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Sensitivity of modern climate to the presence, strength and salinity of Mediterranean-Atlantic exchange in a global General Circulation Model.

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Keywords

Mediterranean Outflow Water; Mediterranean salinity; North Atlantic circulation; North Atlantic climate; Atlantic Meridional Ocean Circulation (AMOC); North Atlantic Deep Water (NADW)

Abstract

Mediterranean Outflow Water (MOW) is thought to be a key contributor to the strength and stability of Atlantic Meridional Overturning Circulation (AMOC), but the future of Mediterranean-Atlantic water exchange is uncertain. It is chiefly dependent on the difference between Mediterranean and Atlantic temperature and salinity characteristics, and as a semi-enclosed basin, the Mediterranean is particularly vulnerable to future changes in climate and water usage. Certainly, there is strong geologic evidence that the Mediterranean underwent dramatic salinity and sea-level fluctuations in the past.

Here, we use a fully coupled atmosphere-ocean General Circulation Model to examine the impact of changes in Mediterranean-Atlantic exchange on global ocean circulation and climate. Our results suggest that MOW strengthens and possibly stabilises the AMOC not through any contribution towards NADW formation, but by delivering relatively warm, saline water to southbound Atlantic currents below 800 m. However, we find almost no climate signal associated with changes in Mediterranean-Atlantic flow strength. Mediterranean salinity, on the other hand, controls MOW buoyancy in the Atlantic and therefore affects its interaction with the shallow-intermediate circulation patterns that govern surface climate. Changing Mediterranean salinity by a factor of two reorganises shallow North Atlantic circulation, resulting in regional climate anomalies in the North Atlantic, Labrador and Greenland-Iceland-Norwegian Seas of ±4 °C or more. Although such major variations in salinity are believed to have occurred in the past, they are unlikely to occur in the near future. However, our work does suggest that changes in the Mediterranean's hydrological balance can impact global-scale climate.

1. Introduction

The exchange of water between the Mediterranean and Atlantic, which today occurs through the Gibraltar Straits, is an important control on Mediterranean water temperature and salinity characteristics (Béthoux and Pierre 1999; Bethoux and Gentili 1999; Bethoux et al. 1999; Cacho et al. 2002; Gómez 2003; Dubois et al. 2011). These in turn affect thermohaline ocean circulation in the Mediterranean basin, redistributing heat and impacting regional climate (Candela 1991; Alhammoud et al. 2010; Sanchez-Gomez et al. 2011). But far from being unilateral, the system feeds back into itself. The geometry of the Gibraltar Straits; which is influenced by processes of erosion, tectonics and changes in eustatic sea level (Bethoux and Gentili 1999; Loget and Vandendriessche 2006; Govers et al. 2009; Alhammoud et al. 2010); governs the volume of water that can physically pass between basins at any one time and so can be described as the primary control on Mediterranean-Atlantic exchange (e.g. Stommel and Farmer 1952, 1953; Bryden and Kinder 1991; Bryden et al. 1994; Rogerson et al. 2012). However, salinity and temperature exchange through the Straits is also regulated by the density gradient across it (e.g. Bryden et al. 1994; Thorpe and Bigg 2000; Mariotti et al. 2002; Somot et al. 2006), therefore providing a feedback system between Mediterranean water characteristics and Mediterranean-Atlantic exchange.

A local surfeit in evaporation over precipitation and runoff causes a freshwater deficit in the Mediterranean of 400-600 mm yr^{-1} (Bryden et al. 1994; Bethoux and Gentili 1999; Jungclaus and Mellor 2000; Tsimplis and Bryden 2000; Mariotti et al. 2002; Gómez 2003) and is responsible for a salinity difference of 2-3 psu between the westernmost Mediterranean and eastern North Atlantic (Boyer et al. 2009). This salinity difference dominates the density gradient across the Gibraltar Straits and so influences Mediterranean-Atlantic exchange, which acts to equalise conditions in the two basins (Bryden et al. 1994; Jungclaus and Mellor 2000). Thus, by affecting local water temperature and salinity properties, regional changes in

Mediterranean climate and circulation contribute towards the strength of Mediterranean-Atlantic exchange (e.g. Bethoux and Gentili 1999; Mariotti et al. 2002).

Interestingly, both model and observational data suggest that due to regional climate warming and the diversion of fluvial runoff for domestic and agricultural purposes, the Mediterranean freshwater deficit has increased by over 10 % in the last 40-50 years, affecting local climate and Mediterranean deep water formation (Bethoux et al. 1999; Mariotti et al. 2002; Skliris and Lascaratos 2004; Xoplaki et al. 2006; Dietrich et al. 2008; Vargas-Yáñez et al. 2010). This trend is expected to persist, possibly increasing towards the end of the 21st Century (Somot et al. 2006, 2008; Christensen and Christensen 2007; Gao and Giorgi 2008; Giorgi and Lionello 2008; Mariotti et al. 2008; Dubois et al. 2011; García-Ruiz et al. 2011; Sanchez-Gomez et al. 2011), implying that changes in Mediterranean-Atlantic flow strength are already afoot and may accelerate.

Further to its influence on the climate and thermohaline circulation of the Mediterranean, flow through the Gibraltar Straits has wider, global significance through its effect on North Atlantic Ocean circulation. For example, it has long been supposed that Mediterranean Outflow Water (MOW) contributes towards the pattern and vigour of the Atlantic Meridional Overturning Circulation (AMOC). The earliest hypotheses (Reid 1978, 1979) suggested that upon leaving the Gibraltar Straits, a core of MOW takes a direct, northward flow path to the northernmost North Atlantic and Greenland-Iceland-Norwegian (GIN) Seas, thus providing relatively saline waters to areas of deep water formation. It was proposed that upon cooling, the relatively high-salinity, MOW-origin waters contribute towards destabilising the water column and thus drive local overturning. However, this deep source hypothesis, so termed by McCartney and Mauritzen (2001), has for the most part been disproved by more recent ocean model investigations (e.g. Stanev 1992; Mauritzen et al. 2001; New et al. 2001) and observational data (e.g. McCartney and Mauritzen 2001; Bower et al. 2002a, 2002b). These newer studies favour a shallow source hypothesis (McCartney and Mauritzen 2001), suggesting that MOW spreads predominantly westwards to precondition the north-eastward flowing North Atlantic Current with relatively warm, saline waters. These waters are thus transported to the northernmost North Atlantic and GIN Seas, providing an indirect, but nevertheless important, source of warm, saline waters to the high latitude sites of NADW formation.

It should be noted that although perhaps more famous for his deep source hypothesis, Reid himself also viewed the westward flow path as the ‘most obvious’ and well documented route of MOW to the North Atlantic, commenting that subsequent mixing with intermediate North Atlantic Central Waters supplies relatively warm, high salinity water to the high latitudes (Reid 1978, 1979). Certainly the global General Circulation Models (GCMs) hitherto used to investigate MOW generally agree on its mainly westward flow into the North Atlantic at depth (Rahmstorf 1998; Bigg and Wadley 2001; Chan and Motoi 2003; Rogerson et al. 2010), broadly emulating the observed spread of relatively warm, saline, Mediterranean-origin water in modern North Atlantic Central Waters (Boyer et al. 2009).

Thus it would seem that questions over MOW’s flow path in the North Atlantic and transportation to the GIN Seas are largely resolved. Yet, uncertainty remains over the extent to which MOW is capable of influencing North Atlantic circulation and global climate. For example, there is strong model-based (Bigg and Wadley 2001; Rogerson et al. 2010) and proxy-based (including Rogerson et al. 2006, 2010; Voelker et al. 2006; Penaud et al. 2011) evidence to suggest that for cold periods in the recent geologic past (50 ka to present), strengthened Mediterranean-Atlantic exchange (and hence enhanced MOW) provided a crucial negative feedback to North Atlantic freshening, boosting NADW formation during intervals of weaker or interrupted Atlantic thermohaline circulation. Penaud et al. (2011) even propose that MOW could be the trigger for switching between stadial and interstadial AMOC modes.

Similarly, others have speculated that anthropogenic influences on the Mediterranean are affecting AMOC strength in the present day and will continue to do so in the coming decades and centuries. For example, Johnson (1997) proposed that recent and near-future amplification of the Mediterranean freshwater deficit (e.g. Somot et al. 2006; Dietrich et al. 2008; Mariotti 2010; Vargas-Yáñez et al. 2010) will raise the Mediterranean-Atlantic density gradient enough to fully deflect the Gulf Stream and thus induce Northern Hemisphere Glaciation. In an equally speculative claim, Gómez (2003) suggested that future Mediterranean climatic changes or outright damming of the Gibraltar Straits (as advised by Johnson 1997) would reduce AMOC strength and stability; a sentiment echoed by Bethoux et al. (1999) who suggest that without MOW, NADW formation could not be continuous throughout the year. To test the hypothesis put forward by Johnson (1997), Rahmstorf (1998) investigated the impact of human-induced changes in Mediterranean salinity on North Atlantic circulation and climate. Coupling his ocean GCM to a simple atmospheric energy balance model in this study, Rahmstorf concludes that the enhanced freshwater-deficit, hailed by Johnson (1997) as the bringer of the next Ice Age, has negligible impact on North Atlantic circulation and hence negligible impact on climate; the increase in Mediterranean salinity and MOW flow strength is simply too small and MOW is unable to bring about Northern Hemisphere Glaciation.

On the other hand, both Rahmstorf (1998) and Chan and Motoi (2003) suggest that MOW contributes 1-2 Sv of flow to the present day AMOC south of the Gibraltar Straits, although it has no direct impact on the overall maximum overturning strength. This is also seen in the ocean modelling study carried out by Kahana (2005). These GCM findings do, to some extent, support the postulations of Bethoux et al. (1999) and Gómez (2003) and furthermore, they show a small climatic signal associated with the existence of MOW in the North Atlantic. For example, Rahmstorf (1998) finds that the presence of MOW enhances the Gulf Stream, which warms the surface North Atlantic by up to 0.3 °C, whilst Chan and Motoi (2003) show that removing MOW for a period of several centuries, reduces meridional overturning in Antarctic Bottom Water formation and cools Southern Ocean air temperatures by up to 6 °C.

