulsby, 1980). It is difficult to calculate accurate shear stresses using this method because the measurements are very sensitive to sensor misalignment. Errors of 10% per degree of tilt out of the vertical are typical (Soulsby, 1983, Huntley 1988).

5.2.1.4 Inertial dissipation method (IDM)

This technique is more theoretical than the previous two methods and uses the velocity spectra from high frequency time series measurements to predict the bed shear stresses. The main advantage of this method over the Reynolds stress method is
that small miss-alignments of the axes have been shown to have little effect on the stress estimates (Huntley & Hazen, 1988).

The IDM calculates $u_*$ from an assumed distribution of turbulent kinetic energy within the inertial sub-range of the scalar wavenumber spectrum, $E(k)$, which has the form (e.g. Grant & Williams, 1984, Huntley, 1988 and Dewey & Crawford, 1988):

$$E(k) = \alpha \varepsilon^{3/4} k^{-5/3}$$  \[5.7\]

where $\alpha$ is a three dimensional Kolmogorov constant, $\varepsilon$ is the dissipation rate and $k$ is the wavenumber. The inertial sub-range is the range of wavenumbers where the production and dissipation of energy are well separated in the kinetic energy spectrum. The relationship defined by Eqn. 5.7 follows from Kolmogorov theory and its assumed form within the inertial sub-range is essential for calculating bed stresses using the IDM. The formulation of Eqn. 5.7 is now described (e.g. Tritton, 1988). The basic principles of this theory are shown below in fig 5.4 and will be described further in the following discussion.

Kolmogorov theory can be used to express the distribution of turbulent kinetic energy (TKE), defined by $E(k)$ over scalar wavenumber, $k$, such that;
\[ \int_0 E(k) \, dk = \frac{1}{2} (u'^2 + v'^2 + w'^2) = \text{total TKE} \tag{5.8} \]

where \( E(k) \, dk \) is equal to the contribution of TKE in an interval \( dk \), where \( k = \frac{2\pi}{\lambda} = \text{wavenumber} \). Turbulent velocity measurements in the ocean are generally in the form of time series so that equation 5.8 needs to be expressed in terms of frequency and not wavenumber. The Taylor concept of frozen turbulence is used to convert wavenumber to frequency (e.g. Huntley, 1988) by assuming that the time scale of the turbulent eddy is much larger than the time for the eddy to pass the measurement point by the mean flow. In terms of the velocity field;

\[ \bar{u} >> \sqrt{u'^2} \tag{5.9} \]

i.e. the fluid moves past the sensor as a frozen field. Therefore;

\[ T = \frac{\lambda}{u} \Rightarrow \sigma = \frac{2\pi u}{\lambda} = k \bar{u} \tag{5.10} \]

The turbulent velocity variation with wavelength can therefore be expressed in one dimension in terms of frequency and visa-versa, i.e.

\[ u'(t) = \phi_u(\sigma), \text{ where } \phi_u(k) \, dk = \phi_u(\sigma) \, d\sigma \]

\[ \phi_u(k) = \phi_u(\sigma) \frac{d\sigma}{dk} = \bar{u} \phi_u(\sigma) \tag{5.11} \]

At high wavenumbers the spectral transfer of turbulent energy, \( \partial E/\partial k \), is primarily from low to high wavenumbers and can be represented by the equation;

\[ \frac{\partial E(k)}{\partial t} = -\frac{\partial \varepsilon}{\partial k} - 2\nu k^2 E(k) \tag{5.12} \]

where \( \varepsilon(k) \) is the amount of energy moving from one energy band to another, \( \nu \) is a molecular viscosity. Term [1] on the right hand side of Eqn. 5.12 represents the rate of change of energy associated with \( k \) and the negative term [2] represents viscosity and energy dissipation. If the total dissipation is in steady state then

\[ \int_{k_o}^{k_e} \frac{\partial \varepsilon}{\partial k} \, dk = -2\nu \int_{k_o}^{k_e} k^2 E(k) \, dk \tag{5.13} \]

where \( k_o = \text{a typical wavenumber of the turbulence energy production} (E=E[k,\varepsilon,k_o]) \).
Kolmogorov theory predicts that there is a universal form of $E(k)$ for high wavenumbers in which the turbulence is homogeneous, isotropic and randomised by cascade. This concept has been summarised in the schematic in fig 5.4.

Within the inertial sub-range ($k_0 < k < k_v$; where $k_v$ describes the high wavenumber limit of the inertial sub-range where viscosity becomes important for the dissipation of energy at molecular scales) the spectral density function $E(k)$ depends only on $\varepsilon_0$ and $k$. Dimensional analysis of $\varepsilon_0$ ($[\varepsilon_0] = (\text{velocity})^2/\text{time} = L^2 T^{-1}$) and $k$ ($[k] = 1/\text{wavelength} = L^{-1}$) and $E(k)$, ($[E(k)] = (\text{velocity})^2/\text{wavenumber} = L^3 T^{-2}$) shows that $E(k)$ must be of the form $E(k) \sim \varepsilon_0^{2/3} k^{-5/3}$, i.e. $E(k) = \alpha \varepsilon^{2/3} k^{-5/3}$ (Eqn. 5.8). $\alpha$ is determined experimentally. This scalar wavenumber spectrum can be represented as one-dimensional spectra which takes the form

$$\phi_i(k_i) = \alpha_i \varepsilon^{2/3} k_i^{-5/3}$$  \[5.15\]

where the $i$ represents a given direction in the horizontal and vertical components of turbulent flow and $\alpha_i$ the corresponding Kolmogorov constant. Within the inertial sub-range of the spectrum, the Kolmogorov -$5/3$ law describes the run-off from low to high wavenumbers.

The method of equating Eqn. 5.7 to the bed stress is described by Grant et al. (1984), Huntley (1988) and Dewey & Crawford (1988) and involves linking the dissipation rate $\varepsilon$ to $\tau_0$. Three assumptions are made when estimating $u_*$ from Eqn. 5.7 (from Huntley, 1988);

(i) There is a local balance between the rate of production of TKE through the Reynolds stresses and the rate of viscous dissipation. The production of TKE is then given by $(\tau/\rho) \partial u/\partial z$ and is equal to $\varepsilon$.

(ii) The measurements are assumed to be taken within the constant stress part of the logarithmic layer. Therefore using (i) and Eqns. 5.1 and 5.3, $\varepsilon = u_*^3 / k z$. Substituting into Eqn. 5.7 gives:
(iii) The turbulent Reynolds number \( (Re) \) is greater than some critical Reynolds number \( (Re_c) \);

\[
Re = \frac{u_* k z}{\nu} > Re_c
\]  

[5.17]

where \( 2500 > Re_c < 4000 \) (Huntley, 1988).

The frozen turbulence hypothesis (Eqns. 5.9-5.11) converts Eqn. 5.16 from wavenumber spectra to velocity spectra.

The algorithm used for calculating \( u_* \) from the EMCM burst velocity spectra is taken from Stapleton & Huntley (1995) and is not described further. For the EMCM data sets, the horizontal \((i = 1, 2)\) and vertical \((i = 3)\) Kolmogorov constants are taken as \( \alpha_1 = 0.52, \alpha_2 = \alpha_3 = 0.69 \) respectively (see Huntley & Hazen, 1988 for further details) and \( z = 0.393m \). The bed shear stress is estimated from the velocity spectra using Eqn. 5.16. The upper and lower limits of the inertial sub-range for each spectra are linearly dependant on (i) the physical dimensions of the EMCM's (Soulsby, 1980) and (ii) by the height of the EMCM's above the bed such that (Stapleton & Huntley, 1995);

\[
(i) \quad f = \frac{2.3 \bar{u}}{2\pi d} \quad \text{(upper limit)} \quad [5.18]
\]

\[
(ii) \quad f = \frac{\bar{u}}{z} \quad \text{(lower limit)} \quad [5.19]
\]

where \( d = 10cm \) is the distance between the electrodes on the sensing head of the EMCM.

These limits are shown opposite in fig 5.5 and show a maximum frequency range of 0.44Hz for typical maximum velocities of 40cms\(^{-1}\).

The results of the shear stress estimates are presented in the following section.
5.2.2 Bed-stress estimates

The vertical rotation and offset calibration procedure for the mean burst EMCM data was described in section 4.2.2. The same corrections were performed on the individual burst EMCM velocity time series prior to analysis. Each port and starboard horizontal velocity component was then rotated so that $\overline{U} =$ mean stream velocity and $\overline{V} = \overline{W} = 0$. For the Reynolds stress method a linear trend was then removed from each time series prior to the stress calculation.

5.2.2.1 Velocity profile method

Unfortunately the logarithmic profile method did not estimate sensible values for $u_\ast$. Regression analysis of rotor pairs (section 4.2.1) showed no detectable shear between rotor 4 (81cm) and rotor 3 (63cm). In fact the general trend in fig 4.18 is for the lower current meter to record flow speeds a few cms$^{-1}$ higher than the current meter 18cm above. Even at times of peak flow, the two rotors record comparable flow speeds of the order of 45cms$^{-1}$ (for the unaveraged data recorded at 1 minute intervals) and 40cms$^{-1}$ (for the 10 minute averaged data). A large shear is then observed between rotors 3 & 4 and 3 & 1 (27cm) where the flow speed is reduced by 50%. It follows from Eqn. 5.3 that:

$$[u_2 - u_1] = \frac{u_\ast}{\kappa} (\ln z_2 - \ln z_1)$$  \hspace{1cm} [5.20]

and

$$z_0 = \exp\left(\frac{u_2 \ln z_1 - u_1 \ln z_2}{u_2 - u_1}\right)$$  \hspace{1cm} [5.21]

and so $u_\ast$ is estimated as $\sim 9.4$cms$^{-1}$ and 7.3cms$^{-1}$ when comparing typical peak values between rotors 3 & 1 and rotors 4 & 1 respectively. As will be demonstrated in 5.2.2.2 these values are very high compared to the Reynolds stress and IDM calculated values. For a typical roughness length, $z_0$, equal to 0.5cm for a sand/gravel bed (Heathershaw, 1981) and for peak flows speeds of $u_{n3}\sim 40$cms$^{-1}$ at 63cm above the bed, one would expect (using Eqn. 5.3) a $u_\ast \sim 3.3$cms$^{-1}$. The corresponding flow speeds at 27cm and 81 cm above the bed would be $u_{n1} = 33$cms$^{-1}$.
and \( u_{m4} = 42\text{cms}^{-1} \) respectively. The logarithmic profile method was clearly inadequate for describing the bed stresses at the STABLE site.

In view of the success that Soulsby & Humphery (1989) and Williams (1995) had in using the profile method from the rotors on STABLE and STABLE II respectively, it remains uncertain as to why the rotors from La Chapelle Bank did not reproduce a characteristic logarithmic velocity profile. Flow shielding from the main body of the rig (which was located approximately 1.2 mab) would seem a plausible explanation for the observed results, but this has not been observed during other STABLE deployments. Variations in the stall speeds of each rotor have previously been discussed in section 4.2.2 and could also contribute to these discrepancies. Rotor calibrations have been tested and proven in the field and so the inertial response of the rotors at higher flow speeds would seem satisfactory (unless the internal mechanisms of the rotor were blocked due to sediment entrainment for example). If the boundary layer vertical velocity structure is as observed in the results presented from the rotors, the three methods used for estimating bed stresses would be invalid.

If we assume that none of the rotors under-estimated the flow speed, i.e. if we assume that the calibration coefficients (which are the same for each rotor) are correct, then the true flow speeds must be at least as large as those measured. Rotor 3 is shown to measure the maximum flow speed in an assumed logarithmic layer when compared to the inferred flow speeds at the other two rotor heights (using Eqn. 5.3 and \( z_0=0.5\text{cm} \)) and so is assumed to be the most accurate.

5.2.2.2 Reynolds stress and IDM

Estimates of the bed stresses using the Reynolds stress method (Eqn. 5.6) and the IDM (Eqn. 5.17) are shown in fig 5.6. Three shear stress estimates have been calculated using the IDM from the spectra of the vertical and horizontal EMCM burst time series.
Fig 5.6 Shear stress estimates using the inertial dissipation method and the Reynolds stress method on the EMCM burst time series.
The two methods are independent since they are based on two different theoretical approaches. There is reasonable agreement between the two, with the Reynolds stress method generally giving smaller values. The IDM time series plots are in good agreement with each other as expected. The IDM results do however periodically show rapid decreases in measured stress, \( \tau_{IDM} \), which are not consistent with the Reynolds stress, \( \tau_{Rey} \), estimates. The agreement between the two methods improves with increased stress values towards the end of the deployment. The increases in the \( \tau_{IDM} \) and \( \tau_{Rey} \) estimates are consistent with fig 4.33 in section 4.2.4, which show a correlation between spring tides, maximum down slope tidal streaming and maximum current variance. The significance of the higher measured stresses is discussed later in this section but on initial inspection the results appear reasonable.

Indeed, Soulsby & Humphery's (1989) comparison of the profile method, Reynolds stress method and turbulent kinetic energy (TKE)* method also found that the largest discrepancies between estimates occurred at time of weakest stress. In the present study there is up to an order of magnitude difference between the two shear stress methods. Future reference to the \( \tau_{IDM} \) results will refer only to those taken from the spectra of the longitudinal fluctuations which are transverse to the mean flow (i.e. the blue line in fig 5.6).

A requirement in the theory of both shear stress methods is that the current measurements are made within the constant stress layer of the boundary layer. Eqn. 5.4 is used to calculate the thickness of the boundary layer, \( \delta \), using the results of \( \tau_0 \) in fig 5.6 and Eqn. 5.1. To recap, for steady near-bed flows, the approximate thickness of the boundary layer can be estimated from turbulence data collected from within the neutrally stratified (i.e. mixed) bottom layer using the equation;

\[
\delta = 0.4 \frac{u^*}{f} \tag{5.4}
\]

where \( f = 1.07 \times 10^{-4} \text{ sec}^{-1} \). For typical peak shear stress values of approximately 1Nm\(^{-2}\) (for Julian days 185-189);

---

* The turbulent kinetic energy method is closely related to the IDM since both are spectral techniques. See Soulsby (1989) for further details.
\[ \tau_0 = \rho u_*^2 = 1 \text{Nm}^{-2} \Rightarrow u_* = \sqrt{\frac{1}{1027.2}} = 3.1 \text{cms}^{-1} \]

for \( \rho = 1027.2 \text{kgm}^{-3} \). This value of \( u_* \) is consistent with the peak flow value inferred from rotor 3 in section 5.2.2.1. It follows that;

\[ \delta = \frac{0.4 \times 3.1 \times 10^{-2}}{1.07 \times 10^{-4}} = 116m \]

assuming no stratification in the bottom 1m. The logarithmic layer height is approximately 10-15% of this value (e.g. Tennekes, 1973) and the constant stress layer thickness, \( z_c \), is approximately half this value again (Huntley, 1988) so that;

\[ z_c = \frac{0.03 u_*}{f} = 8.7m \quad 5.22 \]

which is well above the height of the EMCM's. When the above calculations are repeated for typical minimum values of shear stress, \( \tau \sim 0.1 \text{Nm}^{-2} \), \( u_* = 1 \text{cms}^{-1} \), \( \delta = 37m \) and \( z_c = 2.8m \), which is still above the height of the EMCM's.

