Chapter 20

# Ocean Current Changes as an Indicator of Global Change

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## 1. INTRODUCTION

The high heat capacity of seawater and the relatively slow ocean circulation allow the oceans to provide significant 'memory' for the climate system. Bodies of water that descend from the sea surface may reside in the ocean interior for decades and centuries, while preserving their temperature and salinity signature, before they surface again to interact with the overlying atmosphere. In contrast to that, the residence time of water in the atmosphere is about 10 days and the persistence of dynamical states of the atmospheric circulation may last up to a few weeks. Thus, on long time scales ocean dynamics becomes important for climate, which implies that climate variations and climate change can only partially be understood without consideration of ocean dynamics and the intricate ocean–atmosphere interaction. The El Nino/Southern Oscillation phenomenon in the tropical Pacific is a prominent example of tightly coupled ocean–atmosphere dynamics on interannual time scales, other more weakly coupled interactions exist throughout the system.

The oceans' role in climate and climate change is manifold. Ocean circulation transports large amounts of heat and freshwater on hemispheric space scales which have significant impacts on regional climate in the ocean itself but also noticeable consequences via atmospheric teleconnections on land. What is well known for the seasonal cycle with only moderate temperature changes between summer and winter in marine climates compared with much larger swings within the continents, is also true on decadal time scales. Since 1960 the heat uptake of the oceans has been 20 times larger than that of the atmosphere. Thus the oceans have been able to reduce the otherwise much more pronounced temperature rise in the atmospheric climate. Also, over the last 200 a, the oceans have absorbed about half of the $CO_2$ release into the atmosphere by human activities (fossil fuel combustion, de-forestation, cement production), thereby reducing the direct effect of greenhouse gases on atmospheric temperatures.

## 2. THE VARIABLE OCEAN

Bodies of water circulate throughout the oceans – both horizontally and vertically – as a consequence of physical forces exerted on them according to Newton's Law. The oceanic circulation is not steady in time. Rather motions of water bodies in the ocean are known to vary on a broad range of spatial and temporal scales. The following four examples serve to highlight natural variations of large-scale circulation patterns:

**(i)** Seasonal variations of the strength of the North Atlantic subtropical gyre at 26°N have the amplitudes of 25 Sv[1] (peak to peak). This range is comparable to the time mean strength of the wind-driven, anti-cyclonic basin-scale gyre at this latitude [1].

**(ii)** The Pacific subtropical cells (STC) – a meridional, upper-ocean pattern of circulation that links the subtropical subduction regions north and south of the equator to the equatorial thermocline – has seen a decline of 11 Sv or 30% since the 1950s, however, displaying decadal variations of the same order of magnitude [2]. The observed strong decadal and multi-decadal variations in sea surface temperature (SST) in the equatorial Pacific have been shown to be related to changes in STC strength.

**(iii)** The cyclonic circulation of the North Atlantic subpolar gyre has possibly weakened by 25% and shrunk in size since the mid-1990s [3,4]. This

1 1 Sv = $1 \times 10^6$ $m^3 \cdot s^{-1}$ (unit for volumetric transport, named after Harald Ulrik Sverdrup). For comparison, the Amazon River discharge in the Atlantic is about 0.2 Sv.

has been attributed to the transition of the North-Atlantic Oscillation[2] (NAO, [5]) from a comparably strong phase between 1960 and 1995 (manifesting itself in stronger than average westerly winds at mid-latitudes) to a significantly weaker one after 1995. The gyre's weakening and westward retreat has allowed large quantities saline subtropical upper-ocean waters to flow northward past its eastern flank, as a consequence of which a drastic increase in salinities in the Nordic Seas[3] has been observed [6]. This is thought to have an impact on the sinking of waters as part of the Atlantic Meridional Overturning Circulation (AMOC), the latter being the primary focus of this chapter.

**(iv)** Although not having been observed directly, it is commonly thought that temporal changes in the strength of the AMOC – a basin wide meridional circulation pattern that links upper-ocean net northward flow of warm, saline waters with cold southward return flow below roughly 1000 m throughout the Atlantic – explain large parts of the observed multi-decadal North-Atlantic SST changes [7]. A recent summary and discussion of climate variability and its predictability in the Atlantic sector [8] provides a perspective on the difficulties one has, to distinguish decadal variability from long term, possibly anthropogenic induced trends.

The reason why the different components of the ocean circulation have the potential to change substantially over time is a consequence of the complex forcing at the sea surface (exchange of momentum, heat and freshwater between ocean and atmosphere) on the one hand and internal ocean dynamics on the other. Examples of internal ocean dynamics include advection of water of anomalous density by the mean large-scale ocean circulation, westward energy transfer by off-equatorial planetary waves, the equatorial wave guide, horizontal mixing by meso-scale eddies, deep-water formation due to convection or small-scale vertical mixing, acting to push the cold waters of the oceans' abyss upwards. The large variations that basin-scale circulation patterns may exhibit have the potential to delay the detectability of climate change related shifts in the flow field.

## 3. OCEANOGRAPHERS' TOOLS

Oceanographers have developed direct and indirect techniques for the observation of ocean currents in order to document and analyse the strength of the interior ocean circulation and its changes in space and time. Direct current measurements can be divided in two classes, Eulerian and Lagrangian ones.

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[2] The NAO is the dominant mode of (winter) climate variability in the North Atlantic region ranging from central North America to Europe and into Northern Asia.

[3] Nordic Seas is used as collective term for Greenland Sea, Norwegian Sea and Iceland Sea.

The Eulerian approach measures current velocity and direction at a fixed location. Current meters are typically used in this context that can be mounted on a stationary platform or 'mooring' [9], installed in a vessel's hull [10] or lowered from a vessel [11]. The scope of Lagrangian observations aims at deriving the streamlines of regional or ocean basin-scale flow fields. For this drifters are used, that move passively with the flow at a specified depth horizon and whose displacements are monitored over time [12].

