Stochastic Models of the Meridional Overturning Circulation: Timescales and Patterns of Variability

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The global meridional overturning circulation (MOC) varies over a wide range of space and time scales in response to fluctuating "weather" perturbations which may be modelled as stochastic forcing. This study reviews the effects of climate noise on decadal to centennial MOC variability, on transitions between MOC regimes, and on the dynamics of Dansgaard-Oeschger events characteristic of glacial periods.

Keywords: Meridional overturning circulation, stochastic climate models, decadal- to centennial-scale variability, abrupt climate change, Dansgaard-Oeschger events

1. Introduction

The meridional overturning circulation (MOC), which represents oceanic mass transport in the 2D meridional/vertical plane, is a fundamental diagnostic for understanding the role of the ocean in past, present and future climate change. In particular, sudden changes in the Atlantic meridional overturning circulation (AMOC) are associated with abrupt climate changes prevalent in high resolution proxy records over the last glacial cycle (e.g. Clark et al. 2002, Rahmstorf 2002). The importance of the AMOC to climate lies in its association with much of the total oceanic poleward heat transport in the present-day Atlantic, peaking at about 1.2 +/- 0.3 PW at 24° N (e.g. Ganachaud & Wunsch 2004). Because of the importance of the AMOC in the transport of heat to the North Atlantic, the variability and stability of the AMOC (particularly in response to anthropogenic forcing) is a subject of considerable scientific interest (e.g. Wood et al. 2003; Meehl et al. 2007).

This collection of currents has traditionally been referred to as the thermohaline circulation, a term which is problematic because it emphasises density gradients while implicitly downplaying the importance of mechanical forcing by the winds and interior turbulent mixing. In fact, ocean density gradients are largely set up by the winds via Ekman currents and surface buoyancy fluxes. Furthermore, mechanical forcing provides the energy source driving these circulations (see Kuhlbrodt et al. 2007 for a review), although there is no simple or necessary link between mechanical energy supply and oceanic heat transport (e.g. Saenko and Weaver 2004). Ocean stratification determines the structure and strength of the response to this mechanical driving. Inherent in the use of the terms MOC and AMOC is the fact that wind and buoyancy forcing are inseparable, and that wind and tidal forcing play a fundamental role in providing the energy required for turbulent mixing within the ocean.

The existence and structure of the MOC is fundamentally connected with the locations of deep water formation in the ocean. The two main constituent water masses of the deep North Atlantic Ocean – North Atlantic Deep Water (NADW) at the bottom and Labrador Sea Water (LSW) at an intermediate level – are currently formed in the Greenland-Iceland-Norwegian Seas and the Labrador Sea, respectively. Deep convection also occurs at a number of locations around Antarctica, but the dense bottom water is susceptible to being trapped by topographic sills (as in the Bransfield Strait), or by local circulation patterns (not excluding the Antarctic Circumpolar Current - ACC). In the Southern Ocean, around the southern tip of South America, an enhanced formation of low salinity Antarctic Intermediate water (AAIW) also occurs. AAIW plays a critical role in linking the Pacific and Atlantic Oceans and in particular in determining the stability of the AMOC (Saenko et al., 2003; Weaver et al., 2003). Today's climate has no sources of deep water in the North Pacific.

The various forcings which drive the MOC (e.g. evaporative and heat fluxes, precipitation, sea ice advection and melting, wind stresses, ocean eddy activity) display variability over a broad range of space and time scales, with a substantial concentration on timescales of atmospheric variability (subannual to interannual). When studying the dynamics of the MOC on timescales from decades to millennia, it is often convenient to represent these fluctuations as rapidly decorrelating stochastic processes (e.g. Hasselmann 1976). In this review, the influence of stochastic forcing on the dynamics of the MOC is considered. Section 2 examines the role of atmospheric fluctuations in driving MOC variability on decadal to centennial timescales. The influence of fluctuating forcing on transitions between MOC regimes is discussed in Section 3; Section 4 addresses the importance of these fluctuations in driving millennial-scale variability during glacial periods. Finally, a conclusion is presented in Section 5.

2. Decadal to Centennial Scale Variability

Climate variability on timescales ranging from decades to centuries has been identified in the instrumental and observational record of the North Atlantic, associated with a wide range of atmospheric, oceanic, and biological processes (e.g. Mann & Park 1996; O'Sullivan et al. 2002). A major difficulty in studying decadal to centennial timescales of climate variability is the lack of sufficient data: paleoclimate evidence suggests that the AMOC has been comparatively stable over the past 10,000 years (McManus et al. 2004) with present day variability estimated to be between 5% and 10% of the mean transport (Pasquero & Tziperman 2004), yet instrumental observational data sets are generally shorter than 100 years. While the mechanisms responsible for the low frequency variability in the climate system are not fully understood, the ocean circulation is believed to play a key role because of its large thermal inertia. In particular, the variability of the AMOC has received a great deal of theoretical and modelling attention because of its role as a major transporter of heat from low latitudes to high latitudes. In both simple and complex oceanic models the MOC exhibits variability on decadal and century time scales (e.g. Weaver 1995).