Roughness values, \( z_0 \), are shown to vary over several orders of magnitude using the mean logged EMCM flow speeds and estimated \( \tau_{\text{IDM}} \) and \( \tau_{\text{Rey}} \) values. Mean values of \( z_0 = 5.3cm \) and 4.2cm are found using the two methods respectively which are clearly over estimates when compared to typical \( z_0 \) values (0.5cm for sand/gravel beds) (Heathershaw, 1981). To test whether one of the two shear stress methods gives more accurate results than the other an assumed \( z_0 \) of 0.5cm is used along with the measured \( \tau_{\text{IDM}} \) and \( \tau_{\text{Rey}} \) values to re-calculate the flow speed at 39.3cm above the bed and to then compare these results to the mean EMCM flow speed time series.

Two key results of this test are shown in fig 5.7. Firstly, there is good agreement between the EMCM flow speed and the predicted flow speed using the shear stress estimates from the IDM method during Julian days (JD) 185-189 (indicated by the dotted lines). The flow speeds predicted using the shear stress estimates from the Reynolds stress method are under estimating the flow speed during this period by approximately 10-15 cms\(^{-1}\). Secondly, prior to JD 185 and after JD 189, both methods over estimate the EMCM flow speed by up to 30 cms\(^{-1}\) during periods of
high measured values of shear stress (as shown in fig 5.6). Since the sea-bed photographs taken at the STABLE site showed no evidence that $z_0$ should vary throughout the deployment, the only other plausible explanation is that the EMCM’s failed to some degree during the observed periods of high variance (refer to fig 4.32, section 4.2.4). The shear stress measurements have therefore provided a very useful independant test of the accuracy of the EMCM burst logged data. The mean flow is unaffected by this increased noise in the EMCM bursts. To summarise, during JD 185-189 the shear stress estimates calculated using the IDM compare favourably with the observed EMCM flow speed. During this period, the mean roughness length $z_0$ is 2.9cm. Calculations of $u_*$ using Eqn. 5.3 are relatively insensitive to $z_0$ and so using a $z_0$ value equal to 0.5cm still provides a good fit to the measured flow speed.

The results give added confidence to the mean calibrated EMCM data. This section confirms that to a first approximation, the mean EMCM flow speed, $\bar{U}$, can be used in Eqn. 5.3, with a $z_0$ value of 0.5cm, to estimate shear stress values of the correct
order of magnitude. Since this section has raised concerns over the accuracy of the
turbulent EMCM measurements during periods of high variance / high shear
stresses, and therefore reduced the time period of reasonable shear stress estimates to
4 days (JD 185-189), bedload transport rates (see following section) will be
estimated from $\overline{U}$ and $z_0$ for the complete EMCM record.

5.2.3 Bedload transport

Several bedload transport formulae exist which can be used to estimate bed load
transport of sand/gravel beds. The various formulae assume that the vertical velocity
distribution in the bottom boundary layer is logarithmic. Vertical stratification
effects due to suspended sediment will make this logarithmic profile invalid since $\kappa$
under these circumstances will no longer equal 0.4 (e.g. Adams & Weatherly, 1981).
At the La Chapelle Bank site the sediment will primarily move as bed load or saltate
along the bed at times of peak flows. Previous sediment transport studies have been
conducted on La Chapelle Bank just shelfward of the shelf break in approximately
160m water depth (Heathershaw & Codd, 1985, 1986) and the effect of the internal
tide on cross-shelf sediment transport rates has also been investigated using a two-
layer internal tide model by Heathershaw et al. (1987). The purpose of this study is
to extend those results out from the shelf break (mooring 123) and onto the upper
slope (mooring 124) and critical slope (STABLE) regions.

5.2.3.1 Critical velocity exceedance curves

Evidence of tidal winnowing of the finer sediments at the La Chapelle Bank shelf
break / slope can be observed from comparisons of the threshold velocity for
sediment movement, for a specific sediment size, and the observed near-bed
currents. The threshold for sediment movement can be calculated using Yalin's
(1972, p82) modified shields curve. The threshold velocity $U_{CR}$ is given in terms of
a critical friction velocity $u_{*CR}$ as;
where $Y_{CR}$ is determined graphically, $v$ is the kinematic viscosity of the fluid, $\rho_s$ is the sediment density, $\rho$ is the water density and $D$ is the grain size diameter. These estimates are largely for granular material and the threshold for carbonate material is not well known. Heathershaw (1987) remarks that the CaCO$_3$ material is likely to have a lower critical threshold than granular material and so bedload transport estimates in the La Chapelle Bank region may be significantly under-estimated.

The mean grain size, 16th and 84th percentile grain sizes for the upper slope region (refer to section 4.2.2.2) are shown below in table 6.1. The larger sediment is made up of mainly carbonate material, with some random particle sizes in the range between 1000-3500$\mu$m. These results are consistent with Heathershaw & Codd (1986) at approximately 160m depth.

### Table 6.1 Grain size statistics for the La Chapelle Bank slope sediment samples.

<table>
<thead>
<tr>
<th>Grab sample</th>
<th>84% Coarser than ($\mu$m)</th>
<th>Mean ($\mu$m)</th>
<th>16% Coarser than ($\mu$m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shipek 1</td>
<td>233</td>
<td>870</td>
<td>3249</td>
</tr>
<tr>
<td>Shipek 2</td>
<td>353</td>
<td>860</td>
<td>2144</td>
</tr>
<tr>
<td>Shipek 3</td>
<td>250</td>
<td>595</td>
<td>1414</td>
</tr>
<tr>
<td>Mean</td>
<td>300</td>
<td>775</td>
<td>2270</td>
</tr>
</tbody>
</table>

A histogram of flow speed measured from rotor 3 is shown in fig 5.8 for the unaveraged continuously logged data set. This information can also be plotted as a flow speed exceedance curve as is shown in fig 5.9. The construction of these flow exceedance plots for each near-bed current meter station are now described.

For direct comparison with Heathershaw & Codd (1985, 1986) the exceedance curves for Rotor 3 and the EMCM's at the STABLE site, CM8511 at mooring 124 and CM7643 at mooring 123 have been plotted for the velocity 2m above the bed. These have been calculated using Eqn. 5.3 by assuming a roughness length, $z_0$ equal to 0.5cm and calculating $u_*$ from the velocity time series for each current meter.
Fig 5.8 A histogram of the current speed recorded by rotor 3, 63cm above the bed.

Fig 5.9 Exceedance curves for the flow speeds measured 2m above the bed ($U_{200}$) at the STABLE site (EMCM's and Rotor 3 at 385m), mooring 124 (305m) and mooring 123 (200m). Also shown are the critical threshold velocities $U_{200CR}$ for sediment movement for grain sizes of 329, 506 and 800µm which is consistent with Heathershaw & Codd (1985).
sensor at height \( z(m) \) above the bed and re-evaluating Eqn. 5.3 for \( U_{200} \). The EMCM flow speed time series for \( z_0=0.5\text{cm} \) have reasonably estimated the bed stress at the STABLE site (section 5.2.2.2) and further on-shelf (Heathershaw, 1985). Out of the three Rotors which successfully logged data, Rotor 3 showed the best agreement with the predicted logarithmic vertical flow speed profile (5.2.2.1) and is shown plotted in fig 5.9. To validate the use of this method for the current meters 8m above the bed on moorings 124 and 123 the assumption is made that the current meters are within the logarithmic part of the boundary layer, which we recall is 10-15% of the total boundary layer height. The estimated height of the logarithmic layer on the upper slope region is shown to be approximately 12m at times of peak flow (section 5.2.2.2) but is probably below 10m during other periods. Further on-shelf (~160m), Heathershaw (1985) fitted a tidally averaged logarithmic velocity profile to the measured currents at 2, 5 and 10m above the bed and estimated \( u_* \) with a correlation coefficient of \( R = 0.9998 \). Since tidal currents are larger at the shelf break (200m) than those observed by Heathershaw (1985) it is reasonable to assume a logarithmic profile to 8mab, at least at peak flow rates when bedload transport is significant (see next section). If the flow is periodically outside the logarithmic layer then the \( U_{200} \) values will underestimate the currents 2mab. Finally because of the shortness of the EMCM record, the EMCM exceedance curve is not smooth and is represented in fig 5.9 by a polynomial fit with an \( R^2 \) value of 0.998. The results of this study are now discussed.

In fig 5.9 the exceedance curves are compared to the threshold velocity (\( U_{CR200} \)) for three grain sizes. The difference between the EMCM's and rotor 3 is of some concern but previous results suggest that the EMCM's are the more reliable source of information. Even with this discrepancy the threshold velocities for grain sizes of 329, 506 and 800\( \mu \text{m} \) still show a distinct trend. The threshold of movement for \( D=506\mu\text{m} \) at the shelf break (200m) is exceeded for 63% of the time, reduces to 30% of the time at 305m but then increases to 42-54% (Rotor 3-EMCM) of the time at 385m depth on the critical slope region. For an overall mean grain size of approximately 800\( \mu \text{m} \) for the sediment samples taken on the upper slope region, the exceedance ratio reduces to 42%, 11% and 22-37% respectively. These values for
D=800μm compare with values of 10% at station LCB4 (furthest from the shelf-edge) and 20% at station LCB2 (nearest to the shelf-edge) by Heathershaw & Codd (1986). The results suggest that the critical threshold velocity is maximum at the shelf break and decreases shelfward with increasing distance on-shelf. Immediately off-shelf the threshold decreases, but then increases again at the critical depth (385m). These results are consistent with a larger mean grain size in the vicinity of the STABLE deployment site compared to those reported just on-shelf by Heathershaw & Codd (1986).

5.2.3.2 Bedload transport equation

There are several bed-load transport formulae which are used to predict sediment transport rates and directions for a variety of physical environments. Heathershaw (1981) compares five predicted transport equations to estimates using radioactive tracer techniques. Williams et al. (1989) compares the same five transport equations with estimates using in situ acoustic techniques. The previous bedload transport studies near the La Chapelle Bank shelf edge (Heathershaw, 1985, Heathershaw & Codd, 1985 & 1986 and Heathershaw et al., 1987) have used Hardisty’s (1983) modified excess stress formulation of Bagnold’s sediment transport equation, defined as:

\[ q_{sb} = k_1 \left( U_{100}^2 - U_{100CR}^2 \right) U_{100} \quad \text{(gm cm}^{-1} \text{s}^{-1}) \]

where \( q_{sb} \) is the quantity of sediment transported as bedload, \( U_{100} \) is the velocity \( 1 \text{mab}, U_{100CR} \) is the threshold velocity for sediment movement (defined by Eqn. 5.23) and \( k_1 \) is a dimensional coefficient which depends on grain size, defined as (e.g. Heathershaw et al., 1987);

\[ k_1 = 1.77 \times 10^{-6} D^{-0.5916} \quad \text{(gm cm}^{-4} \text{s}^{-2}) \]

where \( D \) is the grain size in mm. Hardisty’s equation is suitable for \( D_{50} \) grain sizes of 180-1450 μm, for plane beds, fully developed steady flows and a large flow depth \( / D_{50} \) ratio and is therefore generally applicable in the La Chapelle Bank environment. The formula is also shown by Heathershaw (1981) to be the least sensitive to changes in roughness length, \( z_0 \) and particle size, \( D \). The estimated
transport rates using Eqn. 5.24 were also in moderate agreement with the estimates using the radioactive tracer technique described by Heathershaw (1981) for fine sand to gravel sized material (Heathershaw & Hammond, 1978) and with the acoustic measurements made by Williams et al. (1989) for gravel material.

The $U_{100}$ values were obtained from the measured currents at 39.3cm and 63cm from the EMCM’s and rotor 3 respectively at the STABLE site, from 8m at mooring 124 and 8m at mooring 123 using an assumed logarithmic profile and a roughness length for the sand/gravel bed of 0.5cm. The validity of the assumptions made for this transport prediction, particularly at 8mab, were discussed in the previous section. This procedure is consistent with calculations by Heathershaw & Codd (1985) who calculated $U_{100}$ values from measurements 2mab and with estimates by Heathershaw et al. (1987) with their two layer internal tide model. In this instance, $U_{100}$ values were obtained by integrating the logarithmic velocity profile over the lower layer, which when at rest is 120m above the bed (section 2.2.4.3). Direct comparisons with previous results can therefore be made.

5.2.3.3 Bedload transport predictions

To be consistent with previous estimates shelfward of the shelf break, values of $q_{sb}$ were calculated for an overall mean grain size of 506 $\mu$m. The corresponding critical friction velocity $u_{CR} = 1.5$ $\text{cm s}^{-1}$ and $U_{100CR} = 19.9$ $\text{cm s}^{-1}$. The results are shown plotted for JD’s 184-192.5 in fig 5.10 and are compared to the in situ acoustic measurements at the STABLE site and the transmissometer results at mooring 124.

Fig 5.10a reveals that the $M_2$ tidal currents (barotropic + baroclinic) are strong enough to mobilise and transport the sediment at the STABLE site during both neap and spring tides. Peaks in relative ABS concentration coincide with maximum down slope tidal currents (see section 5.1.2, fig 5.2). The predicted $q_{sb}$ rates for the EMCM’s and rotor 3 are in general agreement with the ABS results and suggest that the 506$\mu$m grain sizes will become more mobile during down slope tidal streaming.
Fig 5.10 A comparison of the predicted bedload transport rates, $q_{ob}$, at (a) the STABLE site using the EMCM's and rotor 3and compared with the in situ acoustic measurements, (b) mooring 124 and compared to the turbidity measurements and (c) mooring 123. $q_{ob}$ is calculated from $U_{100}$ values and $z_0 = 0.5\text{cm}$ and $U_{100CR} = 19.8\text{cm/s}$ for a mean grain size of 506$\mu$m. These values are consistent with Heathershaw & Codd (1985, 1986) and Heathershaw et al (1987).
than during on-shelf tidal streaming. Material transported down slope does not therefore return up-slope during the on shelf phase of the M2 tide. The mean 84% (coarser than) grain size is ~300μm (i.e. sand sized material) which has a $U_{100}$ value of 18.5cms$^{-1}$. These flow speeds are exceeded for 60-70% of the time at the STABLE site (fig 5.9) and occur during both down-slope and up-slope tidal streaming. Since the ABS is calibrated for sand sized material in suspension it remains unclear as to why a double peak is not observed in the time series plot which one associates with simple tidal advection. Two possible explanations are:

(i) The ABS threshold for the detection of suspended sediment is only exceeded for the relatively higher ssc's during down slope tidal streaming.

(ii) The sediment is composed of fine material which is advected past the sensor from further up-slope during the down-slope phase of the tide and which does not return during the up-slope phase.

Either explanation shows a net transport down slope with peak $q_{sb}$ rates of 0.2 gm cms$^{-1}$ at spring tide.

