The North Atlantic Ocean and climate change in the UK and northern Europe

Neil C. Wells
National Oceanography Centre, University of Southampton

Introduction

The surface climate of the UK and northern Europe is up to 9 degC warmer than it would be if the North Atlantic Ocean did not transport a large quantity of heat northwards to our shores. This unusual warming has been known since sea temperature records were established in the late nineteenth century and was probably well known to the early seafarers in the last two millennia.

In this article I will first explain the background for this unusual warming, as presently understood by oceanographers; this will be followed by how the ocean heat transport is measured and then a discussion of what has been learnt in the last 10 years from the RAPID array and the associated heat and salt changes in the upper ocean. A remarkable and unexpected event in 2009 will also be discussed.

Unusual warming of North Atlantic Ocean and northern Europe

To understand the unusual warming to the west of the UK we need to appreciate the heat balance of the Earth's surface. Polewards of 40° latitude, in both hemispheres, the Earth's surface is cooling because the outgoing thermal radiation to space exceeds the incoming solar radiation averaged over a year. However, in lower latitudes (40°S to 40°N) the Earth's surface is warming because the solar radiation is greater than the outgoing thermal radiation.

Circulations in both atmosphere and ocean are driven by the imbalance in radiation (incoming solar radiation minus outgoing thermal radiation), and they act to transport heat polewards to cool the lower latitudes and warm the higher latitudes. In the troposphere, outside of the tropics, extra-tropical cyclones and anticyclones associated with planetary waves transport large quantities of heat polewards. The oceans also contribute about 40% of the total heat transport (atmosphere and ocean) polewards from the tropical regions, but, unusually, the global ocean circulation transports additional heat from the Indian and Pacific Oceans into the Atlantic Ocean. The northward heat transport in the North Atlantic at 26°N is 1.3 PW, compared with the North Pacific transport of 0.7 PW (Bryden and Hall, 1980; Johns *et al.*, 2011; Wells, 2012). It is this extra heat from the Indian and Pacific Oceans, together with the heat absorbed in the tropical Atlantic, which is transported into the higher latitudes of the North Atlantic and warms the UK and northern Europe.

It may be surprising to learn that the quantitative measurements of this heat transport could not be determined until oceanographers started taking systematic measurements of the temperature and salinity across ocean basins and from the surface to the ocean floor. These started in the North Atlantic during the International Geophysical Year (IGY) in 1957 and have continued roughly once a decade since that time.

They have established that the mean northward transport of heat is about 1.3 PW at 26°N (Johns *et al.*, 2011). About 60% of this heat is transported across 40°N, where a large proportion of the heat is given up to the atmosphere to warm the westerly winds as they travel from North America to Europe.

Heat transport by the atmosphere and ocean is not the only process important for the Earth's climate. For example, seasonal heat storage released during the winter months has a moderating influence on the climate of northern Europe. This will be discussed later.

Measurement of heat transport

To measure the ocean heat transport it is necessary to measure a complete section from one coast to another coast, usually at a specific latitude, to make sure there is no net mass transfer across the basin[1]. This means the mass of the warm northward flow in the upper ocean (~0–1100m) has to be balanced by the mass of southward colder flow in the deeper ocean (1100–5000m) because the Atlantic is effectively a closed basin. Below 5000m there is a small northward flow associated with a dense bottom water mass which originates from Antarctica.

Full ocean sections were the main aspect of the international World Ocean Circulation Experiment (WOCE) from 1990 to 1997 which surveyed all the main ocean basins including the Southern Ocean.

Analysis of these ocean sections has shown that there has been a 30% decrease in heat transport from 1957 to 2004 (Bryden *et al.*, 2005). However, because these measurements were snapshots of the ocean taken over a period of less than 2 months in a particular year, there were concerns about the variability of this heat transport between the time of the ocean sections. If the true variability was much larger than that estimated then the 30% decrease may not be statistically significant. The results from this study could not be ignored because of the impact of such a change on the climate of the UK and northern Europe, which could see a substantial decrease in its climatological temperature if this heat transport was reduced or even stopped (Srokosz *et al.*, 2012).

The Rapid Array at 26°N

In 2004 the Rapid Array was established at 26°N (a latitude where the heat transport measurements could be made in a regular and robust way every 10 days) by scientists from the National Oceanography Centre and University of Southampton, UK and by US scientists based in Miami, USA (Rayner *et al.*, 2011). The funding for this multi-million pounds project came from Natural Environmental Research Council in the UK and National Science Foundation in the USA.

