

# Mechanisms of Climate Variability from Years to Decades

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## Abstract

This paper discusses and reviews some of the mechanisms that may be responsible for climate variability on yearly to decadal timescales. The discussion is organized around a set of mechanisms that primarily involve the atmosphere, the ocean, or the coupling between the two. We choose an example of each, try to explain what the underlying mechanism is, and set it in the context of climate variability as a whole. All of the mechanisms are in principle deterministic, although in at least one of them we do not care about the details of the process that give rise to the variability and in that case a stochastic description may be the most economical and insightful.

*One person's signal is another person's noise.*

## I Preamble

This is an essay on the mechanisms of natural climate variability on timescales of years to decades. It is meant to serve both as an introductory chapter to the articles appearing later in this book that delve into the mechanisms and modelling in greater depth, and as a stand-alone article for those requiring an overview, or at least a perspective, of the subject. In this preamble we'll discuss rather generally the nature of stochastic and deterministic processes and their role in weather and climate, and in the following sections we'll focus more explicitly on climate variability on timescales from years to decades, emphasizing processes that primarily involve the atmosphere and/or ocean.

Variability of climate — indeed variability of many systems — is often partitioned into two categories, *stochastic* and *deterministic*, each associated with rather different mechanisms. According to one dictionary, stochastic means 'randomly determined, having a random probability distribution or pattern that may be analyzed statistically, but may

not be predicted precisely'. Another dictionary defines stochastic as 'involving a random variable' or 'involving chance or probability'. Deterministic, on the other hand, is usually taken to refer to a phenomenon whose outcome is causally determined, at least in principle, by preceding events in conjunction with the laws of nature; thus, a deterministic sequence is one that may be predicted to a specified degree of accuracy, using appropriate equations of motion, if the initial conditions are given. Now, discounting quantum effects, the laws of nature are wholly deterministic — they may be cast as equations of motion that predict the evolution of objects given their state at some instant. Thus, *nearly all phenomena in climate dynamics are deterministic*. This statement, although true, is however perhaps not the whole truth, for whereas a system may be deterministic in principle in practice we may not be able to predict it for at least two reasons:

- (i) We are unable to compute the details of the evolution of part or all of the system because the system is chaotic. No matter how well we know the initial conditions, if not perfectly, then the future outcome is unpredictable and may be best described statistically.
- (ii) Our *knowledge* of the system is imperfect and so we represent possible outcomes by probabilities, as if the system were stochastic. The probabilities then reflect our uncertain knowledge of the system, rather than an inherent indeterminism.

The weather and the climate, respectively, provide illustrations of these two points. As is well known, the earth's atmosphere is a chaotic system and even if we could know the initial conditions extremely accurately (and had a very good numerical weather prediction model) the details of the future weather would still be unpredictable after a couple of weeks. The climate (as usually defined as some kind of average of the weather, or the statistics of the weather) is more predictable in this sense: for example, if we knew how much carbon dioxide we were to put in the atmosphere then the degree of global warming should be predictable, and the fact that it is not reflects our ignorance of climate dynamics. Roe & Baker (2007) argue that even a small amount of ignorance may lead inevitably to large uncertainties in climate projections, but even if this hypothesis is granted the ensuing probability distribution for a climate projection is still of a somewhat different nature than that of weather forecasts: the climate probability distribution primarily reflects our ignorance of how the laws of physics and chemistry apply to the earth's climate, whereas the weather probability distribution reflects the amplification of small fluctuations by chaos.

Nevertheless, there are similarities in the two cases, in the sense that both reflect an ignorance of some aspect of the system, an ignorance that is amplified either by the chaos of the system in the case of weather, or the feedbacks within the system, in the case of climate, and so lend themselves to probabilistic approaches. But there is another reason — perhaps the main one in our context — for studying stochastic processes; it is that we *don't care about the details of a particular process*. We care only about its statistical properties, and sometimes only its variance. One example of this lies in the small scales of turbulence — by and large we don't care about the path of a small eddy near the viscous scale, but we

do care about the statistical properties of eddies in cascading energy and/or enstrophy to small scales where they may be dissipated. There is a similar aspect to climate variability on the decadal timescale, in that we don't care about the weather that is taking place on timescales of weeks. Now, weather can be explicitly modelled far better than it can be parameterized by a stochastic process, but the details of the weather are generally irrelevant to decadal scale climate variability, only its statistics may matter. We may therefore choose to model weather as some kind of stochastic process, running the risk that it might be improperly modelled, in the hopes of isolating the mechanisms that might give rise to climate variability on longer timescales. Similarly, for those whose interest is variability on timescales of millions of years then even centennial variability might best be treated as noise.

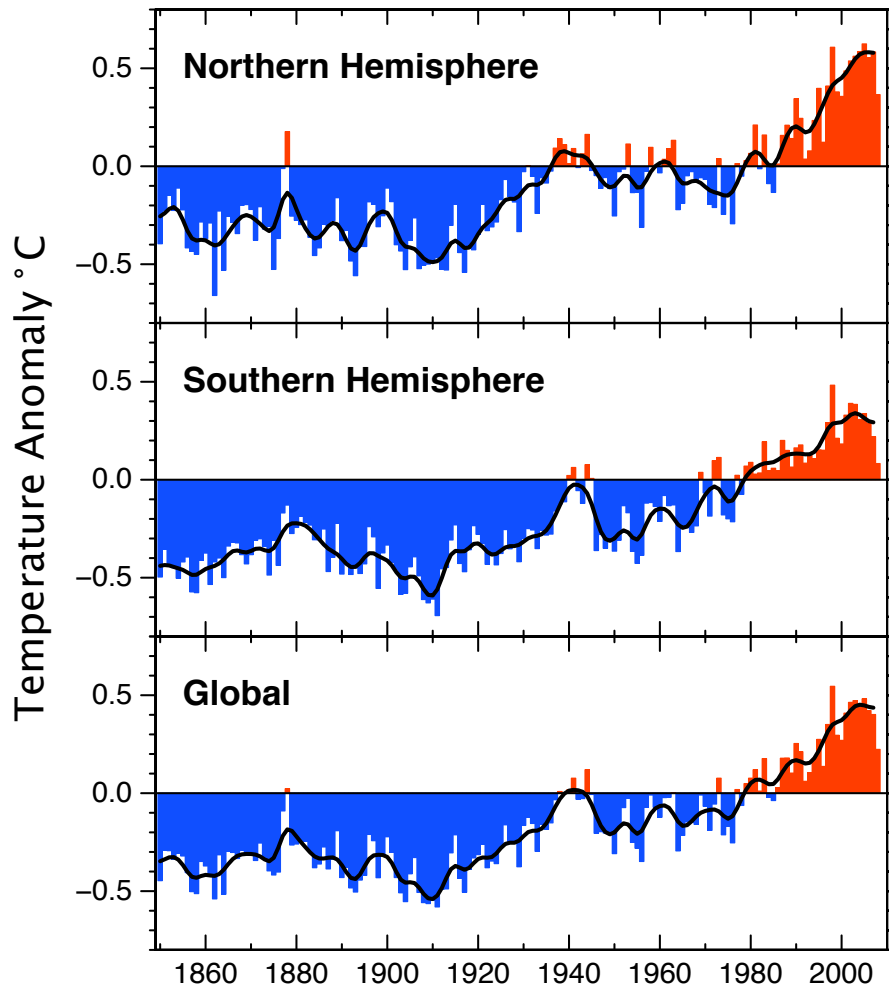
In the rest of this article we focus on climate variability on timescales of years to decades. Our discussion is organized around mechanisms. Thus, following a brief look at some observations, we summarize the general classes of mechanisms that might give rise to such variability. Each of the subsequent sections is then devoted to one type of mechanism, illustrating it with one or two examples.

## **2 Observations and Classes of Mechanisms**

### **2.1 A few observations**

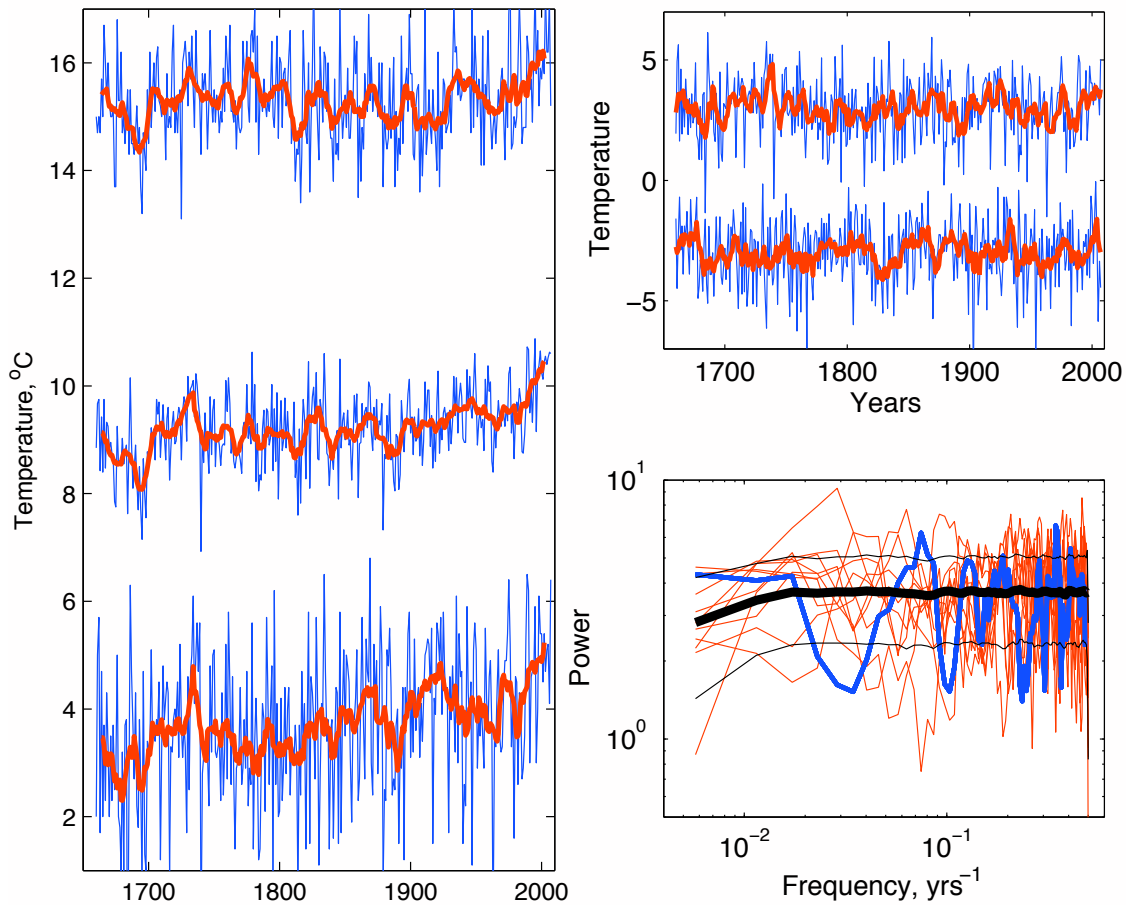
Climate variability exists on timescales of seasons to millennia, but in this article our emphasis will be on timescales of years to centuries, or the decadal time scale. A rough indication of such variability is shown in Fig. 1, where the globally averaged surface temperature is plotted for the period 1850–2007. In addition to the evident general warming trend one seems to see variability on the decadal scale, which also seem evident if one restricts attention to a particular region of the globe, as in the central England temperatures shown in Fig. 2, which is perhaps the longest continuous instrumental record in climate.

