4. Circulation

1. Key points

i. **Introduction**

The Atlantic Meridional Overturning Circulation (AMOC) gives warm water flow ('North Atlantic Drift') past the west of the UK, strongly influencing UK climate by warming the prevailing westerly airflow. Circulation is important to distributions of salt, of deep-ocean heat and hence regional climate, of pollutants and of many species carried by the flow during their lifecycle. Currents affect offshore operations and habitats. In UK shelf seas, the instantaneous observed current comprises an important tidal contribution, and contributions due to winds and to flows driven by differences in density (arising from seasonal heating and salinity differences between locations). 'Residual' flow (after averaging out oscillatory tidal flow) is dominated in many areas by flows forced by winds and by differences in density. Tides, density and winds all change on various timescales, so that observed and residual flows can be very variable. Wind forcing is the most variable factor; transports of water in one storm can be significant relative to a year's integrated transport (e.g. in the Irish Sea). Currents vary strongly with location. Notable persistent flows are polewards along the upper continental slope; from the Atlantic onto the continental shelf around Scotland and into the northern North Sea; and northwards through the Irish Sea to form the relatively fresh northward Scottish Coastal Current west of mainland Scotland (see sections 2.2, 2.6, 4).

ii. **How has the assessment been undertaken?**

Long-term circulation in UK waters has mostly been inferred from distributions of tracers, tracks of drifters and floats, or from numerical hydrodynamic models. Data also come from current-meter measurements (there are a few long-term mooring arrays) and from submarine cables. Circulation can be inferred from hydrographic sections, for components with time-scales longer than a day. HF radar gives spatial coverage for surface currents within a limited range (section 2.3). There are no criteria for status and there is no baseline.

iii. **Current and likely future status of circulation**

- Five sections from 1957 to 2004 suggest AMOC decline but this is within the range of large variability on time-scales of weeks to months. An overall trend has not been determined from the continuous measurements begun in 2004 (new data since *Charting Progress*; Defra, 2005). (section 4.1).
- Deep outflows of cold water from the Nordic seas are likewise too variable to infer any overall trend (section 4.1).
- Strong North Atlantic flow eastwards towards the UK may correlate with positive North Atlantic Oscillation (NAO) Index (i.e. prevailing westerly winds). Enhanced along-slope current around the UK may correlate with a negative NAO Index (section 4.1).
- Climate models' consensus makes it very likely that AMOC will decrease over the next century, but not 'shut down' completely (section 5).
- Similar spatial and temporal variability (arising from complex topography and variable forcing) is likely in future (sections 4.2 to 4.7, 5).

iv. **What has driven change?**

Circulation is directly dependent on forcing by tides, winds and spatial differences in density.

v. **What are the uncertainties?**

Relative to spatial variability, measurements are generally sparse. This lends importance to developing models. Temporal variability complicates the inference of changed conditions unless several years' data are available. Instabilities generating meanders and eddies at irregular intervals are a large part of variability over the continental slope and nearby ocean.

vi. **Forward look**

There is a need for long-term data to elucidate climate-change signals from background variability. The RAPID monitoring to 2014 needs complementary measurements, especially at higher latitudes, to help understand how changes in the AMOC are relayed from place to place and possibly to establish proxies for easier monitoring. In UK shelf seas, measurements are sparse relative to currents' variability; reliance will be placed on models for most distributional information. In view of the importance of currents, better prediction of short-term variability in circulation is needed; this entails model validation and development of observational networks.

2. Introduction

2.1 Types of current

Tides (surface elevations and currents or 'streams') are generated by the varying gravitational attraction of the Moon and the Sun. In the North Atlantic Ocean, tides are predominantly semi-diurnal (two tides a day); in the relatively shallow (< 200 m) UK shelf seas, tides are amplified to give a macro-tidal (> 4 m) regime around much of the UK (particularly large elevations, > 10 m range, occur in the Bristol Channel). Strong tidal currents occur in the southern North Sea, in some areas of the Irish Sea, and in constrictions between islands and around headlands, for example the Pentland Firth.

Meteorologically forced 'surge' currents are due to variations in wind stress and atmospheric pressure. Wind stress is most effective in shallow water whereas the pressure effect is independent of depth. Surge currents have timescales of hours to days according to storm duration, water depth and the extent of the storm. Winds that drive surface waters offshore may induce 'upwelling' of replacement water from below.

Density currents are driven by density gradients due to changes in temperature and/or salinity, arising from the net flux of heat through the sea surface, freshwater inputs from rivers and the atmosphere, and different mixing regimes according to water depth and tidal-current strength. Intense cooling and hence dense-water formation may induce local sinking or 'convection'.

2.2 Circulation

The net movement of water, the circulation, is driven by 'residual' currents which arise from a combination of net tidal action, mean meteorological forcing and the mean density distribution. (Upwelling and convection are components of vertical or 'overturning' rather than lateral circulation.) Traditionally, there has been a tendency to think of circulation as a smooth, wide constant flow, perhaps because such characteristics are most amenable to measurement (see Section 2.3). In reality, and especially in UK waters, most elements of circulation vary strongly on short (daily and monthly) timescales or short spatial scales.

2.2.1 Short-term circulation

Tidal currents are primarily oscillatory and contribute little to daily mean circulation, except where currents are strong and spatially variable, such as the southern North Sea or headlands such as Portland Bill.

During winter in UK shelf seas, the net movement is likely to be determined by the last storm, because surge currents typically exceed density currents and daily-mean tidal currents. Thus daily or monthly mean circulation may even be the reverse of the long-term pattern. An example is flow through the North Channel (Knight and Howarth, 1999); the two largest daily mean flows in February 1994 amounted to 20% or more of the year-long transport along the Channel.

[Animation of flows with reversal through a tidal cycle in Liverpool Bay; in UKMAR OPEG as LivBayU.gif. Courtesy of A. Lane, National Oceanography Centre.]

[Animation of surge levels around UK for 8-9 November 2007 storm (surge) event; in UKMAR OPEG as surge9nov.mpg. Courtesy of K Horsburgh, National Oceanography Centre.]

2.2.2 Seasonal mean circulation

The seasonal mean circulation is mainly due to the strong seasonality in surge and density currents: storms mainly occur in winter, river discharge has an annual cycle and solar input varies seasonally.

Each year, from about May to October, large areas of UK waters stratify as surface waters are warmed. Below the (relatively sharp) seasonal pycnocline, colder bottom water remains from the previous winter. However, the water column remains vertically mixed by tidal stirring all year where the water depth is relatively shallow (typically less than 50 m and nearer to shore) and/or depth-averaged tidal currents are large (Simpson and Hunter, 1974). Between the stratified and mixed areas is typically a sharp horizontal temperature (and/or salinity) gradient with a horizontal scale O(10 km). Because the cold bottom water is largely static (friction tends to bring the velocity to zero at the sea bed), the corresponding density gradient is expected to drive a near surface geostrophic jet or ‘thermal wind’ above the region of maximum horizontal density gradients; the flow direction is cyclonic with the dense bottom water to the left of the direction of flow (northern hemisphere; Hill et al., 2008).