Three primary classes of mechanisms have been proposed to explain natural decadal to centennial scale variability of the MOC: (a) damped, uncoupled ocean modes excited by atmospheric variability; (b) unstable, uncoupled ocean modes that express themselves spontaneously; and (c) unstable or weakly damped coupled modes of the ocean-atmosphere system. Numerous studies have found self-sustained MOC oscillations associated with diffusive, advective, and convective processes (e.g. Weaver & Sarachik 1991; Weaver et al. 1993; Delworth et al. 1993; Greatbatch & Zhang 1995; Chen & Ghil 1995; Rivin & Tziperman 1997; Fanning & Weaver 1998; Huck et al. 1999; Arzel et al. 2006). This section considers the major results of those studies which have found that the maintenance of oscillatory variability of the MOC requires the input of energy from fluctuating atmospheric forcing.

Stochastic forcing has traditionally been used to represent high-frequency variability in surface fluxes. If the climate system has no internal mode of variability then the ocean integrates short-term atmospheric fluctuations, transforming the essentially white noise atmospheric forcing into a red response ocean signal (e.g. Hasselmann 1976). However, if the climate system has preferred (if damped) modes of variability then stochastic forcing results in peaks in the response spectrum at the characteristic timescales of that variability. In general self-sustained MOC oscillations are due to internal nonlinearities in the climate system while oscillations driven by stochastic atmospheric forcing can be largely accounted for by linear dynamics and would not exist without the energy from the external stochastic forcing.

In studies of MOC variability on decadal to centennial timescales, stochastic forcing has been used to model fluctuating freshwater fluxes (e.g. Mikolajewicz & Maier-Reimer 1990; Weaver et al. 1993; Spall 1993; Weisse et al. 1994; Pierce et al. 1995; Aeberhardt et al. 2000), thermal fluxes (e.g. Griffies & Tziperman 1995; Saravanan & McWilliams 1997; Weaver & Valcke 1998; Kravtsov & Ghil 2004) and surface winds (e.g. Holland et al. 2000; Holland et al. 2001; Herbaut et al. 2002). A fundamental challenge in the parameterisation of these fluctuations by stochastic processes is a lack of knowledge of the forcing fluctuation amplitude, decorrelation time, length scale and distribution. The available evidence indicates that the timescales of MOC variability the dominant modes of atmospheric variability are essentially white in time but may not be white in space (e.g. Saravanan et al. 2000).

Identifying spatial patterns of stochastic forcings that effectively excite oscillation in the AMOC was the focus of studies by Tziperman and Ioannou (2002), Capotondi and Holland (1997) and Saravanan and McWilliams (1997). Analysing the linearised dynamics of a 3 box model (representing the polar ocean and the midlatitude surface and deep ocean) using generalised stability theory (e.g. Farrell & Ioannou 1996), Tziperman and Ioannou (2002) determined the optimal spatial structure of the noise that results in maximal variance of the AMOC variability. They found that the optimal forcing induces low-frequency variability by exciting the salinity variability modes of the AMOC. While a three-box model can be useful for theoretical studies, it is too idealised to be quantitatively accurate. In particular, it is impossible to answer the question of whether observations project onto the predicted optimal modes: more complex models are needed to answer this question. Using a three-dimensional ocean model with idealised basin geometry, Capotondi and Holland (1997) analysed decadal variability by considering the spatial pattern of stochastic forcing as a variable of the problem. The period of oscillation of the AMOC was found to be independent of the spatial pattern, leading to the conclusion that the variability at the decadal timescale is an internal mode of the system and not associated with some characteristics of the forcing (although the amplitude of the response was found to depend on the spatial structure of the forcing). Saravanan and McWilliams (1997) found that spatial resonance, defined as the forcing of a system with a spatial pattern that results in oscillations at a preferred frequency, was responsible for exciting the oceanic decadal oscillation in a coupled atmosphere-ocean model. Eliminating the spatial correlations in the forcing was found to substantially reduce the variance associated with the interdecadal oscillation of the AMOC. The spatial pattern of the dominant mode of surface heat-flux interacted with a single oceanic mode to induce AMOC oscillations. Other studies have identified spatial patterns of buoyancy flux variations bearing a strong resemblance to the North Atlantic Oscillation (NAO) and which drive multidecadal to centennial AMOC fluctuations associated with damped internal oscillatory modes (Mikolajewicz & Maier-Reimer 1990; Delworth & Greatbatch 2000; Dong & Sutton, 2005; Bentsen et al. 2004). On the other hand, Spall (1993) found sea surface salinity variance to be a direct response to the stochastic forcing and not an internal mode of variability. As hypothesized by Hasselmann (1976), Spall found that an undisturbed straightforward integration of the white noise freshwater flux anomalies took place in the Labrador Sea and it was the irregularly occurring salinity anomalies here that were responsible for the decadal variability in the North Atlantic.