The equation of motion for a fluid (i.e. Newton's Law applied to a fluid) requires the various different physical forces acting on a unit mass of water to balance one another. The two probably most-widely used indirect methods to study the flow field in the open ocean – (i) the Ekman balance and (ii) the geostrophic balance – rely on well-founded simplifications of the equation of motion, that are valid on time scales longer than 1 or 2 days and spatial scales in excess of a several tens of kilometres.

**(i)** Within roughly the top 50 m of the water column (near-surface Ekman layer), the flow field results from a balance between the stress the atmospheric wind field exerts onto the sea surface and the Coriolis force [13]. To first order, the horizontal flow in the Ekman layer depends on strength of the wind speed but moves at right angles to the direction of the wind as a consequence of the Earth's rotation.[4] The availability of daily wind fields over the global ocean from space borne measurements makes the Ekman balance a very powerful diagnostic tool of large-scale near-surface flows.

**(ii)** In the vast ocean interior below the Ekman layer, the horizontal movement results in long surfaces of constant pressure as result of the near geostrophic balance between the horizontal pressures gradient force and the Coriolis force.[5] This is analogous to atmospheric conditions as depicted in weather charts, where (to first order) wind flows *around* cells of high or low pressure (and not from high to low pressure), again as a consequence of the Earth's rotation. As a result, the strength of the flow across a section between two points is approximately proportional to the difference in pressure at the section's two end points (both in the ocean and in the atmosphere). The geostrophic balance is therefore an effective tool to diagnose the net strength of basin-scale ocean circulation patterns. All that is required are measurements of the pressure field at the section end points, while the actual horizontal structure of the flow in between does not need to be resolved. Practically, the pressure field of the ocean cannot be measured directly in the water column to derive reliable estimates of the strength of the flow.

4 The net flow in the Ekman layer is to the right of the direction of the wind stress in the Northern Hemisphere and to the left in the Southern Hemisphere.

5 The geostrophic flow is directed such that the pressure increases to the right in the Northern Hemisphere and to the left in the Southern Hemisphere.

However, profiles of water density allow the computation of the ocean's pressure and velocity field relative to reference pressure level. This has served oceanographers for many decades to study strength and vertical structure the ocean circulation [14].

Besides observing ocean currents the numerical simulation of the circulation using so-called ocean general circulation models (OGCM) has emerged as an important discipline in physical oceanography. To this end the equation of motion (if applicable also the heat and salt conservation equations) are solved numerically on a pre-defined spatial model grid sequentially for each time step. This is commonly referred to as model 'integration'. For each time step, the so-called boundary conditions have to be prescribed, that drive the model ocean. The boundary conditions mainly include observed fluxes of momentum, heat and freshwater between the ocean and the atmosphere. Typical horizontal resolutions of currently used OGCMs range between 10 and 100 km. The finer the resolution (and the larger the model region) is chosen, the more computationally expensive it becomes to run the model, such that the integration periods that can be reached, become shorter.

Typical integration periods are up to several decades. Besides OGCMs the class of climate models has found widespread use in oceanography. Here, the equations of motion for the ocean and the atmosphere (and for ice sheets, if applicable) are solved simultaneously, and both model components are coupled at the air–sea interface. In climate models, the ocean is not passively forced, but instead can feed back on the atmosphere. Climate models are driven by orbital forcing (i.e. insolation at the top of the atmosphere). As climate models are typically integrated over climate-relevant time scales (centuries and longer) their horizontal resolution is often sparse compared to OGCMs, typically between 100 and 500 km.

The main advantage of numerical model simulations is that self-consistent estimates of the ocean circulation can be obtained for the entire spatial and temporal domain of interest. The degree of their shortcomings, however, is difficult to evaluate. One problem common to all numerical models is their finite resolution. That means that physical processes that take place on spatial scales smaller than the grid size in the real ocean are not included in the model physics and have to be parameterised. Other uncertainties derive from errors in the boundary conditions and the numerical integration itself.

## 4. THE ATLANTIC MERIDIONAL OVERTURNING CIRCULATION

### 4.1. Motivation

For most of this chapter, we will limit our attention to the aforementioned AMOC, which represents a circulation pattern most relevant for marine and terrestrial climates in many ways:

**(i)** The AMOC represents a mechanism of long-term 'memory' in the climate system.

**(ii)** The AMOC is the most important oceanic flow component for meridional redistribution of heat.

**(iii)** The AMOC is an important pathway for the oceans' uptake of anthropogenic greenhouse gases and for the ventilation of the deep ocean interior.

**(iv)** The vigour of the AMOC and the associated heat transport are thought to experience a reduction between 30% and 50% over the next century as a consequence of global warming.

Taken together, highly possible long-term changes in the AMOC are thought both to be indicative of climate change and to contribute to climate change. Because of this the AMOC represents a subject of active ongoing research involving observations and numerical modelling. In the following, we outline the AMOC's the relevance for climate, climate variations and climate change. This is preceded by a description of the underlying pattern of circulation and its potential driving mechanisms. The authors intend to convey that although our knowledge about the AMOC has advanced dramatically over the last few decades, many uncertainties remain yet to be solved.

## 4.2. Circulation, Driving Mechanisms

A striking feature of the temperature distribution in the oceans – Fig. 1 displays a section of temperature along the meridional extent of the Atlantic – is the strong vertical contrast in temperatures at low and mid-latitudes, with warm upper-ocean waters floating on top of cold deep and abyssal waters. The vertical layering of waters of different temperatures (densities) is referred to as stratification. It was already recognised as early as 1798 by Count Rumford that – in the absence of any deep-ocean heat sinks at low latitudes – those cold waters had to originate from high latitudes propagating equatorward at depth.[6] Today, it is well established that the observed temperature distribution is a consequence of the AMOC, that moves roughly 19 Sv of warm, saline waters northward throughout the Atlantic and the same amount of cold water back south at depth ([16,17]; Fig. 2). Carried northward within the Gulf Stream/North Atlantic Current system the near surface waters release heat to the atmosphere and thus become gradually denser. The waters eventually reach the Nordic Seas and the Labrador Sea. Here, deep-reaching wintertime convection (i.e. vertical mixing throughout the upper 2000 m of water column) can occur [4,18,19], when the vertical stratification has eroded after periods of excessive heat loss (Fig. 2). The bulk of the newly formed deep waters – that

[6] Longworth and Bryden [15] give an exciting account of the history of the recovery of the Atlantic Meridional Overturning Circulation.