For a typical horizontal density gradient 0.4 kg/m^3 per 10 km, the jet velocity is 0.2 m/s at 50 m above the bed resulting in a typical transport of the order of 10^5 m^3/s, an important (albeit localized) contribution to persistent transports from late spring to autumn (Hill et al., 2008).

Such flows can transport water over many hundreds of kilometres in areas of the North Sea, Celtic Sea and Irish Sea (see Section 4). The timing of the onset of this seasonal circulation in April or May is dependent on wind mixing, surface heat fluxes and freshwater input and may vary between years by up to a month (Brown et al., 1999, 2003). An animation of the evolution of surface to seabed temperature differences, and hence thermal fronts, is in UKMAR OPEG as dtmp.HRCSrun017.avi . Courtesy of J. Holt, NOC.

The seasonal mean circulation varies through depth: jets are stronger near the surface, and the relative contributions from surge or density currents may vary with depth. This is illustrated in section 4.5 (see Figure 4.9 showing an ‘estuarine / coastal’ type circulation: flow near the bed is towards the coast and flow near the surface is away from the coast). The prime driving force is density for both the outflow and the inflow, although the near-surface is more affected by the wind (M.J. Howarth, NOC, personal communication, 2009).

2.2.3 Long-term mean circulation

When averaged over some years, the long-term or ‘climatological’ mean circulation indicates some persistent features in UK waters (see Figure 4.1). However, there are large uncertainties in estimating the magnitude of mean circulation, and most regions have significant inter-annual variation (see sections 2.3 and 4).

‘Flushing time’ is a concept describing the average time needed to completely replace the waters in a region. However, it depends on the circulation and on the amount of mixing; hence flushing time is not easy to estimate and hides large local variations. For example, the volume of water that enters and leaves the North Sea each year is approximately the entire volume of the North Sea basin (Huthnance, 1997). However, most of this (red and blue arrows in Figure 4.1) only passes through the northern North Sea, which therefore has a flushing time of ~ 1 year or less, whereas flushing time in the central North Sea is thought to be several years.

2.3 Measuring circulation

Long time-series of observed currents are sparse, hindering the determination of long-term circulation and its variability. Long-term circulation in UK waters has mostly been inferred from distributions of tracers (e.g. salinity or radionuclides), tracks of drifters and floats, or from numerical hydrodynamic models, optimised with any available observations.

Observational data on circulation come from current-meter measurements, drifting buoys and floats, submarine and telephone cables (to measure induced voltages across channels) and distributions of ‘tracers’ such as salinity and radionuclides, for example, caesium-137 and technetium-99 (Kershaw et al., 2004). Circulation can also be inferred from hydrographic sections, albeit only for components with time-scales longer than a day and leaving an arbitrary depth-independent component to be determined by other constraints. However, accurate estimation from measurements needs strong and persistent circulation. In the presence of strongly time-varying currents, it is difficult to infer a relatively weak long-term circulation from current meters and Acoustic Doppler Current Profilers (ADCPs), whose point locations may also be unrepresentative. HF radar gives fuller spatial coverage for surface currents within a limited range (typically 50 to 100 km) of its transmitters and receivers ashore. Wider coverage of surface currents can be given by satellites: (1) altimetry can measure surface slopes to infer surface currents, however, measurements are degraded within O(10 km) of the coast and are everywhere infrequent so that only variability on time-scales of weeks is resolved; (2) the movement of features in colour or infra-red imagery can be tracked – evidently the outcome is degraded by the unknown development of the features themselves and by the infrequency of clear images. The motion of floats and drifters is often difficult to interpret in continental shelf seas because of the usual short time of deployment and observation; surface floats are also partly driven by the wind, so that they do not simply track the bulk current. Tracers can help to determine circulation patterns rather than current speed.

Descriptions of the monitoring networks which regularly measure currents and circulation are given in the UK Directory of Marine Observing Systems (www.ukdmos.org). These include RAPID arrays for the North Atlantic circulation, moorings on the West Shetland slope for Atlantic water transport past northern Scotland, a mooring in Tiree Passage, two moorings and HF radar in Liverpool Bay, but little else on a regular basis. This is a sparse array relative to the spatial scales of circulation patterns.

2.4 Models

Charting Progress (Defra, 2005) expressed caution regarding the use of numerical models to predict circulation; inevitably, different models may best reproduce (hindcast) observations in different contexts (Jones, 2002). Insufficient resolution and unsuitable advection schemes were a factor in some comparisons, for example Smith et al. (1996) and NOMADS2 (Delhez et al., 2004); in deeper waters and over steep slopes, effects due to density differences can be important and difficult to model.

More recently, Holt et al. (2005) have assessed uncertainties in a coupled hydrodynamic–ecosystem model of the northwest European continental shelf. The three-dimensional baroclinic circulation model POLCOMS had ~ 7 km horizontal resolution and 20 ‘s-levels’ in the vertical. POLCOMS was coupled with the European Regional Seas Ecosystem Model representing the functioning of the ecosystem with 52 state variables. The period August 1988 to October 1989 of the North Sea Project was simulated. Tidal currents and elevation, and sea surface temperature, were well modelled, with root mean square (RMS) errors of less than 0.4 standard deviations of the data. However, residual current speed and salinity had RMS errors similar to the standard deviation of the data. Improving the comparison for residual currents and salinity requires improvements in the model (forcing and formulation) and in observational data (quantity and quality). Nevertheless, the resolution was found sufficient to model complex biophysical interactions in the horizontal (e.g. enhanced production at fronts) and in the vertical (e.g. mid-water production modulated by the spring–neap cycle). The assessment also suggests what data assimilation might prove most fruitful.

2.5 The significance of circulation

The heat capacity of the seas is large relative to the atmosphere, thus the ocean can store and transport large amounts of heat; the world’s ocean circulation is critical in the global climate system. A meridional (tropics to poles) transport of energy is required for the Earth system to be in global radiative balance; some 30% to 50% of this meridional energy transport is in ocean currents at mid-latitudes and a higher proportion at lower latitudes (Bryden and Imawaki, 2001).

In particular, the Atlantic Meridional Overturning Circulation (AMOC) comprises the large-scale surface and subsurface circulation of the Atlantic basin, including regional currents such as the Gulf Stream and the North Atlantic Drift. The system is responsible for transporting large amounts of heat into the North Atlantic region (Ganachaud and Wunsch, 2000), much of which is transferred to the atmosphere, and then advected into NW Europe. This relatively warm surface water enhances the inherent moderating effect that the North Atlantic Ocean has on the climate of NW Europe and the sub-Arctic. Climate model simulations suggest that without the presence of the AMOC, mean annual temperatures in the UK, for example, may be 2 to 4 °C cooler than is otherwise the case, in particular with much colder winters (Vellinga and Wood, 2002).