However, beyond blocking Mediterranean-Atlantic exchange (Rahmstorf 1998; Chan and Motoi 2003), perturbations to modern MOW strength and salinity in global GCM studies have so far been small (Rahmstorf 1998). So the question remains; how much influence is MOW actually able to exert over global ocean circulation and climate? To address this, we performed a series of idealised simulations of extreme changes in Mediterranean-Atlantic exchange strength and salinity. These were run using a fully-coupled atmosphere-ocean GCM (HadCM3), which, although no longer considered state-of-the-art, enables the simulations to be integrated over a period of several centuries. According to the only other published atmosphere-ocean GCM studies of MOW's impact on global ocean circulation and climate, this is necessary in order to fully capture the climate signal associated with changes in MOW (Chan and Motoi 2003) and to enable the model to reach a near steady state (Ivanovic et al. 2013). Thus, as well as emulating the standard 'on/off' GCM experiment performed by Rahmstorf (1998) and Chan and Motoi (2003), we were able to test more rigorously the ability of MOW to impact modern North Atlantic circulation and global climate.

2. Methods

In this section we briefly describe the global GCM used for this investigation, giving specific details about how the model simulates Mediterranean-Atlantic exchange. Following this, we outline the three experiments used to test the impact of (i) the presence, (ii) the strength and

(iii) the salinity of Mediterranean-Atlantic exchange on modelled ocean circulation and climate.

2.1. Model Description

For this study, we used version 4.5 of the UK Met Office's fully coupled atmosphere-ocean GCM HadCM3. The atmosphere model has a horizontal resolution of 2.5° x 3.75°, 19 vertical layers based on a hybrid vertical coordinate scheme (Simmons and Burridge 1981) and a timestep of 30 minutes. The model includes physical parameterisations for the radiation scheme (Edwards and Slingo 1996), convection scheme (Gregory et al. 1997) and land surface scheme (MOSES-1; Cox et al. 1999). The ocean model grid has half the temporal resolution of the atmosphere model, with a timestep of one hour, but is more finely resolved. It has a horizontal resolution of 1.25° x 1.25° and 20 vertical levels that are distributed on a depth based (z) coordinate system, as given by Table 2 in Johns et al. (1997), to give maximum resolution towards the ocean surface. Level spacing is small (10 m) near the surface to resolve the mixed layer, but increases with depth, reaching 615 m for levels below 1193 m deep.

The ocean model's physical parameterisations include the eddy-mixing scheme of Visbeck et al. (1997), the isopycnal diffusion scheme of Gent and Mcwilliams (1990) and a simple thermodynamic sea ice scheme for ice concentration (Hibler 1979) and ice leads and drift (Cattle et al. 1995). Gordon et al. (2000) show that the model adequately reproduces modern sea surface temperatures without the need for unphysical 'flux adjustments' at the ocean-atmosphere interface. The ocean component of HadCM3 has a fixed lid; in other words, the volume of the ocean grid boxes (and hence sea level) cannot vary. Evaporation, precipitation and river runoff are therefore represented as a salt flux (Gordon et al. 2000).

Once per model-day, the atmosphere and ocean components pass across the fluxes accumulated over the previous 24 model-hours. To accommodate the different horizontal resolutions, the ocean grid is aligned with the atmosphere grid and the data is respectively averaged and interpolated during the coupling. River discharge is implicitly simulated using grid-defined river catchments and estuaries, which instantaneously deliver continental runoff to the coasts. For more details on the model and its constituent components, including improvements on earlier versions, see Gordon et al. (2000) and Pope et al. (2000).

2.2. Mediterranean-Atlantic exchange

Due to the restrictions of global GCM grid resolution, it is a challenge to represent shallow, narrow marine gateways such as the Gibraltar Straits, which is as shallow as 300 m and as narrow as 14 km (Candela 1991; Gómez 2003). Mediterranean-Atlantic exchange can be (and in many cases is) simply prescribed as a boundary condition to the North Atlantic ocean, but this allows for neither a full mechanistic investigation of the dynamical responses and feedbacks to changes in global or regional climates, nor a thorough examination of the response to forcings originating in the Mediterranean Sea or North Atlantic Ocean. To do this in global GCMs, there are two approaches for modelling the exchange. The first is to use an unrealistically wide and deep open seaway to connect the two basins. This is the approach adopted by Bigg and Wadley (2001) and Rogerson et al. (2010) to investigate the effect of changing North Atlantic circulation and water-column characteristics on the intensity of Mediterranean-Atlantic exchange, and the resulting feedback to North Atlantic salinity, temperature and circulation patterns. It is worth noting that in both of these cases, a realistic rate of exchange (1 Sv) is achieved, despite the oversized seaway.

The second approach is to have a land bridge connecting North Africa (Morocco) with Southern Spain in the model and use a physical parameterisation of the net flows, which we will refer to as a pipe. This parameterisation mixes thermal and haline properties across the Gibraltar Straits and in this way emulates the exchange that is achieved with open flow through the narrow, shallow Straits in reality. For example, in their studies of the impact of changing Mediterranean-Atlantic exchange properties on Atlantic circulation and global climate, Rahmstorf (1998) and Chan and Motoi (2003) replace water column temperature and salinity properties in the upper 600 m of the two grid boxes situated either side of their Morocco-Spain land bridge with the mean values for the pair at every time-step, resulting in complete mixing across the Straits.

HadCM3 employs a version of the latter of these two methods, using a pipe to exchange water across the land bridge connecting Morocco and Spain. The parameterisation, described in more detail below, achieves partial mixing across the Straits according to a coefficient of mixing (μ) as well as the comparative characteristics of the Alboran Sea (in the westernmost Mediterranean) and the Gulf of Cadiz (in the North Atlantic), which is more realistic than total mixing. This mixing takes place between the temperature and salinity tracer fields across two pairs of grid boxes situated immediately adjacent to the land bridge; one pair in the Alboran Sea and one pair in the Mediterranean. The depth of the pipe is constrained by the minimum bathymetry of any one of these boxes, which is ~ 1 km, 13 vertical levels. For every level and at every timestep, the mean of the four points is calculated for each tracer field ($\overline{T}$). Then, where (T_j) is the tracer for each of the four grid boxes, the difference between the old (previous timestep) and the new (current timestep) tracer is given as:

$$\left.\frac{\partial T_j}{\partial t}\right|_{pipe} = \mu(T_j - \overline{T})$$

(Gordon et al. 2000), where $\left.\frac{\partial T_j}{\partial t}\right|_{pipe}$ is the tracer tendency for the pipe parameterisation. Thus, the model exchanges temperature and salinity properties based on the temperature and salinity gradients existing at every ocean level and at every timestep of the model run. The constant μ (set at $9.6e^{-5}$) defines the coefficient of mixing; that is, the proportion of each grid-box that will be mixed, thus representing the control of Straits geometry on the exchange. There is no advection of water through the pipe. However, the parameterised exchange in temperature and salinity leads to flow in and out of boxes on either side of the pipe. The water and salinity fluxes in and out of the Mediterranean Sea presented in this study were calculated in the westernmost grid boxes of the Straits, immediately adjacent to the Morocco-Spain land bridge. The mixing constant μ has been set to achieve 1 Sv of transport between the Mediterranean and Atlantic in each direction, which is close to observational values (>0.74 ±0.05 Sv; García-Lafuente et al. 2011). Although the geometry of the pipe remains constrained by the ocean model's resolution, a more realistic rate of flow can be achieved than by having a similarly size-constrained open seaway (as in Ivanovic et al. 2013). Therefore, we chose to use this pipe set up in the experiments presented here.

Even though the modern Gibraltar Straits is only 300 m deep, MOW is observed to rapidly descend the continental slope and settle at around 1000 m by 8.75° W in the modern ocean, from where it spreads westwards into the North Atlantic (e.g. Baringer and Price 1999; Serra et al. 2005; Dietrich et al. 2008; Boyer et al. 2009). Because of the width of the Morocco-Spain land bridge in HadCM3, which is constrained by horizontal grid resolution, this makes 1000 m an appropriate depth for the pipe configuration, enabling the model to simulate the large scale features of MOW's flow path in the North Atlantic. The resulting eastward surface flow of North Atlantic Central Water into the Mediterranean and a deeper westward flow of Mediterranean Outflow Water (MOW) into the Atlantic matches the observed two-layer flow structure for flow through the Gibraltar Straits. The interface, defined as the 'depth of maximum vertical shear' (Tsimplis and Bryden 2000), lies roughly halfway between the surface and sill depth; at around 500 m in HadCM3, where the sill depth is 1000 m (this study), and around 150 m in observations for a sill depth of 300 m (Bethoux and Gentili 1999; Tsimplis and Bryden 2000; Gómez 2003; Boyer et al. 2009).

HadCM3 is not spatially fine-scaled enough to accurately resolve MOW eddies (meddies) or its two-core flow structure observed for the narrow eastern boundary current (e.g. Jungclaus and Mellor 2000; Johnson et al. 2002; Papadakis et al. 2003; Serra et al. 2005). The model therefore probably underestimates shallow-intermediate level mixing of MOW and Atlantic water in the eastern Atlantic. For example, results presented by Ivanovic et al. (2013) suggest that increased mixing of Mediterranean and Atlantic water in the Gibraltar Strait-Gulf of Cadiz region does slightly improve the HadCM3 representation of the MOW plume in the modern North Atlantic. Similarly, our study is limited by its use of a depth based vertical coordinate system (Johns et al. 1997), which incompletely resolves the dense overflow of MOW from the Gibraltar Straits sill. As a result, the model has a tendency to overestimate the mixing of MOW with surrounding water, simulating North Atlantic entrainment, as it descends the continental shelf. These two limitations may affect the model's sensitivity to changes in MOW density, perhaps under-projecting any buoyancy anomalies; a caveat that should be considered when interpreting the results of any such change. However, these effects also partly counteract each other. Thus overall, the model's ability to simulate the large-scale features of MOW in the North Atlantic (e.g. as seen in Boyer et al. 2009) makes the standard pipe set-up in HadCM3 an appropriate control configuration for this investigation.

