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Currents on Continental Margins and beyond

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Currents on continental margins and beyond

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A description is given of the techniques at present in general use for the measurement of currents with particular emphasis on methods used in water depths greater than 200 m. The general characteristics of current motions both in the deep ocean and in continental shelf seas are described and are categorized in terms of their energetics, periodicities and vertical and horizontal length scales.

Introduction

The continental slope is a boundary region between the tidally dominated current régime on the shelf and the generally less energetic but more complex currents in the deep ocean. Currents in this transition region are difficult to characterize since the local variations in water depth are great and thus have a large influence on current patterns and energetics. In this paper we shall consider the characteristics of currents in the deep ocean and on the continental slope, and where relevant compare these with the currents in the shallow continental shelf seas. Apart from a few intensive studies of deep ocean currents, rather little is known of their regional variability, a result largely of the immense horizontal extent of the abyssal ocean and the high cost and sophistication of the instrumentation needed to make measurements in the deep ocean.

Much is now known of the temporal variability of ocean currents and of their vertical structure, and a large part of this paper will be devoted to the types of motion present over a range of periodicities from months to seconds.

Current measuring techniques

In order to understand the problems of making measurements of currents in the ocean it will be informative to consider the techniques now in general use for making such measurements, the history of their development and, above all, their shortcomings.

Eulerian measurements

This class of measurements includes all measurements of velocity–time-series at fixed points in the ocean. Before the mid-1960s there were almost no measurements of currents away from the continental shelf (Bowden 1954). This situation was changed by the development of self-contained, internally recording current meters which could make measurements at any depth for periods of weeks and months (Richardson, Stimson & Wilkins 1963; Aanderaa 1964). These and other similar instruments employed propellers or rotors to measure current speed and a magnetic compass and vane to determine flow direction. The instruments rapidly achieved a high degree of reliability which was unfortunately not matched by the reliability of the mooring systems on which they were deployed. In the deep ocean, moorings capable of withstanding months of exposure have become routine only over the past 5 years or so. The engineering problems encountered by one U.S. laboratory are well documented by Heinmiller (1976).

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The types of mooring now in general use have evolved with the development of new materials and with the understanding of the effects of environmental conditions on the performance of instruments. In the past 3 or 4 years it has been shown that under certain circumstances the records of currents made with conventional rotor/vane instruments on moorings with buoyancy at the sea surface may significantly overestimate the magnitude of the currents. The problem is due to energy at surface-wave frequencies being transmitted down the mooring line and producing motions of a frequency too high for the instrument to respond adequately. Paradoxically

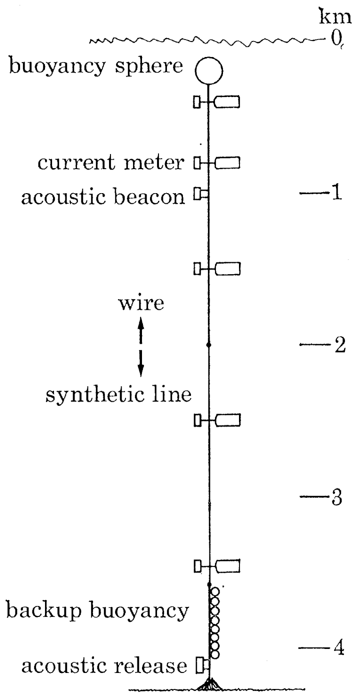

FIGURE 1. Design of current meter mooring for measurements in deep water. Typical main buoyancy would be around 700 kg and anchor mass 1000 kg.

the effects are most marked at the deeper levels, due to the generally low current speeds at those depths, and may not be so severe in the energetic, low-frequency near-surface flows. A comparison of current statistics at the same site from both surface and subsurface moorings (Gould & Sambuco 1975) showed that current amplitudes could be overestimated by as much as 100 % over a depth range from 100 to 2000 m in water of 2600 m depth. As a general precaution the use of conventional rotor/vane current meters on surface buoy moorings should be avoided.