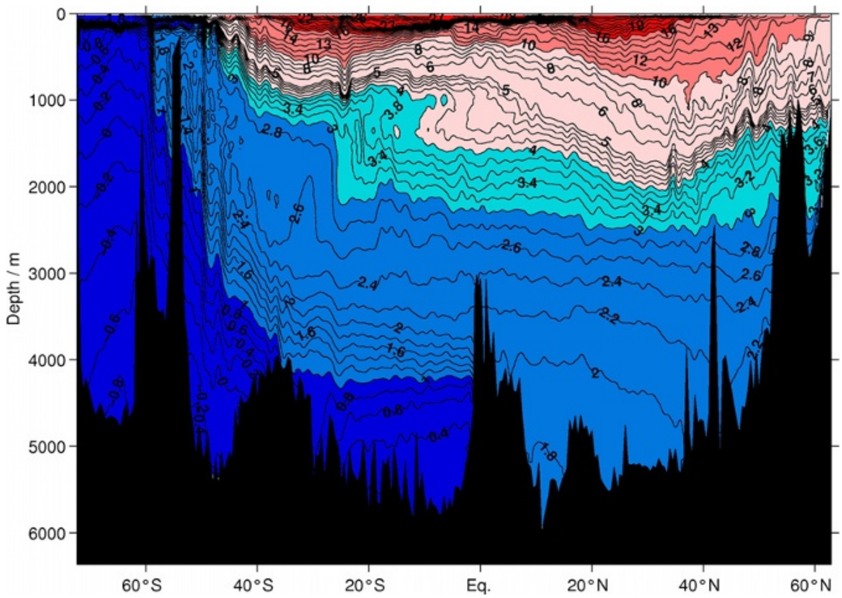


**FIGURE 1** Section of potential temperature along the meridional extent of the Atlantic. For temperatures less than 5 °C and greater than 5 °C, the black contours have a spacing of 0.2 and 1 °C, respectively. The red, indian red, salmon, cyan, light blue and dark blue areas denote temperatures above 16 °C, from 10 to 16 °C, from 4 to 10 °C, from 3 to 4 °C, from 1 to 3 °C and below 1 °C, respectively. Lowered temperature measurements acquired during three research expeditions – aboard *RV Ronald H. Brown* in 2003 (section A16N, PI: Bullister [PMEL]) and in 2005 (section A16S; PIs: Wanninkhof [NOAA]/Doney [WHOI]) and aboard *RV James Clark Ross* in 1995 (section A23; PIs: Heywood/King [NOCS]) – were joined together to compile this figure. Data source: Clivar and Carbon hydrographic data office (http://whpo.ucsd.edu/atlantic.htm). Adapted from a figure of Lynne D. Talley (http://sam.ucsd.edu/vertical_sections/Atlantic.html#a16a23).

are subject to overflow and entrainment processes – constitute the North Atlantic Deep Water (NADW). The NADW is subsequently exported southward, partly confined to the deep western boundary current (DWBC) along the Americas below roughly 1000 m. The intensity of the strongly localised, buoyancy-loss induced formation of NADW at high latitudes (Fig. 2) ‘pushing’ surface waters downwards has long been thought to control the strength of the AMOC.

To close the circulation, the dense NADW needs to return to the upper ocean eventually. This is assumed to be accomplished mainly by two processes. The first process relates to winds and tides that represent the major sources of mechanical energy input into the ocean [20]. Ultimately, this energy input is balanced by dissipation into small scale motions, a process by which turbulent mixing occurs. Dissipation and mixing are ubiquitous in the open ocean; however, they seem most active in the vicinity of rough bathymetry such as exhibited by mid-oceanic ridges [21]. As a consequence deeper (denser) waters from below are mixed with overlying warmer (less

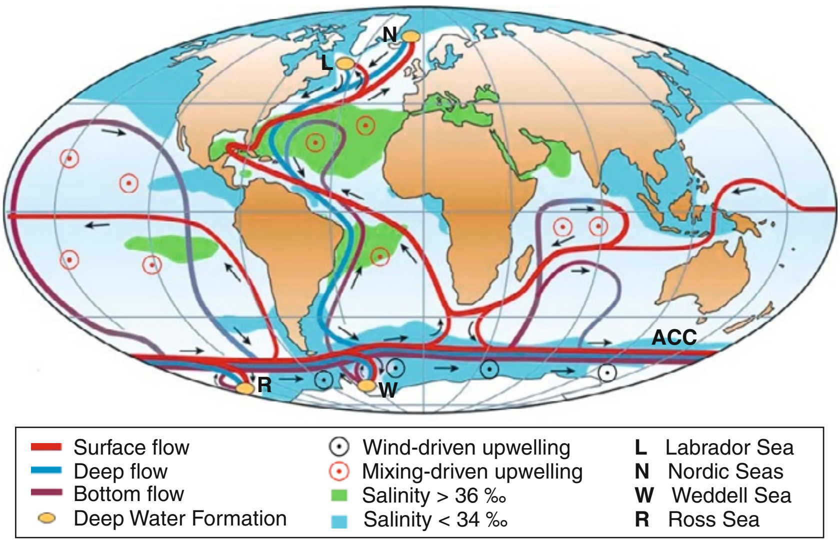


**FIGURE 2** Strongly simplified sketch of the global overturning circulation system. In the Atlantic, warm and saline waters flow northwards all the way from the Southern Ocean into the Labrador and Nordic Seas. By contrast, there is no deep water formation in the North Pacific and its surface waters are fresher. Deep waters formed in the Southern Ocean are denser and thus spread in deeper levels than those from the North Atlantic. Note the strongly localised deep water formation areas in comparison with the wide-spread zones of mixing-driven upwelling. Wind-driven upwelling occurs along the Antarctic Circumpolar Current (ACC). This figure has been published by Kuhlbrodt et al. [17].

dense) waters, thus making deep waters gradually lighter. This allows them to rise and to return to the upper ocean. The fact that as a direct consequence of vertical mixing even at the deep ocean below 1000 m exhibits a notable stable stratification (i.e. water becoming denser with depth, as shown in Fig. 1) has been used to argue that dissipation induced vertical mixing 'pulling' deep water upwards might ultimately have a stronger control on the vigour of the AMOC than the downwards 'pushing' at high latitudes. To move waters vertically across surfaces of constant density, vertical mixing is required and this cannot be generated by high latitude buoyancy forcing [22].