The results further up-slope away from the critical region at mooring 124 show that the sediment is only mobile during the down slope phase of the M2 spring tide (peaks correlate with peaks in the ABS time series). Peak $q_{sb}$ rates also correlate with low transmission (increased turbidity) and suggests net bedload and turbidity transport down slope. It should be noted that by assuming a logarithmic velocity profile from 8mab away from peak flows, the $U_{100}$ values could be underestimated and hence $q_{sb}$ rates could also be underestimated. This would seem to be the case in figs 5.10 and 5.10 where $q_{sb}$ rates are largely undetectable during neap tides and then very peaked during spring tides. $q_{sb}$ rates can therefore only be inferred during peak flows in figs 5.10. The relative $q_{sb}$ rates in fig 5.10c confirm this assumption since peak $q_{sb}$ rates are larger than at the STABLE site at spring tide (0.4 gm cms$^{-1}$) but much smaller at neap tide. In both instances the near-bed tidal currents are known to be maximum at the shelf break (200m). Fig 5.11 repeats fig 5.10 for all of the data collected during the deployment period.
Fig 5.11 A comparison of (a) the in situ ABS measurements at the STABLE site with (b) the in situ turbidity measurements and predicted bedload transport rate, $q_{bol}$, at mooring 124 and (c) $q_{bol}$ at mooring 123. See fig 5.9 for further details.
The main conclusions to be drawn from figs 5.10 are;

(i) Sediments on the bed at the shelf break and upper slopes of La Chapelle Bank are mobile.

(ii) At peak flows, the $q_{sb}$ rates are largest at the shelf break and the critical slope region.

(iii) The critical slope (385m) is a significant export region for the oceanward transport of sediment.

(iv) There is direct evidence of bed-load parting between the shelf break and the upper slope region during summer stratification.

(v) Maximum transports are observed at the shelf break and are shelfward.

5.3 Discussion

A comparison of transport rates from the present shelf/slope study with that by Heathershaw & Codd (1985) are summarised in table 5.2. Average values for all $q_{sb}$ estimates were calculated with a mean grain size of 506μm. It should be noted that the measurements were not taken at the same time or in exactly the same location along the La Chapelle Bank shelf edge. They are similar by virtue that both data sets were collected during summer stratified conditions.

Table 5.2 Calculations of averaged sediment transport rates ($q_{sb}$), transport directions ($\gamma$), residual currents ($u$) and directions ($\theta$) at various locations across on La Chapelle Bank. Stations LCB2-4 are taken from Heathershaw & Codd (1985). Residual currents for Moorings 123-124 and the STABLE site are described in sections 4.1.2 and 4.2.3 respectively. Sediment transport rates are calculated for $D = 506\mu m$ and $z_0 = 0.5cm$.

<table>
<thead>
<tr>
<th>Location</th>
<th>Distance from 200m contour (km)</th>
<th>Average transport rate $q_{sb}$ (gm cm$^{-3}$ s$^{-1}$)</th>
<th>Transport direction $\gamma$ (°T)</th>
<th>Residual current $u$ (cms$^{-1}$)</th>
<th>$\theta$ (°T)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LCB4</td>
<td>17.5</td>
<td>1.19x10$^{-3}$</td>
<td>049</td>
<td>0.98</td>
<td>099</td>
</tr>
<tr>
<td>LCB3</td>
<td>8.0</td>
<td>2.88x10$^{-3}$</td>
<td>175</td>
<td>0.76</td>
<td>153</td>
</tr>
<tr>
<td>LCB2</td>
<td>2.5</td>
<td>7.75x10$^{-3}$</td>
<td>106</td>
<td>4.68</td>
<td>150</td>
</tr>
<tr>
<td>Mooring 123</td>
<td>0.0</td>
<td>5.02x10$^{-2}$</td>
<td>012</td>
<td>4.30</td>
<td>331</td>
</tr>
<tr>
<td>Mooring 124</td>
<td>-4.2</td>
<td>7.80x10$^{-3}$</td>
<td>226</td>
<td>6.70</td>
<td>240</td>
</tr>
<tr>
<td>STABLE - RTR3</td>
<td>-6.0</td>
<td>1.90x10$^{-2}$</td>
<td>241</td>
<td>11.2</td>
<td>261</td>
</tr>
<tr>
<td>- EMCM's</td>
<td></td>
<td>3.76x10$^{-2}$</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
The residual flows on the outer shelf shown in the above table are smaller than those at the shelf break region and in an opposing NW to SE equatorward direction. This flow is not uncommon on the outer shelf of the Celtic Sea region. Pingree & LeCann (1989) also observed residual shelf currents which moved in a south-easterly direction. The residual flows are not always consistent with the prevailing wind direction and Pingree & LeCann (1989) briefly suggested that this may result from enhanced shelfward cross-slope currents at the shelf break.

The residual currents measured at mooring 123 are orientated shelfward. The vertically averaged residual currents at the shelf break are however generally poleward but do show considerable vertical and temporal changes near to the bed at M₂ tidal periods. Barotropic and baroclinic M₂ tides are the dominant source of cross slope flux of energy in this region. Indeed the M₂ tide appears to be responsible both for the strong down-slope residual flows on the upper slopes and the on-shelf component of flow at the shelf break (section 4.1.6). These results show that at the shelf break and on the upper slopes the transport rates, q_{sb}, are in the direction of maximum tidal streaming, i.e. cross-slope (~020°T). Between the upper slope and shelf break the cross-slope transport becomes 180° out of phase and implies bedload parting of sediment. This is in agreement with the predictions of the two layer internal tide model. (Heathershaw et al., 1987, refer to section 2.2.4.3).

The predicted cross shelf sediment transport rates in table 5.2 are compared with the internal tide model (Heathershaw et al., 1987) and are shown in fig 5.12. Due to temporal and spatial variations in these calculations the magnitude of the model predictions should be treated with caution in comparison to the real observations. However significant trends can still be noted. The numerical model results show that the M₂ internal tide can result in a region of divergence in the cross-shelf transport rates q_{sb}. Sediment is transported down-slope (-ve q_{sb}) at the depth of internal tide generation and on shelf (+ve q_{sb}) immediately landward of it, thus giving rise to bedload parting. The magnitude of the on-shelf transport was insensitive to shelf break topography contrary to down-slope transport rates. This is depicted in fig 5.12 for idealised shelf break curvatures of R=50km (the most realistic case) and R=10km. Further down slope the results show continued
Fig 5.12 (a) Predicted averaged transport rates \( (q_w) \) at different locations (b) across the La Chapelle Bank shelf break region (values are resolved from table 5.2). The STABLE deployment site, mooring 124 and 123 are from the present study. The predicted transport rates from the two-layered internal tide model by Heathershaw et al (1987) are also shown at the location where on-shelf and off-shelf predicted transports are maximum. The results from two idealised shelf break topographies are shown with curvatures of \( R=50 \text{km} \) (solid line) and \( R=10 \text{km} \) (dotted line), a layer depth of \( h_l = 60 \text{m} \) and \( \delta p = 5 \times 10^4 \text{ g cm}^{-3} \). The predicted cross-shelf transport rates at stations LCB2, LCB3 & LCB4 (from Heathershaw & Codd, 1985) are also shown. See text for further details.
down-slope transport of sediment with an increase at the critical depth on the upper slope at 385m. At the shelf break itself, observations show large on-shelf transport compared to smaller transports on the outer shelf where near-bed currents are weaker and orientated along slope.

Transport rates have been recalculated for mooring 123, 124 and the STABLE site with a mean grain size of 800μm. This grain size represents the overall mean grain diameter on the upper slope region. The transport rates shown in table 5.3 are comparable with those of Heathershaw et al. (1987) and also reflect the ‘true’ nature of the bed material on the slope.

Table 5.3 Sediment transport rates calculated for D = 800μm and z₀ = 0.5cm. Refer to table 5.2 for further details.

<table>
<thead>
<tr>
<th>Location</th>
<th>Distance from 200m contour (km)</th>
<th>Average transport rate q₀ (gm cm⁻³ s⁻¹)</th>
<th>Residual current (°T)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mooring 123</td>
<td>0.0</td>
<td>3.25×10²</td>
<td>012</td>
</tr>
<tr>
<td>Mooring 124</td>
<td>-4.2</td>
<td>2.30×10³</td>
<td>217</td>
</tr>
<tr>
<td>STABLE - RTR3</td>
<td>-6.0</td>
<td>8.90×10³</td>
<td>238</td>
</tr>
<tr>
<td>- EMCM's</td>
<td>-6.0</td>
<td>2.26×10²</td>
<td>261</td>
</tr>
</tbody>
</table>

The general trends summarised in table 5.3 are consistent with those found in table 5.2, but also suggest that larger bed material can be mobilised periodically.

Fig 5.13 portrays a schematic of present day sediment transport paths from the western English Channel to the ocean margins of the Celtic shelf and incorporates data from Stride, (1963). It is evident, even from this early work, that the Southwest Approaches to the English Channel were considered major export regions for sediment;

“Sand inferred to be issuing from the western end of the English Channel is heading towards the edge of the continental shelf. It seems likely that its subsequent path to the floor of the deep ocean will be determined by the numerous submarine canyons in the continental slope south of 49°N” (Stride, 1963).
Fig 5.13 A schematic showing present day sediment transport paths from the western end of the English Channel towards the deep ocean (from Stride, 1963). The patterned zone represents regions of deposition and the arrows represent sediment transport paths of largely sand sized material. The arrows at the shelf break region of the Celtic Sea are described further in the text and are appropriate for summer stratified conditions.

This diagram suggests that during seasonal stratification the oceanward transport of sediment is blocked, resulting in a build up of sediment just shelfward of the shelf break. It is hypothesised that during winter periods, when the water column is mixed at the shelf break, that this transport continues oceanward. The accumulation of material at the shelf break may imply storm events are acting as an important mechanism in the down-slope transport of large amounts of material i.e. within turbidity currents.

Since the field experiment did not provide any long-slope resolution, the relative contribution of submarine canyons in the direct transport of sediment down-slope can not be ascertained.
To conclude, maximum measured shear stresses of 1Nm$^{-2}$ ($u_* = 3.1\text{cms}^{-1}$) on the critical slope infer that the bed is mobile on the upper slopes and that net sediment transport is down-slope. This is consistent with the ABS mean burst data which suggests that finer sandy material is winnowed away from the bed during peak flows with net transport also down-slope. However, the photographs do not indicate that the bed is particularly mobile. The apparent consolidated nature of the bed suggests that larger stresses are required, than those predicted, to mobilise the bed. The coarse material shown in plate 5.2 reveals little evidence of any movement and therefore the critical threshold velocity was never exceeded for the largest material. The 16% 'coarser than' sediment fraction (refer to table 6.1) had a mean grain size of approximately 2000\(\mu\text{m}\), which for a roughness length of $z_0 \sim 0.5\text{cm}$ has typical threshold velocities exceeding 50\text{cms}^{-1}. Flow speeds of this magnitude were not observed Imab anytime during the deployment of STABLE.
As part of this project, further deployments of STABLE (the Mk II version, here on referred to as STABLE II) had originally been planned on the upper slopes of the Hebrides shelf-edge along 57°N, as part of the 3 year UK NERC funded Land Ocean Interaction Study - Shelf Edge Study (LOIS-SES). The subsequent 2 year postponement of the LOIS-SES programme (to begin 1995) was an initial setback to this NERC funded project.

A second, opportunistic deployment of 'pop-up' STABLE II (see section 6.3 for a short description of improvements since the original STABLE) on the Goban Spur continental slope in January 1994 was a welcomed further study of the near-bed turbulent and mean current-bed interactions and associated sediment transport processes. Funding for this deployment was provided through the European Community Ocean Margin EXchange (OMEX) programme. Consequently, the results of this near-bed study have formed an inaugural part of a much larger multi-disciplinary European Community study of shelf-edge exchange processes on the
Goban Spur. Chapter 6 presents the preliminary results from the 1994 Goban Spur deployment of STABLE II.

6.1 Introduction

The EC funded OMEX project is a multi-disciplinary 3 year programme (1992-1995) to study the processes which influence the shelf-ocean transport of material via the continental margins. The Goban Spur region is part of the northern Celtic shelf and extends from approximately 48-50°N (refer to fig 2.1). The upper continental slopes are comparatively gentler than the Armorican and southern Celtic shelf slopes further to the south, and the Malin and Hebrides shelf slopes further to the north. Consequently, the Goban Spur region is less insulating to cross-slope flow (Pingree, Sinha & Griffiths, 1996) and is therefore a potential export region for the flux of material from the land to the ocean.

The main objective of the extensive OMEX sea time programme was to implement a comprehensive sampling strategy to monitor the biogeochemical fluxes of material along an approximate east-west transect between 11°W and 15°W at approximately 49°N. This transect is shown by the dotted line in fig 6.1 and crosses the 180m contour near the shelf break and the 4800m contour in the deep ocean. As a component of this OMEX programme, sediment trap and current meter moorings were deployed along the OMEX transect and continuously monitored. In June 1993, the NIOZ bottom lander was deployed along this transect in position 49°11.31'N 12°44.00'W at 1296m water depth (Cruise report, Poseidon 200, leg7, 1993) and collected data on sediment transport and near-bed currents for approximately one year.

The Charles Darwin Cruise 84 (Statham, 1994 [CD84 cruise report]) biogeochemical OMEX cruise in the Goban Spur and La Chapelle Bank regions of the Celtic Sea provided the opportunity to deploy the Proudman Oceanographic Laboratory’s (POL’s) Sediment Transport And Boundary Layer Equipment on the northwest European continental slope for a second time. The time constraints of the CD84
Fig 6.1  Bathymetry (in meters) of the Goban Spur continental slope region. Also shown is the OMEX sampling transect (----), the STABLE II deployment site, CTD stations and R.D. Pingree's OMEX current meter mooring site (OM154). See text for further details.
scientific programme, combined with the desire to deploy STABLE II for as long as possible (by deploying early on in the cruise and recovering as late as possible) restricted site selection to the Goban Spur region. To maximise the use of supplementary OMEX data for this near-bed study, and to establish links with other OMEX work groups, the site was further restricted to the main OMEX transect. Site selection is discussed further in section 6.4.

6.2 Objectives

This second European continental slope deployment of STABLE (II) on the Goban Spur continental slope is essentially a continuation of the work undertaken during the first deployment in July 1990 on the upper slopes of La Chapelle Bank. That is, to study the response of the benthic boundary layer to shelf edge dynamics, in particular, to study the near-bed response to local near-bed enhancements of baroclinic tidal energy and subsequent sediment erosion. Such hydrodynamic conditions are considered favourable to the formation of BNL's and INL's (see section 6.4) in the Goban Spur and Porcupine Bank sloping regions. The nature and stability of the poleward flowing slope current is also of primary interest since the main transport mechanism for sediment is most likely through suspension (refer to section 6.5.4). The advective flux of near-bed material is therefore likely to be in the direction of the residual current.

Finally, the opportunity to study regional shelf slope variations in sediment transport mechanisms at two contrasting sites along the Celtic Sea continental shelf slope has added increased stimulus to the project. A third continental slope/shelf-edge deployment of STABLE II (a joint NERC LOIS-SES funded collaborative project between POL and the University College of North Wales, Bangor) on the Hebrides Shelf margin in September 1995 has increased the geographical coverage of observations still further.
6.3 STABLE II: developments.