The potential of the AMOC to undergo a partial or (less likely) a total slowdown is of serious concern given the capacity for the associated cooling to offset some (or in the extreme case, most) of the projected global warming for the UK and wider North Atlantic region. This has obvious direct societal implications with respect to the planning and implementation of climate change adaptive measures. Implications for the UK's seas, and those directly adjacent, are equally serious, with changes in circulation patterns and heat and salinity transport likely to have marked biological effects (see Schmittner, 2005 for example).

Circulation is an agent for flushing of sea areas (and pollutants). It brings Atlantic water on to the shelf; the extent of this influence is limited regarding heat content (hence temperature), but more extensive for salinity and other constituents less subject to exchange with the atmosphere. Ultimately, Atlantic water is the main source of UK shelf-sea nutrients, for example. The circulation also advects carbon from the shelf to the Atlantic.

Locally, cooling water may be a significant contributor to circulation and may exploit coastal flows to disperse the heat. A modern coastal power station discharges heat of the order of 3 GW; assuming excess temperature of 2 °C, this represents a flow of order 400 m^3/s, larger than any UK river's average and comparable with alongshore tidal flow in the first kilometre offshore.

Circulation patterns control the overall movement and distribution of passive objects (eggs, larvae, nutrients, contaminants, flotsam, sediments). Hátún et al. (2009) attributed biogeographical shifts in the NE Atlantic to exchanges of sub-arctic and sub-tropical water masses. Some species exploit the circulation, for example herring, in order to transport larvae from spawning grounds to nursery areas (Turrell, 1992). The flow off NE England provides a direct pathway for material and fish larvae from coastal regions to the northern Dogger Bank and central North Sea (Brown et al., 1999). Density-driven currents provide a continuous transport route from the French coastal region via the Celtic shelf and west of Ireland to the Scottish shelf, potentially a 'conveyor belt' for contaminants and plankton (Hill et al., 2008). On a smaller scale, dispersal of herring larvae in the Blackwater Estuary is dependent on circulation in the area (Fox and Aldridge, 2000). To summarise, advection of marine nutrients, plankton and animals is a critical factor in the lifecycles of many marine species. Movements of objects depend on their density – if neutrally buoyant or dissolved, they move with the water circulation; if particulate or heavier than water, they tend to sink and move less far, with a bias towards the direction of fastest flow; floating objects are driven by winds as well as the water circulation. Areas of retention (material not being advected away) are often characterised by deposition of soft organic sediment, while frontal regions are associated with greater productivity. Frontal boundaries between eco-regions are important when considering the division of shallow seas into biologically relevant management units (applying an ecosystem approach to the sustainable use of marine resources; Hill et al., 2008).

3. Progress since *Charting Progress*

In the open Atlantic Ocean, there are now more Argo profiling floats (which give an indication of circulation from their changing reporting positions), and the RAPID array (especially across 26° N) is giving an estimate of the time-varying meridional overturning circulation. In addition to the 26° N array, RAPID is also funding field work at 38° to 39° N southwest of Cape Cod (six bottom pressure recorders in collaboration with Woods Hole Oceanographic Institution), and at 42° to 43° N off Nova Scotia (moorings comprise combinations of bottom pressure recorders, ADCPs and CTD sensors in partnership with Bedford Institute of Oceanography, Canada). These latter arrays aim to record the variability of bottom pressure across the Atlantic deep western boundary, to provide a robust estimate of the variability of the AMOC itself (Bingham and Hughes, 2008). One-year time series of bottom pressure from the Nova Scotia array will be available by the end of 2009.

The Irish Sea Observatory was established in the eastern Irish Sea in 2002; there are now several years of data to provide estimates of interannual variability; a second site (also with current measurements) was started in 2005; HF radar for surface currents has also been running since 2005. HF radar is able to monitor surface currents hourly or more frequently, with spatial resolution of the order of 1 km out to a range of the order of 100 km, but few systems are operational. Drifters have provided further insight into aspects of shelf-sea circulation as described by Hill et al. (2008). Gliders are a recent development for hydrography with less dependence on ships; they are not yet in routine use but can give an indication of circulation from their changing reporting positions.

Model development has continued, in particular with finer resolution ~ 1.8 km over the UK shelf area (out to the shelf break); this is fine enough to resolve features associated with stratification in summer and in regions of freshwater influence. Forecasts using a 3-D hydrodynamic model (POLCOMS) are now operational at the Met Office National Centre for Ocean Forecasting (NCOF). The UK shelf area is also modelled by several other countries. Models which have been published and well validated are those of MUMM (Brussels, Belgium), BSH (Hamburg, Germany), Ifremer (Brest, France), IMR (Bergen, Norway), Mohid (Portugal) and the Marine Institute (Galway, Ireland). Most of these are run operationally using data assimilation. Nevertheless, strong variability in currents (spatial and temporal) presents a modelling challenge and makes them a poor indicator of any trends in the state of UK seas.

4. Presentation of the evidence

4.1 North Atlantic circulation

The AMOC (sections 2.5, 3) has been estimated from a hydrographic section along 26° N completed in 2004. This updated ship-based estimate of AMOC strength has been compared with previous estimates spanning the past five decades (Bryden et al., 2005). The analysis, based on six-week ‘snapshots’ of the circulation obtained during each cruise, indicates that AMOC strength has declined by around 30% since 1957 (Figure 4.2).

Closer inspection of the circulatory components that comprise the AMOC reveals little change in the actual northwards transport of water associated with the Gulf Stream at the ocean surface (in 1957 to 2004), but an enhanced southwards recirculation of water in the upper 1000 m of the ocean by the subtropical gyre. The effect is to reduce the amount of water entrained into the North Atlantic Drift. The AMOC deep returning limb, comprising dense cold water formed in the Arctic and flowing southwards, is found at 26° N to have declined by ~ 8 Sverdrups (Sv; 1 Sv = 10^6 m3/s; Table 4.1), further implying a slow-down in the overall circulation.

Following the deployment of the RAPID mooring arrays, continuous daily estimates of the AMOC strength at 26° N are now available from spring 2004. Detailed analysis of the first year’s full time series, April 2004 to April 2005, reveals an observed annual mean AMOC strength of 18.7 Sv (Cunningham et al., 2007; Kanzow et al., 2007). Also evident are sizeable high-frequency variations (the standard deviation of the daily series is 5.6 Sv). An extended time series spans 3.5 years (April 2004 to September 2007; Figure 4.3); this shows similar levels of variability and an annual mean AMOC strength for the entire period of 18.5 Sv (± 4.9 Sv). Given the brevity of the time series to date, no attempt is made to comment on any linear trend.

Figure 4.1 illustrates the broad circulation of surface waters around the UK. Warm, salty Atlantic waters flow along the continental slope and further to the west (shown in Figure 4.1 by red arrows). In addition to the surface circulation, deeper flows return water to the Atlantic from Nordic seas; cold dense bottom water ultimately forms the main return limb of the AMOC. Deep 'overflow' water locations include Denmark Strait and the Faroe Bank Channel (prior to this the deep water flows south-westwards through the Faroe-Shetland Channel and is diverted north-westward by the Wyville-Thomson Ridge). There are also intermittent overflows across the Wyville-Thomson Ridge.