The second potentially powerful mechanism to 'pull' deep water back to the upper ocean to close the overturning circulation can be motivated by a careful inspection of Fig. 1. The NADW flows southward away from the regions of its formation and eventually partly reaches the Southern Ocean. While north of 40°S the deep surfaces of constant temperature show only a weak upwards slope towards the south, the situation changes dramatically south of 40°S. This is a direct result of 70% of the global wind energy input into the ocean taking place in this area. Due to the Ekman balance the strong westerly winds over the Southern Ocean push large amounts of near-surface waters northward, which are then replaced by waters being sucked upwards from the deep ocean. The manifestation of this process is the drastic increase in the upwards tilt of the deep temperature surfaces towards the south (Fig. 1). In this scenario, the transition from cold to warm waters (mixing across density surfaces) occurs near the sea surface in the Southern Ocean as suggested from model findings by Toggweiler and Samuels [23]. Whether the Southern Ocean's control on the vigour of the AMOC is stronger than that deep-ocean mixing mechanism is subject to current debate [17], as clear observational evidence is still lacking.

Besides the AMOC, a second major pattern of meridional overturning exists. This involves formation of deep waters by means of convection around Antarctica. The Antarctic Bottom Water (AABW) spreads northward and represents the coldest and therefore deepest water mass in the Atlantic, Pacific and Indian Oceans (roughly represented by the dark blue shaded part of the temperature field in Fig. 1).

In the Atlantic, the waters gradually mix into the lower parts of the overlying NADW, and eventually return southward. Even though the volume of NADW flowing southward and AABW moving northward are comparable in size [24] or possibly larger for the southern cell [25], the contribution of the AABW related meridional overturning cell to meridional heat transport is negligible, as the vertical temperature contrast between its upper and low branches is very small [26].

## 5. THE AMOC's ROLE IN HEAT TRANSPORT, OCEANIC UPTAKE OF CARBON AND VENTILATION OF THE DEEP OCEAN

The Earth's surface takes up heat by absorbing solar short-wave radiation. On a global average this is almost exactly balanced by the Earth's emission of long-wave radiation. Regional budgets of radiative energy fluxes, however, are unbalanced, as they show pronounced heat gain at low latitudes opposing to heat loss at high latitudes [27]. One-quarter of the 5 Pw[7] of maximum global heat transport – that the coupled ocean–atmosphere system is required to transport poleward in the Northern Hemisphere to approximately balance regional energy budgets – is carried by the AMOC in the Subtropical North Atlantic [27]. While most of the remaining three quarters of the heat transport is accomplished by the atmosphere, the AMOC is by far the most important oceanic component of meridional heat transport.

The steady increase in atmospheric $CO_2$ is widely regarded as one of the main drivers of the presently ongoing global warming. Exceeding the atmosphere in terms of carbon storage by more than a factor of 50 [28], the oceans exchange gases with the atmosphere. The $CO_2$ solubility in sea water

[7] 1 Pw = $1 \times 10^{15}$ W; 5 Pw correspond to the output of 5 000 000 power stations.

increases with decreasing temperatures. The NADW formation at high latitudes, acting to increase carbon concentrations at depth is considered a major element in the ocean's carbon uptake ('solubility pump'). Changes in the strength and spatial structure of the AMOC might affect atmospheric $CO_2$ concentrations and thus global temperatures.

The flow of well oxygenated near-surface water to the deep ocean that goes along with the NADW formation at high latitudes and its subsequent export to the world ocean help to maintain the deep ocean basins as habitats for a diverse biota. This is probably the least known component of biodiversity on Earth. Substantial changes in the rate of deep-water ventilation by the AMOC are thus expected to have consequences for deep-ocean habitats.

## 5.1. Simultaneous Changes of the AMOC and Atlantic Climate in the Past

Especially in the Northern Hemisphere the amount of heat carried poleward by the oceans is very much tied to the strength of the AMOC. Analysis of ice cores from Greenland revealed more than 20 so-called 'Dansgaard-Oeschger events' during the last ice age (100 000–10 000 BC) over the course of which Greenland temperatures jumped by roughly 10 °C within a few decades subsequently followed by a gradual cooling on a millennium time scale [29,30]. Based on the analysis of ocean sediment cores these fluctuations are thought to be linked to abrupt changes in the deep-ocean circulation in the North Atlantic [31,32]. The observations are in qualitative agreement with numerical model simulations that associate the climate variations with temporal changes in the vigour of the AMOC [33,34]. The North Atlantic cold phases are generally thought to be linked with a very weak (or inactive) state of the AMOC that goes along with a near-cessation of NADW formation and northward heat transport in the North Atlantic. Warm phases are expected to coincide with a strong state of the AMOC. During the last ice age, the reduction in the NADW formation rates are likely to have arisen from events of massive input of freshwater from the Laurentide ice sheet (covering Canada) into the North Atlantic [35]. The subsequently fresher and thus less dense subpolar upper-ocean waters stabilised the vertical stratification of the water column, hence, heavily impeding deep-water formation.

To simulate the effect of freshwater input on deep water formation Vellinga and Wood [36] carried out a 'water hosing' experiment using a numerical model. They added a sufficient quantity of freshwater to the northern North Atlantic to cause the AMOC to switch off. This resulted in a strong cooling over the North Atlantic peaking at a temperature reduction of 8 °C around Greenland, standing out from patterns of moderate cooling over the entire Northern Hemisphere and warming over the Southern Hemisphere (Fig. 3). Thus, besides its strong importance for climate over the North Atlantic section the AMOC may also have a moderate impact on global climate patterns.