However, we need to be rather careful that we are not deceived by a casual visual inspection of such time series into thinking that there is more decadal variability than might be expected by chance. To illustrate this we perform a Monte Carlo simulation by taking the time series of successive winter temperatures from the central England time series and shuffling the temperatures randomly; in the resulting time series any mechanistic decadal variability has manifestly been removed. After detrending to remove secular changes we plot a realization of a shuffled time series alongside the original time series, and we see that the two series look remarkably similar (upper right panel of Fig. 2). The power spectra of the original series is fairly white for periods from about 1 year to 200 years, with some apparent peaks at about 10 years and 100 years, but the power spectra of the shuffled



**Figure 1.** The global average surface temperature (anomaly from 1961–1990 average) from the HadCRUT3 data, from Brohan *et al.* (2006). In addition to the general warming trend some decadal variability seems apparent, presumably due to natural variability in the system.

time series are not qualitatively different from the original (lower right panel of Fig. 2). In this figure we show the power spectra of the original time series (thick blue line), a sampling of the spectra computed from shuffled time series (thin red lines), the mean of these (thick black line) and the mean plus or minus one standard deviation (thin black lines). Some of the shuffled time series have as much or more decadal scale variability as the original one, although the original time series does just stand out from the noise at long time periods — although the sceptic may certainly argue that decadal variability has not been demonstrated from this time series alone. Other, more complete, observational analyses have detected decadal scale variability, especially when account is taken of the spatial patterns in the data (e.g. Mann & Park 1994; Turre *et al.* 1999). Similarly, Biondi *et al.* (2001) conclude the climate in and around the Pacific region has undergone real



**Figure 2.** Left: Central England temperature from 1650 to 2007, using the HadCET data (Parker *et al.* 1992). The three sets of curves show, from the top, summer (JJA), annual average, and winter (DJF) temperatures. The thin blue curve shows the average over the season or year, and the thicker red curve shows the ten year running mean. Right top: detrended winter temperature anomalies; thin blue line is the actual temperatures, and the thick red line is the temperatures after a six year running mean. Lower curve is the same for a sample shuffled time series, in which the ordering of the years is random. The real and shuffled series are offset from the origin by plus and minus 3° C respectively. Right bottom: Power spectra of real and shuffled central England temperature. Thick blue line is the power spectrum of the actual winter (DJF) temperatures, the thick black line is the mean power spectrum of 1000 shuffled series, with plus and minus one standard deviation marked by thin black lines. The thin red lines are the power spectra of a sample of 10 shuffled time series.

decadal scale variability over the past few centuries. Overall, it is a defensible conclusion to draw that decadal and longer variability is present in the climate system, but by most measures the signal is weak. The weakness of the signal is not a reason for neglecting it, since any skill at prediction on decadal time scales would be enormously important. Rather, it is a warning that the mechanisms of such variability will likely not reveal themselves easily to the investigator.

## **2.2 General mechanisms**

Although climate variability is difficult to define without being either overly general or overly prescriptive, in this article we will regard it as the variability of large-scale atmospheric or oceanic fields (such as surface temperature or precipitation) on timescales of a season or longer. Restricting attention to processes that primarily involve either the ocean or the atmosphere, the source of such variability could arise in the following general ways.

1. Climate variability might arise primarily from the atmosphere. That is, the atmosphere might vary on timescales longer than those normally associated with the baroclinic lifecycle, or have long-lived regimes of behaviour, independent of varying boundary conditions such as sea-surface temperature.
2. Atmospheric variability on short time scales might be suppressed by the presence of an ocean with a large heat capacity, leading to a red spectrum of climate variability. This mechanism, as proposed by Hasselmann (1976) and Frankignoul & Hasselmann (1977), has become a de facto null hypothesis for climate variability.
3. Climate variability might arise via coupled modes, that is via non-trivial interactions between the ocean and atmosphere. The ENSO cycle is one example, perhaps even the only uncontroversial example.
4. Climate variability might have a primarily oceanic origin. Ocean variability might affect the atmosphere, and so the climate, without the need for coupled modes of the kind envisioned in item 3.
5. Secular changes in climate can be caused by changes in forcings external to the ocean-atmosphere system. This includes changes in atmospheric composition (such as carbon dioxide concentration), incoming solar radiation, volcanism, and changes in land surface and distribution.

The next few sections will discuss these mechanisms, excluding the last item which is well documented elsewhere. We do not discuss the El Niño phenomenon for similar reasons. I don't aim to provide a comprehensive review, but nor is my aim to be provocative for its own sake. Rather, the goal is to provide a perspective, to illustrate some of the mechanisms

with results from coupled ocean-atmosphere models, and to see how deterministic or stochastic ideas might fit in with them.

### 3 Atmospheric Variability

In the extra-tropical atmosphere the primary mechanism of variability on large-scales is baroclinic instability, the basic life-cycle of which, from genesis to maturation to decay, is about 10 days (e.g., Simmons & Hoskins 1978). The baroclinic time scale stems from the growth rate of baroclinic instability, and the simplest measure of this is the Eady growth rate,

$$\sigma \equiv \frac{0.3\Lambda H}{L_d} = \frac{0.3U}{L_d} \quad (3.1)$$

where  $\Lambda$  is the shear,  $H$  a vertical scale,  $U$  a horizontal velocity and  $L_d$  is the deformation radius. For values of  $H = 10$  km,  $U = 10$  m s<sup>-1</sup> and  $L_d = 1000$  km we obtain  $\sigma \approx 1/4$  days<sup>-1</sup>. (If  $\beta \neq 0$  the height scale of the instability may be changed — it is no longer necessarily the height of the troposphere — but in practice a similar time scale emerges.) The advective time scale of a baroclinic disturbance can similarly be expected to be about  $L_d/U$ , or a few days, and the total lifecycle, although not exactly an advective timescale, might be expected to be a multiple of it. [In the ocean the baroclinic lifecycle is longer, primarily because the oceanic  $U$  is two orders of magnitude smaller — 10 cm as opposed to 10 m— and so even though the oceanic  $L_d$  is one order of magnitude smaller — 100 km as opposed to 1000 km— the oceanic eddy time scales are roughly ten times longer than the atmospheric ones.]

Of course baroclinic eddies are non-linear, so that time scales considerably longer than the advective scales can in principle be produced. It is known, for example, that baroclinic waves interact with the stationary wave pattern, produced by flow over topography and over large-scale heat anomalies, such as cold continental land masses, to produce slowly varying planetary waves as well storm tracks (e.g., Chang *et al.* 2002) to produce intra-seasonal variability. The zonal index will also vary on intra-seasonal timescales by way of an interaction between the baroclinic eddies and the zonally-averaged flow — this type of interaction is often invoked to explain the variability associated with the North Atlantic Oscillation and with so-called annular modes (Feldstein & Lee 1998; Hartmann & Lo 1998; Vallis *et al.* 2004; Vallis & Gerber 2008). There is evidence that such interactions can involve feedbacks that give rise to timescales longer than those normally associated with the baroclinic lifecycle (Robinson 2000; Gerber & Vallis 2007), although as currently understood they do not give rise to any predictable timescales longer than a few weeks, or months at most. Nevertheless, we can also expect some interannual variability essentially as a residual of the intra-seasonal variability (e.g., Feldstein 2000), but such interannual variability will be weak and unpredictable

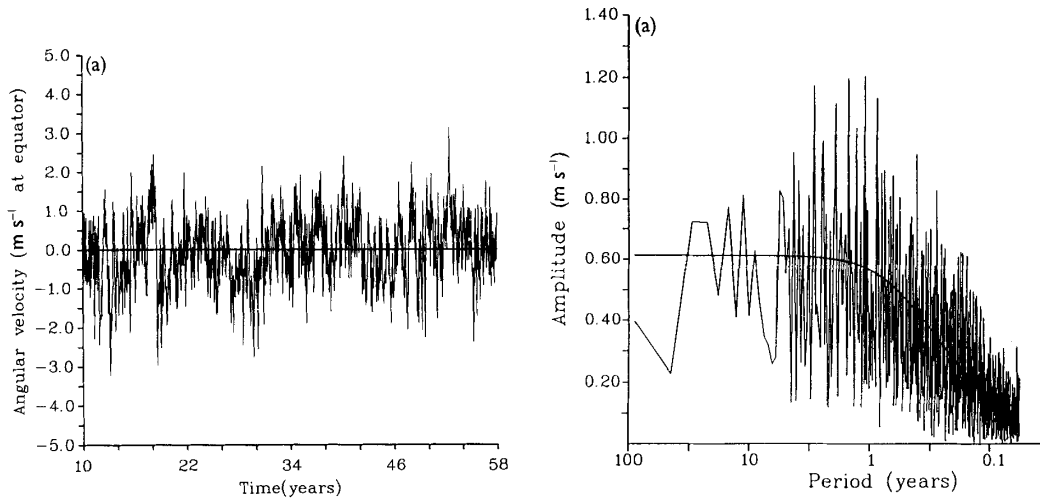
However, for the atmosphere to produce variability on timescales significantly longer

than a few weeks — for example with some peak in the power spectrum at interannual timescales — would likely require there to be some kind of regime behaviour, in which the gross atmospheric behaviour changes on timescales independent of those associated with baroclinic instability or stationary waves. Such behaviour is certainly not impossible, for, to give one example, atmospheric blocks appear to have a timescale not closely associated with baroclinic waves. However, it seems unlikely that the atmosphere alone could give rise to significant, predictable, natural interannual variability, for two reasons:

- (i) No mechanism is apparent that could produce such variability, except as a residual of intra-seasonal variability.
- (ii) Suppose that the atmosphere *were* able to produce regime-like behaviour when steadily forced. But the difference between any two realistic regimes would likely be much smaller than the seasonal cycle, and it would seem likely that a seasonal cycle would disrupt any regime behaviour that persisted beyond a few months.