To measure the northward heat transport at 26°N it is necessary to measure the total meridional flow across the North Atlantic from Florida, USA to Northwest Africa and from the surface to the ocean floor in excess of 5000m depth, as shown in Figure 1, at a

[1]There is a net mass transport across 26°N which has to be removed before the determination of heat transport.

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RMetS Royal Meteorological Society

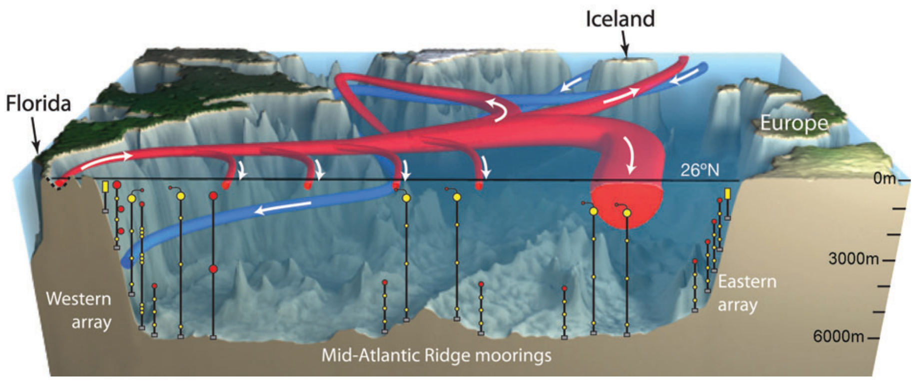

Figure 1. Schematic of the meridional overturning circulation (MOC) in the North Atlantic Ocean. The red arrows show the warm upper limb of the MOC (0–1100m) and the blue arrows show the cold southward lower limb of the MOC (1100–6000m). The measurement array at 26°N is also shown. (Source: Louise Bell and Neil White, CSIRO, Australia.)

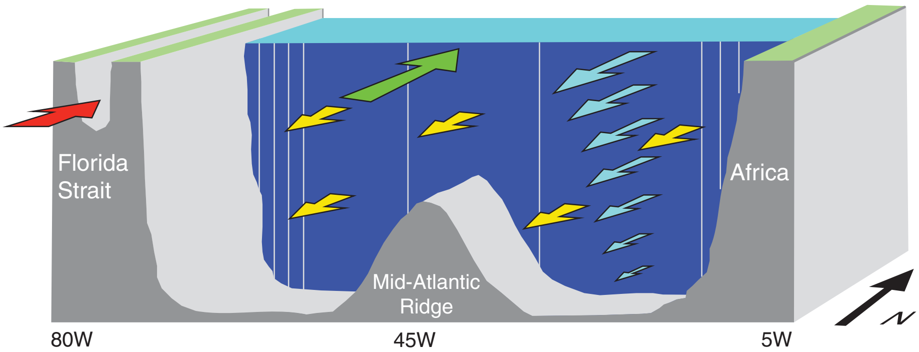

Figure 2. The components of the MOC at 26°N. The red arrow represents the northward Gulf Stream; the green arrow represents the northward surface Ekman flow; the yellow arrows represent the southward return flow. The light blue arrows show the adjusted flow in the ocean interior to conserve mass across the section (see text). (Source: http://www.rapid.ac.UK./rapidmoc/figure 5)