Due to the Mediterranean's net annual evaporation from the basin (e.g. Bethoux and Gentili 1999; Mariotti et al. 2002), there is a net transport from the Atlantic to the

Mediterranean. To conserve salinity in a steady state Mediterranean there is no net export of salt through the Gibraltar Straits. In our fixed-lid model, however, net evaporation from the Mediterranean Sea is represented as a salt flux rather than a true freshwater flux; there is no net water flux across the Gibraltar Strait, but there is a net salt flux across it. This export of salt from the Mediterranean has the same significance as the outflow salinity transport defined by Bryden et al. (1994) as the outflow transport of MOW multiplied by the salinity excess carried by the MOW over the incoming Atlantic water. The pipe parameterisation in our model only allows a mean salt export of around 0.6 psu Sv, less than half of the outflow salinity transport of ~ 1.5 psu Sv (Bryden et al. 1994). Moreover, this salinity flux is controlled by the parameter μ and in the control simulation, the pipe does not conserve salt in the Mediterranean Sea, which has a drift of 9.0 x 10^{-4} psu yr^{-1}. This drift is equivalent to observations of increased Mediterranean salinity by 8.0-9.2 x 10^{-4} psu yr^{-1} over the past 50 years (e.g. Dietrich et al. 2008; Vargas-Yáñez et al. 2010); a rate that is projected to nearly triple by 2100 (Somot et al. 2006).

MOW's influence on the HadCM3 North Atlantic were identified by comparing our control climate to a simulation which has no Mediterranean-Atlantic exchange; see section 2.3, experiment (a). Consequently, we can see that in both the control simulation (Fig. 1) and modern observations (e.g. Boyer et al. 2009), MOW constitutes a distinct plume of relatively warm (up to +6 °C) and highly saline (up to +1.8 psu) water protruding into North Atlantic Central Waters, where it spreads predominantly westwards and mixes with surrounding intermediate waters. In the modern North Atlantic, the MOW plume is centred at 1000-1200 m below the ocean surface (Boyer et al. 2009), whereas in HadCM3, the plume lies a little deeper at 1200-1500 m (Fig. 1). Opening the Gibraltar Straits seaway in the model, rather than using a pipe, does cause a shoaling of the MOW plume by a few hundred meters with respect to the HadCM3 control. It also produces an increase in salt export that more closely matches (though overshoots) observational values. However, the resulting Mediterranean-Atlantic exchange is over three-times too strong compared to observational data, reorganizing shallow-intermediate water circulation in the North Atlantic with significant regional climate repercussions (up to +11 °C and – 7.5 °C) (Ivanovic et al. 2013). Thus, even though the MOW plume in the HadCM3 control is around 200 m too deep and 0.9 psu too fresh, this set up still reaches the best compromise to date for achieving a realistic strength of exchange. Ivanovic et al. (2013) discuss this in more detail and propose ways in which the disparity between MOW plume depth, flow strength and salt export could be overcome in the future.

2.3. Experiment design

To test the sensitivity of modern climate to the presence, strength and salinity of Mediterranean-Atlantic water exchange, we performed three HadCM3 experiments. The simulations represent extremes compared to modern Mediterranean conditions, but are well within the geologic constraints of events such as the Messinian Salinity Crisis (5.96-5.33 Ma), eustatic sea level changes and Mediterranean sapropel formation (e.g. Clauzon et al. 1996; Fleming et al. 1998; Béthoux and Pierre 1999; Milne et al. 2005; Flecker and Ellam 2006; Marino et al. 2007; Roveri et al. 2008; Rohling et al. 2008; Govers et al. 2009; de Lange and Krijgsman 2010; Osborne et al. 2010). These are idealised experiments, and although they are not set-up to be specifically realistic in a geologic, modern, or future context, they directly test extreme cases of Mediterranean variability. Thus, these experiments provide a robust platform from which to interpret further 'realistic' simulations of changes to Mediterranean-Atlantic exchange and the Mediterranean hydrological balance:

(a) The first experiment is a pair of simulations designed to examine how the presence of Mediterranean-Atlantic exchange affects North Atlantic circulation and climate in HadCM3. It consists of no-exch, in which we have turned off the HadCM3

Mediterranean-Atlantic pipe so that there is no exchange between the two basins, and the control.

(b) In the HadCM3 pipe parameterisation, μ is a measure of the restriction of flow between the Mediterranean and Atlantic. As such, it is a crude representation of (i) the geometry of the Gibraltar Straits and (ii) the complex flow of water over the Gibraltar Straits sills, including local meddies. Thus, the second experiment is designed to examine climate-ocean sensitivity to changes in the restriction and mixing of the exchange. In the 'real world', this could be brought about by rising/falling sea levels, tectonic adjustment of the Straits, or changes in Mediterranean circulation. The experiment comprises five simulations, each with different strengths (or restrictions) of exchange for a given temperature and salinity gradient across the Gibraltar Straits: quart-exch, half-exch, control, doub-exch and quad-exch, which have exchange coefficients that are equal to $0.25\mu_c$, $0.5\mu_c$, μ_c, $2\mu_c$ and $4\mu_c$, respectively, where μ_c is the standard (control) coefficient of mixing for the temperature/salinity gradient driven Mediterranean-Atlantic exchange.

(c) The experiment is made up of three simulations: fresh-Med, control and salt-Med. For fresh-Med and salt-Med, we have forced the entire Mediterranean basin, but nowhere else, to have a constant salinity of 19 psu (approximately half the control salinity) and 76 psu (approximately double the control salinity) respectively, at every timestep for the duration of the model run. This means that salt is not conserved and the simulations do not directly represent 'real world' scenarios. However, in this experiment, the volume integral for the global ocean changes by only ~ 0.1 psu over 500 years. Thus, the changes are small (0.2-0.4 %) and so do not present a problematic salt source/sink for understanding the physical mechanisms at work in these idealised simulations.

For all three experiments we have made no alterations to the control, using the standard HadCM3 modern set-up described above, and have also included three conservative ocean tracers with an arbitrary volume-density of 1 m^{-3} to the Mediterranean basin at the very start of the run:

(i) Tracer 1 is the shallow water tracer and was applied to the upper 150 m of the water column.
(ii) Tracer 2 is the intermediate water tracer and was applied to water between 150 m and 500 m deep.
(iii) Tracer 3 is the deep water tracer and was applied to depths below 500 m.

Thus, as the tracer-spiked water exchanges with the Atlantic, we are able to directly track the pathways of MOW and monitor the slow spread and mixing of Mediterranean water in the global ocean through time.

All simulations were run for 500 years using a pre-industrial climate and modern continental configuration. They were initialised using dump files from the HadCM3 public release spin-up simulation, published by Gordon et al. (2000). For all simulations, near steady state was reached within the first 400 years of model run, and the climate means shown in this study were calculated from the remaining 100 years.

3. Results and Discussion

In the following discussion, we present the results of our investigation into modern ocean and climate sensitivity to changes in the presence, strength and salinity of Mediterranean Atlantic exchange. The discourse is divided into three subsections, one for each experiment outlined in section 2.3., and in that order.

3.1. Presence of Mediterranean-Atlantic exchange

The climate differences discussed in this section are given as the climate anomalies achieved in control with respect to no-exch; that is, the effect of having Mediterranean-Atlantic exchange (and hence MOW) in HadCM3 versus no exchange (and no MOW).

In control, MOW spreads westward from the Gibraltar Straits (35° N), between depths of 600 m and 2500 m, reduces net southward flow of NADW at 800-2000 m depth and weakens the AMOC north of the Gibraltar Straits (35° N) by up to 2 Sv (11.1 %, Fig. 2a). Interestingly, this result counters speculation by Reid (1979), Bethoux et al. (1999) and Gómez (2003) that without MOW, modern NADW formation (and hence the AMOC) would be weaker. In fact, South of the Gibraltar Straits (35° N), the AMOC is strengthened by up to 1 Sv (16.7 %; Fig. 2a). This strengthening is in good agreement with the conclusions of Rahmstorf (1998) and Chan and Motoi (2003), whose GCM results suggest that the presence of MOW in the North Atlantic increases the deeper, southward-bound AMOC component by 1-2 Sv. In our model, this reorganisation of the AMOC is due to a change in the North-South density gradient along the Atlantic; the presence of MOW reduces the density gradient between the sites of NADW formation and the Gibraltar Straits and increases the density gradient between the Gibraltar Straits and the South Atlantic (Fig. 2b). This is why AMOC is reduced north of the Straits and increased south of it.

Furthermore, computing the AMOC-associated freshwater transport at the southern boundary of the Atlantic (F_{ov}), as done by Hawkins et al. (2011), we find that the increased export of relatively salty water from MOW to the Southern Ocean in control enhances net freshwater import by 0.03 Sv (an increase of 11 %), compared to no-exch. Others (incl. Rahmstorf 1996; Dijkstra 2007; Huisman et al. 2010; Hawkins et al. 2011) have suggested that such net freshwater import (positive F_{ov}) to the North Atlantic negatively feeds-back to AMOC strength and promotes a mono-stable AMOC regime, whereas net freshwater export to the Southern Ocean (negative F_{ov}) promotes bistability in the AMOC. For example, in the case of positive F_{ov}, a decrease in AMOC strength would result in the accumulation of salt in the North Atlantic, promoting deep mixing, increasing AMOC strength and thereby stabilising NADW formation. Conversely, for a system with negative F_{ov}, AMOC weakening leads to freshening in the North Atlantic, further reducing NADW formation and AMOC strength, pushing it towards a weak (or even an 'off') mode of circulation. Thus it follows that an increase in net freshwater import to the North Atlantic achieved with the presence of Mediterranean-Atlantic exchange in HadCM3 acts to further stabilise the current mode of AMOC, corroborating similar findings by Bigg and Wadley (2001) and Artale et al. (2002).

However, the respective 1-2 Sv weakening and strengthening of different AMOC components that is caused by the presence of MOW in the North Atlantic (Fig. 2), has negligible impact on HadCM3 climate. This is consistent with the findings of Rahmstorf (1998), but contrary to those of Chan and Motoi (2003), who propose that MOW strengthens Antarctic Bottom Water formation, warming Southern Ocean air temperatures by up to 6 °C. We suggest that our results are an inevitable product of the strong, stable, modern AMOC, which at 26.5° N has an overturning strength of 18 ±2 Sv in HadCM3 (this study) and 18.7 ±5.6 Sv in recent observations (Cunningham et al. 2007), but only 10 Sv (approx.) in the control simulation performed by Chan and Motoi (2003). Perhaps the role of Mediterranean-Atlantic exchange is more important for periods of weaker NADW formation or a bi-stable ocean (Broecker et al. 1990; Cacho et al. 2000; Bigg and Wadley 2001; Artale et al. 2002; Voelker et al. 2006; Rogerson et al. 2010; Penaud et al. 2011), but it is assuredly clear that modern HadCM3 climates are insensitive to the presence of Mediterranean-Atlantic exchange, despite the MOW-induced saline and thermal anomalies produced in intermediate to deep North Atlantic waters.