The first argument is rather weak, being an example of what has been called ‘an argument from personal incredulity’. The second argument is a little stronger, for it does propose a mechanism that would prevent long time scales from emerging. Most integrations with atmospheric GCMs do not produce significant variability on interannual timescales (an issue we revisit in later sections), but a notable exception was described by James & James (1992). They performed fairly long (decades and centuries) integrations with a dry primitive equation atmospheric model with very idealized forcing (a Newtonian relaxation), and found a red spectrum of various atmospheric fields, with power increasing as the time period increases from 10 days to 10 years, as illustrated in Fig. 3. James and James call this ‘ultra-low-frequency variability’.

The structure of the variability, as represented by the first EOF of the zonally averaged zonal wind, represents equivalent barotropic (i.e., no tilting in the vertical) fluctuations in the strength, and to a lesser degree the position, of the subtropical jet. The time variations of the first principal component can be modelled fairly well by a first-order Markov, AR (1), process (discussed more in the next section), but the shoulder of the spectrum occurs at a timescale of about 1 year, which is considerably longer than what might be expected to occur as a result of frictional spin-down effects, which would typically produce reddening on timescales of a few weeks or less. In the James and James simulations, the power at very low frequencies appears to come from a transition between a two-jet state, with a subtropical jet distinct from an eddy-driven midlatitude jet, and a single or merged jet state. It seems either state is a near equilibrium state, and that transitions between the two equilibria can occur after rather long intervals in one state. This would fall under the rubric of ‘regime’ behaviour discussed above. Somewhat surprisingly, James and James reported that this persisted even with a seasonal cycle, although the power is somewhat diminished compared to the run with no seasonal cycle and the ambiguity of the result suggests that the work bears repeating.

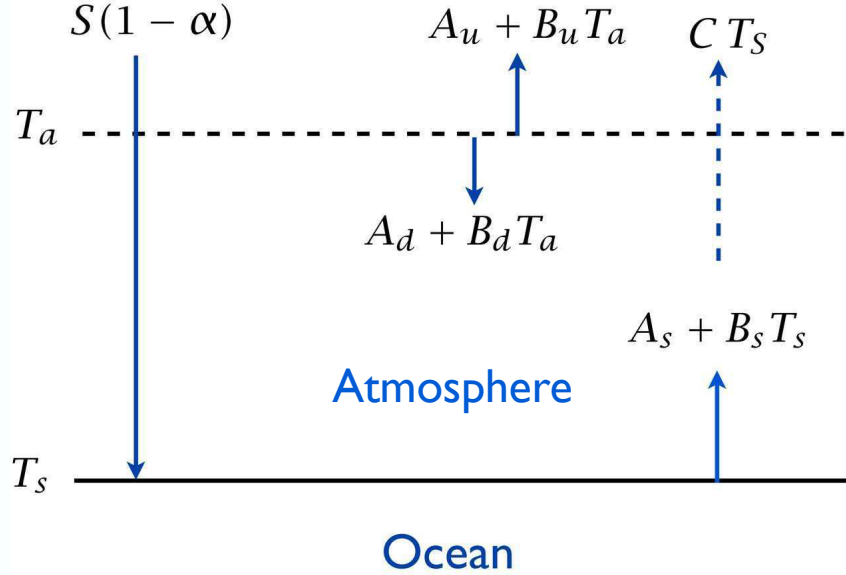


**Figure 3.** (a) Time series of atmospheric angular velocity in an integration of a low resolution (T31) primitive equation model with idealized, time independent forcing. Note the variability at decadal timescales. (b) The power spectra of the first principal component of the zonal mean zonal wind, along with a first-order Markov process ‘red noise’ process (thick smooth line). From James & James (1992)

If we accept, without fully understanding it, the numerical evidence that the atmospheric dynamics can produce, of its own accord, some variability on interannual and even decadal timescales, the question becomes whether such variability is important compared to other mechanisms that can also produce such variability, and we discuss these mechanisms next.

#### 4 The Null Hypothesis: Reddening of the Atmospheric Variability by the Ocean

The most unequivocal mechanism for producing climate variability, if not predictability, on timescales longer than those of atmospheric weather comes by way of the reddening of atmospheric variability by its interaction with the oceanic mixed layer. Climate variability then arises rather in the manner of Brownian motion, as the integrated response to a quasi-random excitation provided by atmospheric weather. The mechanism was first quantitatively described by Hasselmann (1976), although without reference to a specific physical model, and Frankignoul & Hasselmann (1977) subsequently applied the model to the variability of the upper ocean. Our treatment of this draws from the examples of Schopf (1985) and Barsugli & Battisti (1998).



**Figure 4.** A schematic of a simple atmosphere-ocean energy balance model.

#### 4.1 The physical model

The mechanism can be economically illustrated using a one-dimensional climate model with one or two dependent variables, namely the temperature of the atmosphere and the temperature of the oceanic mixed layer, as illustrated in Fig. 4. We will assume that there is no lateral transport of energy, and that the ocean and atmosphere interchange energy by the transfer of sensible and latent heat and by radiation and that very simple linear parameterizations suffice for these.

The physical parameterizations of the model are as follows:

$$\text{Absorption of solar energy at surface: } S(1 - \alpha) \quad (4.1)$$

$$\text{Sensible, latent and radiative flux from surface to atmosphere: } A_s + B_s T_s \quad (4.2)$$

$$\text{Downwards infra-red radiation from atmosphere to surface: } A_d + B_d T_a \quad (4.3)$$

$$\text{Upwards infra-red radiation from atmosphere to space: } A_u + B_u T_a \quad (4.4)$$

$$\text{Upwards infra-red radiation from surface escaping to space: } C T_s \quad (4.5)$$

These are obviously caricatures of the real processes, but they suffice to illustrate the point. The parameters  $A, B$  and  $C$  are constants, as are  $S$  and  $\alpha$ , with typical values given in Table 1; the precise values do not affect the results or conclusions. The sensible and latent energy flux from the surface to the atmosphere might reasonably be taken to have the form  $A_s + B_s(T_s - T_a)$ , so depending on the difference between the atmospheric and surface

Parameter	$B_d$	$B_u$	$B_s$	$C$	$B_2$	$B_3$	$B_4$	$C_a$	$C_s$
Value	11.3	2.83	10.4	0.54	9.86	14.13	3.42	$1.2 \times 10^7$	$3.6 \times 10^8$

**Table 1.** Typical values of radiative parameters for the energy balance model of Fig. 4. The derived quantities are  $B_2 = B_s - C$ ,  $B_3 = B_d + B_u$  and  $B_4 = B_3 - B_2 B_d / B_s = B_u + C B_d / B_s$ . The units of the  $B$  parameters and  $C$  are  $\text{W m}^{-2} \text{K}^{-1}$  and the units of the  $C$  parameters are  $\text{J m}^{-2} \text{K}^{-1}$ .  $C_s$  corresponds to a 60 m deep slab.

temperatures. Here, we have absorbed the  $B_s T_a$  factor in the term  $B_d T_a$ , giving the forms above.

The evolution equations corresponding to the above parameterizations are

$$C_s \frac{\partial T_s}{\partial t} = S(1 - \alpha) + (A_d + B_d T_a) - (A_s + B_s T_s) \quad (4.6a)$$

$$C_a \frac{\partial T_a}{\partial t} = A_s + (B_s - C) T_s - (A_d + B_d T_a) - (A_u + B_u T_a) \quad (4.6b)$$

where  $C_s$  and  $C_a$  are effective heat capacities of the atmosphere and surface (e.g., the mixed layer of the ocean), respectively. Typically,  $C_s \gg C_a$ , and indeed if  $C_s$  represents an oceanic mixed layer of depth 60 m, then  $C_s \approx 30 C_a$ . It is this difference in heat capacity that can lead to the seeming generation of climate variability on long timescales.

The above equations may be written as

$$C_s \frac{\partial T_s}{\partial t} = A_1 + B_d T_a - B_s T_s \quad (4.7a)$$

$$C_a \frac{\partial T_a}{\partial t} = A_2 + B_2 T_s - B_3 T_a \quad (4.7b)$$

where  $A_1 = S(1 - \alpha) + A_d - A_s$ ,  $A_2 = A_s - A_d - A_u$ ,  $B_2 = B_s - C$  and  $B_3 = B_d + B_u$ . We can write (4.7) as anomaly equations by writing  $T_s = \bar{T}_s + T'_s$  and  $T_a = \bar{T}_a + T'_a$ , where  $\bar{T}_s$  and  $\bar{T}_a$  are the steady solutions of (4.7), and if we also introduce a random forcing,  $\sigma \dot{W}$ , on the right-hand side of (4.7b), representing the weather in the atmosphere, the equations of motion become

$$C_s \frac{\partial T'_s}{\partial t} = B_d T'_a - B_s T'_s, \quad C_a \frac{\partial T'_a}{\partial t} = B_2 T'_s - B_3 T'_a + \sigma \dot{W}. \quad (4.8a,b)$$

We will take  $\dot{W}$  to be ‘white noise’, with the same power at all time scales, so that the Fourier transform of  $\dot{W}$  is unity. (Mathematically,  $\dot{W}$  is related to a Wiener process and (4.8) to an Ornstein–Uhlenbeck process.)