Humphery & Moores (1994a) discuss in detail, the enhanced performance, greater accuracy and endurance, and much improved recovery rate that STABLE II can offer over its predecessor. The reader should refer to Humphery & Moores (1994a) for a detailed account of the developments in electronics, data sampling and storage capabilities. Table 6.1 describes the instrumentation fitted to the STABLE II rig for the Goban Spur deployment. A schematic of STABLE II (fig 6.2, taken from Humphery & Moores, 1994a) identifies the individual sensors. Developments aside, the primary consideration for POL in the future development of STABLE II is the capability and space to accommodate extra sampling equipment at the request of the hydrodynamicists and sedimentologists involved in its use.

The main additions to the rig are the two extra pairs of EMCM sensors, all three of which sample at double the frequency (8Hz) for double the time (20 minutes), twice as often (hourly) compared to the old rig (refer back to table 3.1). The vertical array of three pairs of EMCM’s provided a vertical profile of measurements of turbulent currents in the bottom 1m which can be analysed independently to study current shear and associated mixing processes.

Two ABS sensors operating at 1.0MHz and 2.5MHZ, also developed at POL, operate concurrently with the EMCM burst data to provide more information on suspended sediment concentration and sediment size (a single 3.0 MHz transducer on the 1990 STABLE could only accurately predict relative changes in suspended sediment concentration). Humphery & Moores (1994a) reveal that a third transducer is now being developed for STABLE II at POL by P.D.Thorne & P.J.Hardcastle to improve grain size determination. Four (albeit rather crude) settlement traps were also fitted to the main body of the rig (shown in fig 6.2) at two differing heights on the rig in an attempt to collect samples of suspended particulate matter for the calibration of the ABS system and for independent sediment analysis.
### Table 6.1 Summary of STABLE II instrumentation for the 1994 Goban Spur continental slope deployment.

<table>
<thead>
<tr>
<th>Data collected</th>
<th>Instrumentation</th>
<th>Sampling frequency</th>
<th>Height, (z), above bed (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A) Short term 'turbulent' current</td>
<td>3 pairs (arrays) of EMCM's mounted orthogonal to each other, measuring U, V and a comparable W component of the flow.</td>
<td>8Hz for 20 minutes every hour.</td>
<td>Array A - 30.0cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Array B - 60.2cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Array C - 89.9cm</td>
</tr>
<tr>
<td>(B) Tidal currents and current shear.</td>
<td>A vertical stack of 4 Savonius Aanderaa rotor (RTR) current meters, and direction vane.</td>
<td>Continuous 1 minute sampling, (number of counts per minute).</td>
<td>RTR 1 - 22.2cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>RTR 2 - 40.2cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>RTR 3 - 58.2cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>RTR 4 - 76.2cm</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Vane - 90.7cm</td>
</tr>
<tr>
<td>Tidal elevation</td>
<td>(C) Digiquartz pressure transducer.</td>
<td>Continuous 1 minute sampling</td>
<td>Approximately 195cm</td>
</tr>
<tr>
<td>Rig orientation</td>
<td>(D) Compass, pitch &amp; roll sensors</td>
<td>Continuous 1 minute sampling</td>
<td>Approximately 125cm</td>
</tr>
<tr>
<td>Temperature</td>
<td>(C) Digiquartz and (D) Quartz crystal temperature sensors</td>
<td>Continuous 1 minute sampling</td>
<td>Approximately 195cm and 125 cm</td>
</tr>
<tr>
<td>Suspended sediment</td>
<td>(E) Double frequency Acoustic Backscatter System (ABS)</td>
<td>1.0 MHz and 2.5 MHz ABS transducers synchronised with EMCM burst data. Fig 6.2 shows extra housing for a triple frequency ABS which is still under development at POL.</td>
<td>z=128cm looking vertically down in front of the EMCM's.</td>
</tr>
<tr>
<td>Settlement traps</td>
<td>4 simple settlement traps collected samples of suspended sediment throughout the deployment.</td>
<td>1.0 MHz and 2.5 MHz ABS transducers synchronised with EMCM burst data. Fig 6.2 shows extra housing for a triple frequency ABS which is still under development at POL.</td>
<td>1 at 82.5cm (shown in plate 6.1 attached to the front leg).</td>
</tr>
</tbody>
</table>

Note: The suspended sediment concentrations were below the ABS threshold. Peaks in the profiles just showed side echo's from the rig and the sea bed. Consequently the ABS results are not described further.

Letters in brackets are consistent with the labelling of sensors in fig 6.2. Quartz temperature, pitch, roll and compass all housed in the mean logger tube.
Aluminium frame provides support and protection for instruments and sensors.

Transponding release in glass sphere (one of two).

Pressure transducer (waves).

Mechanical release holding foot drop-wire (one of three).

Mean logger tube.

Burst logger and EMCM electronics tube.

Tidal-current direction vane (hidden by frame).

Savonius rotor stack (four rotors).

Three tubes housing triple-frequency ABS transducers.

Disposable ballast foot (weight 145kg) (one of three).

Syntactic foam buoyancy block (one of two).

Pressure transducer (tides).

Tube housing acoustic backscatter system.

Main battery tube.

Three tubes housing triple-frequency ABS transducers.

Plate 6.1 STABLE II on the after-deck of the RRS Charles Darwin prior to deployment on the Goban Spur continental slope during Cruise CD84 on 20th January 1994.
Two mean logging temperature sensors are also fitted to STABLE II. The digiquartz (D/Q) and crystal quartz (Qts) sensors have accuracy's of 20 mdeg C and 12 mdeg C respectively, the latter more sensitive to rapid changes in temperature (a thermal time constant of a few tens of seconds, Humphery & Moores [1994b]). As with the 1990 deployment a vertical array of 4 Savonius rotor current meters, direction vane, digiquartz pressure, pitch, roll and rig heading sensors were also fitted to the rig. The one omission from the previous deployment was the camera system which is no longer available. The camera is a significant source of information regarding the continuous monitoring of the nature of the bed, bedload and suspended transport and bedforms, the latter of which can be used to infer current strength and direction.

6.4 Site selection

On the Goban Spur and Porcupine Bank sloping regions, weaker tidal and slope currents (compared to those further to the north and south) and the fine sandy/silty/muddy nature of the bed (see sections 6.5.3 and 6.5.4) favour the transport of near-bed material in Bottom Nepheloid Layers (BNL's) and/or Internal Nepheloid Layers (INL's). This is in contrast to the bed-load or periodic re-suspended nature of transport of near-bed material on the upper slopes of La Chapelle Bank. Bottom reflection, resonance and breaking of internal waves are thought by and large, to be the mechanisms which cause these erosional features on the slopes. The latter are thought more probable in regions where baroclinic tidal motions are locally enhanced near to the bed.

Although the slopes of the Porcupine Bank are not at an angle critical for internal tide generation on the upper 1000m (Thorpe, 1987a), Dickson & McCave (1986) suggest from CTD and transmissometer observations, that local irregularities (increases) in topographic slope and/or density gradients might depict conditions favourable to the formation of the observed BNL's and INL's.
Bathymetry charts and CTD data collected during the OMEX Poseidon 200/7 Cruise (Cruise report, Poseidon 200, leg7, 1993) in June 1993 suggested that the bottom slope along the OMEX transect did not equal or exceed the semi-diurnal internal tide characteristic on the upper 1500m of the Goban Spur. Preliminary results from the OMEX Pelagia Cruise in October 1993 (OMEX Newsletter, No. 1, January 1994) did however identify the boundary layer along the OMEX transect to be a high energy environment with near bottom current velocities up to 40cms\(^{-1}\) (results were not more site specific at this time and were obtained from the bottom water sampler BIOPROBE which also registers near-bottom current velocity). BNL's and INL's had been previously observed on the Goban Spur from transmissometer profiles and echo sounder returns respectively, in a transect along 50°N during the Frederick Russell Cruise 2/87 in March 1987 (R.D. Pingree, 1987, Frederick Russell Cruise 2/87, Cruise Report, Plymouth Marine Laboratory).

Based on this information it was decided to deploy STABLE II within the permanent thermocline where one might expect local increases in the density gradient near to the bed, thus local increases in the total energy might also be expected. This could be due to an enhancement of baroclinic tidal energy (the semi-diurnal barotropic tide being the dominant signal along these slopes [refer to section 6.5.5.2]) and/or from higher frequency internal waves, which can exist in regions where high density gradients (in this case the permanent thermocline) impinge on the slope.

Post- STABLE II deployment, A. Antia (pers comm.) reported that sediment trap, current meter and transmissometer data from three OMEX sediment trap moorings positioned across the OMEX transect at depths of 670m, 1445m, and 3650m (OMEX Newsletter, No. 1, January 1994) revealed a large load of particles coming into the deep traps (ca. 1100m) laterally. These particles were presumably being 'pinched' off the bottom and transported in INL's at 800-900m water depth. This region of the slope coincides with the depth of the permanent thermocline.
6.5 Field study results

A preliminary study of the Goban Spur deployment of STABLE II has been undertaken and the results are presented in section 6.5.5. These results are preceded by a discussion of the complimentary data collected pre- and post-STABLE II deployment in sections 6.5.2-6.5.4. The following section summarises the data sources utilised.

6.5.1 Introduction

A summary of the data collected and utilised for this benthic study in January 1994 is shown in table 6.2. Results from a near by OMEX current meter mooring, Shipek grab samples of bottom sediment, sea bed photography and finally an across slope transect of CTD stations are all presented in following sections. Each sampling station is also shown in fig 6.1 apart from the Shipek grab sample and camera stations which were in the immediate vicinity of the STABLE II deployment site.

6.5.2 Cross-slope bathymetric and CTD survey

A total of 5 CTD casts were deployed across the OMEX transect (fig 6.1) to within a few meters of the sea bed from the shelf edge at 188m depth (CTD1) to the deep ocean at 4495m depth (CTD7). CTD's 1, 4, 9 & 7 were conducted at continuously monitored OMEX biogeochemical sampling stations and were a matter of course for cruise CD84. CTD3 was conducted as a preliminary investigation of the slope water near the pre-determined STABLE II deployment depth of 800-900m (section 6.4).

The results from CTD stations 3, 9 & 7 are shown in fig 6.3 and show profiles of temperature, salinity, density, attenuance and oxygen. The results from CTD3 (1003m depth) show the presence of a poorly defined nepheloid layer at
Table 6.2  A summary of data collected and utilised in this benthic study

<table>
<thead>
<tr>
<th>Data source</th>
<th>Position (Latitude)</th>
<th>Position (Longitude)</th>
<th>Water depth (m)</th>
<th>Deployment date - recovery date</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>STABLE II</td>
<td>49°23.49'N 11°40.03'W</td>
<td></td>
<td>879</td>
<td>20/01/94 - 31/01/94</td>
<td>CTD3, the Shipek grab and the bed-hopping camera deployments were all conducted pre-STABLE II deployment as a preliminary survey of the site.</td>
</tr>
<tr>
<td>Mooring 154</td>
<td>49°06.48'N 12°10.05'W</td>
<td></td>
<td>996</td>
<td>21/01/94 - 08/06/95</td>
<td>R.D. Pingree, OMEX mooring. Data from near-bed current meter on mooring 154 kindly made available for STABLE II deployment period.</td>
</tr>
<tr>
<td>Shipek Grab</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sample 1</td>
<td>49°23.25'N 11°39.22'W</td>
<td></td>
<td>860</td>
<td>20/01/94</td>
<td></td>
</tr>
<tr>
<td>Sample 2</td>
<td>49°23.20'N 11°40.90'W</td>
<td></td>
<td>890</td>
<td>20/01/94</td>
<td></td>
</tr>
<tr>
<td>Sea-bed Photography</td>
<td>49°23.28'N 11°39.24'W</td>
<td></td>
<td>860</td>
<td>20/01/94</td>
<td>Bed-hopping 35mm camera looks from 120cm above the sea-bed (assuming the camera is hanging vertically at the time the photo is taken). A total of 31 exposures taken in total.</td>
</tr>
<tr>
<td>CTD stations</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CTD 1</td>
<td>49°29.95'N 11°00.22'W</td>
<td></td>
<td>188</td>
<td>20/01/94</td>
<td>OMEX station 5'</td>
</tr>
<tr>
<td>CTD 3</td>
<td>49°22.15'N 11°46.86'W</td>
<td></td>
<td>1003</td>
<td>20/01/94</td>
<td>CTD immediately prior to STABLE II deployment.</td>
</tr>
<tr>
<td>CTD 9</td>
<td>49°13.16'N 12°37.11'W</td>
<td></td>
<td>1183</td>
<td>25/01/94</td>
<td>OMEX station 6'</td>
</tr>
<tr>
<td>CTD 4</td>
<td>49°06.93'N 13°11.94'W</td>
<td></td>
<td>3350</td>
<td>21/01/94</td>
<td>OMEX station 7'</td>
</tr>
<tr>
<td>CTD 7</td>
<td>48°57.26'N 13°39.82'W</td>
<td></td>
<td>4495</td>
<td>23/01/94</td>
<td>OMEX station 8'</td>
</tr>
</tbody>
</table>

* Consistent with the OMEX biogeochemical sampling stations during Cruise CD84 (Statham, 1994)
Fig 6.3 Selected CTD profiles (refer to fig 6.1) across the Goban Spur OMEX transect during the Cruise CD84. The dotted line represents the depth of deployment for STABLE II.
approximately 880m depth (shown by a relative increase in attenuance from the Seatech 660nm 25cm path-length transmissometer) in the oxygen depleted waters of the permanent thermocline (refer to the results for CTD7 in fig 6.3). The temperature at this depth was approximately 9.66°C. The final deployment depth for STABLE II was chosen to coincide with this weak nepheloid layer. After the deployment of STABLE II, CTD9 also showed this layer to be also present at 1171m depth and of a similar thickness (approximately 120m). The Precision Echo-Sounder (PES) fish, bathymetric survey results are shown in fig 6.4(a) as a function of longitude across the slope. The weather deteriorated during this transect and the bathymetric data was unobtainable or suspect oceanward of 2900m. This information has been linearly interpolated from approximately 3000m in fig 6.4(a) using the next reliable source of data at 4500m. Fig 6.4(a) also shows the CTD stations and the STABLE II deployment site. The bathymetric slope, \( \alpha \), is equal to 0.01 at 879m and is maximum at approximately 2750m (\( \alpha = 0.05 \)). Fig 6.4(b) shows the critical period for internal wave generation as a function of longitude for the bathymetric data shown in fig 6.4(a). The inertial period (15.686 hours at 49° 46.25' N) and the \( M_2 \) tidal period are also shown as a guide and show that the slope along the OMEX transect was not critical at \( M_2 \) tidal period during this survey. This calculation is identical to that presented for the La Chapelle Bank field study and the reader should refer to section 3.2.2.1, fig 3.5 for a description of the methodology. The stability frequency estimate, \( N \), is calculated from the averaged density profile from CTD’s 7, 4, 9 & 3 and is shown plotted in fig 6.4(c). The critical period only reduces below the inertial period within the permanent thermocline and at the STABLE II site, where the Brunt-Vaisala frequency is near maximum at 1.6cph, but the slopes are not steep enough to have a significant effect in the critical period shown in fig 6.4(b).