3.2. Strength of Mediterranean-Atlantic exchange

The exchange of water through the Gibraltar Straits governs the temperature and salinity characteristics of the Mediterranean Sea and, to a lesser extent, the North Atlantic Ocean. These in turn control the temperature/salinity gradients across the Straits, as well as affecting large-scale ocean circulation in the two basins, negatively feeding back to regulate the initial change (Candela 1991; Bethoux and Gentili 1999; Mariotti et al. 2002; Rogerson et al. 2011; Sanchez-Gomez et al. 2011). Thus, the effect of changing the coefficient of Mediterranean-Atlantic exchange in HadCM3 is not straightforward. As explained in section 2.2, although there is no advection of water through the pipe, the exchange in temperature and salinity induces flow in and out of boxes on either side of the pipe. What we call the flux of water transported through the Gibraltar Straits, is the flux calculated in the westernmost grid boxes of the Straits, immediately adjacent to the Morocco-Spain land bridge.

For the simulations with a lower coefficient of exchange than the control (quart-exch and half-exch), the flux of water transported through the Gibraltar Straits is initially 0.3 Sv, compared to 1 Sv in control. Because of this reduced mixing between the Mediterranean and the Atlantic, the Alboran Sea in the Mediterranean becomes accumulatively saltier (on average, +1.85 psu for half-exch and +2.40 psu for quart-exch by the end of the run), whilst the Gulf of Cadiz in the North Atlantic becomes accumulatively fresher (on average, -0.05 psu for half-exch and -0.15 psu for quart-exch by the end of the run). The resulting increase in temperature/salinity gradients across the Straits then counteracts the reduced mixing coefficients used in quart-exch and half-exch, elevating Mediterranean-Atlantic exchange from its initial state. Thus, as the temperature/salinity gradients become steeper through the run, so the flow across the Gibraltar Straits increases back towards the control state (Fig. 3) and vice versa for an increased coefficient of mixing. This ‘relaxation’ towards an equilibrium state is not unique to the four perturbed mixing coefficient simulations. We also observe similar behaviour in control (Fig. 3), reaching a steady state exchange of 1 Sv after approximately 100 years of model run. For quart-exch, the transport flux reaches a steady state centred at 0.65 Sv after around 200 years of model run (Fig. 3) and half-exch recovers to a steady state flow of around 0.8 Sv in about half that time (~ 100 years). This negative feedback to flow strength indicates the strong influence of temperature/salinity gradients across the Straits in governing the intensity of exchange and an element of insensitivity to changes in the coefficient of mixing. Notably, the 0.8 Sv of exchange reached by half-exch after 100 years of model run, more closely matches observations of ~ 0.74 ±0.05 Sv (e.g. García-Lafuente et al. 2011) than the 1 Sv achieved by control. However, the Mediterranean salt export of 0.15 psu Sv for half-exch is much less realistic than the already relatively low levels of export in control; 0.6 Sv compared to the observed ~ 1.5 psu Sv (Bryden et al. 1994).

The immediate effect of increasing the coefficients of exchange in doub-exch and quad-exch is to increase the transport of water across the Gibraltar Straits by up to around 0.6 Sv for doub-exch and by up to around 2.4 Sv for quad-exch, with respect to the 1 Sv of flow achieved with the control (Fig. 3). However, by the same feedback-processes that reduced the effect of decreasing the coefficient of exchange, this initial strengthening progressively reduces towards control values. Still, by the end of the run, the Alboran Sea in the Western Mediterranean has, on average, freshened by 2.30 psu in doub-exch and by 3.75 psu in quad-exch, due to enhanced Atlantic inflow. Similarly, due to greater Mediterranean outflow to the Atlantic, average ocean salinity in the vicinity of the Gulf of Cadiz has increased by 0.05 psu in both doub-exch and quad-exch. As a result, Mediterranean-Atlantic exchange in doub-exch and quad-exch generally remains higher than control (Fig. 3). After approximately 150 years of model run, doub-exch reaches a steady state Mediterranean-Atlantic exchange of 1.3 Sv. Quad-exch, on the other hand, takes almost 200 years to settle with a flow of 1.6 Sv across the Gibraltar Straits (Fig. 3). Although the enhanced Mediterranean-Atlantic exchanges in doub-exch and quad-exch settle at almost twice the strength of observed values, the levels of salt export to the Atlantic (0.9 psu Sv in doub-exh and 1.0 psu Sv in quad-exch) are more realistic than the 0.6 psu Sv exported in control.

However, in HadCM3, these changes in flow strength (and salt export) through the Gibraltar Straits have negligible impact on Atlantic circulation and climate, although they do influence AMOC stability. With respect to there being no exchange between the Mediterranean and the Atlantic (no-exch), F_{ov} increases by 0.01 Sv (3.7 %) for quart-exch, and thereafter increases in increments of 0.01 Sv per doubling of the mixing coefficient for half-exch, control, doub-exch and quad-exch. Thus, for quad-exch, F_{ov} is 0.05 Sv (18.5 %) greater than for no-exch, and 0.02 Sv (6.7 %) greater than for control. This further supports the proposition that Mediterranean-Atlantic exchange acts to stabilise the current AMOC mode, as discussed in section 3.1., and shows that even a quadrupling of the coefficient of exchange across the Gibraltar Straits is insufficient to increase the sensitivity of HadCM3 climate to the presence of Mediterranean-Atlantic exchange.

Nevertheless, we do not rule out the possibility that Atlantic circulation and climate are affected by changes in Mediterranean-Atlantic exchange intensity. Mediterranean Outflow Water (MOW) in quart-exch, half-exch, doub-exch and quad-exch remains centred at 1200-1500 m, as it does in the control, and so the perturbations in exchange intensity only impact deeper Atlantic waters. The very presence of MOW at these deeper layers is enough to enhance F_{ov}, suggesting that it acts to stabilise AMOC, but nothing more. We therefore consider that in HadCM3, MOW is some 200-300 m deeper than the plume observed in the modern Atlantic Ocean (Boyer et al. 2009). Should MOW shoal, then changes in Mediterranean-Atlantic flow strength would influence shallower, intermediate North Atlantic currents, and so could significantly impact high northern latitude climates (Ivanovic et al. 2013). As the salinity of the Mediterranean directly affects the salinity and hence buoyancy of MOW in the Atlantic (e.g. Bethoux and Gentili 1999; Mariotti et al. 2002; Rogerson et al. 2011), we propose that changing Mediterranean salinity has a greater impact on Atlantic circulation and global climate than changing the coefficient of mixing across the Gibraltar Straits. Thus, in the following section we investigate the impact of having a fresher (fresh-Med) and a saltier (salt-Med) Mediterranean in HadCM3.

3.3. Mediterranean salinity

As discussed in section 2.2., and as illustrated by the damped effect of changes in mixing coefficients across the Gibraltar Straits (section 3.2.), the temperature/salinity gradients between the westernmost Alboran Sea and easternmost Gulf of Cadiz is the primary control for Mediterranean-Atlantic flow-strength in HadCM3. Therefore, it is unsurprising that halving (fresh-Med) and doubling (salt-Med) Mediterranean salinity has a dramatic impact on flow strength across the Gibraltar Straits in the pipe. In fresh-Med, there is a three-fold strengthening of Mediterranean-Atlantic exchange, settling at 3.0 Sv of flow in both directions (almost double that achieved by quad-exch, Fig. 3), and the effect is even stronger in salt-Med, reaching 5.3 Sv at the end of the run (over three-times that achieved by quad-exch). Also, unlike quart-exch, half-exch, doub-exch and quad-exch, there is barely any relaxation of flow strength in the centuries following the initial ramp-up. This is because Mediterranean salinity is held constant throughout the simulations, which impedes the negative feedback of enhanced mixing reducing the temperature/salinity gradients.

The forced increase in Mediterranean salinity to 76 psu in salt-Med enforces the two-layer Mediterranean-Atlantic exchange structure already in place for control, whereas the freshening to 19 psu in fresh-Med completely reverses this structure. As a result salt-Med and fresh-Med have different effects on North Atlantic Ocean circulation, and so will be discussed separately below.

3.3.1. Halving Mediterranean salinity

In the modern ocean, western Mediterranean salinity is on average ~ 38 psu. For fresh-Med, we forced Mediterranean salinity to be a constant 19 psu across the whole basin and for the duration of the run. This reverses the salinity gradient across the Gibraltar Straits, causing net salt import to the Mediterranean of 3.35 psu Sv, compared to net export of 0.6 psu Sv in control. The Atlantic is now 17 psu saltier, and hence denser, than the Mediterranean, which forces Atlantic water to contribute the lower ~ 500 m of flow between the two basins, with MOW occupying the upper ~ 500 m of exchange in the pipe. This shoals the MOW plume to the surface of the North Atlantic and as a result, it takes a different flow-path (Fig. 4). Thus, in fresh-Med, MOW is routed northward along the Atlantic's eastern boundary, bypassing the North Atlantic subtropical gyre to contribute directly to the more northerly, adjacent subpolar gyre. Consequently, the subpolar gyre is both widened, stretching a further 10° S and 10° E in the North Atlantic, and strengthened, by up to 4 Sv, compared to control. This increases the provision of relatively warm, shallow, more southerly sourced waters to the Greenland-Iceland-Norwegian (GIN) Seas, raising their sea surface temperatures by up to 2.5 °C (Fig. 5). Combined with subsequent heat-release to the overlying atmosphere, this causes a decline in sea ice coverage over these high latitude sites, reducing surface albedo and amplifying the initial warming. With respect to control, annual mean surface air temperatures over the GIN Seas have warmed by up to 1.8 °C (Fig. 6b) by the end of the run and annual mean sea ice concentration has declined by up to 10 %; up to +2.7 °C and -15 % respectively in the boreal winter and spring, when the GIN Seas temperature and sea ice anomalies are greatest.