Some regularly updated time series describe these outflows of deep water from Nordic seas into the subarctic Atlantic basin. Outflow through the Faroe Bank Channel contributes approximately 2 Sv; updated measurements here are provided by Østerhus et al. (2008). These data are continuous from 1995 to 2005 and reveal sizeable inter- (and intra-) annual variability but do not point to any secular change indicative of a major change in AMOC strength.

A slightly larger flux of cold, dense deep water exits the Greenland-Iceland-Nordic Seas via the Denmark Strait, east of Greenland and to the west of Iceland. New data sets assembled here include those by Macrander et al. (2005), based on measurements made at the narrowest point of the channel, and a lengthier record maintained by Dickson et al. (2008) slightly further south. The Macrander et al. (2005) analysis reveals a 20% decrease in deep-water transport through the period 1999 to 2003. The Dickson et al. (2008) record encompasses data from 1996 to 2005 and reveals sizable, additional interannual variability around the Macrander et al. (2005) feature, but no overall trend in transport. The mean flux estimate of deep water transport through the Denmark Strait during the 1996 to 2005 period was 4.0 Sv (± 0.4 Sv).

Interannual changes in the North Atlantic Current and the Subtropical Gyre transport during 1992 to 2002 were found by Pingree (2002) to correlate with the winter NAO Index. Pingree found maximum flow conditions in 1995 and 2000 when NAO was positive and at a minimum in 1996 to 1998 with NAO-negative winters. Conversely, years of extreme negative winter NAO Index correlated with enhanced poleward flow and anomalous winter warming along the west European continental slope (see section 4.7), as measured in 1990, 1996, 1998 and 2001. One explanation is that negative NAO Index associated with weak westerlies reduces the wind-stress curl driving the Subtropical Gyre; in turn, geostrophic balance in the Gyre implies a raising of sea-level at the southern end of the west European continental slope, driving enhanced poleward flow along the slope (Pingree, 2002). An alternative explanation was proposed by Hatun et al. (2005) and is backed up by the observations of warmer water along the shelf edge (Holliday et al., 2008): that tropical gyre water replaces the Subtropical Gyre as the gyre weakens and moves westward. The two mechanisms can work together.

4.2 Circulation in UK waters

As discussed in section 2.1, circulation is variable in time and space and it is therefore difficult to describe any generally persistent circulation patterns in UK waters. There are only a few regions where the long-term circulation has been convincingly measured (usually from the distribution of tracers), for example the north-eastward flow of the North Atlantic to the west of Ireland and Scotland, some aspects of flow in the North Sea, the north-eastward flow from Dover Strait into the North Sea and the mean flow northwards through the Irish Sea (see Figure 4.1). UK shelf-sea circulation is strongly affected by density-driven coastal currents and jets (section 2.2) and by winds which can lead to significant changes and even a reversal of the general pattern for short periods.

Figure 4.4 shows time-mean circulation from the CS3X model used for UK operational storm-surge forecasting; it is two-dimensional (depth-integrated) with a resolution of 1/6° longitude by 1/9° latitude. The time-mean is from 1st September 1957 to 31st August 2002; the model is forced by ERA-40 (the duration and forcing are updates on *Charting Progress*; Defra, 2005). The scale picks out weak, interesting structures and strong, localised currents. However, the model cannot show density-driven flows, and lacks forcing by the oceanic steric height which gives poleward forcing at shelf-sea depths; this model's Celtic Sea time-mean flow to the south-east disagrees with other evidence.

Seasonal circulation is shown in Figures 4.5 and 4.6. Figure 4.5 uses a 45-year run of the POLCOMS (3-D) model with a resolution of ~ 12 km (which has been tested by comparison with UK LOIS Shelf Edge Study drifters for 95/96; see for example, Burrows et al., 1999). Figure 4.6 shows three-month averages of near-surface currents in 2001, using POLCOMS (3-D) with a resolution of ~ 1.8 km and

32 vertical ‘levels’ (Holt and Proctor, 2008). The figures show year-round mean flows locally around prominent headlands (and the Norwegian Coastal Current is present all year). The simulated mean flows are least in the period January to March. Other seasons show additional, typically filamentary flows, especially in the English Channel, Irish Sea and west and north of Scotland (and a stronger Norwegian Coastal Current). These flows are typically associated with the boundaries of seasonally-stratified waters (Hill et al., 2008).

Transports as found by 154 satellite tracked drifting buoys (mostly drogued at 20 to 30 m below the sea surface) between 1994 and 2005 are illustrated in Figure 4.7. These show some concentration and persistence near tidal-mixing fronts (see sub-chapter 2: Sea Temperature and Salinity; also Hill et al., 2008) as reinforced by superposition of the summary tracks on modelled temperature gradients (Figure 4.8).

4.3 North Sea

More than 1 Sv of Atlantic-origin water flows into the northern North Sea, but does not penetrate great distances south. Much flows along the western slope of the Norwegian Trench, recirculates in the Skagerrak and flows out along the eastern side of the Trench underneath the Norwegian Coastal Current (NCC). A smaller inflow of mixed Atlantic and shelf water (including some from the Scottish Coastal Current, see section 4.7) enters around Shetland and between Shetland and Orkney. However, most of this flow is guided eastwards to the Norwegian Trench by the ~ 100 m-depth contour; only a small part flows southwards along the coast of Scotland and England. Less than 10% of inflow to the North Sea enters via the English Channel. Thus most of the transport in the circulation is concentrated in the northern part of the North Sea and in the region of the Norwegian Trench; the one main outflow from the North Sea is ~ 1.3 to 1.8 Sv along the eastern side of the Norwegian Trench (Howarth, 2001). Holliday and Reid (2001) correlated increased oceanic inflows into the North Sea, in 1988 and 1998, positive NAO and strong northward transport of anomalously warm water in Rockall Trough (section 4.7). Although there may be linkage by the slope current and/or local winds, however, causality was not determined.

Tides enter from the Atlantic Ocean, primarily north of Scotland, and progress anticlockwise around the North Sea. Thus tides form the dominant motion in the western and southern parts of the North Sea around to Denmark. Tidally-generated residual currents are relatively small, but are responsible for significant mean circulation in the western and southern parts, especially locally around sandbanks and other features. Wind is the dominant source of energy in the northern and eastern parts (Rodhe, 1998). Wind-driven currents are induced by mostly south-westerly and westerly winds. Surges travel anticlockwise like the tides: southwards along the UK coast and then north-eastwards along the coast of continental Europe. Density-driven currents are important locally in outflows from estuaries (and from the Baltic), and in association with features of summer stratification.

The resulting overall pattern of the mean circulation in the North Sea is broadly anticlockwise around the coasts, with weak and varied circulation in the centre. The mean coastal flow is southward past Scotland and England, into the Southern Bight where there are inputs of salty water through Dover Strait and of fresh water down the main rivers, and on to the German Bight, flowing northward past Denmark in the Jutland current to join the Norwegian Coastal Current in the Skagerrak (Rodhe, 1998; Howarth, 2001). However, this broadly anticlockwise circulation can be reversed by easterly winds (occurring mostly in spring and summer).