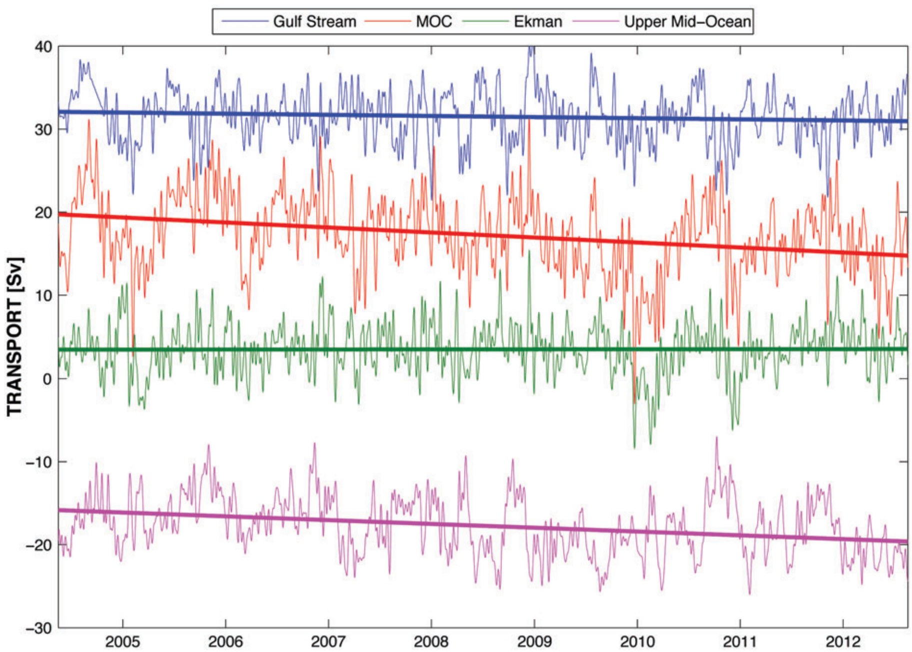

Figure 3. Time series of the components of the MOC and the MOC itself from 2004 to 2012. Positive transport is northward. The Gulf Stream is blue; the surface Ekman flow is green; the upper mid-ocean return flow is magenta. Overturning transport (MOC) is then the sum of the Gulf Stream, Ekman and upper mid-ocean transports and represents the maximum northward transport of upper layer waters on each day. The MOC is red (second from top) and is the sum of the previous three flows. The transport is measured in Sverdrups (Sv) and is equivalent to $10^6 m^3 s^{-1}$. Note that mass conservation is equivalent to volume conservation across the section. (Source: http://www.rapid.ac.uk/rapidmoc/figure 1)

frequency of 10 days. This flow is known as the meridional overturning circulation (MOC). The northward-flowing limb is situated between the surface and a depth of about 1100m, whilst the southward-flowing lower limb occupies the depth from 1100 to 5000m. Below 5000m is another, much weaker, MOC circulation which carries the Antarctic Bottom Water northwards across 26°N.

The 26°N latitude was chosen for the array because over 90% of the heat transport at 26°N is associated with the MOC (Johns *et al.*, 2011) and therefore most of the heat transport variability can be determined from MOC variability. This is also the latitude of the maximum heat transport in the North Atlantic Ocean (Bryden and Hall, 1980).

The remaining 10% of the heat transport is associated with the horizontal circulation of the North Atlantic subtropical gyre. Measurements of temperature and salinity across this latitude–depth section of the North Atlantic allow not only the heat transport to be determined but also the freshwater transport. These latter measurements are very important because they help in understanding the link between the polar oceans, where precipitation, melting glacial ice and rivers are adding fresh water to the oceans, and the tropical oceans, where fresh water tends to be removed from the ocean due to evaporation exceeding precipitation and runoff from rivers and ground water.

The array measures the following components to determine the MOC:

(a) The northward Gulf Stream transport in the Florida Straits (which has a mean and standard deviation of 31.6 ± 3.0Sv ($1Sv = 10^6 m^3 s^{-1}$))
(b) The northward wind driven Ekman[2] transport in the top 100m (which has a mean value of 3.5 ± 3.0Sv)
(c) The southward transport below the Ekman layer and above 1100m over the remainder of the ocean basin which has a mean value of −17.7 ± 2.6Sv. This is known as the Upper Mid-Ocean transport.

The sum of these three components makes up the total northward transport in the upper limb of the MOC, and this is the RAPID measure of the total MOC (Figure 2). The 1100m depth is the level of no motion where the surface northward flow reverses its direction to a southward flow below 1100m. (See Rayner *et al.*, 2011 for a more detailed account of the determination of the level of no motion and the conservation of mass across the section).

[2]The Ekman transport is at 90° and to the right of the direction of the surface wind stress in the Northern Hemisphere. Therefore a westward zonal wind stress will produce a northward meridional flow.

The time series of the MOC and its individual components are shown in Figure 3. These measurements have shown that the mean and standard deviation of the MOC is 17.4 ± 4.5Sv, and that there is variability from weekly to interannual timescales. The RAPID array measurements have also shown that there is a seasonal cycle of the MOC at 26°N, despite the higher frequency variability at weekly timescales. The maximum MOC and heat transport tends to occur in the late summer and the minimum transport in late winter (Kanzow, 2009). This is rather counter-intuitive because if the ocean behaved in a similar way to the atmosphere the largest northward heat transport would be expected to be in the winter.