Model studies have also considered the response of the AMOC to fluctuating surface wind stresses. The random wind forcing field used by Holland et al. (2001) was random in time but had a spatial pattern similar to that of observations. The lack of specific timescales in the forcing indicates that the preferred timescales of the model's response were due to internal model physics and not external forcing. This study showed that the AMOC variability was excited by the stochastic freshwater forcing provided by variable wind driven Arctic ice export and responded linearly to this forcing at interdecadal timescales. Herbaut et al. (2002) described a damped mode of the ocean system requiring stochastic NAO like wind stress anomalies to maintain the oscillation. In this study, wind stress forcing drove anomalous currents; the resulting advection of the mean temperature structure generates temperature anomalies which influenced the strength of the AMOC.

Finally, the character of decadal to centennial-scale variability in the MOC has also been shown to be sensitive to the amplitude of the fluctuating forcing. The propagation of salinity anomalies, which mediate the strength of the AMOC, has been shown to be facilitated by larger random freshwater forcing amplitude (e.g. Weaver et al. 1993; Skagseth & Mork 1998). Furthermore, changes in the strength of the random forcing have been found to cause transitions between different oscillatory states (Aeberhardt et al. 2000).

Because of model uncertainties and limited observations, it is not clear if MOC variability on decadal through centennial timescales is self-sustained or driven by high-frequency variability. Direct comparisons of model results are complicated by variations in model complexity (ranging from box models through two-dimensional models to fully complex three-dimensional models), differences between ocean-only models and coupled ocean-atmosphere models, and the variety of methodologies used to identify the dominant "modes" of decadal to centennial scale variability. Rivin & Tziperman (1997) suggest that the pdf the MOC time series can be used to differentiate between linear noise-forced and nonlinear, self-sustained oscillations.

If the pdf is Gaussian when the stochastic forcing is Gaussian then the oscillations result from stochastic excitation of damped modes, while for nonlinear oscillations the pdf is strongly non-Gaussian. The refinement of tools such as this to distinguish between driven and self-sustained oscillatory variability will be an important component of determining the importance of fluctuating forcing in producing decadal to centennial scale variability in the MOC.

3. AMOC Regimes

The present-day AMOC is characterised by strong NADW formation in the Labrador and Nordic seas, but both paleoclimate and modelling studies suggest that the AMOC can exist in other configurations (e.g. Rahmstorf 2002). There are two sets of feedbacks associated with these rearrangements of the AMOC, involving respectively large-scale and local processes. In its present state, the AMOC transports warm, salty water into the North Atlantic where it is both cooled and freshened. The salt advected northward helps maintain the high densities of water in the North Atlantic and the vigorous formation of NADW. A reduction in deep water formation as a result of surface freshening reduces the poleward transport of salt and amplifies the initial perturbation; GCM simulations demonstrate that if the initial perturbation is sufficiently strong then this large-scale advective feedback can drive the AMOC to another stable steady state in which NADW formation and the overturning circulation are essentially turned off (e.g. Kuhlbrodt et al. 2007). The second set of local feedbacks involve the formation of NADW through deep convection, which homogenises the water column into a convectively neutral state and transports relatively fresh water to depth. If convection is reduced due to surface freshening, the reduced sinking flux of freshwater can amplify the initial surface freshening and (for a sufficiently strong perturbation) shut off convection all together (e.g. Rahmstorf 2001). It is a generic result that the stability properties of nonlinear systems are affected by the presence of noise, and environmental fluctuations that affect the AMOC are ubiquitous (e.g. Weaver et al. 1999). The effects of climate noise on transitions between AMOC regimes will be considered in this section.

The simplest model that captures the AMOC bistability associated with large-scale advective feedbacks is that introduced by Stommel (1961), in which the circulation is associated with the exchange of heat and salt between two well-mixed boxes (representing the high- and low-latitude ocean in a single hemisphere) forced by specified freshwater fluxes and temperature relaxation to an externally specified gradient. Denoting by ΔT and ΔS the interbox temperature and salinity differences respectively, the respective tendencies can be expressed as:

$$\frac{d}{dt}\Delta T = -q(\Delta\rho, t)\Delta T + \Gamma(\Delta T_0 - \Delta T) \tag{3.1}$$

$$\frac{d}{dt}\Delta S = -q(\Delta\rho, t)\Delta S + F(t) \tag{3.2}$$

where ΔT_0 is the externally imposed interbox temperature gradient to which the system relaxes on a timescale of Γ^{-1} and F is the imposed freshwater forcing. The interbox exchange, q, is assumed to depend on the interbox density gradient $\Delta\rho = \alpha_S\Delta S - \alpha_T\Delta T$ (α_S and α_T are respectively the haline and thermal expansivities of