## 4.2 Limiting cases and solutions

First consider the case with no coupling between the atmosphere and ocean. With common configurations of atmospheric GCMs in mind, we might implement this in two possible

ways, the simplest being just to hold the sea surface temperature at its mean value given  $\bar{T}_a$ , whence  $T'_s = 0$  in (4.8), so that the equation for  $T_a$  is just

$$C_a \frac{\partial T'_a}{\partial t} = -B_3 T'_a + \sigma \dot{W}. \quad (4.9)$$

Another way of supposing there are no dynamical interactions between atmosphere and ocean is to suppose that the heat capacity of the ocean is zero, so that the ocean is slaved to the atmosphere; the surface temperature is then given by setting the left-hand side of (4.8a) to zero and  $B_d T'_a = B_s T'_s$ . This model might be best considered as representing an atmosphere overlying a land surface that has a small heat capacity. The equation for  $T_a$  is then

$$C_a \frac{\partial T'_a}{\partial t} = -B_4 T'_a + \sigma \dot{W}. \quad (4.10)$$

where  $B_4 = B_3 - B_2 B_d / B_s$ . This is of the same form as (4.9), but evidently the damping is reduced. Physically, if the atmosphere is warm then the ocean is warmed by the atmosphere, so reducing the damping on the atmosphere. (Note that  $B_4 = B_3 - B_2 B_d / B_s = B_u + C B_d / B_s$ , which is positive, so the system is always damped.) Quantitatively, using the values given in table 1,  $B_3 / B_4 \approx 4$ , so that the reduction in damping is potentially significant, at least on long timescales when the SST can respond, but a more quantitative calculation, using for example a GCM, would be needed to see if the effect is truly important.

Note that (4.9) and (4.10) are closely related to first-order auto-regressive, or AR(1), models. A simple forward-in-time finite differencing of (4.9) gives

$$T_{i+1} = \left(1 - \frac{B_3 \Delta t}{C}\right) T_i + \sigma W_i \quad (4.11)$$

where  $\Delta t$  is the timestep and  $W_i$  a white noise process with zero mean. Equation (4.11) is an example of — indeed, it virtually defines — an AR(1) process.

The power spectrum of (4.9) may be calculated by Fourier transforming, whence

$$i C_a \omega \tilde{T}_a = -B_3 \tilde{T}_a + \sigma, \quad (4.12)$$

where  $\tilde{T}$  is the complex Fourier amplitude of the temperature. The power spectral density may be defined by  $P_a(\omega) = |\tilde{T}_a|^2$ , giving

$$P_a^U = \frac{\sigma^2}{B_3^2 + C_a^2 \omega^2}, \quad (4.13)$$

where we add a superscript (‘uncoupled’) to differentiate it from subsequent atmospheric spectra. Equation (4.13) describes a standard ‘red noise’ power spectrum. At longer times the spectrum of the response is white, with higher frequencies being damped, the shoulder occurring approximately at the frequency  $\omega = B_3 / C_a \approx 1/10^6 \text{ s}^{-1} \approx 1/10 \text{ days}^{-1}$ . Of course, this calculation is very approximate, but it indicates that on timescales longer than a week or so — longer than the spin-down timescale  $C_a / B_3$  — the atmospheric spectrum can be expected to be white.

The power spectrum resulting from (4.10) is of the same form as (4.13), being

$$P_a^{C0} = \frac{\sigma^2}{B_4^2 + C_a^2 \omega^2}, \quad (4.14)$$

differing from (4.13) only in the coefficient in the denominator, with the superscript C0 denoting ‘coupled, zero heat capacity ocean’. The coupling has two related effects. (i) Because  $B_4 < B_3$ , the damping is reduced on long time scales, and the atmospheric variance is stronger in the coupled case. At low frequencies the ratio of the two variances is just  $B_3^2/B_4^2 \approx 16$  using the values given in table 1. (ii) The shoulder of the spectrum occurs at a frequency  $\omega = B_4/C_a \approx 1/40 \text{ days}^{-1}$ , which is lower than that of (4.13).

### 4.3 Finite capacity ocean mixed layer

A more realistic situation is that in which the ocean has a finite heat capacity that is much larger than that of the atmosphere; that is, with  $C_a \ll C_s$ . The Fourier transformed equations of motion are then

$$i\omega C_s \tilde{T}_s = B_d \tilde{T}_a - B_s \tilde{T}_s, \quad i\omega C_a \tilde{T}_a = B_2 \tilde{T}_s - B_3 \tilde{T}_a + \sigma. \quad (4.15a,b)$$

The power spectra of the atmosphere and ocean are then found to be

$$P_a^{C1} = \frac{\sigma^2 |\omega_s|^2}{|\omega_a \omega_s - B_2 B_d|^2}, \quad (4.16)$$

and

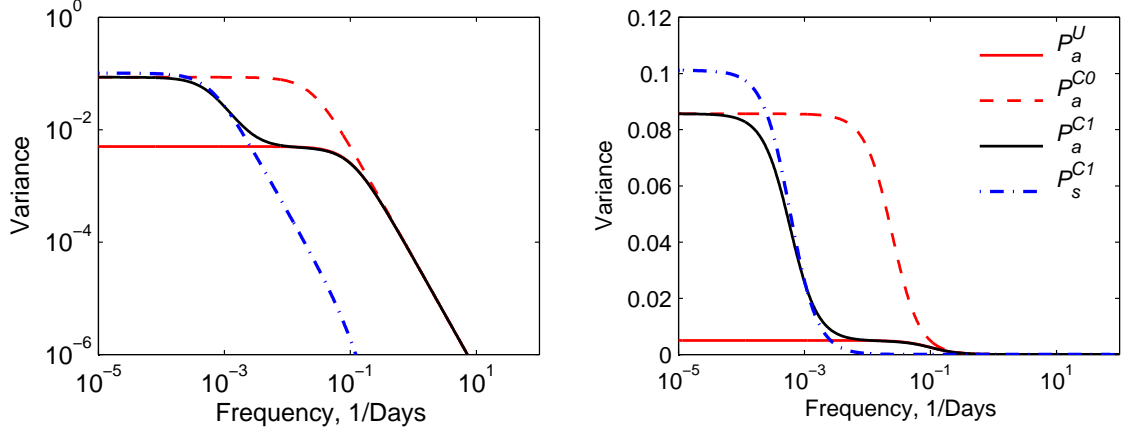
$$P_s^{C1} = \frac{\sigma^2 |B_d|^2}{|\omega_a \omega_s - B_2 B_d|^2} = \frac{|B_d|^2}{|\omega_s|^2} P_a^{C1}. \quad (4.17)$$

where  $\omega_s \equiv i\omega C_s + B_s$  and  $\omega_a \equiv i\omega C_a + B_3$ , and these quantities are plotted in Fig. 5. The atmospheric spectrum  $P_a^{C1}$  (solid black line) lies between the uncoupled spectrum  $P_a^U$  and the spectrum of the atmosphere coupled to a zero heat capacity ocean,  $P_a^{C0}$ , to which it is equal in the low frequency limit. Physically, on timescales sufficiently long enough that the ocean can change its SST, the response of the ocean to the atmosphere reduces damping on the atmosphere (because the ocean warms up if the atmosphere is warm), so increasing variance of the atmosphere. In zero frequency limit, the atmospheric spectrum then displays considerably more power than the power spectra of the process with SST fixed, by the ratio

$$\frac{P_a^{C0}}{P_a^U} = \frac{P_a^{C1}}{P_a^U} = \frac{|B_3|^2}{|B_4|^2} \approx 17. \quad (4.18)$$

As regards the oceanic spectra, we may note that for timescales longer than  $C_a/B_3$  (i.e., longer than 10 days or so) the atmosphere is in a quasi-equilibrium and the left-hand side of (4.8b) may be neglected. The atmospheric spectrum is white and the oceanic spectrum is given by

$$P_s^{C2} = \frac{\sigma^2 B_d^2}{B_4^2 B_s^2 + B_3^2 C_s^2 \omega^2}. \quad (4.19)$$



**Figure 5.** Power spectra of the model ocean and atmosphere obtained from (4.8) in various limits. Both panels show the same spectra, on the left in a log-log plot and on the right in a log-linear plot.  $P_a^U$  corresponds to an atmosphere-only model with  $T'_a = 0$ , and a spectrum given by (4.13).  $P_a^{C0}$  corresponds to a model with a zero-heat capacity surface and a spectrum given by (4.14).  $P_a^{C1}$  corresponds to a model with a finite-heat capacity surface (e.g., an ocean mixed layer) and a spectrum given by (4.16).  $P_s^{C1}$  is the ocean spectrum that accompanies  $P_a^{C1}$ , and is given by (4.17).

This is a red spectrum, with a shoulder at the frequency  $\omega_{sh}$  given by

$$\omega_{sh} = \frac{B_4 B_s}{B_3 C_s} \approx 7 \times 10^{-9} \text{ s}^{-1} \approx 1/1600 \text{ days}^{-1}. \quad (4.20)$$

On timescales longer than  $C_a/B_3$  the above spectrum (not plotted) is almost indistinguishable from that given by (4.17). It is evidently the fact that the heat capacity of the oceanic mixed layer is so much greater than that of the atmosphere that gives the long timescale here, and even given the simplicity of the model and the uncertainty or possible inappropriateness of the parameters, the result suggests that there will be variability in the SST on annual timescales that is forced purely by the ‘noise’ of the weather in the atmosphere.

#### 4.4 Forcing in the ocean component

A variant of the above calculations is to assume that the natural variability arises in the *oceanic* component of the system. If so then it is instructive to put a forcing on the right-hand side of (4.8a). For simplicity we suppose this to be a white noise as before (and because it is a linear equation with additive noise each frequency component is in any case independent) and that there is no noise in the atmosphere. The equations of motion then become,

$$C_s \frac{\partial T'_s}{\partial t} = B_d T'_a - B_s T'_s + \sigma \dot{W}, \quad C_a \frac{\partial T'_a}{\partial t} = B_2 T'_s - B_3 T'_a, \quad (4.21a,b)$$

plainly having the same form as those previously analyzed. There are two particularly relevant limits:

- (i) An ‘uncoupled case’, with atmosphere of fixed temperature and so  $T'_a = 0$ . Such a case roughly corresponds to an ocean model forced by fluxes from a specified, unchanging atmosphere.
- (ii) An atmosphere with zero heat capacity, and so in thermal equilibrium with the ocean beneath with  $B_2 T'_s = B_3 T'_a$ .