Moving focus to the North West Atlantic, the stronger, wider subpolar gyre promotes exchange between the Labrador Sea and the North Atlantic. As a result, there is an increase in flow of relatively cool and fresh, high latitude waters (from the GIN Seas) counter-clockwise into the Labrador Sea, cooling the overlying atmosphere of the southern Labrador Sea (Fig. 6b). This enhanced circulation also boosts the south-easterly expulsion of relatively cool and fresh water from the Labrador Sea, exaggerating the cold-tongue protruding into the North Atlantic, centred at a depth of 40-100 m (Fig. 7). Furthermore, the south-eastward extension of the subpolar gyre limits northward flow of relatively warm, low latitude water to the central North Atlantic 40-50° N. This, combined with the increased injection of colder water from high latitudes (the GIN and Labrador Seas) cools the North Atlantic water column by up to 4.5 °C (Fig. 7), with respect to control. The effect is greatest where these processes coalesce, centred around 45° N and 39° W (Figures 6b and 7), and as a result, the overlying atmosphere cools by up to 2.5 °C (annual mean), leading to a very localised increase in annual mean precipitation-evaporation of up to 76 %, which is equivalent to wetting of up to 1 mm day^{-1}.

In the northernmost North Atlantic, vertical density-stratification in the upper 650 m is increased by the compounded effect of mixing with the relatively fresh MOW and the boosted southward contribution of relatively fresh, shallow to intermediate water from the GIN Seas. In both fresh-Med and control, the core of the AMOC lies at around 800 m depth. Consequently, diminished vertical mixing in the upper 650 m of the water column reduces NADW formation and weakens the AMOC by up to 4 Sv (Fig. 8b). As a result, F_{ov} in fresh-Med decreases by 0.1 Sv (33 %), compared to control. Now at only 0.2 Sv, this reduced freshwater import suggests that not only is AMOC weaker, but that it is closer to reaching a point of bistability (e.g. Hawkins et al. 2011). Also, compared to control, the reduced fresh-Med AMOC transfers less heat polewards from the equator, which further enhances cooling in the shallow to intermediate North Atlantic water column.

Returning to the immediate vicinity of the Gibraltar Straits, freshening the Mediterranean to 19 psu reduces North Atlantic salinity throughout the water column (Fig. 9a). This is by the dual effect of (i) having a fresher, shallow MOW that decreases salinity in the upper North Atlantic and (ii) removing the deeper, relatively saline MOW plume that is present in control. As a result, vertical stratification increases (Fig. 9c) and vertical mixing is reduced. Furthermore, because Mediterranean salinity is held constant (at 19 psu), the Mediterranean basin also becomes highly stratified, with almost no vertical mixing taking place. This restricts warmer surface waters to the upper 15 m, causing the deeper Mediterranean to cool by up to 6 °C. As MOW is predominantly formed from deep Mediterranean water both in the model (Fig. 4) and the modern ocean (Cacho et al. 2000; Gómez 2003; Voelker et al. 2006), this cools the MOW plume in fresh-Med with respect to control. Thus, in the easternmost North Atlantic, a shoaled, relatively cold MOW plume replaces the deeper, relatively warm plume in control, cooling and freshening the entire water column, and increasing vertical stratification (Fig. 9). This cools the overlying atmosphere by up to 4 °C (Fig. 6b), following the south-westerly track of the shallow, relatively fresh, cool MOW plume (Fig. 4b).

3.3.2. Doubling Mediterranean salinity

For salt-Med, we forced Mediterranean salinity to be approximately double (76 psu) the modern average, for the duration of the run. This substantially increased the temperature/salinity gradients across the Gibraltar Straits (similar to fresh-Med, but opposite in direction) and so enhanced Mediterranean-Atlantic exchange by 4.3 Sv at the end of the run, with respect to control. As a result, there is a fifteen-fold increase in Mediterranean salt export to the North Atlantic in salt-Med compared to control, settling around 8.9 psu Sv by the end of the run and raising Atlantic salinity in the intermediate and deep layers. Despite the opposite change from control in salt-Med compared to fresh-Med, and their opposite effects on flow through the Gibraltar Straits and North Atlantic salinity, the trends in climate anomalies achieved in both simulations are notably similar, especially in the GIN Seas region

(Fig. 6). However, any similarities in climate signal are misleading; the salt-Med climate anomalies are brought about through different mechanisms than the fresh-Med anomalies are.

Due to enhanced Mediterranean-Atlantic exchange, the stronger, denser, MOW plume spreads further and deeper in the North Atlantic and there is a stronger, shallow flow of water across the Atlantic into the Mediterranean. This boosts both the southward spread of deep, saline waters in the North Atlantic (Fig. 10) and the northward draw of shallow, tropical waters. As a result, the AMOC is strengthened by up to 3 Sv south of the MOW injection (35° N, Fig. 8c) and F_{ov} by 0.2 Sv (66 %), suggesting an enhancement of the stability of the AMOC. The increased export of relatively dense water to the Southern Ocean also promotes Antarctic Bottom Water Formation, which strengthens by up to 5 Sv (Fig. 8c). Unlike for Chan and Motoi (2003), this does not impact Southern Hemisphere sea surface temperatures or atmospheric climate, but given the depth at which Antarctic Bottom Water formation occurs (below 2500 m in HadCM3), this is not surprising.

Another effect of the increased exchange is to enhance the northward spread of relatively warm, saline Mediterranean-origin waters, from around 1000 m deep and below (Fig. 11). This reduces AMOC circulation North of the Gibraltar Straits (35° N) and, combined with the decrease in poleward transport of shallow waters (induced by the strong eastward draw of water into the Mediterranean, which weakens the Gulf Stream and North Atlantic Drift), reduces NADW formation by up to 4 Sv (Fig. 8c). Furthermore, where this relatively warm, salty water upwells at around 47° N 46° W, annual mean ocean temperatures and salinities increase throughout the overlying water column (Fig. 12a), reducing density gradients in the intermediate layers and warming the upper 200 m by up to 5.3 °C. The subsequent increase in heat release warms the overlying atmosphere by up to 1.1 °C (annual mean, Fig. 6c), with respect to control. In the north-easternmost North Atlantic, intermediate to deep MOW-warmed waters rise over the Greenland-Scotland Ridge to flow into the interior of the GIN

Seas, increasing shallow water salinity and vertical mixing, and warming the upper 400 m by up to 2.5 °C by the end of the run (Fig. 12b). This reduces sea ice formation in the region, and increases heat exchange with the atmosphere. Additionally, the local decrease in sea ice cover reduces surface albedo, positively feeding back to the initial surface warming to amplify the effect. As a result, salt-Med reaches a steady state annual mean loss in sea cover of up to 30 % and surface air warming of up to 5.1 °C (Fig. 6c). These anomalies are heightened in the boreal winter and spring, reaching up to -50 % and +10.0 °C respectively.

It is important to point out that particularly in salt-Med, the depth coordinate scheme employed by HadCM3 could result in a rather more diffuse spread of the highly saline MOW than is physically realistic (e.g. Griffies et al. 2000). In this case, the modelled MOW plume may be more interactive with Atlantic Intermediate and Deep Water than it should be, and the effect on AMOC and Antarctic Bottom Waters could be biased in this respect. However, it is difficult to be sure of this, given also the reduced turbulent mixing that occurs in the modelled vicinity of the Gulf of Cadiz compared to the real ocean.

Returning our focus to the shallow ocean, the strong draw of North Atlantic water through the upper 500 m of the Gibraltar Straits pipe increases eastward- and, to a lesser extent, southward-flow across the Atlantic into the Mediterranean, as discussed above. This constricts the subpolar gyre northwards and westwards, reducing the circulation of relatively warm, subtropical water to higher latitudes. Consequently, the northernmost North Atlantic and the Labrador Sea freshen by up to 1 psu and cool by up to 3.5 °C (both annual means), also becoming more stratified in the upper 200 m (Figures 12c and 12d) with reduced vertical mixing. Similar to the GIN Seas, but opposite in direction, the resulting increase in sea ice cover and associated albedo feedback enhances this effect, which is greatest in the Labrador Sea (Fig. 6c). By the end of the run, Labrador Sea annual mean sea ice cover has increased by up to 25 % and annual mean surface air temperatures have cooled by up to 3.5 °C (Fig. 6c).

Again, these climate anomalies are greatest in the boreal winter and spring, when sea ice cover reaches +60 % and surface air temperatures plummet by up to 8.25 °C in the Labrador Sea, with respect to control.

4. Summary and conclusions

In HadCM3, the presence of Mediterranean-Atlantic exchange acts to strengthen the export of North Atlantic waters at a depth of 1500 m to 2500 m to the Southern Ocean by up to 1 Sv, in good agreement with Rahmstorf (1998) and Chan and Motoi (2003). At the same time, NADW formation between 35° N and 58° N is weakened by 1-2 Sv, contrary to the suggestions of Reid (1979), Rahmstorf (1998), Bethoux et al. (1999) and Gómez (2003). Respectively, these changes are caused by the south- and north-westward spread of Mediterranean Outflow Water (MOW) 1200-1500 m deep from 35° N. The net effect is an 11 % increase in F_{ov}, which may have a stabilising effect on the current AMOC regime (Rahmstorf 1998; Dijkstra 2007; Huisman et al. 2010; Hawkins et al. 2011). However, these small perturbations in AMOC strength are insufficient to affect the HadCM3 present-day surface climate. Similarly, neither a quadrupling nor a quartering of the Mediterranean-Atlantic mixing coefficient impacts North Atlantic circulation enough to induce a climate signal in the surface ocean or atmosphere, although they do increase and reduce $F_{ov,}$ and hence possibly AMOC stability, respectively.

The only statistically significant surface climate signals arise from a change in Mediterranean salinity (>95 % confidence, using student t-test). Raising (lowering) Mediterranean salinity by an approximate factor of two increases (decreases) MOW salinity, and hence affects the buoyancy of MOW in the North Atlantic. It also enhances (reverses) the two-way Mediterranean-Atlantic flow structure by amplifying (and reversing) the salinity gradient across the Gibraltar Straits. This not only affects AMOC strength and the pattern of

NADW formation, but is even more influential on shallow- and intermediate-water circulation in the North Atlantic, impacting the subtropical and subpolar gyres, as well as upwelling in the central North Atlantic and over the Greenland-Scotland Ridge. The GIN Seas, the Labrador Sea, the north and central North Atlantic, and the region immediately south-west of the Gibraltar Straits are most sensitive to the resulting changes in North Atlantic salinity and circulation, achieving regional climate anomalies in annual mean surface air temperature of ±4 °C or more.