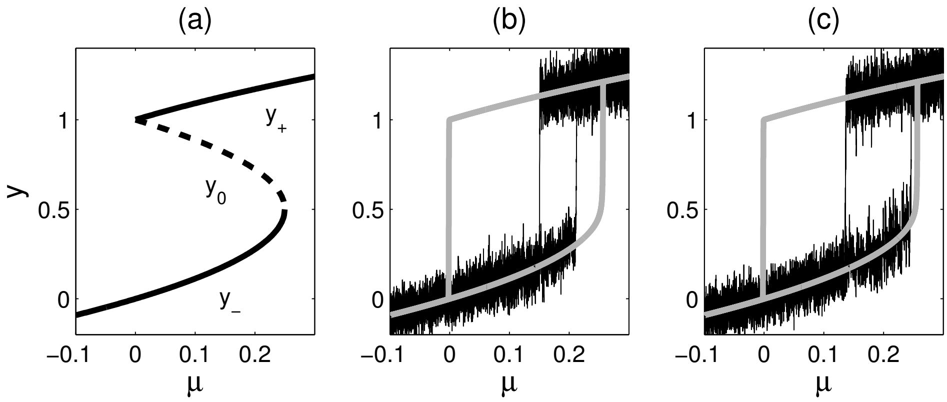

Figure 1. (a) Bifurcation structure of the AMOC box model Eqn. (3.3) with $\eta = \xi = 0$. The solid (dashed) lines are stable (unstable) solution branches. (b) Grey curve: hysteresis loop traced out by deterministic model as freshwater forcing μ is moved increased from -0.1 to 0.3 and then back to -0.1. Black curve: realisation of hysteresis curve for the stochastic model with $\eta = \sigma_1 \dot{W}_1$, $\xi = \sigma_2 \dot{W}_2$ where $\sigma_1 = \sigma_2 = 0.075$ and $\dot{W}_i$ are independent white noise processes. (c) As in (b), for a second realisation of the stochastic model. With very high probability, the stochastic hysteresis loops are smaller than the deterministic ones.

seawater) as $q(\Delta\rho, t) = \beta(t) + f(|\Delta\rho|)$ where $\beta(t)$ is a "diffusive" exchange (including interbox fluxes mediated by gyre circulations; Monahan 2002c) and the function $f(\Delta\rho)$ models advective exchange by the overturning circulation (e.g. Stommel 1961; Cessi 1994). As in Monahan (2002), we will focus on the model $f(|\Delta\rho|) = c|\Delta\rho|$; the following results are not qualitatively sensitive to this parameterisation.

This already highly idealised system can be further simplified by taking the temperature relaxation timescale to be much faster than the interbox exchange timescale (a reasonable approximation), so to leading order $\Delta T \simeq \Delta T_0$ and we obtain a differential equation in ΔS alone. The influence of high-frequency fluctuations on the overturning will be modelled by taking $\beta(t)$ and $F(t)$ each to be the sum of a fixed mean and random fluctuations. Non-dimensionalising the resulting equation as in Monahan (2002a), we obtain the stochastic differential equation (SDE)

$$\frac{d}{dt} y = -(b_0 + |1 - y|)y - y \circ \eta + \mu + \xi \tag{3.3}$$

where y is the non-dimensional salinity gradient, b_0 and η are respectively the mean and fluctuations of the non-dimensional diffusive exchange parameter, and μ and ξ are respectively the mean and fluctuations of the freshwater flux (the open circle indicates that for η modelled as white noise the SDE is interpreted in the Stratonovich sense; Gardiner 1997). This highly idealised model of the AMOC (or ones very similar) have been considered in a number of studies (e.g. Stommel & Young 1993; Cessi 1994; Bryan & Hansen 1995; Lohmann & Schneider 1999; Timmermann & Lohmann 2000; Vélez-Belichí et al. 2001; Monahan et al. 2002; Monahan 2002a,b; Kleinen et al. 2002).