Most of the central and northern North Sea becomes thermally stratified during April/May, due to increasing solar heat input, with a well-mixed layer about 30 to 40 m deep (Howarth, 2001). In autumn, surface heat loss causes the surface mixed layer to cool and deepen until the bottom is reached in October-December. In contrast, tidal energy in southern and shallow coastal regions is strong enough to keep the water column well mixed most of the year. Some coastal regions stratify because of freshwater river discharge; the fresher water tends to form a thin surface layer, up to about 30 km wide, which stays close to the coast (Howarth, 2001). Off Sweden and Norway, the Baltic Sea outflow forms low-salinity water in the Norwegian Coastal Current, which is typically broader. A summer front in the central North Sea separates the thermally-stratified water to the north from the well-mixed water to the south. The front (Figure 4.8) is off Flamborough Head, bifurcates around Dogger Bank and passes to the north of the Frisian Islands (Howarth, 2001). Some fronts in the southern North Sea are related to freshwater from rivers, but most are tidal fronts (Rodhe, 1998).

An associated persistent and narrow (10 to 15 km) near-surface flow extends continuously for ~ 500 km along the ~ 40 m depth contour from the Firth of Forth to Dogger Bank (Figures 4.6, 4.8). It is an example of flow associated with a strong bottom front bounding dense bottom water isolated below the summer thermocline (section 2.2; Hill et al., 2008). This flow is the main component of the (otherwise very weak) summer circulation in the central North Sea.

The flushing time, for the complete renewal of the water, is about one to three years (Simpson, 1998). *Charting Progress* (Defra, 2005) tabulates mean transport across sections in the North Sea for the period 1987 to 1993 from three numerical models (typical values 1 to 1.6 Sv into and out of the northern North Sea at 59.5° N).

4.4 English Channel and Celtic Sea, including the Bristol Channel

Along the English Channel, mean circulation is from west to east, driven by winds (prevailing south-westerlies), non-linear tides (due to strong tidal forcing from the Atlantic) and density currents. The latter are primarily due to freshwater discharge from rivers. Much of the Channel has strong tidal flow and is well mixed, however, a tidal mixing front occurs in the western Channel where the stratified waters bordering the Celtic Sea meet the well-mixed regime; summer time flows at the edge of the region are anticlockwise (i.e east to west). Net west-to-east flow in the bulk of the Channel has been modelled (Salomon and Breton, 1993; Salomon et al., 1993) and confirmed from distributions of radionuclides released from the nuclear fuel reprocessing plant at Cap de la Hague on the NW French coast (Gueguenіat et al., 1995). Prandle et al. (1993) estimated net flux north-eastwards through Dover Strait as 0.11 Sv. However, long-term net flow patterns are complex, including for example an anticlockwise gyre off Cap Gris Nez (Prandle and Player, 1993) and other gyres in bays, off headlands and around the Channel Islands (e.g. Mardell and Pingree, 1981; Pingree and Maddock, 1985; Salomon and Breton, 1993).

In the Celtic Sea, Pingree and le Cann (1989) analysed an extensive set of current meter data and found a generally weak mean circulation, albeit tidal currents are strong. During winter (November to April) the Celtic Sea is vertically mixed and residual circulation is largely controlled by wind forcing. In summer, most of the Celtic Sea has strong thermal stratification (where tidally-generated turbulence is insufficient to mix the solar heat input (near the surface) throughout the water column). Then summer circulation is dominated by anticlockwise jets associated with bottom fronts bounding a cold saline pool (Brown et al., 2003; Hill et al., 2008; see Figure 4.8). On the eastern side of St George’s Channel, a jet transports water northwards from the mouth of the Bristol Channel towards the Irish Sea. Tides induce local circulation around the Scilly Isles (Pingree and Mardell, 1986).

In the Bristol Channel, tidal currents are strong but residual flows are weak and the estimated flushing time is 150 to 300 days (OSPAR, 2000). Prevailing south-westerly winds drive a flow northwards along the Cornish coast. Along the northern coast of the Bristol Channel, between Carmarthen Bay and Nash Point, flow is also into the Channel. As wind piles up water in the Channel, however, an adverse pressure gradient is created, driving a depth-mean flow westwards along the central axis of the Bristol Channel. This flow is then steered northward around St David’s Head and into the Irish Sea. Density gradients also contribute to the weak circulation; when freshwater input is large, these flows are significantly enhanced, although no direct measurements have been made. There are local residual circulations: closed eddies, arising primarily as water flows past headlands, bays and islands; however, they contribute little to the overall mean circulation. As for the Celtic Sea there is flow across the mouth of the Channel at about 5° W.

4.5 Irish Sea

Surge- and density-driven currents both contribute significantly to the overall long-term mean circulation of the Irish Sea. Density-driven currents are particularly important in the eastern Irish Sea where the differences between the saline oceanic inflows and freshwater input from the Rivers Dee, Mersey, Ribble and Lune cause density changes in Liverpool Bay. The long-term effect in circulation offshore at the surface and onshore below is illustrated in Figures 4.9 and 4.10. These flows are strongest in winter and spring but can be overwhelmed during periods of strong winds.

The distribution of caesium-137 discharged from Sellafield has been used to infer the mean surface water circulation in the Irish Sea (Jefferies and Steele, 1989; Irish, 2003). The main input of water is from the Atlantic, flowing south to north through St. George's Channel. The general shape of the isopleths suggests that the main flow veers towards the Welsh coast as it moves north, with a weaker flow, generally northward, to the west of the Isle of Man. A minor component of the flow enters the eastern Irish Sea to the north of Anglesey and moves anticlockwise around the Isle of Man before rejoining the main flow to exit through the North Channel. The overall flushing time for the Irish Sea as a whole is about one year (Knight and Howarth, 1999).

Most regions of the Irish Sea are continuously mixed because tidal currents are strong. However a deep basin region in the western Irish Sea (centred at 53°40□ N, 5° W) and part of Cardigan Bay experience strong seasonal stratification in the summer and are separated from the well-mixed areas by tidal mixing fronts. In the western Irish Sea, a dome-shaped pool of cold water sits below the thermocline and is separated from surrounding waters by strong temperature fronts. These fronts drive strong narrow (~ 10 km) jets that dominate the circulation in the region during summer months, forming a closed circulation; the lack of through-flow enables material to remain in the region (Hill et al., 1997, 2008). Following the breakdown of stratification in autumn, the mean flow is then weakly northwards until the following spring.

Considerable changes in Irish Sea flow conditions are suggested by models. In order to obtain a reasonable model fit to observed caesium-137 concentrations, McKay and Baxter (1985) found that the coastal flow conditions from the NE Irish Sea to western Scottish coastal waters had changed considerably since 1977, with a further change in Irish Sea outflow during 1981. Jefferies and Steele (1989) had to infer a factor-of-two change in the Irish Sea circulation in the mid-1970s, doubling the flow rate out of the North Channel from the end of September 1976; they also inferred a change in flow during 1980/81, after which the flow pattern returned to that of 1977-1980. However, substantial transports can take place in one storm (more than 20% of the Irish Sea volume during two days in February 1994; Knight and Howarth, 1999) and affect the inference of longer-term flow from constituent distributions.