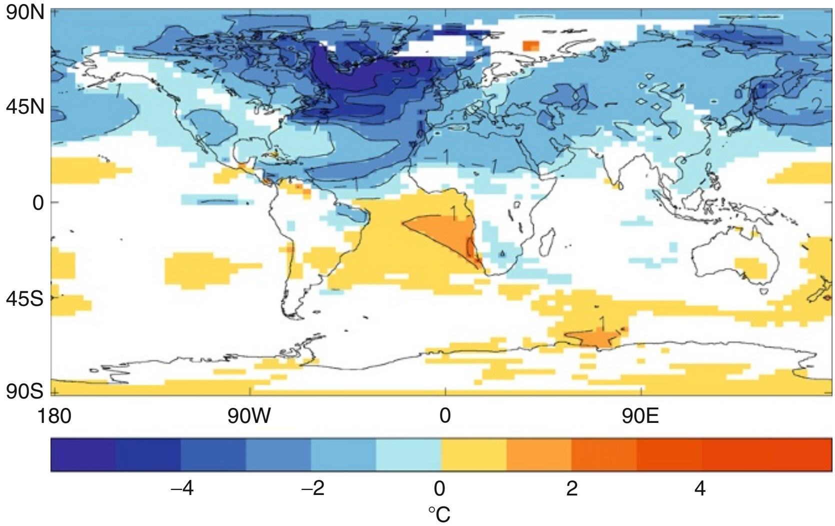


**FIGURE 3** Change in surface air temperature during the years 1920–1930 after the collapse of the AMOC in a water hosing experiment using the HadCM3 climate model. Areas where the anomaly is not significant have been masked. This figure has been published by Vellinga et al. [36].

Additionally, palaeoclimate records suggest that changes in the global circulation involving the AMOC during the early last deglacial period (19 000–14 500 a ago) went along with a significant net transfer of $CO_2$ from the ocean to the atmosphere, leading atmospheric $CO_2$ concentration to rise by about 50 ppmv [37].

## 5.2. Why should the AMOC Change as Part of Ongoing Climate Change?

In their fourth comprehensive climate assessment the Intergovernmental Panel of Climate Change (IPCC) considers it 'very likely' that the AMOC will have gradually slowed down by the end of the twenty-first century as a consequence of the Greenhouse climate [38]. Climate model projections – all of which are based on the greenhouse gas emission scenario A1B [38] – predict a reduction between 0 and 50% by the year 2100 [39], such that a complete (and possibly irreversible) AMOC shutdown is considered 'unlikely'. The future evolution of the AMOC in several selected climate model projections is shown in Fig. 4.

Future greenhouse gas emission scenarios carry a high level of uncertainty as they depend on parameters such as economic and population growth, technology development and basic political and social conditions, all of which are difficult to predict. Also, none of the present-day climate models have a sufficiently fine spatial resolution to resolve the processes that govern either the

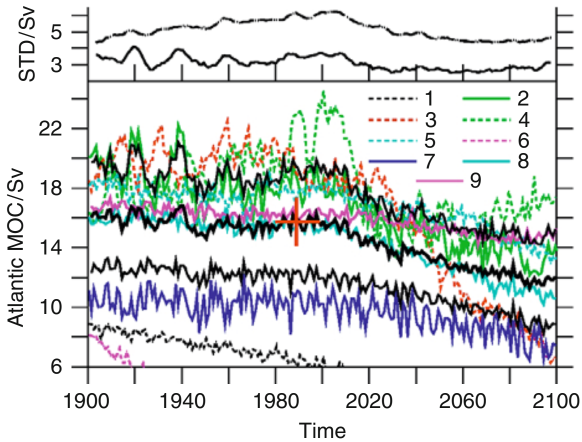


**FIGURE 4** Evolution of the AMOC as defined by the maximum overturning at 24°N for the period 1900–2100 in nine different climate models forced with the greenhouse gas emission scenario A1B. The AMOC evolutions of integrations with a skill score larger than one are shown as solid lines, those from models with a smaller skill score as dashed lines [39]. The weighted ensemble mean is shown by the thick black curve together with the weighted standard deviations (thin black lines). This figure was published by Schmittner et al. [39].

sinking or the rising, and have to rely on parameterisations instead. Both aspects may significantly add to the uncertainty in the prediction of the long-term AMOC evolution.

Reasons for a long-term greenhouse gas induced reduction of the strength of the AMOC include straightforward effects, such as warming of surface waters, melting of continental ice sheets acting to reduce high latitude salinity (a mechanism not included in many climate models), and intensification of the hydrological cycle [40]. All of these act to impede deep-water formation.

More complex feedbacks (that either stabilise or de-stabilise the AMOC) also involve wind field changes in the deep-water formation regions leading to buoyancy flux anomalies [41] and oceanic teleconnections driven by changes in the freshwater budget of the tropical Atlantic and South Atlantic [36,42–45].

Huang et al. [46] found a significant increase in wind stress and its energy input into the Southern Ocean between 1950 and 2000, which may have been caused by decreasing stratospheric ozone concentrations [47]. Using a climate model, Shindall and Schmidt [48] predicted the positive wind stress trend over the Southern Ocean, which will prevail until 2100 as a consequence of anthropogenic Greenhouse gas induced global warming. While oceanographers have not yet been able to establish a relationship between the multidecadal wind stress increase and changes in the ocean circulation – which may partly be a consequence of unavailability of a sufficient number of suitable observations over the last 50 years – one might speculate that in the future, beyond the end of this century, in a then different climate 'pulling', the AMOC in the Southern Ocean might gain in importance relative to 'pushing' it in the North Atlantic [49].