In these two cases we obtain, respectively,

$$C_s \frac{\partial T'_s}{\partial t} = -B_s T'_s + \sigma \dot{W}, \quad C_s \frac{\partial T'_s}{\partial t} = -B_5 T'_s + \sigma \dot{W}, \quad (4.22a,b)$$

where  $B_5 = B_s - B_2 B_d / B_3 = B_4 B_2 / B_3$ . The parameter  $B_5$  is, like  $B_4$ , always positive, but it is smaller than  $B_s$  so that the system is damped less than in the fixed atmosphere case. Physically, if the ocean is anomalously warm because of the forcing (or more generally because of its internal dynamics) then the atmosphere warms up, and the damping of the ocean is subsequently less than it otherwise would be.

The corresponding power spectra (with superscript  $U$  denoting ‘uncoupled’) are

$$P_s^U = \frac{\sigma^2}{B_s^2 + C_s^2 \omega^2}, \quad P_s^{C2} = \frac{\sigma^2}{B_5^2 + C_s^2 \omega^2}. \quad (4.23)$$

These are both red noise spectra, with more power in the case with the coupled atmosphere — on long timescales the ratio of their respective variances is given by

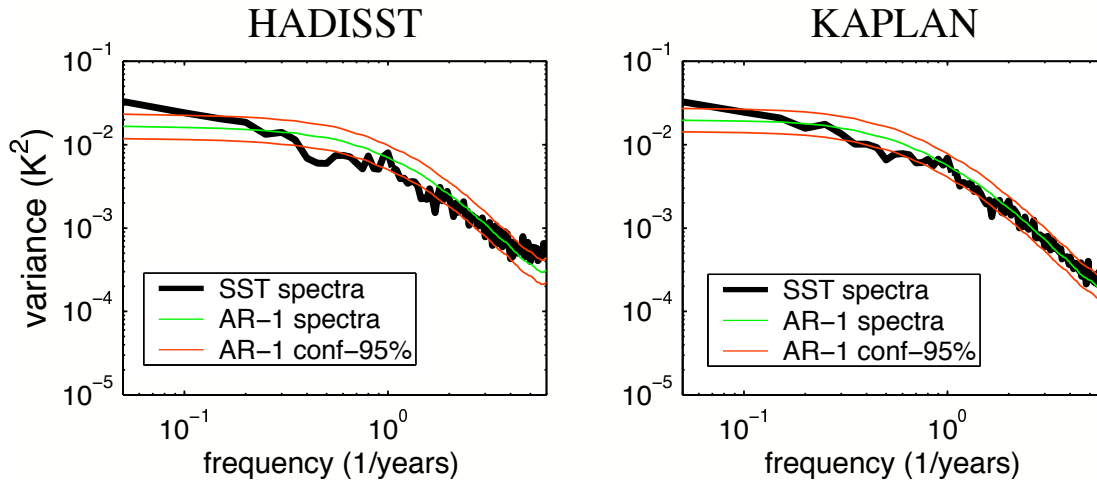
$$\frac{P_s^{C2}}{P_s^U} = \frac{B_s^2}{B_5^2} \approx 20. \quad (4.24)$$

This is potentially significant, but its importance in reality can probably be best assessed by performing experiments with coupled ocean-atmosphere models.

## 4.5 Observations and predictability

If the ultimate cause of variability in the coupled system described above is weather noise, and not the longer term dynamics inherent in the ocean, then there is little predictability in the system on timescales longer than the predictability inherent in weather, even though variables may appear to vary on long timescales; it is simply that the high frequencies have been filtered out leaving only the slow variations. Another way to see this is that one need only note that ‘white noise’ is essentially equivalent to a sequence of random numbers with each value independent of all the others; we plainly cannot predict future values by smoothing the sequence, which is more-or-less what reddening the spectrum by damping is equivalent to.

It is the nature of a null hypothesis that the mechanisms upon which it relies certainly do exist — otherwise it would not be a null hypothesis. But this is not to say that mechanisms are the only ones that operate, or that the predictions of the hypothesis are borne out by



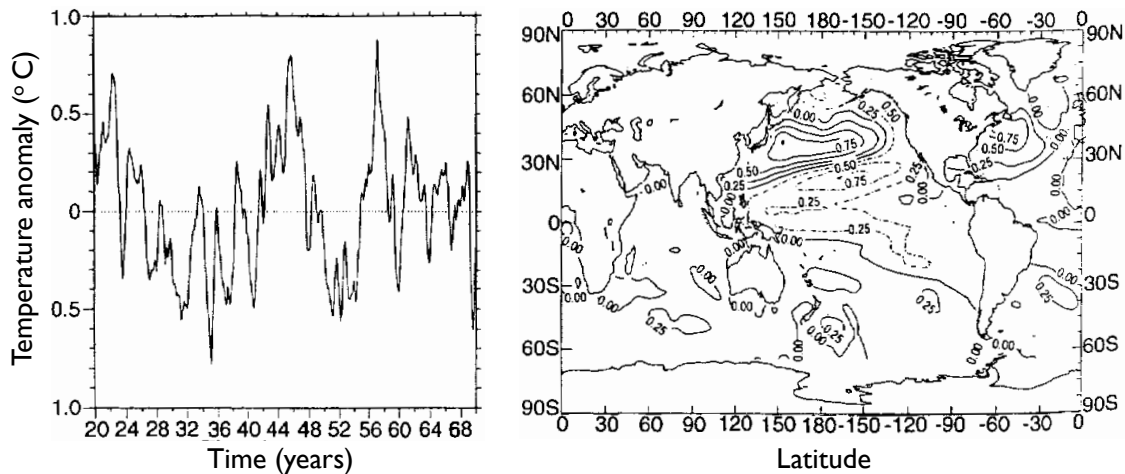
**Figure 6.** Mean spectra of midlatitude SST anomalies of the HADISST and Kaplan SST data sets (thick lines), along with the best fit spectra from an AR(1) process (thin central line) with 95% confidence levels (thin outer lines). Adapted from Dommengeset & Latif (2002).

reality. In the case at hand we may ask, do observations indicate that the integration of the atmospheric variability by the oceanic mixed layer so producing a red spectrum in the oceanic mixed layer is ‘all there is’? The HADISST set (Folland *et al.* 1999) and the Kaplan set Kaplan *et al.* (1998) both extend over a hundred years, and Fig. 6 shows their mean spectra as computed by Dommengeset & Latif (2002). Neither of the spectra conform very well to AR(1) spectra with 95% confidence limits. The deviations do not occur through a single spectral peak indicating some periodic oscillation, but the general shape of the spectrum is different, having a shallower slope than is predicted at seasonal to interannual timescales, but at the same time the spectrum fails to flatten into a white spectrum at long timescales; rather, it continues to redden at decadal timescales, suggesting perhaps that there are dynamical processes that can directly produce variance on these long periods. We discuss what these might be in the next few sections.

## 5 Coupled interactions modes of interaction

Let us now look at the evidence for dynamically coupled modes of interaction between the ocean and atmosphere, but omitting discussion of the single unambiguous example in the climate system, namely El Niño and the Southern Oscillation (ENSO). Our reason for such a seemingly egregious omission is that the ENSO phenomena is well documented, reasonably well understood, and discussed at great length elsewhere.

A striking example of apparent mid- and high-latitude ocean-atmosphere coupling was described by Latif & Barnett (1996). Using a then state-of-the art coupled ocean atmosphere



**Figure 7.** Left: Time series of the SST in a coupled ocean-atmosphere model. The SST is averaged over the region from 150° E to 180° E and 25° N to 35° N, and smoothed with a 9 month running mean. Right: the regression of the SST on the time series shown in the left panel. From Latif & Barnett (1994).

model, in conjunction with observations, they posited a cycle of unstable air-sea interactions involving the North Pacific subtropical gyre and the Aleutian low-pressure system. This cycle appeared to be responsible for about a third or the variability of the climate system and, unlike the null-hypothesis of the previous section implies some predictability within the system. The mechanism they suggested was roughly as follows. Suppose that, for whatever reason, the subtropical gyre is anomalously strong. The western boundary current (the Kuroshio in the Pacific) transports more warm water polewards and this produces a positive SST anomaly in the North Pacific. This heats the atmosphere leading to a weakened Aleutian low, and this in turn leads to air-sea fluxes that reinforce the initial anomaly. At the same time, according to Latif and Barnett, the storm track is weakened and the anomalous Ekman heat transport also contributes to maintaining the positive SST anomaly. Their model produced examples of SST variability varying on time and space scales as illustrated in Fig. 7.

The cycle above is mechanistic and deterministic and, in principle, predictable, at least on the timescale of a single cycle — that is, on the decadal timescale. The memory of the system resides in the ocean, and it is the ocean dynamics that determine the timescales on which anomalies come and go, as in Fig. 7.

#### *Other coupled mechanisms*

Other possible mechanisms exist for the coupling between ocean and atmosphere that might give rise to climate variability, and one such was identified by Cessi (2000). In a numerical study, she noted the presence of a strong thermal front at the boundary between the

subtropical and subpolar gyre in the ocean, associated with the weaker oceanic meridional heat transfer at the gyre boundary. This creates, on average, a sharper atmospheric gradient and a stronger meridional atmospheric heat transport by midlatitude baroclinic eddies, compensating in part for the weaker ocean heat transport, resulting in a well-defined atmospheric storm track. The atmospheric surface winds are, of course, responsible for the location of the inter-gyre boundary, for at zeroth order this is determined by the latitude at which the wind-stress curl changes sign. Changes in the position of the gyre boundary lead to changes in the storm track, and this may feed back onto the location of the gyre boundary and so on, and Cessi found that, because of the delayed adjustment of the gyres to the wind stress, the whole system oscillated with a period of about 18 years.

### **5.1 Air-sea interaction: evidence and models**

Although the coupled mechanisms described above are quite striking they have not been robustly reproduced by other coupled atmosphere ocean models. Such a coupled mechanism requires, broadly speaking, the following sequence of events:

1. The generation of some large-scale pattern of sea-surface temperature anomalies.
2. The SST anomaly pattern must imprint itself on the atmosphere.
3. The atmospheric dynamics must then act in such a way as to reinforce or maintain the SST anomaly pattern.