A plot of steady state solutions of Eqn (3.3) for $b_0 = 0$ in the absence of fluctuations ($\eta = \xi = 0$) is given in Figure 1. For a range of values of the freshwater forcing parameter $0 \leq \mu \leq 0.25$, the idealised model admits three steady states. Two of these are stable, y_+ and y_-, with respectively weak (strong) overturning circulations and strong (weak) interbox density gradients. The third steady state y_0 is unstable. This interval of bistability is bounded by fold bifurcations, beyond which only a single steady state solution exists. If the parameter μ is increased from below the lower bifurcation point to above the upper bifurcation point and then decreased again to below the lower bifurcation, the hysteresis loop displayed in Figures 1(b), (c) is traced out. Because the steady states are equilibrium solutions, deterministic transitions between solution branches for an evolving freshwater forcing $\mu(t)$ will generally occur somewhat beyond the bifurcation point; that is, there will be a slight overshoot (e.g. Berglund & Gentz 2006). The fact that 3-dimensional coupled ocean-atmosphere models produce analogous hysteresis structures (e.g. Weaver & Hughes 1994, Kuhlbrodt et al. 2007) suggests that this AMOC model captures the essential physics of the advective feedback bistability (although its predictions cannot be expected to be quantitatively meaningful).

In the absence of climate noise, AMOC regime transitions can only occur if μ is moved beyond one of the fold bifurcations. However, in the presence of noise, transitions between regimes can occur within the region of bistability; if the noise is unbounded (as with Gaussian fluctuations), then transitions are possible everywhere both states exist. For η and ξ white noise processes, the mean transition time from y_- to y_+ can be computed analytically (e.g. Cessi 1994, Monahan 2002a) and takes the form $\tau(y_- \to y_+) \sim \exp\left(-V(y_-, y_0)/\Sigma^2\right)$ where Σ is a measure of the noise level and $V(y_-, y_0)$ is a measure of the "potential barrier" the system must overcome to pass from y_- to y_+ (an analogous expression holds for the reverse transition). These transition rates are highly sensitive to the strength of the noise forcing; a small change in Σ can change τ by orders of magnitude. If the transition time out of an AMOC regime is longer than any physically meaningful timescale, then the system will remain (with very high probability) in this state effectively forever (the limit considered by Bryan & Hansen, 1995). If on the other hand the residence times of both regimes are smaller than the longest physically meaningful timescales, the AMOC will pass back and forth between regimes, exploring thoroughly the available set of states. In this case the signature of multiple regimes will be multimodality of the stationary pdf (the long-term equilibrium pdf). For η and ξ white noise processes, the Fokker-Planck equation for the stationary pdf associated with Eqn. (3.3) can be solved analytically (Cessi 1994; Timmermann & Lohmann 2000; Monahan 2002a). Finally, if the noise strength is so large that typical excursions of $y(t)$ are much larger than the separation between y_- and y_+, then the system will not feel the presence of the different deterministic equilibria and the two peaks of the stationary pdf will not be well-separated (the limit considered by Stommel & Young 1993); this high-noise case is not relevant to the AMOC.

If the intensity of the climate noise is independent of the state of the AMOC (that is, if the noise is *additive*), then the primary effect of stochastic fluctuations is inducing transitions between AMOC regimes (or exciting oscillations, as discussed in the previous section). However, if the intensity of one or more of the noise processes is dependent on the state of the system (that is, if the noise is *multiplicative*) then the noise itself can alter qualitative aspects of the dynamics (e.g. Gardiner 1997). In Eqn. (3.3), the fluctuations in diffusive exchange η enter multiplicatively (when η is taken to be white noise), and their intensity has an effect on the multimodality of the stationary AMOC pdf. As the intensity of η increases, the range of freshwater forcings μ over which the pdf is bimodal shifts and contracts, eventually vanishing (Timmermann & Lohmann 2000; Monahan 2002a). The domain of bistability is also altered if η and ξ are correlated (as they might be expected to be physically; Monahan 2002b). In this way, the structure of the AMOC regimes (rather than just their occupation statistics) is influenced by the stochastic climate forcing.

Shifts in the domain of bimodality produced by multiplicative noise persist when η is allowed to have a nonzero autocorrelation time (i.e. to be red noise). In this situation, then the dynamics of the vector (y, η) is governed by a 2-dimensional SDE with an associated Fokker-Planck equation (not analytically tractable). This stationary pdf is multimodal outside of the range of freshwater forcing values μ for which the deterministic part of the dynamics has multiple equilibria (Timmermann & Lohmann 2002; Monahan et al. 2002; Monahan 2002b). Bimodality occurs when there is a neighbourhood of the (y, η) state space without a fixed point but with a local minimum of the magnitude of the deterministic tendency. The system lingers in this neighbourhood when driven in by random fluctuations, building up probability mass; the resulting pdf can be bimodal (Monahan 2002b).