Although the changes in Mediterranean salinity in fresh-Med and salt-Med may be considered extreme in a modern context (Rahmstorf 1998; Somot et al. 2006; García-Ruiz et al. 2011), they are well within proxy-reconstructed fluctuations for the Mediterranean in the geological past (e.g. Clauzon et al. 1996; Krijgsman et al. 1999; Flecker and Ellam 2006; Rohling et al. 2008; Roveri et al. 2008; Govers et al. 2009; de Lange and Krijgsman 2010). Furthermore, recent modelling work identifies the Mediterranean as being particularly vulnerable to future climate trends (Mariotti et al. 2002; Gao and Giorgi 2008; Dubois et al. 2011; Sanchez-Gomez et al. 2011), suggesting that current climate models under-project 21st Century changes in Mediterranean salinity and temperature. Also, on a much longer timescale, ongoing tectonic restriction of the Gibraltar Straits may eventually culminate in a salinity crisis akin to that of the Late Miocene, 5.96-5.33 Mya (Krijgsman et al. 1999).

However, these suppositions aside, this work shows that the presence of MOW in the North Atlantic acts to enhance F_{ov}, suggesting that it also stabilises AMOC (Rahmstorf 1996; Dijkstra 2007; Huisman et al. 2010; Hawkins et al. 2011); an effect that is amplified with increasing Mediterranean-Atlantic flow strength. However, in the model, the correlation between F_{ov} and MOW strength under present day conditions is weak. In support of Bigg and Wadley (2001), Rogerson et al. (2006, 2010), Voelker et al. (2006), Penaud et al. (2011) and others, we propose that Mediterranean-Atlantic exchange has the propensity to play a much more important role in maintaining AMOC during periods of weaker NADW formation, whether in the geologic past, or anthropogenic future. Furthermore, our results provide strong evidence that fluctuations in Mediterranean-Atlantic water exchange have the greatest affect on North Atlantic circulation and global climate when they instigate a combined change in MOW buoyancy and flow strength.

Our findings may be influenced by the simulation of an overly diffuse MOW core in the ocean interior. For example, a less diluted plume descending the continental shelf would probably not interact with intermediate and deep Atlantic Ocean circulation as significantly as in these simulations. On the other hand, the results may also be affected by the relatively deep injection of MOW to the North Atlantic (1200-1500 m) in our control. A shallower MOW plume that more closely matches observations could alter North Atlantic circulation more significantly and thus have a greater climatic impact than our control does compared to no-exch, as suggested by Ivanovic et al. (2013). Similarly, with a shoaled MOW plume in the North Atlantic, it is likely that increased interaction of relatively warm, saline, Mediterranean-sourced water with the northward flowing components of the AMOC and the shallower North Atlantic gyres would enhance the effect of fluctuations in Mediterranean-Atlantic flow strength shown here. Future work will examine this more closely.

In short, our atmosphere-ocean GCM results suggest that deeper ocean circulation (including the AMOC and Antarctic Bottom Water formation) is sensitive to changes in Mediterranean-Atlantic exchange, mainly through the provision of relatively saline water to the deeper, exporting branches of the AMOC south of the Gibraltar Straits, as per Rahmstorf (1998), Chan and Motoi (2003) and Kahana (2005). However, in the current regime of relatively strong AMOC, it is the influence of Mediterranean-Atlantic exchange on MOW buoyancy and strength that has the greatest effect on climate. By controlling the way MOW interacts with shallow and intermediate circulation currents; including the North Atlantic

Drift, subtropical gyre and subpolar gyre; a large (factor of two) change in Mediterranean salinity exerts a climate control of several degrees on shallow water and atmospheric temperatures in the North Atlantic, GIN and Labrador Seas. Therefore, for projections of future climate change, it will be important to consider the effect of regional climate trends over the Mediterranean, as well as human-controlled changes in river flow into the basin, within the global (or at least North Atlantic) context, especially if there are concurrent changes in AMOC strength and stability. However, over the course of the next century, we would not expect to see such large changes in Mediterranean or Gibraltar Straits conditions as have been modelled here without direct, catastrophic, human interference with the Mediterranean's hydrological budget; such as damming the Gibraltar Straits or enhanced Mediterranean freshwater consumption. Small changes in Mediterranean salinity conditions alone are unlikely to noticeably impact North Atlantic circulation or climate.

Moreover, we have shown that in such GCM research, careful consideration must be given to the model-specific representation of Mediterranean Outflow in the North Atlantic. This is because the buoyancy of MOW affects the extent to which it interacts with and contributes towards shallow and intermediate North Atlantic currents, which in turn govern regional surface climate.

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Figures

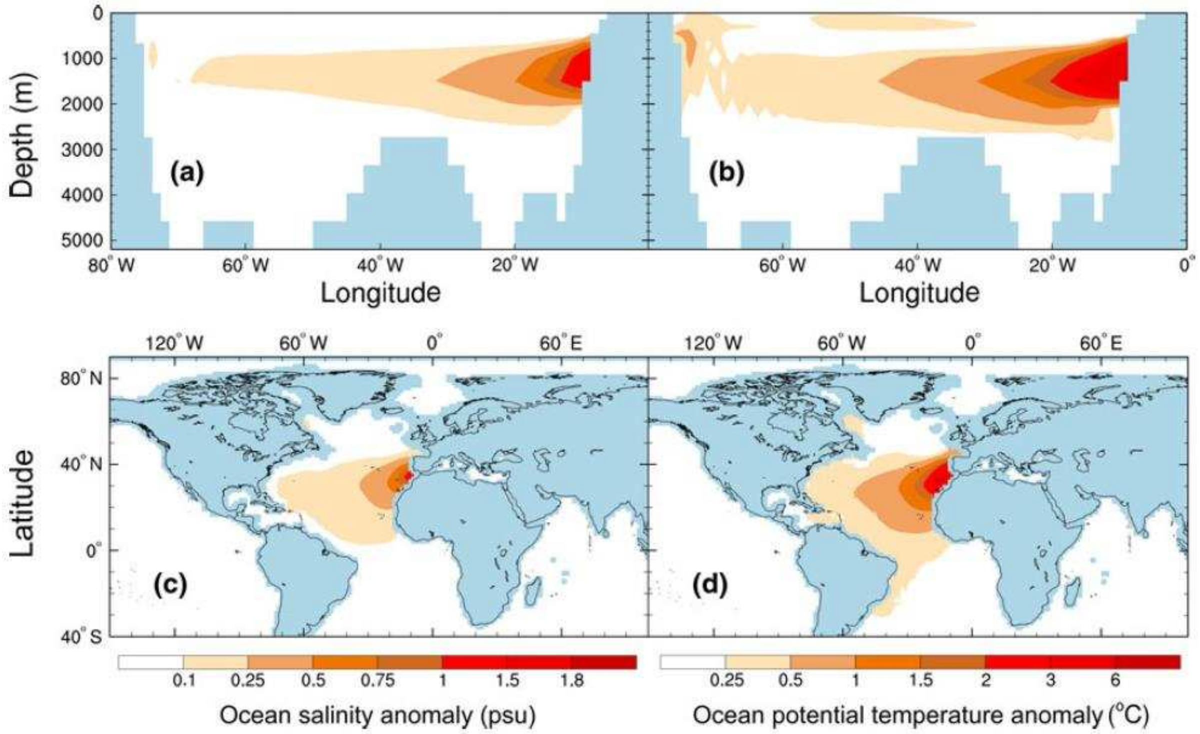

Figure 1. The direct effect of the presence of Mediterranean-Atlantic exchange on HadCM3 Atlantic annual mean salinity and potential temperatures characteristics in HadCM3. The anomalies shown are for (a) and (c) salinity, and (b) and (d) potential temperature, as produced in control with respect to there being no Mediterranean Outflow Water in no-exch. Projections (a) and (b) are cross-sections across the North Atlantic at 35° N, which is the latitude of the Gibraltar Straits. Projections (c) and (d) are taken at an ocean depth of 1501 m.

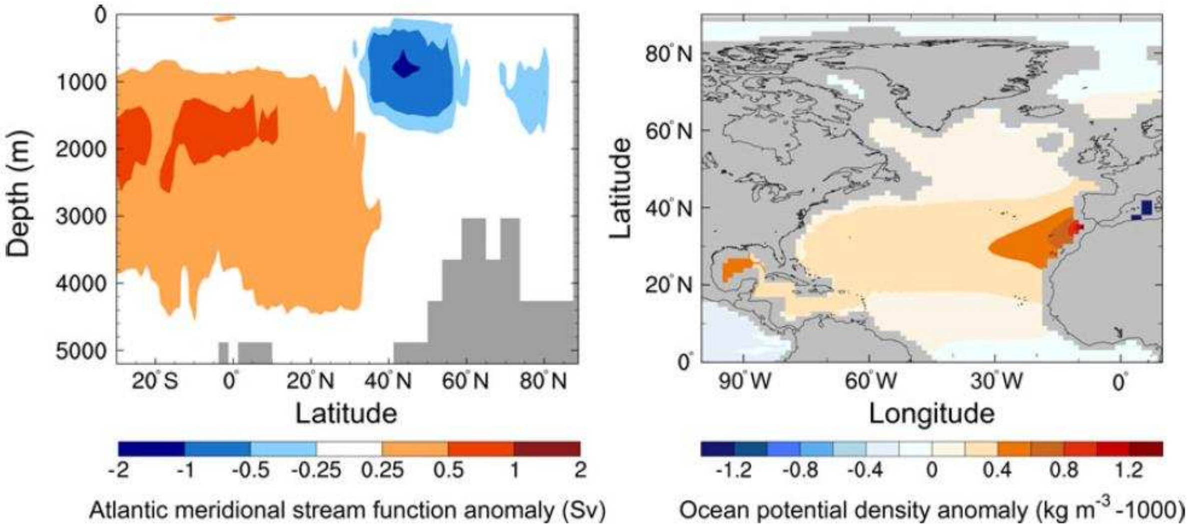

Figure 2. Annual mean anomalies for control with respect to no-exch in (a) Atlantic Meridional Overturning Circulation (as given by the stream function) and (b) North Atlantic Ocean potential density at 1501 m deep (given as the difference from 1000 kg m^{-3}). In control Mediterranean-Atlantic exchange occurs at 35° N. Using a student t-test, there is >95 % confidence in the significance of the anomalies shown.