Although all of these steps seem possible, there are difficulties with each, and so the likelihood of entire chain of events occurring becomes somewhat delicate, as we discuss in the next few subsections.

### **5.2 Generation of SST anomalies**

The first issue is the generation of SST anomaly patterns. Although there are small-scale SST patterns associated with oceanic mesoscale eddies, observations show that, at least on timescales shorter than a year, most of the variance in the SST is forced by the atmosphere, and not vice versa. One of the first demonstrations of this was that of Davis (1976), using 28-year records of SST and sea-level pressure (SLP) anomalies in the North Pacific. Davis tested three hypotheses: (i) that SST anomalies can be predicted, in part, from previous SST anomalies; (ii) that SLP anomalies can be determined, in part, from contemporaneous SST observations; (iii) that (i) and (ii) may be combined to allow a prediction of SLP from present and past SST observations. Davis found that, although (i) and (ii) are true to some degree, indicating that SLP and SST are related, future SLP anomalies *cannot* be predicted from present SST data. He concluded that, at least as regards SLP and SST, the atmosphere drives the ocean and that the ‘back-forcing’ of the ocean on the atmosphere is very weak. That is, in a nutshell, that the observed SST anomalies on seasonal time scales are the

result of, not the cause of, SLP anomalies. Such a study is broadly consistent with a physical model of the SST as being governed by the null hypothesis of section 4 and equations qualitatively similar to (4.8), as well with the model of Frankignoul & Hasselmann (1977), that SST anomalies represent the integral response to short time-scale atmospheric forcing. The higher heat capacity of the oceanic mixed layer, compared to that of the atmosphere, gives SST anomalies a longer time scale than the atmospheric anomalies that cause them. Later studies have also supported this picture — for example Cayan (1992) essentially showed that SST anomalies are positive when there is a heat flux into the ocean, implying an atmospheric driving.

This picture does not, however, preclude the ocean driving the atmosphere on still longer timescales, and it should be said (because it is often forgotten) that the atmosphere must and does respond to the ocean on very long timescales, including the infinitely-long, statistical equilibrium state. This is because the ocean transports heat polewards, especially in the Northern Hemisphere where in the extra-tropics it transfers about one third of that of the atmosphere (Trenberth & Caron 2001). Without such a transfer the pole–equator temperature difference would almost certainly be larger than it in fact is, and the atmosphere and climate would ipso facto be different; indeed Winton (2003) found using a GCM that an Earth-like planet without an ocean became largely ice-covered: a ‘snowball Earth’. In the unlikely event that the atmosphere compensated entirely for the ocean’s lack of transport and kept the pole–equator temperature gradient the same, then again the atmosphere would have responded to the ocean. And since the ocean’s transport is made apparent to the atmosphere primarily through the sea-surface temperature, clearly the sea-surface temperature does influence the atmosphere.

### **5.3 Effect of SST anomalies on the atmosphere**

The question therefore becomes, by what mechanisms and on what timescale do *variations* in the SST affect the atmosphere? The problem is, at least in part, one of signal to noise. On the infinite time scale the ocean does significantly affect the behaviour of the atmosphere; if, for example, we could engineer a large-scale SST anomaly covering the North Pacific for several years we could certainly expect that it would influence the atmosphere, including the sea-level pressure, because on a timescale shorter than the lifetime of the SST anomaly atmosphere would move into a new quasi-equilibrium state consistent with the new thermal boundary conditions. However, on timescales of a year or less the response to an SST anomaly seems likely to be smaller than the natural variability of the atmosphere (as we discuss further below), regardless of whether or not the atmosphere is responsible for producing SST anomalies. We conclude that SST anomalies likely have a weak but pervasive influence on atmospheric dynamics.

The mechanisms and degree of influence of SST anomalies on the atmosphere is a matter on which there is a large literature but less consensus, at least in midlatitudes. [For

a discussion of mechanisms, experiments with general circulation models and observations and reviews, see among others Webster (1981); Lau & Nath (1994); Kushnir & Held (1996); Lunkeit & von Detten (1997); Czaja & Frankignoul (2002); Deser *et al.* (2004); these papers constitute but a tiny sampling of the literature, and Frankignoul (1985) and Kushnir *et al.* (2002) provide more comprehensive reviews.] Most GCMs show a rather *small* response to the imposition of SST anomalies, perhaps somewhat stronger in the summer when the mean flow is small. For example, using a fairly coarse resolution atmospheric GCM Lau & Nath (1994) and Kushnir & Held (1996) found that their GCM exhibited quite weak sensitivity to midlatitude SST anomalies, with results that are generally consistent with linear quasi-geostrophic theory (e.g., Hoskins & Karoly 1981; Held 1983; Vallis 2006) with a shallow heating anomaly. In this case the heating anomaly at the surface will be balanced locally by zonal horizontal advection, producing a baroclinic response in the vicinity of and slightly downstream of the SST anomaly, with a warm ocean leading to a warm atmosphere and the response decreasing with height. In the far-field quasi-geostrophic theory suggests that the flow should be dominated by the homogeneous solution, and equivalent barotropic wave trains propagating around the globe may be generated.

If a sea-surface temperature anomaly is able to generate a *deep* heating source in the atmosphere then the response may be larger. In this case theory suggests that the heating source may be balanced by meridional advection, with a positive heating anomaly producing equatorwards advection, and a warm low downstream of the heating region. The far-field response is the same as that for shallow heating, with equivalent barotropic wavetrains. Although the fields produced by GCMs seem consistent with this general picture the response is not strong for realistic values and scales of SST anomalies. The equivalent barotropic downstream response is typically not robustly found in GCMs, and generally the response is more typical of that of a shallow anomaly. It may be that GCMs response to SST anomalies is unrealistically weak because of inadequacies in their boundary-layer formulation (perhaps confining the response too much to the near surface region) but such a remark is speculative.

Even though the response of GCMs to SST anomalies may be weak, some GCMs (Rodwell *et al.* 1999; Robertson *et al.* 2000) do seem to show that the North Atlantic Oscillation (NAO) is influenced by sea-surface temperature when sea-surface temperature is imposed, and there is also some observational support for this (e.g., Czaja & Frankignoul 2002). Such results should probably not be regarded as contradicting or being wholly inconsistent with the GCM studies because the response to SST anomalies remains rather weak, but the pervasive nature of the anomalies does bring about systematic atmospheric response over the course of a season. Nonlinear effects likely play a role in the response also, as seems have been found by Lunkeit & von Detten (1997)

Such a response does suggest a degree of predictability to the NAO, presuming of course that SST can be predicted. However, it is this last phrase that indicates the weakness in the argument: as discussed above, on the seasonal timescale SSTs seem likely to be a

response to the atmosphere, and not a driver of the atmosphere. Thus, the SSTs themselves are likely not predictable over the course of a season, and so any skill in an atmospheric forecast dependent on such prediction is lost; Bretherton & Battisti (2000) discuss this interpretation further.

#### **5.4 Effect of the atmosphere in maintaining SST anomalies**

On the seasonal to yearly timescale SST anomalies are, as mentioned, the consequence of atmospheric anomalies. For example, the North Atlantic Oscillation, whose dynamics are primarily tropospheric, will produce anomalous heat and momentum fluxes, both of which may generate SST anomalies, on seasonal timescales. Such anomalies may be generated both by anomalous winds — and so anomalous Ekman transports and mixed layers — and by anomalous sensible and latent heat fluxes. However, given the evident weakness on the seasonal timescale of the atmospheric response to SST anomalies, the subsequent dynamical feedback of the atmosphere to the ocean, maintaining and reinforcing the initial SST anomaly, seems unlikely except possibly in one regard: the reduced local thermal damping provided by the atmosphere if the ocean is anomalously warm. This effect is captured by the null hypothesis of section 4: in section 4.4 we noted that having an atmosphere, even with zero heat capacity, coupled to an ocean provides a reduced damping compared to a ocean underneath a constant-temperature atmosphere, and, as in (4.24), more variance at low frequencies. Even if this effect is not large in the context of maintaining seasonal SST anomalies, it may be an important effect in producing and maintaining decadal and longer climate fluctuations, as we discuss in the next section.

### **6 Climate Variability with an Oceanic Origin**

In this section we discuss the degree to which climate variability on decadal to centennial timescales might have a predominantly oceanic origin, possibly modified by the atmosphere. Most of the results come from coupled GCMs and so that is where our focus lies.

#### **6.1 Some results from coupled GCMs**

Coupled ocean-atmosphere models, when integrated for long periods of time after having reached equilibrium, often do display oscillations on timescales of decades to centuries. An early example showing this in a full GCM is the integration described by Delworth *et al.* (1993), in which a fully coupled, then state-of-the-art, ocean-atmosphere model was found to have oscillations in its meridional overturning circulation (MOC) with a period of about 50 years. In these simulations, the mean overturning circulation had an amplitude of about 18 Sv and the variability was about  $\pm 1$  Sv. These oscillations led to variations in SST of about  $0.5^\circ\text{C}$ , mainly in the Atlantic sub-polar gyre, and similar differences in surface air temperature. The surface air temperature anomalies are similarly concentrated at high

latitudes, although they spread over a larger longitudinal range on either side of the North Atlantic. The patterns of both SST variations and the surface air temperature variations seemed qualitatively consistent with observations, or at least what observations there are of the difference in temperature between warm decades (e.g., the 1950s) and cold decades (e.g., the 1970s). Making a connection with observations is especially difficult given that in the real world natural variability is conflated with anthropogenic change (global warming), and even the general cooling trend from the 1940s to the 1970s may have had an anthropogenic origin in the emission of aerosols; thus, any agreement or disagreement between model and observations regarding decadal variability must be regarded with caution.