For moderate values of the noise intensity, the residence times of the two regimes generally differ by orders of magnitude. It follows that for such noise levels the probability of being in one regime is orders of magnitude greater than that of being in the other, so while the pdf is technically bimodal it is effectively unimodal. This phenomenon and its consequences were discussed in Monahan (2002a,b), where it was referred to as regime stabilisation by noise. One particularly significant consequence is that if the freshwater forcing approaches the fold bifurcation where the current AMOC regime vanishes, then a transition to the other regime will occur before the bifurcation point is reached (Figure 1b,c). The point at which the transition occurs is random, with a mean value that depends on both the noise level and the rate at which μ is changing (e.g. Berglund & Gentz 2006). This effect is seen in both idealised models (Monahan 2002a) and more comprehensive coupled atmosphere-ocean models (Wang et al. 1999; Knutti & Stocker 2002). These transitions become less predictable as the bifurcation point is approached. In the presence of climate noise, AMOC regime shifts become less predictable as they become more likely.

The response of the AMOC to climate noise does provide a potential "early warning system" for regime transitions (Kleinen et al. 2002; Held & Kleinen 2004). As the bifurcation point is approached, the feedbacks driving the system towards the equilibrium state weaken. In consequence, the autocorrelation timescale and variance of AMOC fluctuations both increase. Increasing trends in either of these statistics could herald the approach of a bifurcation point (as Carpenter & Brock (2006) have also noted in an ecological context), and these trends could potentially be measured using operational AMOC monitoring networks. The practical utility of such an early warning system would depend on the ability to statistically distinguish anthropogenic trends from variability intrinsic to the climate system.

The influence of fluctuating climate forcing on AMOC regime dynamics in more sophisticated ocean models indicates that the conclusions drawn from box models are qualitatively robust. In the zonally-averaged 2D ocean model used to study the response of centennial-scale AMOC variability to fluctuating freshwater forcing, Mysak et al. (1993) find noise-induced transitions between three distinct AMOC regimes in the limit of large horizontal diffusivity and small vertical diffusivity. Eyink (2005) obtains analytical solutions for the stationary pdf of the AMOC in a similar model. The deterministic equilibria and response to stochastic forcing in this model are broadly consistent with the box model results presented above, despite the considerably greater sophistication of the model. Furthermore, the AMOC transition early warning system originally characterised in a box model by Kleinen et al. (2002) was shown to be characteristic of a coupled ocean-atmosphere model of intermediate complexity (Held & Kleinen 2004).

The discussion of AMOC regimes has so far focused on those associated with large-scale advective feedbacks. In the absence of stochastic forcing, local convective feedbacks lead to either multimodal or oscillatory behaviour (e.g. Weaver et al. 1993; Cessi 1996; Rahmstorf 2001) . For the former case, the effects of climate noise are broadly the same as those associated with the advective feedback multiple equilibria (Kuhlbrodt et al. 2001; Kuhlbrodt & Monahan 2003). For the latter case, stochastic forcing can modify the character of the oscillations and the parameter range over which they occur (Weaver et al. 1993; Cessi 1996) as well as driving the AMOC between different oscillating regimes (Aeberhardt et al. 2000). Furthermore, diffusive processes within the ocean have been shown to produce millennial-scale oscillatory responses whose timescale depends strongly on fluctuating freshwater forcing (e.g. Weaver & Hughes 1994).

4. Stochastic Resonance and Dansgaard-Oeschger Events

Evidence from paleoclimate records, particularly high-latitude ice and deep-sea cores, demonstrates that the climate of the last glacial period was characterised by a succession of abrupt shifts between relatively cold (stadial) and relatively warm (interstadial) states (e.g. Rahmstorf 2002). These transitions, known as Dansgaard-Oeschger (DO) events, are evident in the records of ice oxygen isotopic composition (δ^{18}O; a measure of high-latitude temperature) and calcium concentration (a measure of atmospheric dustiness) from the Greenland Ice Core Project (GRIP) presented in Figure 2. In particular, the joint distribution of δ^{18}O with the logarithm of the Ca concentration is manifestly bimodal with clearly-separated stadial and interstadial regimes (Fuhrer et al. 1999). It is evident from Figure 2 that while DO events occur with millennial timescales, they do not simply reflect a regular sinusoidal oscillation of the climate system.

While feedbacks in many different components of the climate system are involved in DO events (as external periodic forcing might be; see below), the AMOC is found to play a central role in stadial/interstadial transitions as a result of (i) the importance of the AMOC for the transport of heat to the North Atlantic, and (ii) the potentially nonlinear response of the AMOC to buoyancy forcing (e.g. Clark et al. 2002; Rahmstorf 2002). Paleoclimate data indicate that the climate system is much more variable during glacial periods than during interglacials, and modelling studies suggest that the stability properties of the AMOC are also considerably different between glacial and interglacial periods (e.g. Ganopolski & Rahmstorf 2001; Schmittner et al. 2002). During glacial periods, a stadial circulation state with NADW formation in the subpolar North Atlantic is stable (or weakly metastable), while an interstadial state with NADW formation in the Nordic seas is unstable