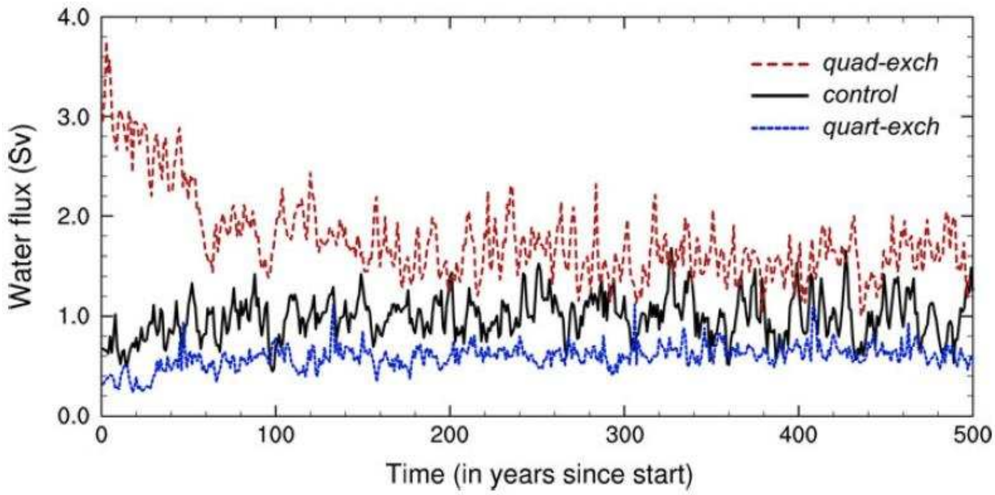

Figure 3. Mediterranean-Atlantic easterly and westerly water transport fluxes from the start of each run through time for quad-exch, control and quart-exch. doub-exch and half-exch show the same behaviour as quad-exch and quart-exch (respectively), but with more muted differences from the control, as described in section 3.2.

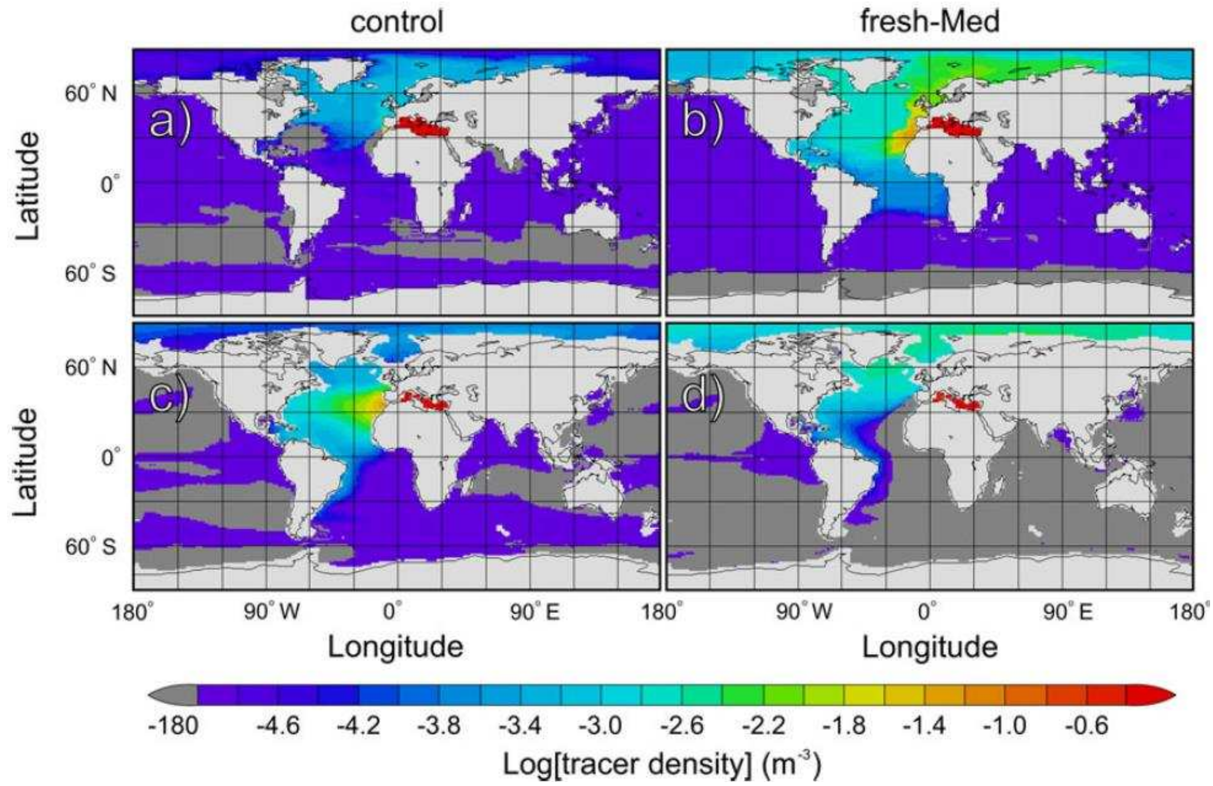

Figure 4. Log scale of the Mediterranean deep level tracer after 50 years of model run showing the difference between (a) and (c) control, and (b) and (d) fresh-Med. Projections (a) and (b) show the tracer at a water depth of 5 m, whereas (c) and (d) are at a depth of 1501 m. The tracer was input to the Mediterranean below 500 m deep and has an arbitrary volume density of 1 m^{-3} at the start of each simulation. After 50 years have elapsed, enough Mediterranean water has exchanged with the North Atlantic to track the flow path of Mediterranean Outflow Water (MOW) in the global ocean. An example of the deep level tracer (>500 m) is illustrated because it indicates MOW's flowpath in the Atlantic. Only a very small amount of the shallow and intermediate level tracers (<150 m and 150-500 m, respectively; neither shown) leaves the Mediterranean; thus little information about MOW's flowpath in the Atlantic is provided by these other tracers. However, from the combined tracer results (not all shown), it is clear that most MOW consists of water originating below 500 m, even in fresh-Med, when MOW is a shallow current.

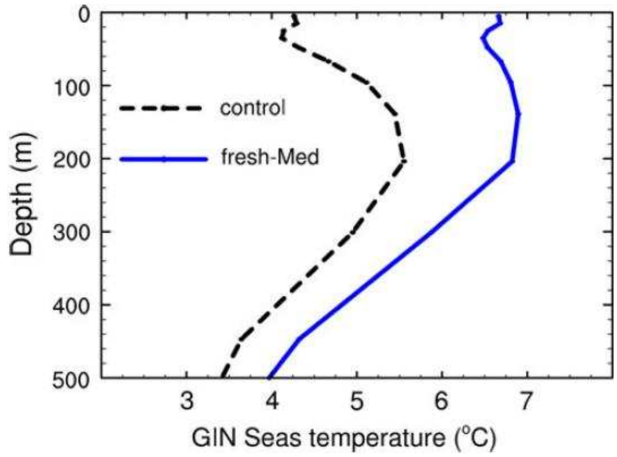

Figure 5. Greenland-Iceland-Norwegian (GIN) Seas annual mean potential temperature through the upper 500 m of the water column at 69° N 14° W for normal control and fresh-Med.

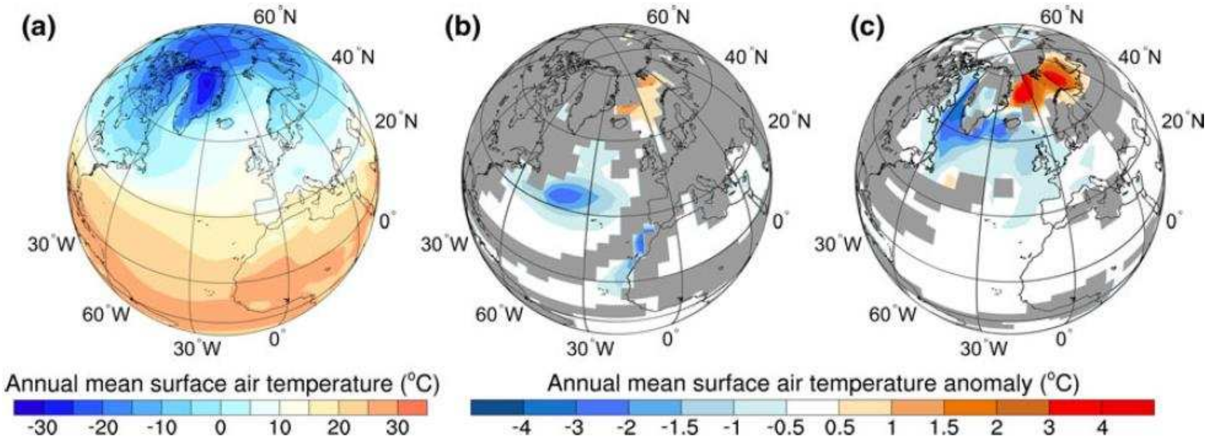

Figure 6. (a) Annual mean surface air temperatures produced in control. Annual mean surface air temperature anomalies produced in (b) fresh-Med and (c) salt-Med with respect to control. Areas with <95 % confidence in significance (using student-test) are shaded dark gray.

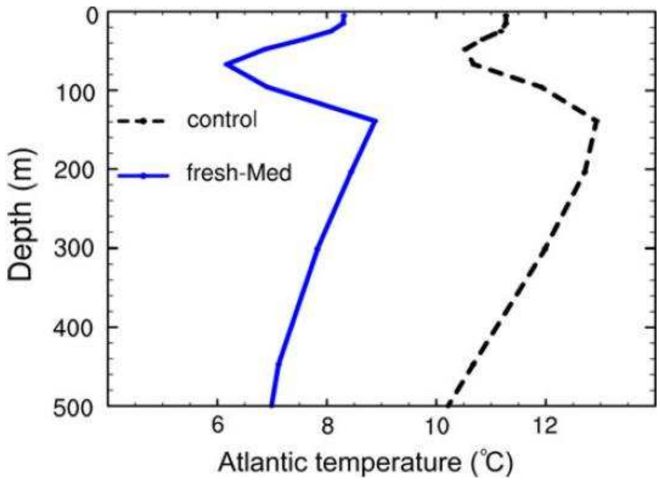

Figure 7. North Atlantic annual mean potential temperature through the upper 500 m of the water column at 45° N 39° W in control and fresh-Med.