Finally, using the Brunt-Vaisala frequency, \( N \), profile shown in fig 6.4(d), the internal \( M_2 \) tide characteristic, \( c \), profile is shown in fig 6.4(d) by the solid line. At any depth, if the bathymetric slope, \( \alpha \), was to equal or exceed this characteristic slope, \( c \), then the slope would be critical or reflective for internal tides. The dotted vertical line shows the value of the bathymetric slope at 879m, the STABLE II site
Fig 6.4 Results from the bathymetric and CTD cross-slope survey along the OMEX transect showing (a) the bathymetric profile as a function of longitude, CTD stations and the STABLE II deployment site, (b) the critical period for internal wave generation as a function of distance (longitude) cross-shelf, (c) the Brunt Väisälä frequency profile and (d) a profile of the internal tide characteristic slope, c, (solid line) compared to the bottom slope, $\alpha$, at the STABLE II deployment depth of 879m (dotted line) and at 2750m (dashed line). See text for further details.
and shows that \( \alpha_{870m} \neq c \) at any time. On the other hand, the steep slopes at 2750m depth show that this gradient would be sufficient to cause critical internal tide reflection at 400m and 1100m depth. At 2750m depth the internal tide characteristic is not critical however because of the reduced stratification below the permanent thermocline. We can conclude from fig 6.4(d) that local increases in topographic slope, \( \alpha \), may be sufficient to cause the slope to be critical at \( M_2 \) period. Periodic increases in the stability frequency, \( N \), (particularly at 600-1000m and approximately 2100m depth) may also produce conditions which are critical since the characteristic slope of the internal tide is inversely proportional to \( N \) (equation 2.5, section 2.2.3).

6.5.3 Sea-bed photographs

The POL bed-hopping camera was deployed in 860m water in position 49°23.28’N 11°39.22’W at 20:17 hours on 20th January, one hour before the deployment of STABLE II. A total of 31 exposures were taken of which 11 were completely obscured by clouds of sediment. These clouds of re-suspended material were caused by the impact of the camera-mounted frame on the sea-bed. A selection of these photographs are shown in plates 6.2A-C, which have dimensions of approximately 1.3x2.0x2.0m across the front, back and from front to back respectively. All three plates show the bed to be of a silt/mud composition. Plate 6.2A is a typical undisturbed image of the sea bed and shows little indication of current-induced features. There is evidence of bioturbation in plates 6.2A & B which show some surface evidence of feeding tracks. In plate 6.2A, the orientation of bedforms shown directly above the compass (on the near side right of all three photos) suggests a possible current induced modification, but there is no consistency with other features seen elsewhere in the image. The abundance of feeding tracks and burrow holes suggests that any such current induced modification to these bedforms has not occurred recently.

It is worth noting at this point the complexity of sediment-current interaction for cohesive sediments and the implications for sediment erosion. For non-cohesive
Plate 6.2  Photographs of the sea bed at 860m water depth on 20th January 1994, in position 49°23.28N 11°39.24W on the Goban Spur continental slope.
sediment (such as that described at the La Chapelle Bank site in Chapter 5) the critical velocity for grain movement is proportional to grain size. For silt/mud sized sediment this relationship does not hold true. The cohesive nature of the grains infers a far more complex relationship. The particles have a large surface area to mass ratio and have exteriors which are often electrostatically active. This encourages the particles to flocculate and therefore increase in size (whilst decreasing in relative density), which itself is determined by a highly complex interaction of factors such as sea water chemistry, suspended sediment concentration, organic content and near-bed shear. For example, organic material derived from bacterial activity and absorbed in suspension will increase the ionic charge of the particle, enhance flocculation and increase the effective particle size, whilst also increasing the flocs resistance to de-flocculation. Flocculation and de-flocculation are largely dependent on the degree of current shear, with large flocs being able to exist in environments of low shear. The actual flocculation process itself however requires the frequent collision of individual particles/flocs, which is more probable where there is an environment of higher shear.

The cohesive properties of sediment are beyond the scope of this present discussion but are nevertheless important considerations in the study of sediment transport rates for fine grained material. One implication is an increase in the critical threshold for the re-suspension of bed material when compared to fine sand, i.e. significantly more energy is required to erode material from the bed for silty mud material. If the material is well consolidated then the energy required increases still further. Once the material is in suspension however the behaviour of the sediments is essentially the same as that for non-cohesive sediment and is primarily governed by its size.

Two important considerations result from this discussion. Firstly in the context of the OMEX study as a whole, the large surface area to mass ratio and electrostatic forces of the suspended particles absorb and transport many contaminants and are therefore an important mechanism for the oceanward flux of pollutants in this region. Secondly, for the purposes of this localised near-bed study (but also in a more global context) it is notable that the threshold velocity required for sediment
erosion is considerably more than the velocity required to keep the sediment in suspension for silt/mud material.

Plates 6.2A & B show no evidence that this threshold velocity has been exceeded locally in the STABLE II deployment region. Biological activity is therefore an important mechanism for sediment erosion. Evidence of bioturbation, burrow holes, animal mounds and feeding tracks are all observed in Plates 6.2A-C, all of which encourage the ejection of material into suspension. A sea urchin is shown in plate 6.2B (top right) along with what initially appears to be two parallel feeding tracks in the top centre. Plate 6.2C reveals that these tracks are most likely made by this large flatfish, probably a skate, which is about 2m long from nose to tail. The fish was probably taking off from the sea bed after being disturbed by the bed-hopping camera.

To conclude, the photographs portray a low energy environment with no current induced features and therefore no recent evidence of current induced sediment erosion at this site at the time of deployment. It is more likely that biological activity is a more important re-suspension mechanism in this region. Once in suspension however, sediment is likely to be transported in the residual current for long periods.

6.5.4 Shipek grab samples - analysis and results.

Based on the visual evidence from the bed-hopping camera, the two Shipek grab samples taken in the local vicinity of the STABLE II site (see table 6.2), are thought to be reasonably representative of the surface sediment. Box cores were also taken at the site and are being investigated by Prof. Nick McCave at Cambridge University. Each sample was immediately stored in an ethanol based solution for later analysis back in Plymouth.

Each Shipek grab sample was stirred and the dispersant Calgon (sodium hexametaphosphate) was then added before being put into a micro-sonic bath for 5
minutes. The sediment was then wet sieved using a particle shaker into 500μm, 250μm, 125μm and 63μm size fractions. Each fraction was then oven dried for 24 hours and weighed. For the diluted sample, <63μm, five small sub-samples of homogeneously mixed suspended sediment medium (of known volume) were extracted, oven dried and weighed and the average weight calculated for that size fraction. The weightings for the Shipek grab samples 1 & 2 are given in table 6.3(a) and show the relative percentage weight distribution of mud/silt:sand to be 52:48 and 58:42 respectively.

The analysis of the silt/mud fraction of the material (<63μm) was undertaken using the PML Malvern Instruments 2200 particle sizer system. This technique was particularly relevant since the grab samples had been stored in an ethanol based solution since the cruise. The main advantage of this laser diffractive technique is that it precludes calibrational variations with changes in refractive index, and also density, temperature, and ionic strength of the sample suspending medium (Bale, 1984). Calibration runs confirmed there were no significant changes in particle size distribution using an ethanol and distilled water medium.

Briefly, the Malvern particle size analyser works by sending a parallel beam of coherent monochromatic light through the continually mixed suspended matter. This light source is diffracted by the particulate matter and is detected by an array of photodiodes mounted normal to the beam. A micro computer scans this array, and by measuring the variation of light intensity with each scan it is able to calculate the particle size distribution that correspond to the measured diffraction pattern. A 300mm focusing lens with a particle size range of 5.8-504μm initially showed results which were skewed towards the finer material and so it was replaced by a 100mm lens with a particle size range of 1.9-188μm.

To reduce errors in the results a total of 15 samples were analysed from each Shipek grab sample fraction <63μm using the Malvern Instruments particle sizer. The mean results of cumulative weight distribution (larger than) were combined with the cumulative distribution results >63μm to produce the cumulative weight distribution
Table 6.3 Shipek grab sediment analysis results.

<table>
<thead>
<tr>
<th>(a) Percentage weight distribution</th>
<th>Shipek 1 (860m)</th>
<th>Shipek 2 (890m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weight (% total)</td>
<td>Weight (% total)</td>
<td></td>
</tr>
<tr>
<td>&lt;63μm (mud-silt)</td>
<td>16.20 (52.19)</td>
<td>15.42 (58.61)</td>
</tr>
<tr>
<td>63&lt;μm&lt;250 (medium sand-silt)</td>
<td>10.29 (33.18)</td>
<td>7.66 (29.11)</td>
</tr>
<tr>
<td>&gt;250μm (medium sand +)</td>
<td>4.54 (14.63)</td>
<td>3.23 (12.28)</td>
</tr>
<tr>
<td>Total</td>
<td>31.04 (100)</td>
<td>26.31 (100)</td>
</tr>
</tbody>
</table>

(b) Statistical analysis in phi units

Mean: $M_\phi = \frac{(\phi_1 + \phi_2)}{2} = 5.0$  
Deviation: $\sigma_\phi = \sqrt{\frac{(\phi_1 - \phi_0)^2 + (\phi_2 - \phi_0)^2}{2}} = 2.8$  
Skewness: $\alpha_\phi = \frac{(M_\phi - \phi_0)}{\sigma_\phi} = 0.18$

*Statistical parameters from Inman (1952)

Fig 6.5 A cumulative weight frequency curve for Shipek grab samples 1 & 2. The Wentworth scale is shown on the x-axis (from Buller & McManus, 1979).
plot in fig 6.5. The mean Phi value, deviation and skewness statistics are shown in table 6.3(b). The mean sediment size, $\phi=5.1$ (0.029mm) is representative of a coarse silt. Any sediment $<0.1$mm will exhibit cohesive forces; once the critical threshold for erosion is exceeded in this region the sediment will pass directly into suspension. The sediment is also poorly sorted (shown by the large deviation, $>1\phi$, in table 6.3(b)) since the samples are composed of both sand, silt and mud. Consequently the bed will appear extremely smooth when communicating with the boundary layer flow and increase the effective threshold velocity for the initiation of sediment erosion. For example Heathershaw (1988) cites typical values for the roughness length, $z_0$, of sediment from 18 various sources, 13 of which are also cited in Heathershaw (1981). These values include 0.02cm for mud, 0.07cm for mud/sand, and 0.005cm for a silt/sand combination sediment. A sand/silt/mud sediment will probably have a $z_0$ between the latter two values cited. Once in suspension the finer material is likely to be advected with the residual current for long periods since the settling times for silt is an order of several minutes and that of clay an order of several hours (Buller & McManus, 1979).

6.6 STABLE II - preliminary results.

Data acquisition was highly successful for this STABLE II deployment. Upon recovery, the data was downloaded, fully calibrated, formatted and archived onto optical disk for subsequent analysis. This work was kindly undertaken by colleagues at the POL and was accompanied by full details of sensor calibrations and data format (Humphery & Moores, 1994b). Axis rotations (required to minimise the dominant $M_2$ tidal signal in the vertical component of flow) and EMCM offset drifts have been estimated using the same techniques described for the previous deployment of STABLE in section 4.2.2. These sources of error have been corrected for in the results presented below and are not described further.

* Heathershaw (1988) is referenced because it cites more recent and unpublished values of $z_0$ which include the sand/silt combined sediment type.
6.6.1 Introduction

The first useable mean data began at 0000hrs on Julian day (JD) 21 and continued logging for 10.4 days until recovery on JD 31. A total of 251 files of bursts data were collected from the EMCM arrays and with 9600 scans in each burst, the subsequent data analysis was a lengthy task using PC based software. The low concentrations of fine sediment were below the threshold for suspended sediment detection by the ABS and the results are not discussed. The settlement traps collected very small samples («lg) of sediment and were also analysed using the PML Malvern Instruments particle sizer. The results were very similar in size and distribution to the Shipek grab samples results, i.e. bed derived material, and provided further evidence of very little sediment erosion at the STABLE II site.

Pitch, roll and heading sensors were unchanged throughout the deployment and confirm that STABLE II remained in a 'stable' position on the sea bed before recovery. The local gradient was calculated to be \(1.1^\circ\) (\(\alpha=0.019\), assuming STABLE II lies flat on the sea bed) in direction 269\(^\circ\)T (corrected for a magnetic deviation of 8.5\(^\circ\)), thus defining the long-slope and cross-slope co-ordinate axes to be orientated northerly and easterly respectively. This orientation agrees well with that shown in fig 6.1 for the deployment region. The local slope compares to \(\alpha=0.010\) when compared to the results from the Precision Echo Sounder [PES] fish but this local increase in slope angle is still not critical for internal tide generation [refer back to fig 6.4(d)].

The complete time series of the calibrated results for the mean logger are shown in fig 6.6. The pressure record confirms the 879m deployment depth and a dominant \(M_2\) tidal signal which covers a neap-spring period over the 10.4 day deployment. Time series plots of current speed are shown for rotors 4, 3, 2 & 1 at heights of 76.2cm, 58.2cm, 40.2cm & 22.2cm above the bed respectively. Maximum current speeds of 24\(\mathrm{cm/s}\) are observed for all three rotors during JD 27 at spring tide. Average flows only exceed 20\(\mathrm{cm/s}\) for 1% of the time in the bottom one meter.
Fig 6.6 Time series plots of the mean logged data.
This compares to 12% for flows exceeding 15cms\(^{-1}\). Regression analysis of pairs of rotor current meter data (averaged over 10 minute periods) revealed no increasing trend of current speed with distance above the bed and therefore no detectable current shear in the bottom 1m using the mean logged profiles. This is consistent with the mean burst horizontal current speed time series calculated from the three pairs of EMCM's at heights of 89.9cm, 60.2cm & 30.0cm above the bed.

6.6.2 Evidence of residual slope current reversal.

Accurate calibrations of both the rotor current meters and EMCM's have enabled an accurate description of the mean currents close to the bed. Fig 6.7 is a progressive vector plot for rotor 4 after applying a 10 minute running mean to the original data. The deployment is seen to coincide with a seasonal reversal in the poleward slope current. The residual current for the 10.3 day period was persistently southward, with a mean flow of 1.33cms\(^{-1}\) in direction 183\(^{0}\)T, i.e. along slope and equatorward, which was consistent in magnitude and direction (± 0.2 cms\(^{-1}\)) for each rotor current meter record (refer to table 6.4 in section 6.6.3) and each mean EMCM burst record. This coincided with a mean decrease in temperature of 0.02\(^{0}\)C per day as is shown in fig 6.8, which itself is further evidence of a southerly advection of slope water. On initial deployment the near-bed temperature at 879m depth was 9.57\(^{0}\)C and was 0.2\(^{0}\)C cooler upon recovery.