Where measurements are not required in the uppermost layers of the ocean a design of subsurface mooring similar to that shown in figure 1 has generally been adopted. The mooring is recovered via an acoustic release between the instrumented part of the mooring line and the anchor. The upper 1500 m is made up of 6 and 8 mm diameter torque balanced wire terminated with swaged steel fittings. The wire is galvanized and may also be jacketed by a plastic coating. In very deep water (greater than 2000 m) the weight of an all-wire mooring is often prohibitive and in this case the lower sections of mooring line may be made up of braided man-made fibres. This is a compromise since the large cross-sectional area of the 'synthetic' line leads to

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high drag and the poorly defined stretch characteristics may mean that instruments cannot be placed at precisely the correct depth.

The aim is to make the mooring as 'stiff' as possible without reducing the safety factor of the mooring line. A 'stiff' mooring will remain upright and not be displaced significantly from the vertical by the currents which pass it. A typical I.O.S. deep-water mooring in 5000 m of water experiences vertical excursions at its upper end of the order of ± 10 m at tidal frequencies and is subject to stretch of the order of 10 m over an exposure period of four months.

A solution to many of the problems of mooring design may lie in the use of the new aramid fibre materials (marketed under the trade name Kevlar). These have similar densities to other man-made fibres but have strengths and stretch characteristics similar to steel. Present drawbacks to such an approach lie in the high cost and the need to armour such materials in the upper ocean owing to the possibility of damage caused by fish bite (Stimson 1965; Haedrich 1965).

In the majority of long exposure moorings in the deep sea, buoyancy is inserted in the mooring line immediately above the acoustic release. In the event of a mooring failure in the upper part of the line, sufficient buoyancy remains to bring the instruments back to the sea surface and enable the cause of failure to be investigated.

Moorings of the type described above have been used extensively in water depths between 200 and 5000 m. The problems of making current measurements in the uppermost, wave-affected depth range and from moorings with surface buoyancy will most probably be overcome by the introduction of electromagnetic and acoustic velocity sensors which can measure directly the orthogonal components of current velocity (Tucker 1972).

A rather different set of mooring design problems has led to another technique for deploying current meters in the shelf seas. Although the instruments are basically the same as those used in deep water, they are not in general used on single-point moorings. Poor acoustic conditions make the use of subsurface buoyancy coupled with acoustic releases difficult and instead a U-shaped design is employed. This consists of an instrumented leg with subsurface buoyancy connected by a ground line to a surface marker buoy (Howarth 1975*a*). The relatively shallow water depths make the use of large diameter wires possible and there are thus no problems in lifting the full anchor weight when recovery is undertaken.

The major hazard to such installations comes from fishing activity. In the event of loss of the surface marker buoy the instruments are recovered by dragging for the ground line. Largely as a result of fishing activity the typical deployment periods of shallow water moorings in U.K. waters are of the order of 2 months or less.

Lagrangian techniques

Into this category fall all the methods that involve the tracking of a drogue or float intended to tag a body of water. At the same time that recording current meters were being developed, the first measurements of currents in the deep ocean were being made by tracking neutrally buoyant floats (Swallow 1955). These devices are designed to be less compressible than seawater. They are ballasted to sink at the sea surface but as they sink they gain buoyancy until at some predetermined pressure level the density of the float is equal to the density of the surrounding water. They remain in stable equilibrium at this level and are tracked by their emission of acoustic signals. In the earlier floats these signals were emitted continuously at a regular repetition rate and the float position was determined by manoeuvring the ship overhead of

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the float. This seriously limited the number of floats that could be tracked at any time and the area over which measurement could be made.

The development and use of a transponding float, in which float ranges from the attendant ship or from a fixed interrogator on the seabed, made possible the tracking of floats over distances of up to 70 km (Swallow, McCartney & Millard 1974).

As with all ship-based tracking systems the duration of the measurement is limited by the endurance of the attendant ship (typically 30 days or less). Using low-frequency sound acoustic ranges of many hundreds of kilometres are attainable and lead to the possibility of tracking neutrally buoyant floats from fixed shorebased or moored listening stations (Rossby & Webb 1970). The sound signals are channelled by a minimum in the vertical profile of sound velocity at a depth generally near 1000 m (the Sofar channel).

Tracks of many months duration from a large area of the western North Atlantic SW of Bermuda have been collected since 1973. For the most part these have been from depths close to the Sofar axis but it now appears that reception conditions may be adequate over a range of depths from 500 to 3000 m.