## 6. CAN WE DETECT CHANGES IN THE AMOC? IS THE AMOC CHANGING ALREADY?

Previously, direct estimates of the vigour of the AMOC have been obtained from transatlantic hydrographic (density profile) sections, assuming the geostrophic balance to hold on the ocean interior. Five such sections have been carried out along 24.5°N in the Atlantic over the last 50 years [50]. Taken together the five snapshots, each of which is assumed to be representative of the annual mean strength of the AMOC of the year in which they where taken (i.e. intra-seasonal variations assumed to be small), implies an AMOC slowdown of 30% (or 8 Sv) since 1957 [50]. Other measurements, focusing on single components of the AMOC gave rather inconclusive results regarding long-term AMOC changes. Using a combination of direct and indirect transport measurement techniques a gradual 1–2 Sv decrease in the amount of cold, dense inflow of deep waters from the Nordic Seas through the Faroe Bank Channel (feeding the NADW) has been found since 1970 [51], implying a long-term AMOC weakening. However, the continuation of the direct measurements showed an increase over the last few years back to the levels of the mid 1990s [52,53].

At the same time measurements in the Deep Labrador Current – which represents a major pathway for the export of NADW from the deep water formation regions – seems to have strengthened by 15% when comparing the 1996–1999 to 2000–2005 periods [54]. However, measurements in the DWBC further south off Grand Banks gave no significant change over roughly the same period [55]. It is uncertain how representative the strength of the DWBC off Grand Banks is, for the basin wide AMOC. Hydrographic measurements in the mid and high-latitude North Atlantic suggest that a substantial part of the southward export of NADW might be accomplished along a pathway in the ocean interior that feeds into the DWBC only in the subtropical North Atlantic [56]. Kanzow et al. [57] showed from observations and model simulations that fluctuations in the strength of the DWBC may not be a good indicator of AMOC changes in the tropical North Atlantic either, due to the presence of time-variable deep offshore recirculations.

A pilot system to measure the strength of the AMOC continuously at 26.5°N (i.e. the zonally integrated meridional transport profile between Florida and Morocco) has been operating since April 2004 [9,16,58]. Figure 5 shows a 1-year long time series of the AMOC between April 2004 and April 2005, exhibiting a time mean of 18.5 Sv and a rms variability of ±5.6 Sv [16]. The range of values the AMOC assumed within one year spans roughly 30 Sv (varying between 5 and 35 Sv). The observed intra-seasonal variability raises concerns

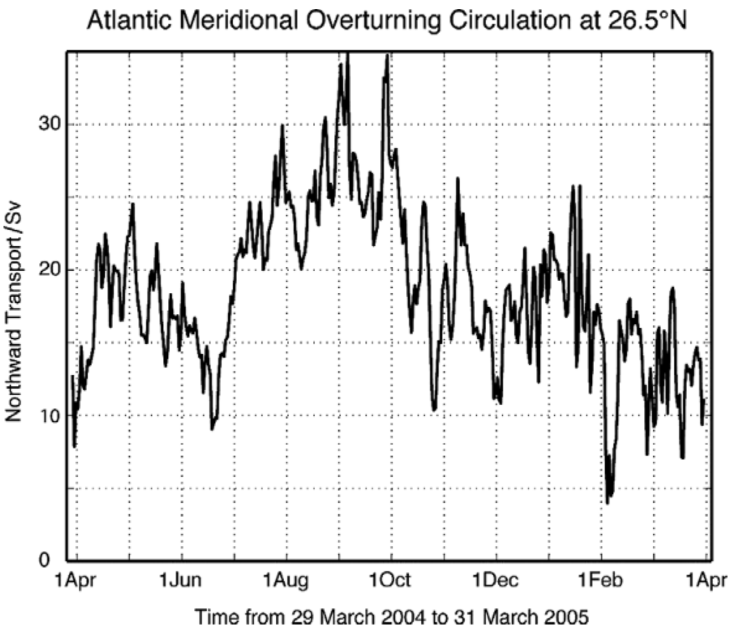


**FIGURE 5** Time series of the strength of the Atlantic meridional overturning at 26.5°N, based on the continuous transport measurements within the RAPID/MOCHAexperiment [16], defined as the vertical integral of the transport per unit depth down to the deepest northward velocity (~1100 m) on each day. It represents the sum of the Florida Current, Ekman and upper mid-ocean transports [16]. This figure was published by Kanzow et al. [59].

whether the hypothesised 30% slowdown of the AMOC [50] may represent aliasing effects (as a consequence of not resolving the large intra-seasonal variations) rather than a sustained change of the ocean circulation [16,59].

While oceanographers have not yet been able to document a statistically significant trend in the strength of the AMOC, it is worth asking, how much time it would take to detect a possible long-term trend from continuous measurements at 26.5°N. Making assumptions about the short term noise level of the AMOC, Baehr et al. [60] concluded from the analysis of an AMOC future projection, that a 0.75 Sv per decade decline could be detected after three decades. A more abrupt (than currently expected) AMOC change would be detectable earlier. The detectability could most likely be shortened significantly if several continuously observing AMOC monitoring system were operated simultaneously at different latitudes.

## 7. CONCLUSION

Observations have revealed that patterns of present-day regional and large-scale ocean circulation may display strong changes on intra-seasonal to multi-decadal time scales. Physical oceanographers have developed a variety of tools to quantify circulation changes, which involve direct and indirect measurement techniques and numerical simulations. While most of the documented present-day circulation changes are believed to fall within the class of natural (ocean climate) variability even at decadal and longer time scales, it is a non-trivial task to disentangle climate variability from presently possibly ongoing climate shifts.

The AMOC has been in the focus of climate change research. The interpretation of palaeo-climate records in the light of findings from numerical climate models reveals that the AMOC has undergone large changes in the Earth's past and that these went along with climate shifts in the North Atlantic sector and beyond. In the present day climate, the AMOC represents the major oceanic mechanism of meridional heat transport. The AMOC moves volumes of cold waters (having sunk at high latitudes) southward throughout the Atlantic at depth and keeps them out of contact with the atmosphere for centuries, until the waters rise to the upper ocean eventually. Thereby the AMOC ventilates the deep ocean with oxygen rich waters. The sinking of waters in the Nordic Seas and the Labrador Sea (push) and their eventual rising (pull) are necessary ingredients for the existence of the AMOC, both of which are thought to change in a changing climate.