A direct link with the circulation of the Irish Sea and the NAO has not been established but it is reasonable to expect a degree of correlation. A positive NAO Index is associated with westerly (veering to north-westerly) winds over the Irish Sea, hence more frequent surges in the eastern Irish Sea and Liverpool Bay, enhancing the contribution of surge currents to the overall circulation. Changes in storm tracks may also modulate the circulation and flushing of the region.

4.6 Minches, west Scotland and Scottish continental shelf

Circulation on the shelf west of Scotland (the Scottish Coastal Current; SCC) is mainly northwards.

Tiree Passage currents have been measured since 1981 (see UK Directory of Marine Observing Systems (www.ukdmos.org). They are constrained by the Passage orientation (to the north-east) and narrowing (between Tiree and Mull) and are dominated by semi-diurnal tidal species (Figure 4.11 and Table 4.2). Hourly time series of northward and eastward velocity have been resolved into components *Ua* along the channel (057°T-237°T) and *Ux* across the channel. *Ua* exhibits much greater variance (standard deviation = 0.368 m/s) than *Ux* (standard deviation = 0.054 m/s), consistent with the alignment of the semi-major tidal ellipses along the channel (Table 4.2). *Ua* is offset from zero by a mean northward flow 0.108 m/s. This is a clear manifestation of the northward Scottish Coastal Current, and significantly larger than previous summer-only estimates; for example 0.108 m/s equates to 9.3 km/d, compared with previous values of 2 to 5 km/d (McKay et al., 1986; McCubbin et al., 2002). The corresponding mean volume flow through the Passage is calculated as 0.067 Sv, a similar value to that for the North Channel outflow, although this is not the same water. Caesium-137 studies

(McKay et al., 1986) indicate that North Channel water is joined by Atlantic water; the average ratio in the Tiree Passage is approximately 3:1. The Atlantic water proportion in the Scottish Coastal Current increases steadily northwards to be a majority past Cape Wrath, where the Scottish Coastal Current total volume transport is correspondingly greater.

The tidal currents are almost rectilinear (semi-minor axis much smaller than the semi-major axis), aligned with the Tiree Passage (direction near 57°T) and strong (> 0.5 m/s at spring tides).

There is significant seasonality in *Ua* but not in *Ux*. After removing the tidal constituents listed in Table 4.2, the residual flow time series clearly shows this seasonal variation in the along channel residual flow (Figure 4.11). Monthly mean residuals (not shown) display only one negative value throughout the entire time series, September 1984, when *Ua* = -2 mm/s. This flow reversal can be attributed to an anomalously strong and persistent northerly air flow during that month.

The direct and indirect influences of atmospheric pressure gradients on the residual flow through the Tiree Passage lead one to view the interannual flow variability in the context of the major mode of North Atlantic atmospheric interannual variability, the North Atlantic Oscillation (NAO). The NAO Index, after Hurrell (1995), and the residual transport through the Tiree Passage, similarly averaged over winter months December to March (DJFM), are presented in Figure 4.12. 35% of the variance in the along channel residual is explained by changes in the NAO (r = 0.59).

In summary, Tiree Passage currents are predominantly semi-diurnal, and constrained by the Passage. A strong northward residual varies significantly with season, correlated to winter storm activity. Interannual variation correlates significantly with the NAO index.

4.7 Continental shelf edge, including Rockall Trough and Faroe Shetland Channel

The steep bathymetry of the continental slope acts as a barrier between oceanic regions and the shelf sea systems, reducing the amount of water that can travel from the deeper waters of the North Atlantic into the shallower waters on the continental shelf. Nevertheless, wind forcing and tides enable some North Atlantic water to flow onto the Scottish shelf (and into the North Sea, between Orkney and Shetland and around Shetland) as well as along the slope into the Norwegian Trench.

Observations at the continental shelf edge indicate a poleward along-slope current, flowing along most sectors of the ocean-shelf boundary from Portugal to Norway (there is less evidence around Biscay). The flow is forced by the combined effect of steep topography and the mutual adjustment of shelf and oceanic regimes to meridional density gradients (Huthnance, 1984; Simpson, 1998), and is enhanced or modified by wind stress. The current is an important source of heat, nutrients and plankton to the waters around Scotland.

Currents and transports along the continental slope from the Celtic Sea to the Faroe Shetland Channel were summarised by Huthnance (1986). Estimated transports between the shelf break and the 2000 m depth contour (probably the great majority) were fairly consistently poleward in the range 1 to 2 Sv (1 Sv = 10^6 m^3/s) from the Celtic Sea to the Wyville-Thomson Ridge. Mean current speeds quoted are typically 0.05 to 0.2 m/s, but more variable than the transport as the flow may be locally 'squeezed' between depth contours. More recent information in the Celtic Sea region (Pingree and le Cann, 1989; Pingree et al., 1999; Huthnance et al., 2001) suggests some seasonality: weaker flow in spring and stronger flow in autumn.

West of Scotland, using detailed year-round measurements for 1995/96, Souza et al. (2001) found that the slope current at 56° to 57° N had a maximum mean flow of ~ 0.15 m/s, with greater flow variability in winter. In summer there was a maximum flow at about 200 m depth whereas in winter the flow was more nearly uniform in depth. The fastest mean flow was where water depth is 500 m or more, but in winter the mean flow was broader and extended onto the shelf. A mean transport of about 2 Sv is suggested by combining these measurements with tracked drogues (Burrows et al., 1999).

Holliday et al. (2000) and Holliday (2003) calculated the mean transport though the Rockall Trough as 3.7 Sv, but the flow fluctuates on interannual timescales. There was unusually strong northward transport in the Trough during 1988/89 and 1998, peaking at 7.9 Sv in 1989 and 7.5 Sv in 1998. These values include the along-slope current and flow in the main width of Rockall Trough. The interannual fluctuations might be related to the position of boundaries between different North Atlantic water masses and hence the balance of inflows to Rockall Trough, but the ultimate cause is not known; there is no correlation with the NAO (Holliday et al., 2000, 2008).

At the Wyville-Thomson Ridge (near 60° N with typical depth 400 to 500 m) there is a complex exchange of flow. Some of the deeper slope current from the Hebrides slope is probably diverted to the north-west. However, the upper-slope current continues to the west Shetland slope (the Faroe-Shetland Channel). Here it is joined by a broader flow of warm, saline North Atlantic water across the Ridge from Rockall Trough. Further on, it is also joined by water that has circulated clockwise around the Faroe Islands to the Faroes side of the Faroe-Shetland Channel. These additions result in increased speeds (0.15 to 0.3 m/s or more) and increased transport along the west Shetland slope, on average about 4.5 Sv centred approximately over the 400 to 500 m isobath. Around Scotland, the slope current can be stronger in winter than in summer.