Length and time scales of current variability

Both in shelf seas and in the deep ocean a measurement of current samples a range of periodicities spanning from a few seconds to hundreds of days. Some of the frequency components are discrete such as tidal motions, the majority are not so and fall in a range of frequencies. Similarly the vertical structure of the current profile and the horizontal scales of current variability are dependent on the dominant motions present and on factors such as the stratification of the water column and the local variation of water depth.

We shall attempt to characterize the various types of motion known to exist in the ocean, to identify their causes and to identify their dominant time scales and vertical and horizontal length scales.

Table 1 gives a summary of these characteristics for the majority of known current motions. The figures quoted for periods, scales and amplitudes are intended only to give orders of magnitude. There is, for instance, no lower limit on the amplitude of signals but in most cases typical values have been quoted.

The mean circulation

Undoubtedly if a very long term average of the circulation of the ocean were obtainable a mean value would emerge. In much of the ocean the magnitude of the mean flow would be small. The strongest and best-defined mean flows in the ocean are found in the western boundary currents (e.g. the Gulf Stream and the Kuroshio) and in the deep flows between ocean basins (e.g. the outflow from the Norwegian Sea into the North Atlantic Ocean and the in and out flows between the Mediterranean Sea and the Atlantic Ocean).

Even these well-defined flows, although they may be energetic, are far from steady. The Gulf Stream varies its position and strength from day to day (Robinson 1971) and similar variability is found in other western boundary currents.

The direct measurement of mean flows in the deep ocean far removed from major currents has only recently been possible with the accumulation of very long direct current measurements. Schmitz (1976*a*, *b*, 1977) has computed the mean values from current meter records in the deep western North Atlantic. The records are several hundred days in length and from them

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it appears that estimates of the mean current begin to stabilize only after an averaging period of the order of 200–300 days.

In continental shelf seas the lowest-frequency circulation patterns are constrained by topography and affected by seasonal changes much more than is likely in the deep ocean. It is unlikely that the record-duration needed to define the mean is appreciably different from that found by Schmitz. There are additional problems in defining the mean circulation of shelf seas owing to the presence of the very strong tidal oscillations.

TABLE 1

motion	period	horizontal wavelength	vertical wavelength	typical amplitude cm/s	geographical area
mean flow	—	various	h†	0–10	all seas and oceans
climatic change	10 a	global	h	?	all seas and oceans
seasonal variability	1 a	global	h	?	all seas and oceans
mesoscale activity	30–100 d	100–500 km	h	5–100	all deep oceans
meteorological disturbances	aperiodic	basin wide in shelf seas	h	up to 100	shelf seas and upper layer of ocean
shelf and edge waves	2–10 d	10 km across slope 500 km along slope	h	5–10	on topographic features, e.g. seamounts, shelf edges, etc.
inertial oscillations	12 h/sin L† (0.5–5 d)	tens of kilometres	100 m	5–50	all seas and oceans
tidal oscillations	0.5–1 d	basin width	h	10–20, deep sea 10–200, shelf seas	all seas and oceans
internal waves	inertial period to 10 min	1 km	100 m	5	throughout the stratified ocean
surface waves	1–20 s	100 m	limited to upper 30 m of water column	up to 200	all seas and oceans

† L = latitude; h = water depth; a = year.

The overflows from deep channels between ocean basins may retain their identity far from the strait from which they issue. No direct measurements have been made which support the existence of such flow but they may be inferred from water mass properties and from sediment distributions. Roberts, Hogg, Bishop & Flewellen (1974), in investigating sediment ridges in the Rockall area, infer the presence of a mean southward flow of water in that area originating from the Norwegian Sea outflow. The flows may well retain some of the energetic characteristics of the contained channel flows.

Climatic change and seasonal variability

The very long period changes associated with the changes in world climate are not, as yet, identifiable in the relatively short records of oceanographic variables that are available and certainly not in the direct measurements of current. Large changes, with periods longer than a year but shorter than the periods of climatic change, are known. The best documented is that of the El Niño phenomenon off the coast of Peru which is connected with long-period, atmospherically coupled changes in the equatorial currents (Wyrtki 1975).