Model projections imply that the AMOC might slow down between 0 and 50% by the end of the twenty-first century. This is thought to be due to an increase in vertical density stratification at high latitudes (both due to warming and freshening of surface waters) as a result of global warming. However, none of the present-day climate models have a sufficiently fine spatial resolution to resolve the processes that govern either the sinking or the rising, and have to rely on parametrisations instead. Additionally, the climate model projections that produced the range of 0–50% in AMOC decline all rely on the same greenhouse gas forcing scenario, which will inevitably differ from the actual one. Thus, the true range of uncertainty of the future evolution of the AMOC is even larger. There is clearly a need to monitor the state of the AMOC continuously over coming decades.

To date there is no clear evidence that the AMOC has started to decrease in strength, partly because it has only very recently been a subject of continuous monitoring. Indeed a reliable time series of the strength of the AMOC spanning the last 50 year (or so) does not exist. The recent continuous measurements at 26.5°N suggest that the amplitude of intra-seasonal variations of the strength of AMOC is larger than previously thought. This makes it doubtful whether reliable estimates of long-term AMOC changes can be inferred from the few sporadic attempts to estimate the strength of the AMOC, that have been done in the past. Measurements that have focused on the observation of one particular aspect rather than the whole AMOC (such the strength of the DWBC or the deep Nordic Sea inflow into the Atlantic) do not show clear, uniform trends either. In addition it is difficult to assess how a change in one of the components translates into a change of the whole

AMOC. However, even with the recently started suitable continuous AMOC observations now well under way, the detection of a possible ongoing, global-warming-induced decline in the vigour of the AMOC may still be decades away, unless more observing systems are put into place. This will strongly depend on how fast the decline actually comes about, if at all.

The major difficulty for scientists to document ocean circulation changes on climate relevant time scales arises from the sparseness of historical in situ observations both in space and time. Over the last decades it has been (and partly still is) a technical, logistical and financial challenge to maintain ocean observatories at key locations continuously for more a few years and/or to repeat measurement campaigns at a frequency that is sufficient to detect trends with a high level of confidence. However, the awareness that understanding the processes that govern ocean circulation changes may be vital for present and future societies has triggered dedicated, internationally coordinated field programmes and along with them technical developments (such as autonomous in situ profilers or advances in remote sensing). As a consequence physical oceanography is currently undergoing a step change in capacity, capability and understanding, from which future generations will certainly profit.

## REFERENCES


1. T.N. Lee, W.E. Johns, R. Zantopp, E.R. Fillenbaum, J. Phys. Oceanogr. 26 (1996) 962–963.
2. D. Zhang, M.J. McPhaden, Ocean Model. 15 (2006) 250–273.
3. S. Häkkinen, P. Rhines, Science 304 (2004) 555–559.
4. M. Bersch, I. Yashayaev, K.P. Koltermann, Ocean Dyn. 57 (2007) 223–235, doi:10.1007/s10236-007-0104-7.
5. J.W. Hurrell, Y. Kushnir, G. Ottersen, M. Visbeck, in: J.W. Hurrell, Y. Kushnir, G. Ottersen, M. Visbeck (Eds.), The North Atlantic Oscillation, Geophys. Monogr. Ser., vol. 134, American Geophysical Union, Washington, DC, 2003, pp. 1–36.
6. H. Hàtùn, A.B. Sandø, H. Drange, B. Hansen, H. Valdimarsson, Science 309 (2005) 1841–1844.
7. J.R. Knight, R.J. Allan, C.K. Folland, M. Vellinga, M.E. Mann, Geophys. Res. Lett. 32 (2005) L20708, doi:10.1029/2005GL024233.
8. J.W. Hurrell, M. Visbeck, A. Busalacchi, R.A. Clarke, T.L. Delworth, R.R. Dickson, W.E. Johns, K.P. Koltermann, Y. Kushnir, D. Marshall, C. Mauritzen, M.S. McCartney, A. Piola, C. Reason, G. Reverdin, F. Schott, R. Sutton, I. Wainer, D. Wright, J. Clim. 19 (2006) 5100–5121.
9. W.E. Johns, L.M. Beal, M.O. Baringer, J.R. Molina, S.A. Cunningham, T. Kanzow, D. Rayner, J. Phys. Oceanogr. 38 (2008) 605–623.
10. T.M. Joyce, D.S. Bitterman, K.E. Prada, Deep-Sea Res. 29 (1992) 903–913.
11. J. Fischer, M. Visbeck, J. Atm. Ocean Technol. 10 (1993) 764–773.
12. R. Davis, W. Zenk, in: G. Siedler, J. Church, J. Gould (Eds.), Ocean Circulation and Climate, Acadamic Press, New York, 2001.
13. V.W. Ekman,Arch. Math. Astron. Phys. 2 (1905) 1–53.