These mean statistics are also consistent with near-bed current meter data collected from the long term OMEX mooring 154 (R.D. Pingree) due south-west of the STABLE II site (see fig 6.2). Mooring 154 was also deployed during CD84 cruise in position 49\(^{0}\)06.48'N, 12\(^{0}\)10.85'W in 996m water, on Julian day 21 and collected data for 16½ months. The residual current 50m above the bed for the 16½ month period was poleward with a mean flow speed of 2.13cms\(^{-1}\) (Pingree, Sinha & Griffiths, 1996). For the initial 17 days however the slope current was equatorward until Julian day 39, when it reversed poleward. Temperature measurements were also consistent with an equatorward advection of slope water. For the period that the
current meter 50m above the bed on mooring 154 and STABLE II was collecting reliable data simultaneously (i.e. JD 22.8 - JD 31.4) the residual currents were 1.55cms\(^{-1}\) in direction 172.7\(^{0}\)T and 1.16cms\(^{-1}\) in direction 177.7\(^{0}\)T respectively showing considerable shear between the boundary and 50mab. If it is assumed that the long-slope ordinate at the mooring site is also orientated approximately north-south then the results not only confirm the poleward current reversal of the mean flow in the Goban Spur region but may also suggest a lack of boundary layer structure close to the bed (such as Ekman veering) at the STABLE II site. The strong seasonality/variability of the slope currents in the Goban Spur region are discussed in detail by Pingree, Sinha & Griffiths (1996).

Fig 6.7 A progressive vector diagram for Rotor 4 (76.2cm above the bed) for the 10.4 day deployment period. The residual flow direction is shown by the arrow but is not scaled.
6.6.3 Tidal and sub-tidal variability of mean logged data sets.

The dominant signal in all the mean logged current meter time series records is that of the semi-diurnal tide. Fig 6.9 shows that the maximum flow speeds at spring tide are of the order of 24cms$^{-1}$ in the along slope and equatorward direction, which compares to maximum poleward flows of 22.5cms$^{-1}$. A typical temperature excursion for a semi-diurnal tidal cycle at spring tide is of the order of 0.12°C but at can be as much as 0.25°C (fig 6.8). The current spectra plot in fig 6.10(a) confirms that there is significantly more semi-diurnal tidal energy contained in the north component of flow compared to the east component. For the first and second harmonics however the reverse is true, with more energy contained in the cross-slope component. The DQ temperature spectra shown in fig 6.10(b) also shows a dominant peak at semi-diurnal period. Cross-spectral analysis reveals that the temperature signal at semi-diurnal and quarter diurnal frequencies is phase locked with the along-slope current and 90° out of phase with the cross-slope current. This is consistent with the up-slope advection of cooler water during up-slope tidal streaming and the down-slope advection of warmer water during down-slope tidal streaming, i.e. a measure of the vertical excursion of the isotherms up and down slope per tidal period.

TIRA analysis has been performed on each rotor current meter time series, the results of which are shown in table 6.4 below. It should be noted that these TIRA results should be interpreted with caution. The record was a little over 10 days
Fig 6.9 Time series plots of cross-shelf velocity (east component) and long-shelf velocity (north component) for rotor 4 and the 'C' array of EMCM's.

Fig 6.10 (a) Cross-slope (east) and long-slope (north) velocity spectra for rotor 4. The Brunt-Väisälä frequency, N, is also shown. (b) Temperature power spectra from the digiquartz temperature sensor.
duration and therefore four days short of being able to accurately separate the S\textsubscript{2} & M\textsubscript{2} constituents. However, the results presented seem generally in good agreement with the results 50m above the bed on mooring 154 (Pingree, Sinha & Griffiths, 1996).

### Table 6.4 STABLE II tidal analysis (TIRA) results.

<table>
<thead>
<tr>
<th></th>
<th>Mean flow</th>
<th>( M\textsubscript{2} )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mab* (cm)</td>
<td>speed (cm/s)</td>
</tr>
<tr>
<td>RTR 4</td>
<td>76.2</td>
<td>1.3</td>
</tr>
<tr>
<td>RTR 3</td>
<td>58.2</td>
<td>1.5</td>
</tr>
<tr>
<td>RTR 2</td>
<td>40.2</td>
<td>1.2</td>
</tr>
<tr>
<td>RTR 1</td>
<td>22.2</td>
<td>1.3</td>
</tr>
<tr>
<td>Mean</td>
<td>49.2</td>
<td>1.3</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>S\textsubscript{2}</th>
<th>M\textsubscript{4}</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>mab (cm)</td>
<td>a (cm/s)</td>
</tr>
<tr>
<td>RTR 4</td>
<td>76.2</td>
<td>5.5</td>
</tr>
<tr>
<td>RTR 3</td>
<td>58.2</td>
<td>5.4</td>
</tr>
<tr>
<td>RTR 2</td>
<td>40.2</td>
<td>5.6</td>
</tr>
<tr>
<td>RTR 1</td>
<td>22.2</td>
<td>5.6</td>
</tr>
<tr>
<td>Mean</td>
<td>49.2</td>
<td>5.5</td>
</tr>
</tbody>
</table>

Most of the energy is contained at M\textsubscript{2} and S\textsubscript{2} frequencies, with a much smaller contribution from M\textsubscript{4} and M\textsubscript{6} (not shown). The mean amplitude of the major axis of the M\textsubscript{2} tide is 9.39 cm\textsuperscript{s\textsuperscript{-1}} in direction 169.4°T which is aligned along shelf. The ellipticity of the M\textsubscript{2} constituent is 0.33 which is consistent with near-bed values at rig 154. It also suggests that there is no significant change in the ellipticity of the M\textsubscript{2} current near-bed at the STABLE II site and 50m above the bed at mooring 154. Also of significance is that there is no decreasing trend in the amplitude of each tidal constituent with distance above the bed in the bottom one meter.
6.6.4 Inter-burst variance

Out of a total of 251 bursts only the first 217 are described in this section. There are numerous noise spikes in the time series data in these latter bursts which increase in number towards the end of the deployment. Although the mean EMCM velocity profile seems unaffected by these spikes, the bursts have artificially high variances and are not included here. The problem of failing EMCM's seems to be common to both STABLE deployments described here and in Chapter 5.

There is a low level of turbulent energy within each burst time series. Standard deviation time series plots are shown in fig 6.11 which are typical examples from the 3 arrays. Variance (standard deviation) plots of the (a) horizontal port and (b) horizontal starboard EMCM bursts in array A are shown plotted and compared to the mean burst velocity. The mean variance is generally very low (<2cms⁻¹) but still shows a high degree of variability. There are however notable trends in the time series plots, particularly in fig 6.11(b) for the horizontal starboard EMCM. During neap tide conditions (periods of weak long-shelf flow) the mean variance is 0.8cms⁻¹ (JD 23.5-25) which increases to 1.5cms⁻¹ during spring tide conditions (JD 27-28.5). Peaks in flow variance are generally associated with maximum equatorward and poleward currents in fig 6.11(b) with generally minimum turbulence during periods near slack water. Both figures (a) and (b) are indicative of a low energy environment with predominantly steady tidally driven near-bed flows.

6.6.5 Shear stress estimates - Implications for sediment transport.

The three methods which can be used with the present STABLE II data set to estimate shear stresses in the turbulent boundary layer, namely the log profile method, the inertial dissipation method (IDM) and the Reynolds stress method, have been described previously in section 5.2.1. The log profile method showed no detectable current shear in the bottom one meter using either the mean logged rotors or the mean EMCM data. As with the previous STABLE deployment, to increase
Fig 6.11  Time series plots of mean burst velocity and standard deviation for (a) the horizontal port component and (b) the horizontal starboard component of the flow for array A. The dashed line is plotted through the origin of the mean velocity y-axis.
the accuracy of estimates using the Reynolds stress method, a linear trend was removed from each EMCM velocity record and the orthogonal components of the flow rotated such that the co-ordinate system orientated along the mean streamline, i.e. $\bar{U} =$ mean flow speed and $\bar{V} = \bar{W} = 0$. The Reynolds stresses are shown plotted in red in fig 6.13. For each array, two shear stress estimates are shown plotted, one for each of the rotated and de-trended vertical velocity bursts, i.e. port (plots a, c & e) and starboard (plots b, d & f) components. For example, the top two plots (a & b) are shear stress estimates for array C (furthest from the bed). Fig 6.13a uses the vertical port time series as the vertical component and fig 6.13b uses the vertical starboard time series.

For the IDM the spectrums of the vertical time series data are used to estimate the turbulent stresses. We therefore take the vertical Kolmogorov constant $\alpha_3 = 0.69$ (refer to section 5.2.1.4). For each array the theoretical limits for the inertial sub-range varies with sensor height above the bed as shown in fig 6.12.

To recall from section 5.2.1, in the frequency domain, the upper limit varies linearly with the free stream velocity at all depths such that $f = 2.3 \frac{U}{d}$ (Eqn 6.1) where $d=10\text{cm}$ = distance between the electrodes on each sensor head. The lower limit depends on the free stream velocity and the height of the sensor above the bed such that $f = \frac{U}{z}$ (Eqn 6.2). These limits are shown in fig 6.12 for each array for the observed near-bed flow speeds and reveal narrow spectral gaps (particularly for array a) where we can theoretically estimate turbulence within the inertial sub-range. The shear stress results using the IDM are also shown plotted in black in fig 6.13.
Fig 6.13 Shear stress estimates in Nm⁻² using the inertial dissipation method (black) and the Reynolds stress method (red). Results are shown for both port and starboard vertical bursts in each EMCM array. See text for further details.
The results of the independently calculated shear stresses in fig 6.13 show the Reynolds stress method is generally over-estimating the shear stress by up to an order of magnitude for arrays A & B. There is good agreement however in the magnitude of the peaks in stress using both methods which, as will be shown later in this section, generally occur at times of maximum current velocity. It would appear therefore that the Reynolds stress method does have an added background contribution to the calculated stresses at times of weakest stresses when compared to the IDM. The mean angle rotation required for each EMCM head about the vertical axis was 2 degrees and the subsequent axis rotation corrected the miss-alignment to within 1 degree, which itself can incorporate errors of up to 10% (Soulsby, 1983). This discrepancy between the two methods warrants further investigation in future work. The mean and maximum statistics for each method are shown in table 6.5.

A general trend for the mean shear stress values shown in table 6.5 is one of increasing stress with increasing distance above the bed. Although these values infer a deviation from an assumed constant stress, the trends are consistent with other field estimates using the IDM (Grant et al., 1984 and Huntley & Hazen, 1988) which have not been modified using Huntley’s (1988) modified IDM. The IDM for calculating $u_*$ can be violated in low energy environments when the initial assumptions are violated (i.e. that measurements must be made within the constant stress layer and above a critical Reynolds number; Huntley, 1988; refer to section 5.2.1.4 for further details). The modified IDM corrects for such measurements and essentially increases the low values of $u_*$ near to the bed and brings them in line with the values higher in the water column (Huntley, 1988). This implies that the measurements highest in the water column are the most accurate. For future analysis this correction should be applied to the present data set to see if the increase in shear stress with height above the bed can be accounted for.

Typical peak values of $0.05 \text{Nm}^{-2}$ are observed for the shear stress estimates using the IDM which compare to typical peak values of $1 \text{Nm}^{-2}$ at the La Chapelle Bank site (described in section 5.2.2.2). The increase in the shear stresses calculated on La Chapelle Bank are consistent with observations of current velocity and bed
Table 6.5  Mean and maximum shear stress values in Nm$^{-2}$ using the inertial dissipation (IDM) and Reynolds stress (Rey) methods. Results are shown for both port and starboard vertical bursts in each EMCM array.

<table>
<thead>
<tr>
<th>Method</th>
<th>Array C (89.9cm)</th>
<th>Array B (60.2cm)</th>
<th>Array A (30.0cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Vert Port</td>
<td>Vert Star</td>
<td>Vert Port</td>
</tr>
<tr>
<td>IDM - Mean</td>
<td>0.017</td>
<td>0.015</td>
<td>0.014</td>
</tr>
<tr>
<td>- (Max)</td>
<td>(0.18)</td>
<td>(0.15)</td>
<td>(0.13)</td>
</tr>
<tr>
<td>Rey - Mean</td>
<td>0.069</td>
<td>0.038</td>
<td>0.071</td>
</tr>
<tr>
<td>- (Max)</td>
<td>(0.19)</td>
<td>(0.12)</td>
<td>(0.16)</td>
</tr>
</tbody>
</table>

composition and so the magnitude of the shear stress estimates at the present site are also interpreted with confidence.

The thickness ($\delta$) of the bottom Ekman layer can be estimated from the velocity shear ($u_*$) as (refer to Eqn. 5.4 in section 5.2.1.1);

$$\delta = 0.4 \frac{u_*}{f}$$

where $f = 1.11 \times 10^{-4}$ sec$^{-1}$. In more stratified conditions the boundary layer height is modified to the form (refer to Eqn. 5.5b in section 5.2.1.1);

$$\delta = \frac{1.3 u_*}{f(1 + N^2 / f^2)^{\frac{1}{4}}}$$

Eqn 5.5b more accurately reflects the more stable oceanic environment of this second deployment at 879m depth on the Goban Spur. It therefore follows that for typical peak values of shear stress (using the IDM) of 0.05Nm$^{-2}$ the boundary layer height can be estimated as;

$$\tau_b = \rho u_*^2 = 0.05 \text{ Nm}^{-2} \Rightarrow u_* = \sqrt{\frac{0.05}{1027.5}} = 0.7 \text{ cms}^{-1}$$

$$\therefore \delta = \frac{1.3 \times 7 \times 10^{-3}}{1.11 \times 10^{-4} \left(1 + \left(2.79 \times 10^{-3}\right)^2 / \left(1.11 \times 10^{-4}\right)^2\right)^{\frac{1}{4}}} = 16 \text{ m}$$
The boundary layer thickness at peak flow is therefore approximately 15m and the thickness of the constant stress layer thickness, $z_t$, is approximately 7% of $\delta$ (section 5.2.2.2) such that;

$$z_t = 1.1m,$$

which is still well within the physical limits for these methods. Between tidal peaks, fig 6.13 shows much smaller estimates of bed stresses. Arrays A, B and C would be outside the constant stress layer if $D < 4m, 8m$ and $12m$ respectively, i.e. $\tau_b = 3.4 \times 10^{-3} \text{Nm}^{-2}, 1.4 \times 10^{-2} \text{Nm}^{-2}$ and $3.1 \times 10^{-2} \text{Nm}^{-2}$. The mean stress values for the IDM in table 6.5 are the same order of magnitude but do not very often fall below these values.