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Seasonal variations in the ocean are most readily observable in the upper ocean thermal structure and in the effect of runoff from the land on surface salinities. It is not unreasonable to suppose that these factors also affect circulation patterns in a seasonal way but the effects are for the most part small. A very striking example, however, is the effect of the seasonal monsoon winds of the Indian Ocean on the Somali current off the coast of east Africa. The Somali current (the Indian Ocean analogue of the Gulf Stream) reverses direction seasonally in response to the monsoon winds (Leetmaa & Truesdale 1972; Leetmaa 1973). This and the El Niño are, however, unique events and seasonal and climatic fluctuation are in general not nearly so dramatic.

Mesoscale variability

The earliest observations of deep currents in the ocean with the use of neutrally buoyant floats were expected to reveal the slow mean circulation pattern. The measurements, however, showed that velocities of many centimetres per second were common and, above all, that the flows were not unidirectional over large distances but rather showed a dominant spatial scale of several tens of kilometres (Crease 1962).

There was a gap of many years before full-scale experiments were mounted to investigate further these energetic mesoscale motions. The Russians conducted a series of Polygon experiments in various seas and oceans (Fofonoff 1976) culminating in a seven-month experiment in the North Atlantic in 1970 with 17 current-meter moorings in a 2° square (Brekhovskikh *et al.* 1971). This was followed in 1973 by the joint U.S.A./U.K. Mid-Ocean Dynamics Experiment (MODE-1). This latter experiment used both moored current meters and ship- and Sofar-tracked neutrally buoyant floats (MODE Scientific Council 1973; Wunsch 1976), together with a wide variety of other oceanographic instruments (Gould 1976). (The main results of the MODE-1 experiment are presented in Simmons *et al.* (in preparation).)

The observations in both the Polygon and MODE-1 experiments revealed features in the low-frequency circulation patterns with velocities as high as 50 cm/s in the upper ocean and with horizontal dimensions between 100 and 140 km. The term 'eddies' has been generally applied to such features and is in common use; however, the implication that the features are closed and eddy-like is not necessarily valid. In the deeper layers the eddies are found to have smaller horizontal dimensions and to be generally less energetic.

It is often difficult to visualize the structure of deep ocean eddies from the discrete and often randomly distributed current velocity vectors from both neutrally buoyant floats and current meters. A technique known as 'objective analysis' has been used by Freeland & Gould (1976) to produce a stream function map from these discrete vectors. Two examples of these maps are shown in figure 2. They represent the flow fields at 500 and 1500 m for a 3-day period during the MODE-1 experiment. The flow fields at the two levels are clearly different and an analysis by Freeland, Rhines & Rossby (1975) shows that the circulation pattern at 500 m and at the deeper levels progress westwards at different speeds, the lower levels having the higher phase velocities.

It is now becoming clear that a range of eddy-like features of different sizes and energetics exist in different parts of the deep ocean. The smallest and most energetic are the cyclonic and anticyclonic rings formed by the 'pinching off' of meanders in the Gulf Stream (Richardson, Cheney & Mantini 1977); some of these may penetrate into the interior of the ocean (Parker 1971) but their distribution seems to be restricted to certain well-defined areas. Larger but less energetic features have been identified in the interior of the Atlantic Ocean, for the most part by

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the isotherm displacements seen in hydrographic or expendable bathythermograph (x.b.t.) sections.

Data from x.b.ts have been used by Dantzler (1977) to identify the most energetic eddy regions of the North Atlantic. The results show large (factor of 4) changes in eddy potential energies over very small horizontal distances in some parts of the ocean. It is possible that these could be associated with major topographic features.

It is to be expected that in situations where energetic mid-ocean mesoscale features impinge upon mid-ocean ridges, sea mounts or continental slopes, the reduction in water depth would lead to an enhancement of velocities. An example of this has been documented by Freeland & Dow (1976) who monitored the track of Sofar floats as they executed a series of violent oscillations over the Blake Bahama outer rise. The floats made roughly circular tracks 50 km in diameter with periods of between 17 and $3\frac{1}{2}$ days (speeds of 12–30 cm/s), before following the depth contours to the south at speeds as high as 45 cm/s. The floats were at 2000 m depth.

It is unlikely that mesoscale eddies could penetrate onto the continental shelf without their energy being largely dissipated but there are examples of Gulf Stream rings penetrating well up the continental slope off the eastern seaboard of the United States (Bisagni 1976).