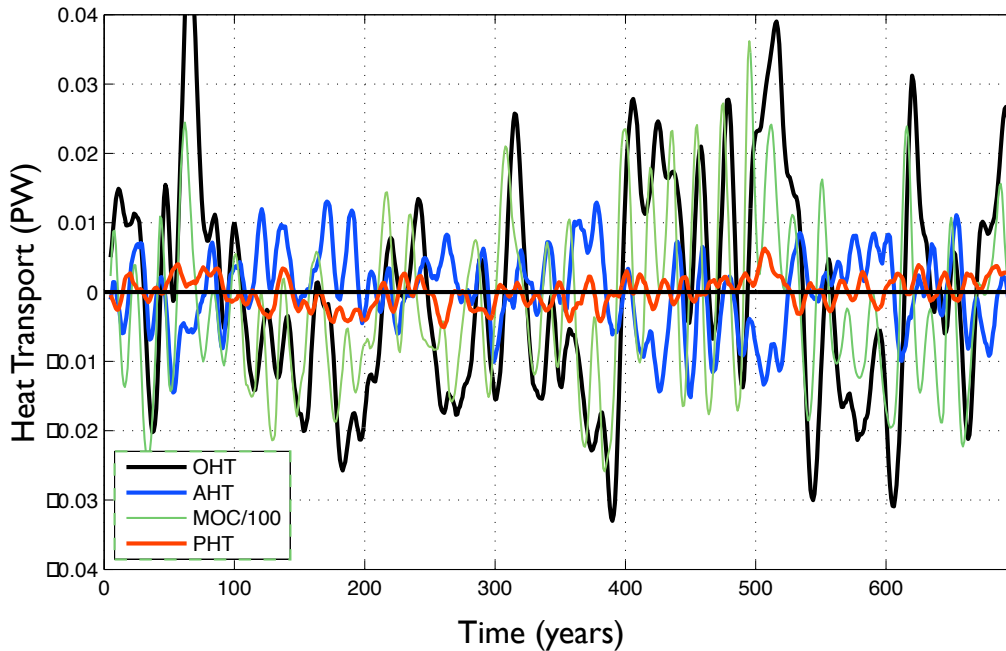
Other coupled climate models have displayed a qualitatively similar variability. For example, using the Hadley Centre coupled ocean-atmosphere Climate Model (HadCM3) Shaffrey & Sutton (2006) also found decadal climate variability associated with, and possibly caused by, fluctuations in the MOC of the Atlantic. Shaffrey and Sutton noted that the meridional heat transport in the atmosphere was, on decadal timescales and in mid- and high-latitudes, significantly anti-correlated with the fluctuations in transport in the ocean, suggesting that the atmosphere may be compensating for the long-term changes in the ocean heat transport, a phenomenon also found by Magnusdottir & Saravanan (1999) and an issue we revisit in the next subsection.

## 6.2 An idealized coupled model

The difficulty of understanding decadal-scale oscillations in the ocean-atmosphere system is essentially as follows. Fully coupled GCMs, which one might expect to provide a reasonably comprehensive and perhaps realistic climate simulation, are very complex and are difficult to understand in themselves. If anything, they have become experimental tools, and rather cumbersome experimental tools at that, rather than theoretical tools that provide some conceptual understanding. The main other line of investigation has been to use quite simple models, including ocean-only models, linear analyses, and (with somewhat more complexity) ocean models coupled to energy-balance atmospheric models. However, these are all potentially unrealistic in their response. This duality of approaches is not ideal, and ideally one would like there to be a continuous spectrum of models from the very simple to the very complex; indeed, the notion of a hierarchy of models has become something of a cliché, often talked about but rarely implemented. However, ideas and sayings often become clichés because they are *good* ideas and sayings; a bad idea is usually ultimately forgotten, at least in science.

In an attempt to fill in part of the spectrum, that part adjacent to but simpler than comprehensive GCMs, Farneti & Vallis (2008a,b) constructed a coupled ocean-atmosphere-ice model with two general simplifications.

- (i) The geometry was much simplified, having a single, two-hemisphere sector ocean



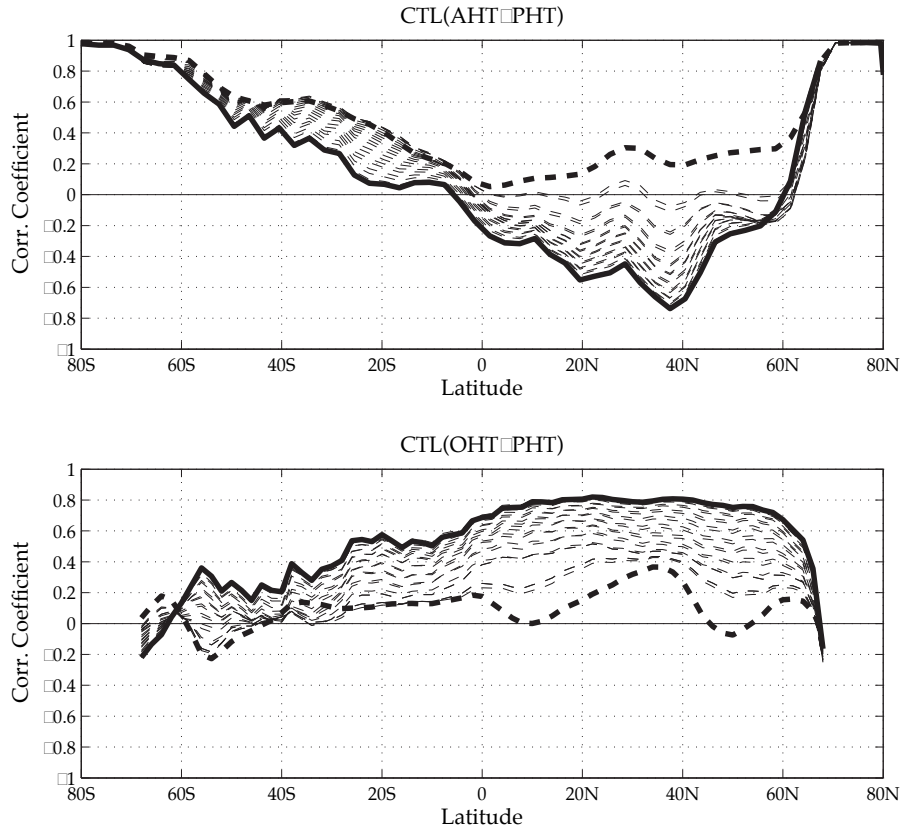
**Figure 8.** The meridional energy transport of the Farneti–Vallis model, averaged from  $20^{\circ}$  N to  $70^{\circ}$  N, and low-pass filtered to show only decadal variability. OHT is ocean heat transport, AHT is atmospheric transport (sensible plus latent) and PHT is the planetary (atmospheric plus oceanic) heat transport. The fainter green line is the strength of the MOC in Sverdrups/100.

with a periodic channel near the southern edge, crudely representing the Antarctic Circumpolar Channel.

- (ii) The ‘physics’ of the atmospheric model was much simplified, having no interactive clouds and a simple, semi-grey radiation scheme. The vertical resolution (seven levels) is also less than is typically used in comprehensive GCMs.

Both the ocean and atmospheric model solve the three-dimensional primitive equations, and include salinity and moisture, with the only external forcing being the incoming solar radiation at the top of the atmosphere. The advantages over a comprehensive GCM lie in both the computational economy (so allowing multiple numerical experiments in clean settings to be performed), and in the relative ease of interpretation of the results. Nevertheless, such a model remains very complex and difficult to understand.

Consistent with results from a number of other more comprehensive GCMs, Farneti and Vallis found decadal-to-multidecadal oscillations involving the overturning circulation; some illustrative results are given in Fig. 8. They find are that, first, decadal oscillations are primarily oceanic, in that the timescales, memory and predictability arise from the ocean



**Figure 9.** The correlation between atmospheric heat transport and planetary (atmospheric plus oceanic) heat transport (top panel) and the correlation between oceanic heat transport and planetary heat transport (bottom panel), calculated using a multi-century integration of the idealized coupled model of Farneti and Vallis. The solid line in each panel shows the correlation after the time series has been filtered to allow timescales of 20 years and longer, and the dashed lines the correlation at successively shorter timescales. The end member of the dashed lines allows timescales of one year and longer.

dynamics, and second that this oceanic variability is indeed imprinted on the atmosphere. Two lines of evidence support the oceanic origin of the variability:

- (i) The meridional heat flux of the ocean-atmosphere system is positively correlated with that of the ocean, and negatively correlated with that of the atmosphere, on decadal timescales. This does suggest that the ocean is the driver and the pacemaker for the variability, and the atmosphere is compensating. On annual and shorted timescales the heat flux is correlated with that of the atmosphere, reflecting the variability ultimately caused by baroclinic eddies (including NAO-type phenomena) in the atmosphere. Of course, these conclusions should be tempered with the usual caveats that correlation does not imply causation. Nevertheless, the interpretation

that the ocean is the pacemaker of variability timescales longer than a year or two, but the atmosphere itself is primarily responsible for interannual and shorter term variability, is the most straightforward one to draw in the absence of other evidence.

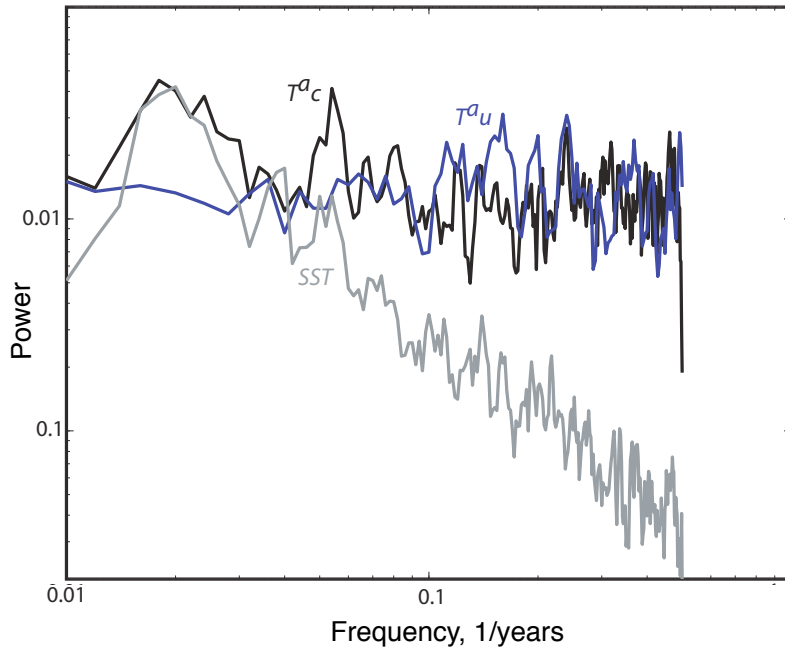
- (ii) Atmosphere-only models do not display significant variability at time scales longer than the interannual (Fig. 10). Thus, any process similar to that found in James & James (1992) either does not operate in this model or is insignificant compared to the variability arising from the ocean.