As previously mentioned, the mean rotors did not detect any appreciable decrease in current speed with distance towards the bed. From this we can conclude that the rotors were either (i) above the logarithmic boundary layer, or ii) were unable to detect the small scale stresses. In the constant stress part of the logarithmic layer

$$\frac{\partial u}{\partial z} = \frac{u_\ast}{\kappa z},$$

so at $z = 0.5m$ with $u_\ast = 7 \times 10^{-3} \text{ms}^{-1}$, $\partial u / \partial z = 3.5 \times 10^{-2} \text{s}^{-1}$. Thus over a vertical separation of 0.54m between rotors 1 and 4 the difference in flow speed, $\Delta u = \partial u / \partial z \cdot \Delta z = 1.89 \times 10^{-2} \text{ms}^{-1}$. This velocity shear should be measurable, at least at times of peak flows. It therefore remains unexplained why the rotors did not see the shear at the Goban Spur site, or at the La Chapelle Bank.

The implications for sediment transport are now briefly discussed. A time series plot for $u_\ast$ using the IDM for array B is compared to the flow speed in fig 6.14. The observed tidal modulation of the friction velocity is a very encouraging result which is shown to vary over a neap-spring cycle as well as over $M_2$ periods. Maximum values of the friction velocity of approximately $1 \text{cms}^{-1}$ are observed at spring tide and reduce to $0.1-0.2 \text{cms}^{-1}$ at times of weak flow.

The sediment distribution and observed near-bed stresses are indicative of smooth flow near to the bed during the entire deployment. A composition of fine sand, silt
and mud allows the sea bed to appear extremely smooth when communicating with the near bed flow ($Re = u_* \cdot D/v < 3.5$).

![Graph](image)

**Fig 6.14** EMCM time series plots of (a) the port component of friction velocity, $U_*$, using the IDM and (b) the flow speed, for array B.

Each of the shear stress methods used for the STABLE II deployment use the assumption of a constant stress layer, i.e. that the stress measured some distance above the bed also the stress which will actually be exerted on the bed. For hydrodynamically smooth flow, skin friction dominates the Reynolds stresses and determines sediment transport conditions within the constant stress layer of the boundary layer. In a similar study in conditions of hydrodynamically smooth flow, Chriss & Caldwell (1982) estimated shear stresses both within the viscous sub-layer and in the logarithmic layer in a region where biological activity had produced small mounds of sediment, and was therefore subject to form drag. The results showed that the measured bed stresses (using the dissipation and logarithmic profile methods) which were more than a few cm's above the bed were more than four times the stresses measured within the viscous sub-layer. The present results are in agreement with those by Dewey & Crawford (1988). One could speculate therefore that estimated mean shear stresses of the order of $0.01 \text{Nm}^{-2}$ using the IDM technique
could be four times larger than that which occurs in a region within cm's of the bed where sediment erosion is important. Further work should be undertaken regarding the implications of these results.

6.7 Discussion

The threshold velocity, $U_{cri00}$, for a typical $z_0$ value of 0.01 cm s$^{-1}$ for a sandy/muddy bed (Heathershaw, 1981) and a mean grain size of 29 µm (section 6.5.4) can be shown from Yalin's (1972) modified Shield's curve never to be exceeded during the deployment period for the measured values of $\tau$ shown in fig 6.13. A minimum likely flow speed approximately 30 cm s$^{-1}$ required to initiate sediment mobilisation of this bed material confirms that the bed was not being eroded at this location at any time during the study. Any fine sediment that is eroded from the bed either periodically or intermittently (due to biological activity for example) will be transported as suspended load in a semi-permanent state and transported with the residual current for long periods. During the deployment period the direction of this residual flow was along slope and equatorward. This is not the norm in this slope region and long term sediment flux estimates would be inaccurately predicted from these results. Pingree, Sinha & Griffiths (1996) showed that the mean flow is predominantly poleward and therefore the mean flux of near-bed suspended particulate matter will also be poleward.

To conclude, the results described here do not suggest that the boundary layer is significantly energetic during this ten day period. Indeed the complete opposite would seem a more plausible description of this slope region. Sediment traps and current meter data suggest that large re-suspended loads of particles were caught in the traps at approx. 1100 m at this time and it is assumed they were being transported in INL's after being 'pinched' off the slope at 800-900 m (A. Antia, pers comm). I can find little evidence of an environment sufficiently energetic enough to form a BNL at the STABLE II site. The settlement traps on STABLE II also showed little evidence of high concentrations of suspended particulate matter advecting past the
rig. It is therefore suggested that if INL's are being formed along this slope transect then it is most probably further up-slope.

Finally, it is worth noting that an initial discrepancy between the mean burst current velocity and the mean logged current velocity resulted from an incorrect sign in the starboard EMCM calibration programme. A considerable amount of time was spent trying to identify this error since it was not known a priori if the compass heading and rotor vane were inaccurate, or whether the port and/or starboard EMCM calibrations were inaccurate. In addition, the initial resolution of the calibrated EMCM data (the order of 0.01 ms\(^{-1}\) resolution) was not of a high enough precision to study the turbulent nature of the bed in detail. The recalibration of the raw data files proved time consuming. The author therefore is strongly in favour of the uncalibrated data sets also being made available to scientists following future deployments. These problems could have probably been overcome if more advise had been given to the engineers at POL regarding the required format of the useable data. This is by no means a criticism of the POL engineers, whose commitment and helpfulness is beyond question, but it highlights a necessity for scientists and engineers to work closely together with regard to the exact requirements of data collected in the field.

6.8 Further work.

From the processes point of view the field study has successfully negated the hypothesis that a INL was being formed at the STABLE II deployment depth at 879m during the 10.4 day period. However, future preparations for the refereed publication of these results would benefit from a comparison of the inference of these results with other benthic studies within the OMEX program. In particular, it would be advantageous to compare the STABLE II current meter results with results collected simultaneously from the cross-slope array of OMEX moorings. From a theoretical and modelling viewpoint the shear stress estimates could be examined in more depth. The estimates of shear stresses near to the bed are unique to the Goban
Spur slope region, and present a theoretical and physical environment in which existing models of bed-flow interaction can be compared. The shear stress results presented in this section are preliminary. For example, results have not been recalculated using Huntley’s (1988) modified IDM and the turbulent spectra from the horizontal burst components of flow could also be used as a comparison for this method. The individual shear stress estimates have also not been extensively compared to time series and wavenumber spectral plots.
Chapter 7

Conclusions

Two highly successful field campaigns on the continental margins of the Celtic shelf have collected unique and valuable data at contrasting sites. The scientific value of the results presented in chapters 4 and 5 have exceeded initial expectations. The EMCM measurements and in situ ABS results are believed to be the first of their kind on the European continental margins. The turbulent current data has inferred bottom stresses which accurately describe the physical and geological nature of the near-bed, deep sea environment. The 1990 La Chapelle Bank STABLE deployment is believed to have been the first intensive sediment transport and boundary layer flow investigation on critical M₂ slopes.

This thesis represents a systematic study of flows affecting the near-bed regime at the Celtic Sea shelf edge. A summary of the main results from the La Chapelle Bank and Goban Spur field programs have been described in detail in the final discussion sections of chapters 4 & 5 and chapter 6 respectively. This final chapter draws together the main findings of the thesis in a short and precise manner and discusses
the implications of these results. Finally, a list of conclusions are included at the end of each section and outline the specific contributions this study has made to our current knowledge of near-bed flows and sedimentation processes on continental slopes.

7.1 La Chapelle Bank field study

A near-bed field study of the hydrodynamic and sedimentation processes on the upper slopes of the Celtic shelf has been undertaken. The 23 day deployment of the benthic lander STABLE in July 1990, was at a depth of 388m and at the head of a canyon, where steep slopes (α=0.10), stratification (N=0.9cph) and strong cross-slope tidal currents generate large internal tides. These internal tides had previously been observed by Pingree & New (1989 & 1991) as beams of energy propagating oceanward from this critical slope. Further observations by Pingree (1988) and model results by New (1987) suggested that near-bed currents would be locally enhanced at the generation site. Pingree & LeCann (1989) observed that within canyons, bottom currents are further amplified, with maximum currents orientated down-slope.

At the STABLE site, local down-slope enhancement of the M2 tidal currents with maximum near-bed flows of 40cms⁻¹ orientated down-slope were observed. The bottom sediment was composed of a poorly sorted gravely sand with a mean grain size of 870μm, which is consistent with the observed flow rates.

The ABS sensor measured peaks in suspended sediment concentration (albeit low concentration) of sand sized material during peak down-slope flows. This is also consistent with near-bed transmissometer data collected 8mab at 305m depth, which showed reduced optical beam transmission during down-slope flow. At the STABLE site this single peak can not be interpreted as a simple advective process back and forth because this would give two peaks per tidal cycle. A single pulse would be produced if the tripod was at the extreme of tidal excursion, but such a
pulse would occur at maximum excursion, i.e. zero current, and not at maximum current. The results therefore suggest a net transport of the finer material down-slope. Two possible scenarios justify these observations;

(1) Local re-suspension of material is taking place on the stronger down-slope phase of the tide but not on the up-slope phase. Typical peak shear stress values of 1Nm$^{-2}$ were measured from the EMCM turbulence measurements and infer a corresponding friction velocity of 3.1cms$^{-1}$. Peak values of shear stress also only occurred during down-slope tidal streaming.

(2) The sediment advected down-slope past STABLE and settles out before the up-slope phase begins.

The threshold for sediment movement for the coarse shell material was never exceeded during the deployment and is in agreement with the STABLE photographs. However the threshold for movement of finer material was exceeded for 63% of the time and suggested that the finer sediment was being winnowed out of the bed at peak flow rates. It was therefore not fully understood why peaks in suspended sediment were not observed during the smaller up-slope flows. It is suggested that the consolidated nature of the bed (from the photographs) imply that a larger critical threshold is required to mobilise the sediment than that predicted for the known grain size distribution. This would suggest that re-suspension only occurs during the down-slope phase of the tide (as observed) when bed stresses are largest. In either case the net transport of sediment is always down-slope.

Further up-slope (305m) the strong asymmetry in the near-bed Eulerian $M_2$ tidal currents also imply a net down-slope transport of sediment. At the shelf break (200m), the asymmetry in the near-bed Eulerian $M_2$ tidal currents are reversed, implying a shelfward transport of sediment and a region of bedload parting at the shelf break. This bedload parting was successfully predicted from near-bed current measurements at the shelf break and on the upper slope using Hardisty's (1983) bedload equation.
It has not been possible to separate the baroclinic and barotropic components of the M$_2$ tide from the present data set and so it is difficult to implicate these observations to internal tide/wave activity. However internal wave activity at the bed can be inferred from:

(i) Near-bed enhancement of flows at tidal period on the upper slope (as previously mentioned). These maximum down-slope flows are smaller than those reported by Pingree (1988), measured in the immediate vicinity of the STABLE site at 548m depth. The residual current 33mab was 15cms$^{-1}$ and compares to 11.2 cms$^{-1}$ at the STABLE site (385m) in 39.3cmab. Both of the residual currents are orientated down-slope. At mooring 124 (305m) the residual current 8mab is 6.7cms$^{-1}$ and also orientated down-slope.

(ii) The development of non-linear effects where internal waves become large and even break. This would be apparent through the development of motions at higher than semi-diurnal frequencies. This seems to be evident from three independent sources:

[a] The sharp cooling ‘front’ which is observed to propagate up-slope, phase locked with maximum on-shelf tidal streaming, suggests something other than sinusoidal motion. This observation is similar to a surging breaker on a beach and is most prominent at spring tide. This feature is not due to the simple advection of a permanent front for the same reasons as those described for the ABS results, and because the mean down slope flow would also quickly advect the front down-slope past the thermistors. Instead, there is a period of more gradual re-stratification during the down-slope phase of the semi-diurnal tide.

[b] Internal tide current surges within the boundary layer at the shelf break. These ITCS’s are essentially forced depressions of the seasonal thermocline and propagate on-shelf consistent with observations by Heathershaw (1985).

[c] The burst EMCM turbulence data (39.3cmab) shows that peaks in current variance and near-bed shear are associated with peak down-slope flows. This implies that increased bed shear and mixing is associated with maximum down-slope flows and not with the enhanced internal wave activity which is associated with the up-slope phase of the semi-diurnal motion (as described in [a] above). Since the temperature sensor on board STABLE failed to collect data, there is no...
means of knowing whether the frontal feature also advected past the STABLE rig. This observation is contrary to observations by Thorpe (1987) and Thorpe & White (1991) and to modelling studies by Slinn & Riley (1996) and Ivey et al. (1995) which suggest increased mixing during the up-slope phase. Gardner (1989b) also observes increased levels of turbulence on the up-slope phase of the tide which is responsible for the periodic re-suspension of sediment on the upper slopes of Baltimore Canyon. Sediment was then seen to be transported down-slope during the off-shelf phase of the semi-diurnal tide. Sediment at the STABLE site is of a coarser composition however and the STABLE photographs suggest that it will only be re-suspended during peak flows.

All of these observations have pointed to enhanced internal wave activity as the mechanism for the down-slope transport of sediment on the upper slope. The results have shown that close to the bed on the upper slopes of La Chapelle Bank the residual poleward current is veered strongly cross-slope due to an enhancement of the baroclinic energy. However, particulate matter which is held in suspension away from the bed will be transported along-slope and poleward (with the slope current). If sediment flux measurements are estimated from residual current measurements then the current meters should be located as near to the bed as possible in this region.

Finally, this systematic study of the near-bed flow and sediment transport regime on the upper slopes of La Chapelle Bank has provided a distinct scientific contribution to knowledge in several respects. These contributions are summarised in the list below;

- **Documentation of phase-locked internal tide and higher frequency internal solitons near their region of generation at the shelf break.**

Near-bed internal wave current surges are observed at the shelf break approximately 3 1/2 hours before maximum on-shelf flow, phased-locked with the semi-diurnal barotropic tide, with a maximum peak to trough amplitude of 25cm/s. These current surges are a manifestation of forced oscillations within the seasonal thermocline.
• Documentation of the mean flow becoming greater with an increasing off-shelf component near to the sea bed on the upper slope. Although the depth averaged Eulerian residual currents are along-slope (being primarily driven by the poleward decline in dynamic height, i.e. density driven), there is clear evidence of vertical (i.e. baroclinic) modulations to the observed slope current and temporal variations over a spring-neap cycle. Near-bed flows are locally enhanced and orientated cross-slope near to the bed.

• Documentation of a steep 'front' of cooler water moving up-slope each semi-diurnal tide which is more prominent at spring tide. Rapid decreases in temperature are observed from near-bed thermistors (0.7°C decrease in 1 hour) near maximum up-slope tidal streaming. A rate of advance of 65cms⁻¹ is estimated from near-bed temperature observations on the upper slope and shelf break region. These 'fronts' are indicative of the near-bed baroclinic tidal response of the isopycnals impinging on the upper slope and have been previously observed by Gardner (1989b), Thorpe et al. (1990) and White (1994). Near-bed transmissometer observations on the upper slope region show that the up-slope advection of this frontal feature is associated with an advection of clearer water up-slope, which is followed by the down-slope advection of more turbid water during the down-slope phase of the M₂ tide.