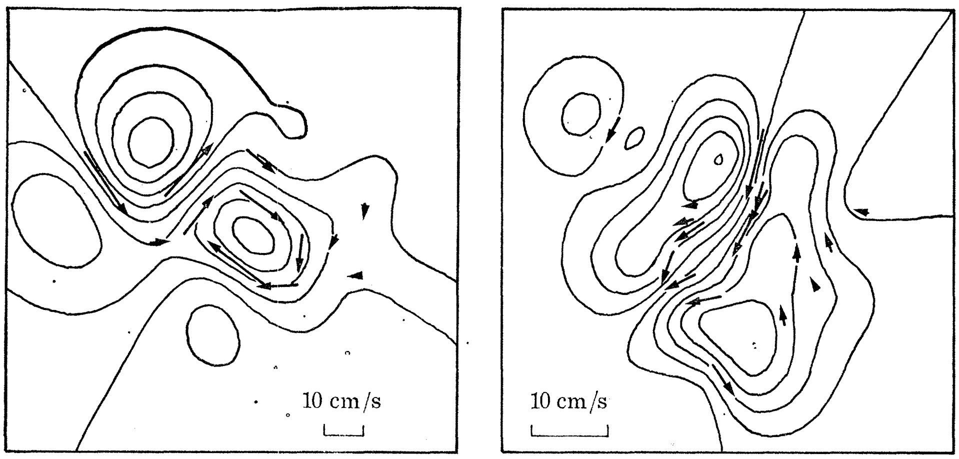

FIGURE 2. Streamfunction maps for a 3-day period of the MODE-1 experiment derived from current observations shown by arrows. Each box is 600 × 600 km. Data are from 500 m (left) with contour interval of 200 cm km s^{-1} and 1500 m (right) with contour interval of 50 cm km s^{-1}. Source data are from both current meters and neutrally buoyant floats, low-pass filtered and averaged over 3 days.

Meteorological disturbances

The passage of atmospheric depressions across areas of continental shelf can, by the action of pressure and wind stress on the sea surface, give rise to anomalously high (and low) water levels (storm surges). Associated with the elevation of the sea surface there are often large perturbations of the generally tidally dominated flow patterns. The mechanism of surge generation is quite well understood and from forecast meteorological conditions computer models can be used to predict both surge heights and current strengths. The validity of the model may be assessed by hindcasting the observed surge heights at coastal stations (Flather & Davies 1976).

There seem to be very few well-documented examples of directly measured current disturbances associated with meteorological forcing. Howarth (1975*b*) presents current data from the southern Irish Sea in which a departure of over 50 cm/s from the normal tidal flow is attributed to meteorological forcing. Caston (1976) documents currents measured in the

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southern North Sea and shows the correlation between flow and surface winds. The strong effect of local bottom topography is evident in this data.

R. A. Flather (personal communication) has compared measured currents in the North Sea during a severe storm in January 1976 with the predicted surge currents from his computer model. Good agreement has been found. It is clear that such meteorologically induced flows are of great importance in shelf seas but the number of cases in which adequate current measurements are available to test computer models is very small.

In addition to the direct forcing the passage of meteorological disturbances may initiate resonant motions trapped along depth discontinuities. The continental shelf edge is an obvious area in which these effects could be found and a considerable number of workers, particularly those involved in the field of coastal upwelling, have studied this phenomenon. Once again the number of examples of direct current measurements in which these oscillations are detectable is small; for the most part the trapped waves are seen in the spectra of sea levels and are forced by changes in the surface wind stress. Kundu, Allen & Smith (1975), Huyer *et al.* (1975) and Kundu & Allen (1976) have all investigated this phenomenon along the coasts of Oregon and Washington State where additional current measurements are available. Clarke (1977) has made an extensive review of all the evidence for the existence of edge waves.

In all cases the typical period of oscillation was in the range 3–6 days and, where measured, the amplitude of the meteorologically forced current was small (*ca.* 10–15 cm/s).

Inertial oscillations

Inertial oscillations can be regarded as the free response of the ocean to an impulsive force at the surface and controlled by the rotation of the Earth. The motions are for the most part circular, horizontal and with a characteristic period (in hours) $T = 12/\sin\lambda$, where λ is the angle of latitude.