The concentrated flow at the shelf edge and the effective separation of the shelf and oceanic regimes by the topographically steered flow is illustrated by the behaviour of tracked drogues. Released into the narrow slope current, drogues have a strong tendency to remain in it and move rapidly along the slope, in contrast to those released on the shelf or in the oceanic regime, which show much more variable behaviour unless they are entrained into the slope current (Simpson, 1998; Burrows et al., 1999). *Charting Progress* (Defra, 2005) includes a link to an animation of the along-slope current, as measured by tracked drogues.

5. What the evidence tells us about environmental status

The accumulation of observed data pertaining to the AMOC since the publication of *Charting Progress* (Defra, 2005) has led to a vociferous debate within the scientific community as to whether or not the AMOC is undergoing a significant slow-down. Considered in isolation, the series of ship-based estimates presented by Bryden et al. (2005) provides compelling evidence that there is a change underway and that the rate of this change is markedly greater than the gradual slow-down predicted for the 21st century by coupled climate models (Meehl et al., 2007). A number of considerations exist, however, which may lead to lower confidence in this assessment or, at the least, to highlight the need for further data and analysis.

First, high-frequency variability evident in the continuous mooring records now available raises the possibility that historical hydrographic sections of short duration may have sampled the background variability of the system, rather than a longer-term trend: the range of daily strengths in the first year's mooring data alone is 4 to 35 Sv, while the range of the five ship-based estimates spanning five decades is ~ 23 to ~ 15 Sv (however, the latter are sampled over six weeks, which may suppress some of the high-frequency variability).

Second, a number of studies make use of an implied connection between the AMOC strength and patterns of Atlantic sea surface temperature (SST) anomalies to reconstruct past AMOC changes and to diagnose the present strength (e.g. Knight et al., 2005; Latif et al., 2006). These studies highlight the recent phase of warm North Atlantic SST anomalies as indicative of a relatively strong AMOC condition. However, these techniques and their associated conjectures depend critically on the anomaly pattern chosen to act as an AMOC proxy. For example, the anomaly index preferred by Latif et al. (2006) implies a sustained weakened AMOC phase from the 1920s to the 1970s which is entirely absent from the Knight et al. (2005) calculations.

Third, if the evidence presented at 26° N is representative of a full ocean-basin slow-down in the AMOC, then it is reasonable to ask why there is no lucid signal in Arctic-Subarctic transports. The evidence presented above shows variability but little by way of an overall trend, even in the longest records. This might be because the signal has yet to propagate internally within the overturning system, or because there is no bona-fide signal at all.

With respect to the future state of the AMOC, new co-ordinated climate-change modelling experiments were conducted for the Fourth IPCC (Intergovernmental Panel on Climate Change) Assessment Report (Meehl et al., 2007), indicating possible changes to the AMOC during the 21st century. In all plausible future greenhouse-gas emissions scenarios, the AMOC undergoes some weakening in most climate models. Figure 4.13 depicts the AMOC strength in 19 IPCC-class models under scenario A1B (medium greenhouse-gas emissions). The inter-model spread of strengths is large even during the observational period (i.e. the 20th century) although most models (apart from three or four) are able to reproduce strengths comparable with observed estimates of the AMOC strength (shown by black bars to the left of the vertical axis). Changes during the 21st century range from a 0% to a 50% decrease in simulated AMOC strength due to changes in the thermal properties and density structure of the Atlantic and Arctic Oceans.

These results lead the IPCC to conclude that a slow-down of the AMOC during the 21st century remains very likely (i.e. chance of occurrence > 90%). None of the models that maintain a realistic observational AMOC strength undergo a complete shutdown before 2100. Hence the consensus is that a complete shutdown is very unlikely (i.e. chance of occurrence < 10%). However, there are uncertainties relating to these projections: in particular, if outflow from the Greenland Ice Sheet accelerates (not accounted for in most climate models), additional AMOC weakening may result.

It is important to remember that the cooling (and indeed reduced transport of salt) associated with any slow-down in the AMOC is accounted for in the IPCC climate change projections (e.g. Meehl et al., 2007). In other words, the reduction in heat transport related to any AMOC slowdown is 'competing' with the direct warming effects of anthropogenic climate change which continue to dominate. Thus UK waters are warming and this is projected to continue despite any counter-effect of AMOC slow-down.

In UK shelf seas, models and specific local measurements have confirmed current understanding of the variable forcing and space-time variability of currents and circulation. Factors include tides, winds and the density field (resulting from freshwater inputs and atmospheric exchanges) as elaborated in section 2. In some locations (e.g. Tiree Passage) the relation to atmospheric forcing is manifested in a significant correlation with the NAO. In shallow areas, tidal and directly wind-driven currents on short time-scales tend to prevail; coastal currents from freshwater input give net transports. The large variability in space and time (on short and long scales in each case), together with knowledge of the variable factors, enables a confident statement that circulation is a derivative quantity and not showing a significant trend relative to shorter-term variations.

The summary table (Table 4.3) includes an assessment of trend but not status ('traffic-light') because (1) no accepted criteria apply for circulation giving significant risk of adverse effects; (2) the UK (government), or even the EU, cannot itself take measures to improve the status.

Circulation is subject to a wide range of natural variability on many time-scales. There is no reason to suppose that this variability or circulation intensity has changed significantly since *Charting Progress* (Defra, 2005) for any present impact on the environment or human health. Local construction has the potential to introduce local changes and significant impacts on the spatial scale of the construction, such as in harbours and around wind turbine pylons. There is little basis to distinguish between CP2 Regions except in respect of such local activities.

6. Forward look and need for further work

The influence of the Atlantic Meridional Overturning Circulation on NW European climate implies a pressing need to increase our understanding of the AMOC system, in particular its present status and sensitivity to change, as well as validating the coupled ocean-atmosphere climate models used to generate future predictions. Coverage and availability of observational data and monitoring capability are continually increasing, allowing limited assessment of the state of the AMOC. However, spatial and especially temporal coverage remains well short of that required to make any such assessment with high confidence. In particular, it is as yet impossible to ascertain with confidence whether or not the AMOC is undergoing change – or at what rate.

Some change, although not an abrupt collapse, can be expected this century, as indicated in climate model experiments, and such changes will be detectable, given sufficient time, by the array of monitoring equipment deployed in recent years. Any resulting effect on UK shelf-sea circulation is not yet predicted or resolvable in climate models, but is likely to be related most closely to any changes in the most adjacent along-slope flow and associated cross-slope fluxes.

The primary knowledge gap remains a lack of high-quality data sets of sufficient temporal length with which to elucidate climate-change signals from background variability. The RAPID-funded monitoring projects have recently acquired funding to extend observations until at least 2014, in order to assemble a continuous decade-long time series of AMOC strength. This extended record will enable clarification of interannual variability and will form the basis of a statistically significant time series with which to begin to comment on longer-term change. A research priority must be to ensure that supplemental measurements, especially at higher latitudes, are maintained simultaneously. Without the latter, it will not be possible to understand how changes in the AMOC are relayed from place to place – an understanding which, ultimately, will feed back into the monitoring strategy.