However, the atmosphere does play a role in the oscillations, in two regards.

- (i) A catalytic aspect, which seems the most important, is that an active atmosphere reduces the damping felt by the ocean, as described in section 4.4. That is, an anomalously warm ocean leads to an anomalously warm atmosphere, which damps the ocean less than would a fixed atmosphere and allows it to oscillate. The effect is not required in all parameter regimes; thus, with a high diapycnal diffusivity the ocean spontaneously oscillates even with a constant atmosphere, whereas at very low diapycnal diffusivity the oscillations are smaller, even with an interactive atmosphere.
- (ii) An active aspect is that stochastic forcing from the atmosphere seems, in part, to enable the oceanic oscillations. Thus, given an ocean-only simulation that evolves into a steady state, the addition of daily fluxes taken from a coupled-simulation (but with the long-time variability removed so that the fluxes do represent only ‘weather noise’) gives rise again to decadal scale variability in the ocean’s MOC. This is suggestive of a damped oscillator that may be excited by noise, much as in the box model of Griffies & Tziperman (1995). However, the oceanic variability is relatively weak compared to that in a fully coupled model.

### **6.3 Robustness and mechanisms**

The mechanisms of decadal oceanic oscillations remain somewhat obscure, except in a broad sense. Ultimately, oscillations seem to arise because of a delay in the response of the MOC and the associated transport of heat to the meridional temperature gradient (Huck *et al.* 2001). If one posits that the heat transport of the MOC will fully respond only after some years to a change in temperature gradient, and that the change in MOC will then transport sufficient heat meridionally to compensate, and potentially over compensate, for the changes in temperature gradient, then it is not difficult to construct, albeit on a somewhat ad hoc basis, ‘delayed-oscillator’ type equations that produce oscillations on the decadal timescale. Nevertheless, the detailed manifestation of the oscillation seems to differ somewhat from model to model, with varying roles for salinity playing an important role in some simulations and less so in others. For example, Delworth *et al* described their



**Figure 10.** Power spectra of atmospheric temperature at 450 mb in a fully coupled integration (black) and an integration in which an atmosphere is coupled to a mixed-layer ocean (blue), using the Farneti–Vallis model. The grey curve shows the spectrum of the SST in subpolar region in the coupled model. The atmosphere-only simulation shows much less variability on decadal timescales than does the coupled model.

mechanism by supposing first there is a weak MOC; the reduced heat transport associated with this leads, some years later, to a cold North Atlantic extending from the surface to about 1 km. This appears to generate an anomalous quasi-horizontal cyclonic circulation, and this results in an enhanced salinity transport into the region. The enhanced density that results leads to increased sinking and a strengthening of the circulation that, by transporting heat into the sinking region, develops a warm pool and reduces the density anomaly. This decreases the salt advection into the region, producing a warm, fresh and therefore light pool. Sinking is inhibited, and once more the MOC weakens. Selten *et al.* (1999) also described a mechanism in which salinity anomalies were important, using the ECBILT model. In their model the forcing by the atmosphere, in particular the NAO pattern, was responsible for producing an initial SST anomaly, following which ocean dynamics were responsible for producing a decadal oscillation. Their mechanism seems to involve shallower dynamics than those in the GFDL model but, like the GFDL result, the subsequent response of the atmosphere to SST anomalies does not seem essential in maintaining the oscillation and the mechanism may be thought of as essentially oceanic. The role of salinity in these oscillations suggests a connection to the so-called ‘Great Salinity Anomalies’ (GSAs Dickson *et al.* 1988; Walsh & Chapman 1990), which, through their affect on the meridional overturning circulation may affect the path of the Gulf Stream and so have an influence on

climate on decadal timescales (Zhang & Vallis 2006), but the mechanisms of the genesis of observed GSA events remains obscure.

In the more idealized integrations of Farneti & Vallis (2008a) the dynamical mechanism is also primarily oceanic, except that by having an interactive atmosphere the reduced damping allows the ocean to oscillate more readily than it might if the forcing on the ocean were held fixed. Indeed, Huck & Vallis (2001) suggest that the oscillations might be the manifestation of a linear instability of the three dimensional overturning circulation, and depending on the boundary conditions applied to the ocean model, they find that the structure of the linearly unstable modes resembles that of the oscillation found in the forward integration of a fully nonlinear model.

These oscillations do seem fairly robust to model formulation and parameters, having appeared across a variety of models from the comprehensive to the very idealized (Winton 1996; Greatbatch & Zhang 1995; Huck *et al.* 2001), but the details vary considerably from model to model. The presence of realistic bottom topography also seems to have a damping effect (Winton 1997), and as noted previously the oscillations are also sensitive to the value of the diapycnal diffusivity and to a thermally interactive atmosphere. As a consequence, no clearly accepted single explanation has emerged that is akin to baroclinic instability as the cause of weather in the atmosphere. Can we expect one to do so? I am optimistic; we are now rather at an analogous the stage development as occurred in the 1950s with regard to the understanding of rotating annuli, mapping out regimes of flow. Increasing computer power will make the mapping easier, although the theoretical challenge of understanding such a complex system remains daunting.

## 7 Discussion

Natural climate variability remains a rather poorly understood phenomena, compared for example with our understanding of weather variability for which the underlying mechanism — namely baroclinic instability and its nonlinear consequences — are relatively well understood. In what follows I'll set out some conclusions about the matter, trying to be clear about what is generally accepted and what is not.

First of all, over the past few decades of research it has become clear that on timescales of a year or less the atmosphere is the dominant driver of variability. The dominant extra-tropical modes of variability — the North Atlantic Oscillation and related phenomena such as the variability in the zonal index and the 'annular modes' (both in the Northern Hemisphere and Southern Hemisphere) — are uncontroversially primarily atmospheric in origin. They are related to the variability of the major storm tracks and to the stationary wave pattern, and although the mechanisms of such variability are still not fully understood [see Wallace (2000); Feldstein (2003) and Vallis & Gerber (2008) and references therein for discussions], and may be influenced by both the ocean and the stratosphere, the troposphere itself is almost certainly the primary origin of the variability. There are still nevertheless a

number of weaknesses in our understanding, notably in the precise mechanisms that seem to give rise to variability on timescales rather longer than that of a baroclinic lifecycle and the possible influences of the stratosphere and the ocean, as well as the mechanisms that give rise to an oscillation (or at least a variability) that is in some ways more prominent in the North Atlantic than the North Pacific.

It seems likely that it is this atmospheric variability that gives rise to variability in the sea-surface temperature on timescales of up to a year, as we discussed in section 5.2, and if so it is likely that there is very little predictability in this process. What to the meteorologist is a signal, namely the atmospheric variability on daily to monthly timescales, becomes to the oceanographer a source of noise. The oceanic mixed layer integrates this noise, generating sea-surface temperature anomalies that might be quite long lived and of large scale. This integration is perhaps more usefully described as suppression of fast-time variability, so reddening the spectrum of the near surface temperature but providing no genuinely new predictable source of information on longer timescales aside from that due to persistence. That there is no additional information coming in on long times is apparent when one realizes that white noise is essentially the same as a sequence of random numbers, each uncorrelated with its neighbour; subsequently applying a low-pass filter to give red noise plainly cannot induce predictability.

Whether significant variability on timescales of much longer than a year can arise purely from an atmosphere, whether with fixed SSTs, an oceanic mixed layer or still more idealized forcing seems, on balance, unlikely. In order for this to happen the atmospheric dynamics would have to produce regimes of flow that persisted longer than any natural timescales in the system. Of course, such behaviour *is* known to happen in some dynamical systems, and it does seem to happen in the simulations of James & James (1992), so the situation is not wholly settled. Jet merger and separation is one process that can demonstrably produce regimes with long timescales (Vallis & Maltrud 1993; Panetta 1993), especially when dynamics lie on the border between one and two eastward jets, but if this were the case it would seem that the seasonal cycle would still prevent multi-annual timescales from emerging. Evidently, further investigation is called for.

The ocean certainly has dynamical timescales of decades, centuries and more, and ocean models seem to be able to oscillate, fairly robustly, on decadal and longer timescales. Whether the ocean is responsible, in whole or in part, for climate variability on these timescales depends on two things: (i) the existence of variability in the sea surface temperature on these timescales; (ii) this variability affecting the atmosphere. Regarding the latter, results from most GCM experiments seem to show a weak but persistent effect; the subsequent feedback from the atmosphere to the ocean seems weak, effective mainly in reducing the damping felt by the ocean. (In this sense the oscillations may be thought as weakly coupled, with the timescales and memory residing in the ocean.) Still, the results are not wholly uniform from GCM to GCM and it may still be possible that GCMs underestimate the effects of SST anomalies if their boundary-layer schemes are inadequate.

In any case, it is on the decadal and longer timescale that the oceanic effects are more likely to be felt, because now the oceanic signal may be felt over the atmospheric noise. On these timescales the main robust atmospheric response may simply be a warming at latitudes of anomalously warm ocean, so affecting the storm tracks and producing weak but potentially predictable behaviour in such indices as the NAO or Southern Annular Mode.

Finally, let us make a comment on the relation of natural variability on the decadal–centennial timescale to anthropogenic global warming. Two lines of evidence argue *against* natural variability being the cause of the global warming seen in the last century. The first is that the variability produced by most models of the climate system is rather smaller than the warming observed. Over a given decade or conceivably over a few decades the natural variability may match or even outweigh the trend, and so we should not be surprised if in the future sometime a cool decade occurs. But over the course of a half century or longer, the natural variability, assuming that its variance will be in the future what it has been in the past, is too small to overcome predicted global warming. The second is that there is no evidence that there has been a flux of heat from the ocean to the atmosphere (Levitus *et al.* 2001; Barnett *et al.* 2001); if anything, there has been a flux of heat *into* the ocean, implying that the ocean is warming *because* of a radiative imbalance and not that the ocean is producing the warming by fluxing heat into the atmosphere.

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