Perhaps more remarkably, Figure 3 shows that the MOC can change significantly from one year to the next. In 2009 the MOC decreased by 30% from its mean transport over the period 2004–2008. This decrease in the MOC transport in 2009 was unexpected by the RAPID scientists (McCarthy *et al.*, 2012). Previous measurements were not frequent enough to detect this short timescale event. Furthermore, this was similar to the decrease found from 1957 to 2004 (Bryden *et al.*, 2005) and similar to the predicted decrease of MOC over the twenty-first century from IPCC AR5 climate prediction models (IPCC, 2013), but over only 1 year. The reasons for this large change in the MOC in 2009 are not fully understood, and they have not yet been seen in any of the IPCC climate models.

The downward trend in the MOC from 2004 to 2012 (Figure 3) is related to a decrease in the southward Upper Mid-Ocean transport, whilst the Gulf Stream shows little or no trend and the Ekman transport shows a small increase in the northward direction (Smeed *et al.*, 2014).

Heat storage

Variations of the northward ocean heat flux at 26°N will have implications for the ocean heat storage in the North Atlantic. If an ocean box is imagined between 26°N and 70°N, then the heat can either go into the atmosphere, the ocean storage between 26°N and 70°N, or into the Arctic Ocean. The latter is small compared with the first term. If, over a long period of time, the heat storage remained constant, then all the heat transport would go into the atmosphere.

However, this heat transport is not in a steady state on seasonal to interannual scales, and, therefore, fluctuations in the ocean heat storage as well as the heat exchange with the atmosphere will take place.

On an annual timescale the seasonal ocean heat content is dominated by the seasonal cycle of solar radiation and vertical heat exchanges with the atmosphere. For example, at 40–50°N the seasonal cycle of net heat gain (solar radiation gain is moderated by heat lost to the atmosphere by thermal radiation, latent heat of evaporation, and sensible heat) is in the range of 150–200Wm^{-2} into the ocean from May to August, with similar magnitudes of heat loss from the ocean between November and February (Hadfield *et al.*, 2007).

By contrast, if all the heat transported across 26°N was lost to the atmosphere between 30°N and 70°N, the heat flux out of the ocean would be about 45Wm^{-2}. This number is not negligible, but it is significantly smaller than the seasonal cycle of heat exchange between the ocean and atmosphere.

On the seasonal timescale the atmosphere–ocean heat exchange and the heat transport appear to be varying in a complex way. However, a large part of the atmosphere–ocean heat exchange concerns the upper 100m in the tropical latitudes and most of the mid-latitudes. In the higher latitudes there can be deeper mixing, particularly in the Irminger Sea, Greenland Sea and the Labrador Sea, and this mixing may extend to over 1000m in late winter (Wells *et al*., 2009).

To obtain an idea of the importance of the heat storage on the climate system, consider a hypothetical situation where the heat transported across 40°N was not lost to the atmosphere but instead changed the temperature of the upper 1000m of the Northeast Atlantic to the west of the UK between 40°N and 60°N. Let us assume the anomalous warmth of this region is 5 degC above expected for this latitude without the ocean heat transport. If we assume the ocean loses 45Wm^{-2} to the atmosphere, then it would take 14 years to cool this volume of water by 5 degC. In a single year the ocean could cool by a few tenths of a degree.

This estimate suggests that if the MOC and its associated heat transport stopped, this part of the ocean could cool by a few degrees in one or two decades.

In 2009 it was observed that the MOC at 26°N (Figure 3) reduced by 30% from its mean transport over a year and the northward heat transport was reduced by a similar amount.

This resulted in a cooling of the region from 30 to 40°N in the upper ocean (surface to 1000m including the main thermocline) (Cunningham *et al.*, 2013). Figure 4 shows the changes of upper ocean heat content from 2008 to 2010 based on two different methods for estimating ocean heat storage, mainly from Argo. An interesting aspect of this event was that the ocean cooling was limited to subtropical latitudes (20–50°N). At both lower latitudes of 10–20°N and higher latitudes of 50–70°N estimates indicate a warming of the upper ocean. The warming at the lower latitudes is consistent with the reduction in the northward heat transport at 26°N.

However, the warming at the higher latitudes suggests that changes in the MOC at 26°N and its associated change in the heat transport are not necessarily reflected in a change in the ocean heat transport at higher latitudes. In this case there could be a change in ocean heat content between the two latitudes which may persist for a few years. This gives an indication of the thermal inertia of the ocean and the longer timescale that it adds to the climate system (Wells, 2012).