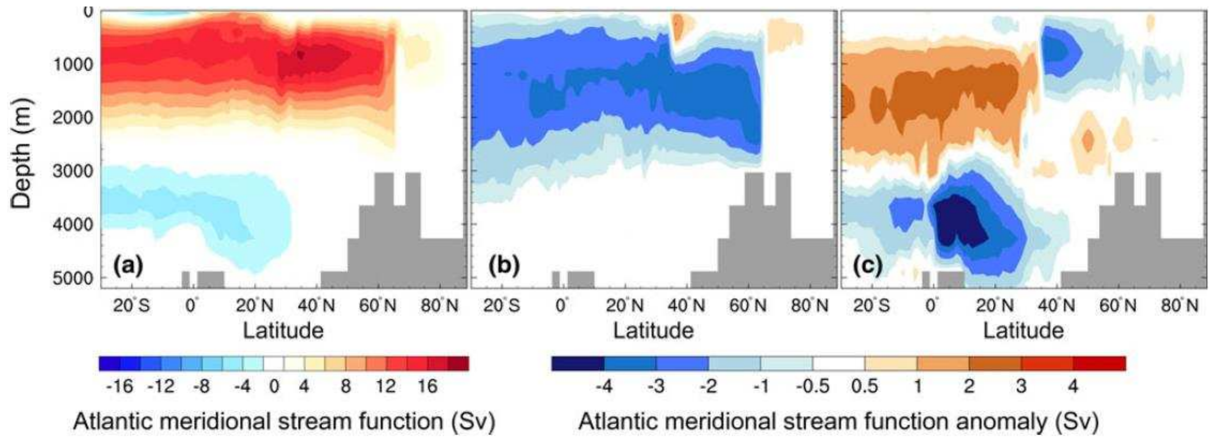

Figure 8. (a) Annual mean Atlantic Meridional Overturning Circulation (AMOC) strength in control. Annual mean AMOC strength anomalies for (b) fresh-Med and (c) salt-Med, with respect to control. Mediterranean-Atlantic exchange occurs at 35° N. The anomalies are given with >95 % confidence in their significance using a student t-test.

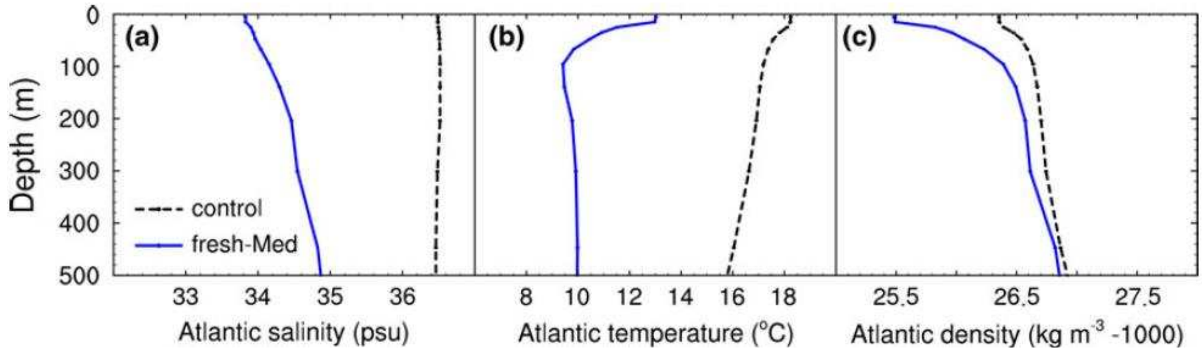

Figure 9. Eastern North Atlantic annual mean (a) salinity, (b) potential temperature and (c) potential density (given as anomalies from 1000 kg m^{-3}) in the upper 500 m of the water column in control and fresh-Med at 33° N 10° W; the vicinity of the Gulf of Cadiz.

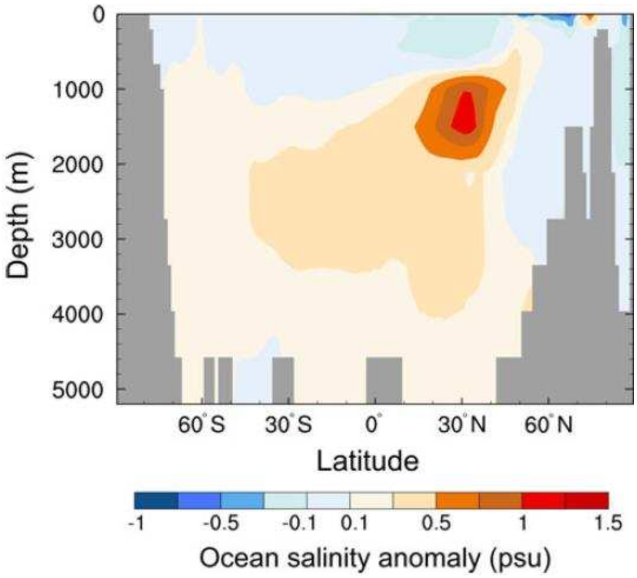

Figure 10. Annual mean salinity anomalies from North Pole to South Pole, averaged over 60° W to 10° W to capture the Atlantic and adjoining Southern Ocean, achieved for salt-Med with respect to control.

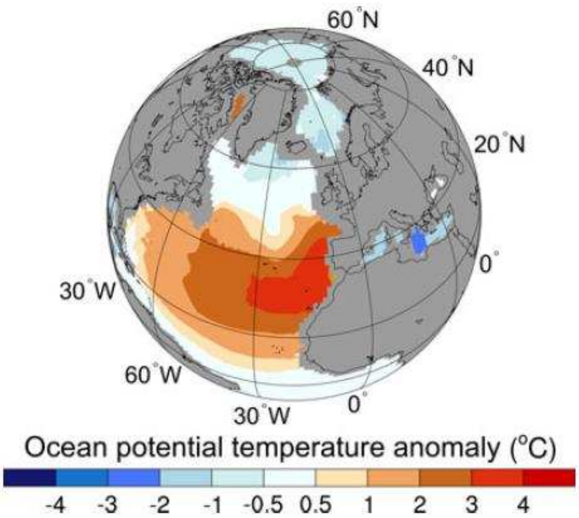

Figure 11. Annual mean ocean potential temperature anomalies at a depth of 1501 m for salt-Med, with respect to control.

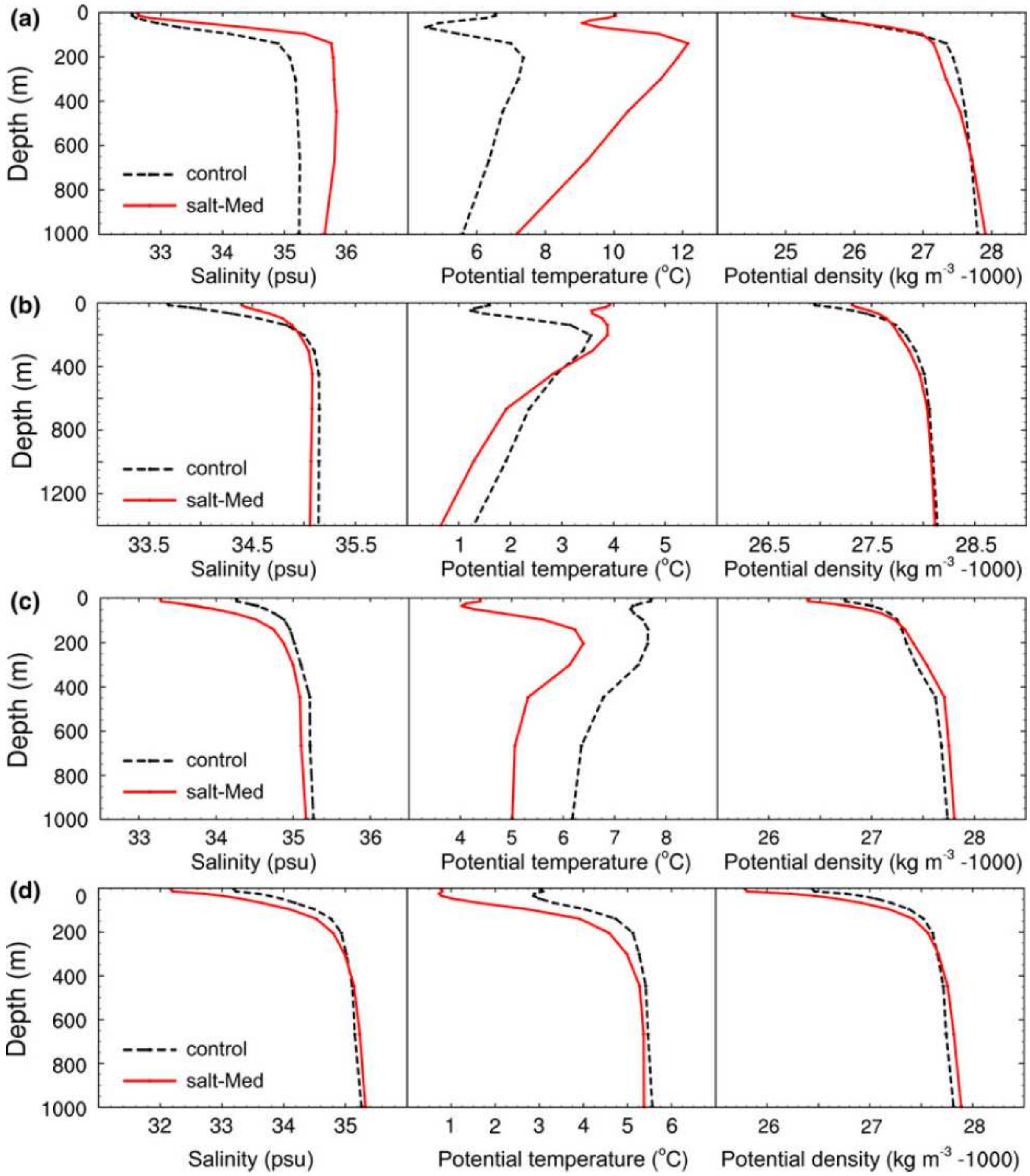

Figure 12. Ocean annual mean salinity, potential temperature and potential density (given as anomalies from 1000 kg m^{-3}) in control and salt-Med at (a) 47° N, 46° W in the North Atlantic; (b) 72° N, 5° W Greenland-Iceland-Norwegian (GIN) Seas; (c) 63° N, 28° W in the northernmost North Atlantic and (d) 60° N, 60° W in the Labrador Sea.