In many cases the motions are energetic and since they have, at any position, a well-defined frequency which is in general well separated from tidal frequencies they have proved a subject of considerable interest when current records of a few days duration are available.

Webster (1968) in a review of the observations up to that date concluded that the phenomenon was transient and of very thin vertical extent and remarked that the generation mechanism was not probably related to the passage of storms. A considerable advance was made by the work of Pollard (1970) and Pollard & Millard (1970). In these papers a simple model of inertial current generation by a changing wind stress is used to hindcast the measured inertial oscillations on two buoys at which wind observations were also available.

The model predictions are in good general agreement with the observations. The intermittent nature of the inertial signal is reproduced with energetic oscillations persisting for between 2 and 5 days before decaying.

The model does not account for the relatively high level of inertial activity found at depth. Pollard analysed records of 200 days duration from depths between 10 and 2000 m in 2600 m. He found amplitudes of inertial oscillations were approximately uniform over a depth range of 50–500 m, that at 10 m the amplitude was approximately twice as large, and at 1000 and 2000 m approximately half as large. Amplitudes at the deepest two levels were 10 cm/s maximum and for 10 % of the time the amplitude exceeded 3–4 cm/s.

Perkins (1972) analysed records of approximately 50 days duration from a site in the western Mediterranean. Five records were obtained at levels every 500 m from 200 to 2200 m.

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Persistent oscillations at a frequency 3 % above the local inertial value were found at all levels. Almost all the energy had a clockwise polarization. Energies varied very little over the deepest levels but were much less (factor of 4) than at 200 m. This simple deep structure was attributed to the very weak stratification below about 500 m.

Recent techniques of current profiling (Sanford 1971) have led to more detailed studies of the vertical structure of inertial oscillations (Leaman & Sanford 1975). Vertical profiles obtained one-half an inertial period apart are averaged and then subtracted from the original profiles. This leads to the elimination of lower wavenumber signals and enables a study of the wavenumber spectrum of the inertial oscillations to be made. The calculations have been done in terms of 'stretched' coordinates since the local Brunt–Väisälä stability frequency has the effect of changing the vertical scale and also the amplitude of the signals.

Analysis of several profiles from the western North Atlantic shows again a dominance of clockwise polarization. The wavenumber spectrum has a maximum energy density at low wavenumbers corresponding to stretched wave lengths between 100 and 500 m and at higher wavenumbers a spectral slope of -2.5 on a log–log plot.

A further analysis of these data is presented by Leaman (1976). He concludes that the behaviour of inertial oscillations in the deep ocean is consistent with internal wave theory. The data show a downward flux of energy from the upper ocean at the inertial frequency. This flux persists into the deepest parts of the profiles (up to 5000 m) but its magnitude cannot be determined accurately owing to the uncertainty in measuring the exact frequency of the observed oscillations.

Inertial oscillations certainly exist in continental shelf seas but their occurrence is less well documented than for deep water. This may partly be due to the fact that shallow water records are in general dominated by tidal signals which may mask the inertial oscillations.

Currents of tidal period

The astronomical forcing of the attraction of the sun and moon induce surface elevations and currents in the ocean at well defined frequencies. These are almost entirely in the diurnal (period 24–25 h) and semidiurnal (period 12–12½ h) bands.

Energy at these tidal frequencies if found in virtually all current records and in shelf seas the tides represent by far the most energetic motions. In shelf seas the pattern of tidal currents is usually fairly simple and is controlled by the geometry of the adjacent land mass and the variations in bottom depth.

Perhaps the most detailed analysis of tidal currents at a fixed point in shallow water is that of Pugh & Vassie (1976). They have analysed year-long data of both tidal heights and current values from the Inner Dowsing Light tower. The current meter measurements are from approximately mid-depth in 20 m of water.

The analysis of the currents reveals that over 96 % of the variance is accounted for by the principal tidal constituents. The remaining 4 % is shared between the higher-frequency shallow-water tidal components and the residual aperiodic motions. The mean current over the total duration of the record has a magnitude of only 4 cm/s. No energy is seen in the current record at inertial frequencies.

In the Pugh & Vassie observations particularly, the currents behave in a regular and predictable way and are almost entirely deterministic. The currents in the deep ocean at tidal frequencies are quite different.