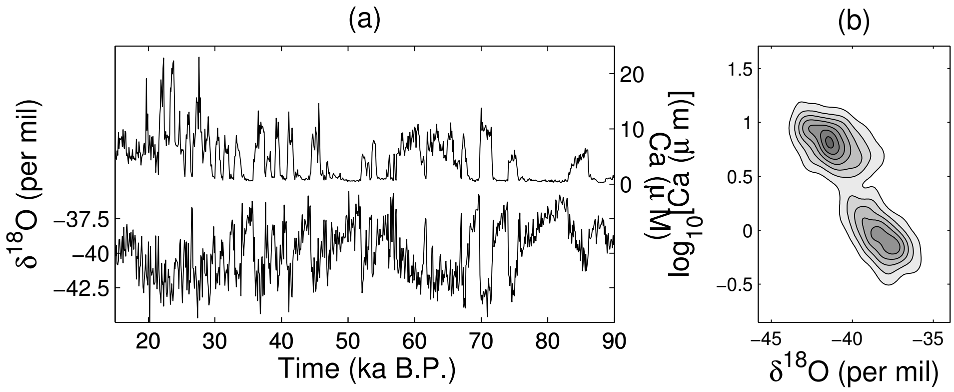

Figure 2. Oxygen isotope and Ca concentration data from the GRIP ice core. The δ^{18}O record measures the excess of O^{18} over O^{16} in the ice and is a measure of high-latitude temperature (higher δ^{18}O corresponding to warmer temperatures), while the Ca record is a measure of atmospheric dustiness (Fuhrer et al. 1999). (a) δ^{18}O and Ca concentration from 15,000 to 90,000 years before present, linearly interpolated to a nominally uniform temporal resolution of 100 years. (b) joint pdf of δ^{18}O and $\log_{10}$Ca.

but easily excited from the stadial state by freshwater perturbations to the North Atlantic. While the interstadial state is not steady, the trajectory of the system is relatively slow in its immediate neighbourhood. This picture is consistent with the observed evolution of DO events: a rapid transition from stadial to interstadial is followed by a gradual relaxation of the AMOC towards the stadial state with a final rapid shift. Note that simple models of the AMOC discussed in the previous section demonstrate that it is possible for the pdf of the stochastic system to be bimodal (as seen in the paleoclimate records; Figure 2) even when the deterministic component of the system has a single fixed point, if the deterministic tendency takes a local minimum in some neighbourhood. Evidence for a third circulation state with essentially no NADW production (the so-called "Heinrich mode") is also found in paleoclimate records and in climate models (e.g. Rahmstorf 2002; Schmittner et al. 2002). The North Atlantic climate during a Heinrich mode is indistinguishable from that during a stadial period.

Analyses of North Atlantic paleoclimate records (both glacial and deep-sea) of the last glacial period suggest the presence of a $\sim$1500-year periodic signal associated with the sequence of DO events (e.g. Mayewski et al. 1997). The existence of a well-defined spectral peak in the time series, however, depends on both the dataset (e.g. Ditlevsen et al. 2005) and the time period (e.g. Schulz 2002) under consideration, and has been suggested to be a spurious signal resulting from aliasing of the annual cycle (Wunsch 2000). Furthermore, there is evidence (still controversial, as discussed below) that DO events do not occur with strictly regular periodicity, but are separated by time intervals which are approximately integer multiples of 1470 years (Alley et al. 2001; Schulz 2002; Rahmstorf 2003). Such behaviour is the hallmark of the phenomenon of *stochastic resonance* (SR), in which the addition of noise to a periodic sub-threshold signal results in approximately periodic crossings of some particular threshold. On occasion, a crossing will be missed, so intervals between successive crossings will cluster together around integer multiples of the period of the forcing; the distribution of these crossing times will decay exponentially (Alley et al. 2001). Stochastic resonance was originally introduced as a model for glacial/interglacial transitions in response to Milankovitch forcing (Benzi et al. 1982), and although it does not appear to be relevant in this original context SR has since been found to be characteristic of a broad range of physical and biological systems (e.g. Gammaitoni et al 1998).