The atmosphere in 2009–2010 also showed some interesting behaviour. For example, Buchan *et al.* (2014) have shown that there were exceptional cold winters in northern Europe associated with extremely low values of the North Atlantic Oscillation (NAO). The coincidence of the low NAO with the reduction in the MOC is remarkable. The

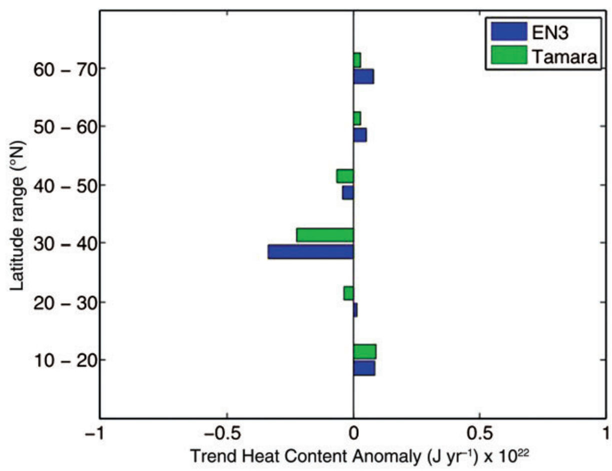

Figure 4. The trend in zonally integrated ocean heat storage (Joules per year), in the 0–1000m layer, in 10° latitude bands, from 2008 to 2010. The MOC at 26°N decreased by about 30% during this period (see text). TAMARA and EN3 are two different estimates of the heat storage from the University of Southampton and the Met Office respectively.

causal links between the atmosphere and ocean in this period are still being discussed. Some evidence suggests the atmosphere is driving this event (Roberts *et al.*, 2013) whilst others suggest the signals for this event may come from the ocean (Bryden *et al.*, 2014).

Salt storage and freshwater transport

Estimates of salt storage in the North Atlantic can also be made from the Argo array. In the North Atlantic there are about 200 Argo floats which are sufficient to obtain reasonably good estimates of the heat and salt storage every month, from the start of the RAPID array in April 2004 to the present day. These Argo floats sample the temperature and salinity between the surface and a depth of 2000m using consistent and accurate sensors. They sample at regular intervals of about 10 days and therefore 200 floats can provide 7200 profiles each year (Hadfield *et al.*, 2007; Wells *et al.*, 2009; Ivchenko *et al.*, 2010). Unlike heat storage there are generally only small changes in salinity over the seasonal cycle, but it can change on decadal and longer timescales.

Monitoring of the freshwater transport[3] is important because of its influence of the vertical stability of the ocean, particularly in the region of the North Atlantic subpolar gyre and Greenland-Iceland-Norwegian Seas. Lighter freshwater (lower salinity) at the ocean surface will increase the stability of the ocean and reduce the occurrence of deep convection. This in turn reduces the strength of the MOC (National Research Council Committee on Abrupt Climate Change, 2002). In the North Pacific Ocean the overturning circulation is very much weaker because of a surface layer of fresher water which provides a stable vertical density profile (Wells, 2012).

The IPCC models (IPCC, 2013) have suggested that melting of glaciers and the Greenland ice sheet will add more fresh water into the North Atlantic. Furthermore, higher precipitation at the polar and subpolar latitudes due to global warming will further increase the freshwater input to these latitudes in this century.

Summary

This article has indicated that the North Atlantic Ocean is showing changes in its circulation as represented by the MOC at 26°N in the last 10 years. The changes in the MOC are associated with heat transport which has a direct effect on the upper ocean heat storage northwards of the RAPID array. The event in 2009 caused a cooling of the subtropical ocean between 20 and 40°N but did not appear to influence the region poleward of 50°N. The role of the atmosphere in the changes in the MOC in this region, in particular on inter-annual and decadal timescales, is still not well understood.

The IPCC AR5 scenarios (IPCC, 2013) for the world climate under global warming strongly suggest that we can expect changes to the ocean circulation in this century and future centuries.

In the North Atlantic the MOC is very important for maintaining an equable climate in the UK and northern Europe. The modelling studies and observational evidence suggest that the MOC will decrease by at least 30% by the end of this century and that this will reduce the heat transport to the UK.

The continuous measurement of the MOC at 26°N provides a valuable insight into the changes in circulation within the North Atlantic. It is a scientific experiment which has a direct consequence to our climate in the UK and northern Europe, and it is imperative that we continue to measure the MOC circulation and its trends, and understand the causes of its observed variability.

[3]The total salt content of the ocean only changes on geological timescales and therefore net evaporation of fresh water will increase the salinity whilst precipitation will decrease the salinity.

There is a freshwater transport from regions which have an excess precipitation to regions where is there an excess of evaporation.

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Correspondence to: Neil C. Wells

ncw@noc.soton.ac.uk

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