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To date the most detailed analysis of tidal currents in the deep ocean is that by Hendry (1977). He has taken both current and temperature observations from the moored instruments in the MODE-1 experiment to investigate the vertical and horizontal structure of motions in the semidiurnal tidal frequency band.

Tidal currents in the deep ocean may be considered to have a component which is uniform and in phase over the entire water column (the barotropic component) and a component which varies with depth (baroclinic component). The baroclinic component can be resolved into motions with particular vertical modal structures having 1, 2, 3,... zero crossings (the first, second, third, etc., baroclinic modes). The depths of the zero crossings are a function of the vertical stratification of the water column.

Hendry analysed data from the MODE-1 field experiment to investigate the distribution of energy between the barotropic and baroclinic (internal) tide in the semidiurnal band of frequencies (periods between 12.00 and 12.86 h).

The tidal currents did not exceed an amplitude of *ca.* 5 cm/s at any depth and both the barotropic and baroclinic currents were found to be around 1 cm/s.

The baroclinic tides have short horizontal wavelengths and are generated by the interaction of the barotropic tide with the slopes of major features such as mid-ocean ridges and continental slopes. Their dependence on the stratification of the water column and their propagation into the interior of the ocean from a variety of source regions tends to make the baroclinic tide non-deterministic.

Internal waves

The internal tides are a particular case of internal waves. These can be shown to exist over a range of frequencies bounded at the lower end by the local inertial frequency and at the high frequency end by the Brunt–Väisälä frequency, N.

Values of N in the ocean vary between those corresponding to periods of 2–3 h in the deep ocean to a few minutes in the thermocline.

Freely propagating internal waves can only exist in a stratified ocean and within the specified range of frequencies. Although their vertical amplitude may reach values of tens of metres the associated horizontal current velocities, except perhaps in the tidal band, are generally no more than 1 or 2 cm/s.

A very detailed study of internal wave activity in the deep ocean was carried out in 1973 (Briscoe 1975) and involved the measurement of current velocity and temperatures along the legs of a tetrahedral, three-legged mooring in the Sargasso Sea. The horizontal kinetic energies at various depths in the internal wave band were found to be proportional to the Brunt–Väisälä frequency at the depth of measurement, i.e. the most energetic internal wave motions exist where the stratification of the water column is strongest.

The method of generation of internal wave activity in the ocean is most probably a combination of forcing induced at the sea surface and the interaction of currents with topographic features. Internal wave activity has characteristic signatures which can be recognized by the relation between particular spectral qualities. This ordered nature distinguishes the wave-like activity from random motions which exist in the ocean both in the internal wave band and at higher frequencies due to general turbulent motions. These are of such low amplitude as to be of no importance in most practical offshore activities.

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Surface waves

In the uppermost layers of all oceans and seas, surface waves may provide the most energetic current component. The direct measurement of the wave orbital velocities in the open ocean poses considerable difficulties not only in the design of instruments which can adequately measure the high frequency motions but also in the design of stable platforms from which the measurements can be made.

The presence of these energetic, high-frequency motions in the upper layers has been found to have a serious effect on the measured values of current velocity at all depths on a surface following buoy when conventional rotor/vane current meters are used (Gould & Sambuco 1975). As a general rule, unless current meters with an adequate high-frequency response are available, the measurement of currents on moorings using a surface following buoy should be avoided. This problem means that there are few reliable measurements of currents in the uppermost, wave-affected layers of the ocean and results in a lack of adequate statistics of the currents likely to be encountered by surface piercing structures.

Particular problems of continental margins

The range of ocean depths between 200 and 2000 m occupy the boundary between the distinctly different continental shelf and deep ocean current régimes and as such are subject to motions characteristic of both zones.

The continental slopes may be subject, for example, to both the energetic tidal motions originating on the shelf and the intermittent and unpredictable inertial and mesoscale motions from the deep sea. They may additionally, in some specific areas, be swept by deep overflowing currents from channels between ocean basins. The zone is thus one of energetic but largely unpredictable (except in a statistical sense) currents. The prospects of offshore engineering activity in the 200–2000 m depth zone will probably highlight the lack of sufficiently long current records from this zone of the ocean.

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