Vélez-Belchí et al. (2001) used a stochastic box model (such as in Section 3) to demonstrate SR in the AMOC in response to periodic forcing; although this study focused on Milankovitch forcings rather than the millennial timescales characteristic of DO events, it demonstrated that SR occurs over a broad range of driving periods. Stochastic resonance on millennial timescales was demonstrated by Ganopolski & Rahmstorf (2002) in a more sophisticated coupled atmosphere-ocean model forced by boundary conditions appropriate for the last glacial period. In this study, a small sinusoidal freshwater perturbation with a period of 1470 years was applied to the North Atlantic, along with stochastic freshwater fluxes (with amplitudes comparable to present-day interannual variability). In response, the simulated AMOC displayed SR, alternating between stadial and interstadial circulation states with a distribution of transition times that compared favourably with those of DO events in the GRIP ice core. Stochastic resonance could be achieved with realistic noise levels because the interstadial state is easily excited from the stadial state by freshwater perturbations to the North Atlantic (Ganopolski & Rahmstorf 2001; Schmittner et al. 2002) . In Ganopolski & Rahmstorf (2002) the 1470-year periodic forcing was some external perturbation of unknown provenance; using the same model, Braun et al. (2005) suggest that the forcing may in fact arise through the superposition of 87- and 210-year Gleissberg and DeVries solar cycles. When forced by a linear combination of North Atlantic freshwater forcings with these periodicities (or a more realistic modulation of the 11-year solar cycle by the Gleissberg cycle), the model responds with a 1470-year periodic alternation between stadial and interstadial states (for certain forcing parameter ranges). That the AMOC should respond on millennial timescales to this shorter timescale forcing was attributed to a combination of strongly nonlinear dynamics and the long intrinsic AMOC adjustment timescales.

Using an idealised coupled atmosphere-ocean-sea-ice model, Timmermann et al. (2003) suggest a variation on the SR mechanism for driving cycles of DO events. Instead of being driven by an external periodic signal, DO events in this model occur through a combination of noise and a periodic limit cycle internal to the climate system itself in a phenomenon known as *coherence resonance* (CR) or *autonomous stochastic resonance* (also discussed in Ganopolski & Rahmstorf 2002). An essential (and testable) difference between SR and CR is that in the former the phase of the external driving signal and the resulting transitions is fixed, while in the latter the phase of the oscillation can drift as a result of internal climate system interactions.

Stochastic resonance is a meaningful candidate mechanism for driving DO event cycles only to the extent that the distribution of observed inter-transition times clusters around integer multiples of a single (1470-year) timescale, with transitions that are phase-locked to this external periodic forcing. An exponential distribution of transition times without this clustering would be suggestive of stadial-interstadial transitions being driven by climate noise alone, without a periodic forcing (e.g.

Ditlevsen et al. 2007), while a looser clustering of transition times without locking to a periodic signal of fixed phase would be suggestive of CR. Distinguishing between these alternatives is a statistical problem complicated by (i) the difficulty of defining precisely when a transition has occurred, (ii) problems with ice core chronology, and (iii) the diversity of null hypotheses against which the data can be compared. Time series analyses by Roe & Steig (2004) and Ditlevsen et al. (2005,2007) suggest that a model with stadial/interstadial transitions driven by climate noise without a preferred periodic forcing provides a better fit to the observations than a stochastically resonant model, particularly for the newly-obtained NGRIP ice core (Ditlevsen et al. 2007). As well, the influence of solar variability on the onset of DO events has been questioned (Muscheler & Beer 2006). Because of the numerous uncertainties involved in the reconstruction of past climates, SR remains a controversial mechanism for the pacing of DO event cycles.

5. Conclusions

Variability of the oceanic meridional overturning circulation is an important component of variability in the climate system on timescale from decades through centuries to millennia. Modelling studies suggest that MOC variability on these "climate" timescales may be strongly influenced by fluctuations in surface forcing on much shorter "weather" timescales. Such high-frequency forcing has typically entered models of the MOC through stochastic processes parameterising unresolved atmospheric processes, representing fluctuations in surface buoyancy and (less often) mechanical fluxes. A significant gap in our understanding of the importance of stochastic forcing comes not through its role in external forcing but rather, through the means that the effects of this external forcing is parameterized in ocean and climate models. Mechanical forcing provides the energy necessary to drive the thermohaline circulation both via direct mixing (through both wind- and tidally-generated internal wave breaking) or through wind-driven upwelling in the Southern Ocean (e.g. Kuhlbrodt et al., 2007). A natural question arises as to whether or not the circulation in large-scale ocean models is sensitive to random fluctuations in mixing associated with the internal wave field, which is patchy and episodic.

Recent observations of MOC transport at 26.5°N from the Rapid Climate Change (RAPID) mooring array measure variability in the maximum meridional overturning with a standard deviation of 5.6 Sv around the mean value of 18.7 Sv on subannual timescales from 2004-2005 (Cunningham et al. 2007). The existence of such variability has important consequences for predictability of the MOC, both because it can make detection of trends difficult and because these fluctuations themselves may influence the timing of transitions between circulation regimes. Understanding the sources and consequences of this variability is of more than just theoretical interest.

The ice core data was generously provided by Eric Wolff (British Antarctic Survey), Sigfus Johnsen (University of Copenhagen) and Bernhard Stauffer (University of Bern). AHM and AJW acknowledge support from Natural Sciences and Engineering Research Council of Canada and from the Canadian Foundation for Climate and Atmospheric Sciences (CFCAS). JA acknowledges support from CFCAS. AHM also acknowledges support from the Canadian Institute for Advanced Research Earth System Evolution Program.

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