14. G. Wüst, Wissenschaftliche Ergebnisse der Deutschen Atlantischen Expedition. “Meteor,” 1925–1927, vol. 6, 1935, pp. 109–288 (In English, W.J. Emer y (Ed.), The Stratosphere of the Atlantic Ocean, Amerind, New Delhi, 1978, 112 pp.).
15. H. Longworth, H.L. Bryden, in A. Schmittner, J.C.H. Chiang, S.R. Hemmings (Eds.), Geophys. Monogr. Ser., vol. 173, American Geophysical Union, Washington, DC, 2007.
16. S.A. Cunningham, T. Kanzow, D. Rayner, M.O. Barringer, W.E. Johns, J. Marotzke, H.R. Longworth, E.M. Grant, J.J.-M. Hirschi, L.M. Beal, C.S. Meinen, H.L. Bryden, Science 317 (2007) 935–938.
17. T. Kuhlbrodt, A. Griesel, M. Montoya, A. Levermann, M. Hofmann, S. Rahmstorf, Rev. Geophys. 45 (2007) RG2001, doi:10.1029/2004RG000166.
18. J.M. Lilly, P.B. Rhines, M. Visbeck, R. Davis, J.R.N. Lazier, F. Schott, D. Farmer, J. Phys. Oceanogr. 29 (1999) 2065–2098.
19. J. Marshall, F. Schott, Rev. Geophys. 37 (1999) 1–64.
20. C. Wunsch, R. Ferrari, Ann. Rev. Fluid Mech. 36 (2004) 281–314.
21. K.L. Polzin, J.M. Toole, J.R. Ledwell, R.W. Schmitt, Science 276 (1997) 93–96.
22. F. Paparella, W.R. Young, J. Fluid Mech. 466 (2002) 205–214.
23. J.R. Toggweiler, B. Samuels, J. Phys. Oceanogr. 28 (1998) 1832–1852.
24. A.H. Orsi, W.M. Smethie, J.L. Bullister, J. Geophys. Res. 107 (2002) 3122, doi:10.1029/2001JC000976.
25. G.C. Johnson, J. Geophys. Res. 113 (2008) C05027, doi:10.1029/2007JC004477.
26. G. Boccaletti, R. Ferrari, A. Adcroft, D. Ferreira, J. Marshall, Geophys. Res. Lett. 32 (2005) L10603, doi:10.1029/2005GL022474.
27. H.L. Bryden, S. Imawaki, in: G. Siedler, J. Church, J. Gould (Eds.), Ocean Circulation and Climate, Academic Press, San Diego, 2001, pp. 455–474.
28. J.L. Sarmiento, N. Gruber, Ocean Biogeochemical Dynamics, Princeton University Press, Princeton, 2006.
29. S.J. Johnson, H.B. Clause, W. Dansgaard, K. Fuhrer, N. Gundestrup, C.H. Hammer, P. Iversen, J. Jouzel, B. Stauffer, J.P. Steensen, Nature 311 (1992) 313.
30. W. Dansgaard et al., Nature 364 (1993) 218–220.
31. M. Sarnthein, K. Winn, S.J.A. Jung, J.-C. Duplessy, L. Labeyrie, H. Erlenkeuser, G. Ganssen, Paleoceanography 9 (1994) 209–267.
32. J.F. Adkins, E.A. Boyle, L. Keigwin, E. Cortijo, Nature 390 (1997) 154–155.
33. A. Ganopolski, S. Rahmsdorf, V. Petoukhov, M. Claussen, Nature 391 (1998) 351–356.
34. A. Ganopolski, S. Rahmsdorf, Nature 409 (2001) 153–158.
35. S. Hemming, Rev. Geophys. 42 (2004) RG1005, doi:10.1029/2003RG000128.
36. M. Vellinga, R.A. Wood, J.M. Gregory, J. Clim. 15 (2002) 764–780.
37. M. Sarnthein, P.M. Grootes, J.P. Kennett, M.-J. Nadeau, in: A. Schmittner, J. Chiang, S. Hemming (Eds.), Geophys. Monogr. Ser., vol. 173, American Geophysical Union, Washington, DC, 2007, pp. 175–196.
38. IPCC, Climate Change, The Scientific Basis: Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel of Climate Change, Cambridge University Press, New York, 2007.
39. A. Schmittner, M. Latif, B. Schneider, Geophys. Res. Lett. 32 (2005) L23710, doi:10.1029/2005GL024368.
40. J.M. Gregory et al., Geophys. Res. Lett. 32 (2005) L12703, doi:10.1029/2005GL023209.
41. P. Gent, Geophys. Res. Lett. 28 (2001) 1023–1026.
42. M. Latif, E. Roeckner, U. Mikolajewicz, R. Voss, J. Clim. 13 (2000) 1809–1813.
43. R.B. Thorpe, et al., J. Clim. 14 (2001) 3102–3116.

44. A.X. Hu, G.A. Meehl, W.M. Washington, A. Dai, J. Clim. 17 (2004) 4267–4279.
45. U. Krebs, A. Timmermann, J. Clim. 20 (2007) 4940–4956.
46. R.X. Huang, W. Wang, L.L. Liu, Deep-Sea Res. II 53 (2006) 31–41.
47. D.W. Thompson, S. Solomon, Science 296 (2002) 895–899.
48. D.T. Shindell, G.A. Schmidt Geophys. Res. Lett. 31 (2004) L18209, doi:10.1029/2004GL020724.
49. M. Visbeck, Nature 447 (2007) 383.
50. H.L. Bryden, H.L. Longworth, S.A. Cunningham, Nature 438 (2005) 655–657.
51. B. Hansen, W.R. Turrell, S. Østerhus, Nature 411 (2001) 927–929.
52. B. Hansen, S. Østerhus, Prog. Oceanogr. 75 (2007) 871–865, doi:10.1016/j.pocean.2007.09.004.
53. S. Østerhus, T. Sherwin, D. Quadfasel, B. Hansen, in: R.R. Dickson, et al. (Eds.), Arctic-Subarctic Ocean Fluxes, Springer Science, Business Media B.V., Berlin, 2008, pp. 427–441.
54. M. Dengler, J. Fischer, F.A. Schott, R. Zantopp, Geophys. Res. Lett. 33 (2006) L21S06, doi:10.1029/2006GL026702.
55. F.A. Schott, J. Fischer, M. Dengler, R. Zantopp, Geophys. Res. Lett. 33 (2006) L21S07, doi:10.1029/2006GL026563.
56. S. Lozier, Science 277 (1997) 361–364, doi:10.1126/science.277.5324.361.
57. T. Kanzow, U. Send, M. McCartney, Deep-Sea Res. I 55 (2008) doi:10.1016/j.dsr.2008.07.011.
58. T. Kanzow, S.A. Cunningham, D. Rayner, J.J.-M. Hirschi, W.E. Johns, M.O. Baringer, H.L. Bryden, L.M. Beal, C.S. Meinen, J. Marotzke, Science 317 (2007) 938–941.
59. T. Kanzow, J.J.-M. Hirschi, C. Meinen, D. Rayner, S.A. Cunningham, J. Marotzke, W.E. Johns, H.L. Bryden, L.M. Beal, M.O. Baringer, J. Oper. Oceanogr. 1 (2008) 19–28.
60. J. Baehr, H. Haak, S. Alderson, S.A. Cunningham, J.H. Jungclaus, J. Marotzke, J. Clim. 20 (2007) 5827–5841.