• Analysis of the measurements of near-bed turbulence on the critical M₂ slope. The results find increased turbulence during the down-slope phase of total tidal flow, which is not in accord with most previous observations or model predictions. Estimates of bed stresses using the inertial dissipation method (τ IDM) show a consistent correlation between maximum down-slope tidal flow (~40cms⁻¹), maximum current variance (~30cm²s⁻²) and maximum bed shear stress (~1Nm⁻²).
• Acoustic backscatter (strength) correlated with maximum down-slope flow on the critical slope region where there is expected (and observed) to be an amplification of the M$_2$ tidal current near to the bed. Suspended sediment concentration derived from acoustic backscatter is tidally modulated with neap-spring and individual M$_2$ tidal cycles clearly evident. Increased ABS concentrations occur at times of maximum current (during the down-slope phase of M$_2$ tidal flow) and not maximum advective excursion, suggesting that the sediment is eroded locally on the critical M$_2$ slope. Sediment transport predictions confirm that near-bed stresses are large enough to erode the sand sized material at the shelf break during maximum down-slope flow and infer a net transport of sediment down-slope.

• Calculation of bed-load parting on the upper slope below the shelf break. Near-bed current measurements on the upper slope and shelf break suggest that the upper slope region (below the shelf break) is a region of bed-load parting during summer stratified conditions. Predicted sediment transport rates suggest that transport is largest at the shelf break, is orientated strongly shelfward, and infers an accumulation of sediment. These observations of bed-load parting are consistent with Heathershaw et al.'s., (1987) internal tide/ sediment transport model. This sediment may be transported oceanward during large intermittent transport events during winter mixed storm conditions. This last statement is purely speculative.

7.2 Goban Spur field study

This second opportunistic deployment of STABLE II in January 1994 has enabled a study of the along-slope variability of near-bed flows and sediment transport processes on the Celtic Sea continental slope. The specific objectives of this second STABLE II deployment on the Goban Spur was to study the hydrodynamic conditions which are considered favourable for the formation of BNL's and INL's, and the nature and stability of the slope current. Some 'truncations' to this work are acknowledged in chapter 5 (such as the correction to the bed stress estimates using
the inertial dissipation method, and further current analysis and comparisons with other available OMEX data). This section briefly summarises the main findings of the study to date but is subject to further analysis (as described in section 6.8).

The 10.4 day deployment on the Goban Spur (49°20'N) was at 879m depth with a reduced bottom slope (α = 0.01), increased stratification (N=1.6cph) and reduced tidal (maximum tidal currents of 24cms⁻¹) and residual (1.55cms⁻¹) currents in comparison to the previous deployment site. The semi-diurnal tidal flows and the residual flows were both orientated along-slope, but the gentler slopes are generally thought to be less insulating to cross-slope flow (Pingree, Sinha & Griffiths, 1996). The residual current was orientated equatorward for the entire deployment and coincided with a seasonal reversal in the poleward slope current. This flow reversal was consistent with long term current meter measurements by Pingree, Sinha & Griffiths (1996) 50mab.

The analysis of the Shipek grab samples and the series of bed photographs taken prior to deployment showed that the bed was of a fine sand-mud composition and would therefore be transported in suspension if the critical velocity was exceeded. STABLE II was deployed in the permanent thermocline where it was hypothesised that the enhancement of baroclinic M₂ tidal energy would locally re-suspend this fine bed material. Once in suspension the sediment would subsequently be transported in INL's or BNL's (e.g. Dickson & McCave, 1986, Thorpe & White, 1988). Estimated shear stresses of 0.05Nm⁻² and a corresponding friction velocity of 0.7cms⁻¹ were shown and observed to be far smaller than that required to exceed the critical threshold for initiation of motion. Unfortunately the camera system was not available on STABLE II and so the mobility of the bed at peak flow rates were not observed in situ.

7.3 STABLE instrumentation and developments in data analysis techniques

The camera system used to monitor the mobility of the sea-bed during the La Chapelle Bank deployment was invaluable for the correct interpretation of the
suspended sediment measurements and predicted transport calculations. For future STABLE deployments it would seem imperative that a new camera system is fitted (the old system is no longer available). The accurate interpretation of the residual, tidal and turbulent currents, as measured by the EMCM's, were heavily dependent on the rotor current meter data being available for the calibration of the EMCM offsets. The Aanderaa rotor current meters should be included as part of the instrumentation in all future STABLE deployments for the EMCM calibration. This offset discrepancy is discussed further, later in this section. It is also notable that during both STABLE I & STABLE II deployments the EMCM's failed before recovery (the mean logged currents were not effected by this failure however). It would seem that the present system can only operate for limited periods when deployed in the deep waters off the continental shelf. This limitation should be addressed for future deployments.

A systematic procedure for the analysis of the STABLE data has been described in detail in section 4.2. Particular attention has focussed on the EMCM analysis since field data collected from these sensors are notoriously difficult to accurately calibrate. The two STABLE deployments demonstrated that the EMCM offset values varied significantly between laboratory calibration and deep sea deployment. The deployment offsets have been shown to be consistently smaller than the laboratory offset calibrations with a maximum offset discrepancy of \(-8\text{cms}^{-1}\) observed for the starboard EMCM during the 1990 La Chapelle Bank deployment. The long term offset drift in the mean vertical burst EMCM data also suggests that the offsets vary during the deployment. The calibration procedure described in section 4.2 is summarised below and is recommended for future STABLE deployments on the northwest European shelf and continental margin.

Pre-analysis checks and calibration procedure for EMCM data collected from the STABLE I and II deployments;

(i) axis alignment

- Check that the individual EMCM sensor heads are correctly positioned in the \(U,W\) plane relative to the sea bed.
Apply the laboratory calibrations to the raw data, calculate the mean value of each horizontal and vertical burst and de-mean each mean burst EMCM time series. If the EMCM's are truly aligned in the vertical relative to the sea bed, the mean of each vertical port and starboard burst will be zero. In tidally energetic environments, a tidal periodicity in either vertical mean burst time series is evidence of sensor mis-alignment and should be corrected for prior to implementing the offset calibration procedure.

- **Axis rotation.**

  If the individual EMCM sensor heads are incorrectly positioned in a U,W plane relative to the sea bed, the ratio of the variance at the $M_2$ spectral peaks (or other statistically significant spectral peaks of the higher tidal harmonics) in the vertical and horizontal spectra can be used to estimate the degree of mis-alignment, $\phi$, of the U,W plane such that:

  $$\phi = \tan^{-1} \sqrt{\frac{PSD_{M2\text{ Vertical}}}{PSD_{M2\text{ Horizontal}}}},$$

  where PSD is the power spectral density.

  The axes can then be rotated using the equation:

  $$\begin{bmatrix} U' \\ W' \end{bmatrix} = \begin{bmatrix} \cos \phi & \sin \phi \\ -\sin \phi & \cos \phi \end{bmatrix} \begin{bmatrix} U \\ W \end{bmatrix},$$

  where $U'$,$W'$ is the new coordinate system.

  Re-calculate the PSD of the mean vertical burst time series to check that the magnitude of the variance at $M_2$ tidal period has been minimised.

(ii) Offset calibration

- **Calibrate the EMCM offset values for the port and starboard horizontal components of flow using the rotor flow speed time series.**

  Identify discrete mean values from the continuously logged rotor flow speed time series which are coincident with the mean EMCM burst sampled data. Compute the optimal R-squared correlation between the rotor flow speed and EMCM flow speed time series using a combination of port and starboard offset pairs. Add this offset correction to each horizontal EMCM time series and compute the cross spectral density of the horizontal flow speed for the rotors and EMCM's. An accurate offset calibration will show a high level of coherence across the
entire frequency domain. Little, or no coherence at the low frequency end of the spectrum may suggest EMCM offset drift during the deployment.

(iii) Check the accuracy of the high frequency inter-burst EMCM measurements.

- Compare the predicted EMCM flow speed time series (using estimates of the bed shear stress from the turbulence EMCM measurements) to the measured EMCM flow speed time series.

Calculate the bottom stresses ($\tau$) from the turbulence measurements using the inertial dissipation method (IDM) and Reynolds stress method (Rey). Estimate the thickness of the constant stress layer to check the validity of the methods used. If the turbulence measurements were collected in the constant stress layer then re-calculate the EMCM mean horizontal flow speed from the estimated $\tau_{\text{IDM}}$ and $\tau_{\text{Rey}}$ values using the assumption that the vertical distribution of velocity varies logarithmically with depth, where:

$$\tau = \rho u^2 (\text{Nm}^2) \quad \text{and} \quad U(z) = \frac{u_s}{\kappa} \ln \left( \frac{z}{z_0} \right) \text{(ms}^{-1})$$

(refer to section 5.2.2.2 for further details). A comparison of the EMCM flow speed time series with the flow speed calculated using $\tau_{\text{IDM}}$ and $\tau_{\text{Rey}}$ provides a good independent test of the accuracy of the EMCM burst logged data.

7.4 Future work and recommendations for future STABLE deployments.

The La Chapelle Bank region of the Celtic Sea is a major export region for the transport of non-cohesive material from the Western Approaches out into the ocean, via the many submarine canyons which incise the shelf break and upper slope region. The transport of bed material onto the upper slope is likely to be prohibited during summer stratified conditions but will move oceanward during winter mixed conditions.

The Goban Spur is a region of sediment deposition. The bed is typically composed of fine silt/mud sized material and if eroded from the bed will remain in semi-
permanent suspension for long periods. Suspended sediment can then peel off the sloping region along isopycnals and subsequently be advected with the residual current. The primary transport mechanism in this region is not therefore restricted to the turbulent boundary layer.

The geological and hydrodynamical variability of the two separate STABLE deployments have highlighted major difficulties in the accurate prediction of shelf-ocean fluxes along the whole of the northwest European continental slope. Before flux estimates can be confidently made, a more global data base is required from benthic lander deployments, such as those described, which are fitted with a suite of sediment transport and boundary layer equipment. On the La Chapelle Bank, the cross-slope instrument transect was not able to resolve the relative importance of submarine canyons as a conduit for sediment transport in tidally energetic environments (Gardner, 1989b). The La Chapelle Bank upper slope region is the perfect field laboratory at which to study this phenomenon further and future work should be conducted here, in preference to other locations along the slopes of the northwest Europe.

Recommendations for future deployments of STABLE on the La Chapelle Bank continental slope, and the necessary ancillary measurements are now described. The primary objective of any further study should address the question “what is the turbulent and sedimentary response of the sloping benthic boundary layer to barotropic and baroclinic tidal forcing?” This present field study has suggested (but is by no means conclusive) that the biases in the flow that cause enhanced down-slope sediment transport on critical M2 slopes, could be caused by periodic variations in the size of the bottom mixed layer over a tidal cycle. The asymmetric form of the near-bed temperature time series suggests evidence of an enhancement of baroclinic tidal energy during the up-slope phase of tidal flow, which may (through enhanced mixing) increase the thickness of the bottom mixed layer. To have a net zero cross-slope temperature flux of water the flow will move more slowly during the up-slope phase of tidal flow when the height of the bottom mixed layer is largest. During the down-slope phase of the M2 tidal cycle, the boundary layer is observed to
re-establish itself and would therefore be confined to a narrower region near to the bed. The near bed flows would therefore be strongest during the down-slope phase of the tidal flow. Bed induced shear stresses will then also be largest during down-slope flow and the net transport of sediment will also be down-slope for the sand/gravel sized bed material which is observed at this location. An intensive measuring strategy is required at one location to address these uncertainties/hypothesis more fully. Temporal resolution and near-bed vertical spatial resolution is favoured at the expense of along-slope and cross-slope spatial coverage from the resources available. In what follows, two successive STABLE deployments should be implemented. One deployment should be at the head of a submarine canyon and one slightly along-slope, so that the relative significance of the focusing of barotropic and baroclinic tidal energy and higher frequency internal wave energy in canyons can be addressed. A seasonal study should also be undertaken to study the relative importance of seasonal stratification for transporting material from the shelf break onto the upper slope.

(A) Recommendations for future STABLE deployments;

1. An array of EMCM’s are essential for the comprehensive study of the boundary layer flow and turbulent regime on the continental slope. The previous two continental slope deployments of STABLE have demonstrated that the mean logged rotor current meter data is not a reliable source of data for calculating shear stresses using the velocity profile method. Estimates of the shear stress using the turbulence EMCM measurements (i.e. the inertial dissipation method and Reynolds stress method) accurately reflect the nature of the near-bed physical environment, at least to a correct order of magnitude. The shear stress estimates from the array of three pairs of EMCM’s on STABLE II (for the Goban spur deployment) have also revealed a potential variability of the magnitude of the stress with distance above the bed. The general trend was one of increasing stress with increasing distance above the bed. Although this infers a deviation from a constant stress layer (which affectively violates the initial assumptions of the shear stress methods used, and is subject to further analysis) the data provides an invaluable, independent comparison of the variability of these
estimates. The rotors play an important role in the accurate calibration of the EMCM offsets, which all of the shear stress methods are dependant upon. The rotor and EM current meters should therefore not be treated simply as a duplicate set of measurements. It would be interesting for future deployments to position pairs of EMCM's and individual rotor current meters at the same height. A direct comparison of results would enable a thorough assessment of the limitations and advantages of each system.

2. For an intensive near-bed, sediment transport and boundary layer study, the emphasis should be on maximum temporal resolution of the burst sampled data and not on a prolonged deployment, where large discontinuity's are required between individual bursts. One 14 day neap-spring tidal cycle is a sufficient sampling period and the length of each burst record (at 8hz sampling frequency) should be adjusted so that the maximum temporal resolution is obtained over this period.

3. A high resolution temperature sensor should be included in the instrumentation package to study turbulent energy dissipation within the boundary layer. The baroclinic response of the isopycnal impinging on the upper slope region could also be more extensively investigated.

4. A camera unit should be included as part of the STABLE instrumentation if a comprehensive investigation of the sedimentation processes is anticipated.

5. The single frequency ABS on STABLE I and the double frequency ABS on STABLE II provided limited and no information respectively, on the near-bed flux of suspended sediment at the STABLE deployment sites. A triple frequency ABS system, which could be fitted to STABLE II, is under development at the Proudman Oceanographic Laboratory. This system should provide clearer information regarding suspended sediment size and absolute concentration in the near future.

(B) Ancillary measurements

1. It is recommended that a CTD yoyo station be maintained near the STABLE deployment site for a complete tidal cycle at spring tide, continuously profiling from the sea bed to 100mab. This will enable a comprehensive study of the
response of the isopycnals to the \( M_2 \) tidal flows near to the bed on the upper slope region. A transmissometer and nephelometer should both be included on the CTD to continuously monitor the turbidity of the bottom water over the \( M_2 \) tidal cycle. The Rosette bottles should be fired at 1 hourly intervals near to the bed during this continuous CTD profile, to enable an accurate calibration of the turbidity measurements.

2. A bottom mounted ADCP should be deployed at a location near to STABLE, and at a similar depth, enabling a comprehensive investigation of the tidal currents throughout the bottom boundary layer.

3. A thermistor chain should also be deployed in the near vicinity with an increasing vertical spatial resolution as the sea bed is approached. The sampling interval should be the shortest possible for a 14 day spring-neap period. The CTD yoyo station can be used to calibrate individual thermistors to a high level of precision.

4. Shipek grab samples should also be taken from the immediate vicinity for later analysis in the laboratory.
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