In UK shelf seas, current understanding of processes gives medium confidence in extrapolating from areas of specific process studies to comparable locations. There is also confidence in models to represent many processes. Challenges are still presented by (1) the large variability and random element in adjacent deeper-water flow impacting on the shelf and by (2) shorter-scale internal waves and their contribution to turbulence, hence stratification and associated depth-dependent flow. Measurements are still sparse relative to currents' large variability in space and time (on short and long scales in each case); reliance is placed on models for most distributional information. In view of the importance of currents to offshore operations and habitats, and of circulation to many species' lifecycles (for example), better understanding and prediction of short-term variability in circulation is needed. This entails further work on model validation and the development of observational networks. Circulation is strongly tied to the details of the atmospheric forcing. Predictions for future climate therefore depend on regional scenarios for which there is not yet confidence (e.g. regarding storm tracks and intensity). However, it seems unlikely that there will be large changes in rates of overall exchange and flushing in the coming decades.

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Figure captions

Figure 4.1 Circulation of surface waters in the North-East Atlantic. Red arrows represent the flow of warm, salty Atlantic waters along the continental slope and further west, while the yellow, blue and green arrows represent the flow of coastal waters. The yellow arrow indicates the path of the Scottish Coastal Current, while the blue arrow indicates the inflow of mixed coastal/oceanic water past Fair Isle (the Fair Isle Current) and the green arrow indicates average anti-clockwise flow in the southern North Sea. Supplied from *Scotland's Seas: Towards Understanding their State* (Scottish Government, 2008). Courtesy of S. Hughes, Marine Scotland.

Figure 4.2 Estimates of the strength (dots plus error bars) of the Atlantic Meridional Overturning Circulation at 26° N from five ship-based transects (Bryden et al., 2005), 1957 to 2004. Courtesy of the National Oceanography Centre.

Figure 4.3 Continuous Atlantic Meridional Overturning Circulation strength from the RAPID monitoring array at 26° N, April 2004 to September 2007. Data are expressed in Sverdrups (1 Sv = 10^6 m^3/s). A low-pass filter has been applied to the daily strengths, suppressing much of the variation on time scales less than 10 days. Courtesy of S. Cunningham et al., National Oceanography Centre.

Figure 4.4 Residual currents, for the period 1957 to 2002, from the CS3X model used for UK operational storm-surge forecasting. Speed is illustrated by shading (pixel-by-pixel, units 0.01 m/s, showing the speed at each grid cell). Note the bi-linear scale (0-10 in 0.5 units; 11-20 in 1 unit intervals). Streamlines show direction (where speed > 0.005 m/s). Courtesy of C. Wilson, National Oceanography Centre.

Figure 4.5 Seasonal mean current for the period 1960 to 2004 from the Atlantic Margin Model version of POLCOMS. (a) January-March; (b) April-June; (c) July-September; and (d) October-December. Courtesy of J. Holt, National Oceanography Centre.

Figures 4.6. Modelled near-surface currents for 2001 with a ~ 1.8 km resolution version of POLCOMS. (a) January-March, (b) April-June, (c) July-September and (d) October-December respectively. Courtesy of J. Holt, National Oceanography Centre.

Figure 4.7. Summary of 154 tracks of drifting buoys (from Hill et al., 2008).

Figure 4.8 Horizontal near-bed temperature gradients (°C per km) from a 3-D model (Holt and Proctor, 2008) with observed summary drifter tracks overlaid (from Hill et al., 2008). DB – Dogger Bank, FI – Frisian Islands, FlHd – Flamborough Head, FoF – Firth of Forth.

Figure 4.9 Progressive vector diagrams from the full records of two ADCPs in Liverpool Bay, from 2.5 m above the seabed (blue) to 18.5 to 21.5 m above the seabed (black/grey). Site A shows the average for the period August 2002 to May 2008, while site B shows the average for the period April 2005 to June 2008. Courtesy of J. Howarth, National Oceanography Centre.

Figure 4.10 Mean near-surface currents from HF radar in Liverpool Bay for the period August 2005 to August 2008. Courtesy of J. Howarth, National Oceanography Centre.

Figure 4.11 Residual flow through Tiree Passage, resolved (positive) along 57°T. The tidal constituents listed in Table 4.2 have been removed, a low-pass filter applied (zero response for periods < 72 hours) and the time series sub-sampled at three days. Courtesy of M. Inall, Scottish Association for marine Science.

Figure 4.12 The December-to-March NAO Index and December-to-March-averaged along-channel residual flow from Tiree Passage. Positive flow is directed northwards through the Passage. Courtesy of M. Inall, Scottish Association for marine Science.

Figure 4.13 Strength of the Atlantic Meridional Overturning Circulation in the IPCC suite of climate models, forced with observed greenhouse gas concentrations followed by future 'medium'-scenario A1B emissions. Each coloured line depicts strengths in a different model. 20th century estimates for the observed AMOC are shown by black bars on the left-hand axis. From Meehl et al. (2007).

Table 4.1. Transport estimates of Lower Atlantic Deep Water (originating in the Arctic basin and occupying 3000 to 5000 m depth) across 26° N. Units are Sverdrups (1 Sv = 10^6 m3/s of water crossing 26° N). Negative units indicate a southwards transport. Source Bryden et al.. (2005).

Year	**1957**	**1981**	**1992**	**1998**	**2004**
Transport strength (Sv)	-14.8	-11.8	-10.4	-6.1	-6.9

Table 4.2. 11-constituent tidal analysis of the Tiree Passage current time series, 1981-2008.

Constituent		**Semi-major (cm/s)**	**Semi-minor (cm/s)**	**Align (degrees)**	**Phase (degrees)**
Fortnightly	MSF	0.619	0.167	56.96	284.61
	O_1	3.722	0.367	62.43	137.46
Diurnal	P_1	1.621	0.130	60.14	296.38
	K_1	5.614	0.303	59.92	301.59
	MU_2	0.420	0.050	54.50	328.60
	N_2	8.065	-0.075	56.98	80.15
	NU_2	1.672	-0.072	55.01	86.72
Semi-	M_2	41.855	0.194	57.03	104.14
diurnal	L_2	1.519	-0.085	54.75	123.53
	S_2	14.391	0.372	57.66	141.74
	K_2	5.341	0.106	58.97	155.80

Table 4.3. Summary assessment of trends.

Parameter	**CP2 Region**	**Key factors and impact**	**What the evidence shows**	**Trend**	**Confidence in assessment**	**Forward look**
Circulation	1, 2, 3, 4, 5, 6, 7 (UK shelf-sea areas)	Climate change. Advects sea contents	Short-term variability, especially from tides and winds	Not determined	High (i.e. trend definitely not determined; shorter-term variability is greater)	Continued short-term variability
Circulation	8 (Deeper adjacent Atlantic)	Climate change. Advects sea contents	Short-term variability	Not determined	High (i.e. trend definitely not determined; shorter-term variability is greater)	Reduced AMOC