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# Mechanisms of climate variability from years to decades

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This chapter discusses and reviews of some of the mechanisms that may be responsible for climate variability on yearly to decadal time scales. The discussion is organised 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 we may not always 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.*

## 1.1 Preamble

This is an essay on the mechanisms of natural climate variability on time scales 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 I'll discuss rather generally the nature of stochastic and deterministic processes and their role in weather and climate, and in the following sections I will focus more explicitly on climate variability on time scales from years to decades, emphasising 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 analysed 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; both then 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 time scale, in that we don't care about the weather that is taking place on time scales of weeks. Now, weather can be explicitly modelled far better than it can be parameterised by a stochastic process, but the details of the weather are generally irrelevant to decadal-scale climate variability – only its statistics likely 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 time scales. Similarly, for those whose interest is variability on time scales of millions of years then even centennial variability might best be treated as noise.

In the rest of this chapter I focus on climate variability on time scales of years to decades, with the exclusion of the El Niño phenomenon for that deserves an article unto itself. The discussion is organised around mechanisms; thus, following a brief look at some observations, I summarise 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.

## 1.2 Observations and classes of mechanisms

### *1.2.1 A few observations*

Climate variability exists on time scales of seasons to millennia, but in this article the emphasis will be on time scales of years to centuries, or the decadal time scale. A rough indication of such variability is shown in Fig. 1.1, 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 seems evident if one restricts attention to a particular region of the globe, as in the central England temperatures shown in Fig. 1.2, 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, a Monte Carlo simulation is performed by taking the time series of successive winter temperatures

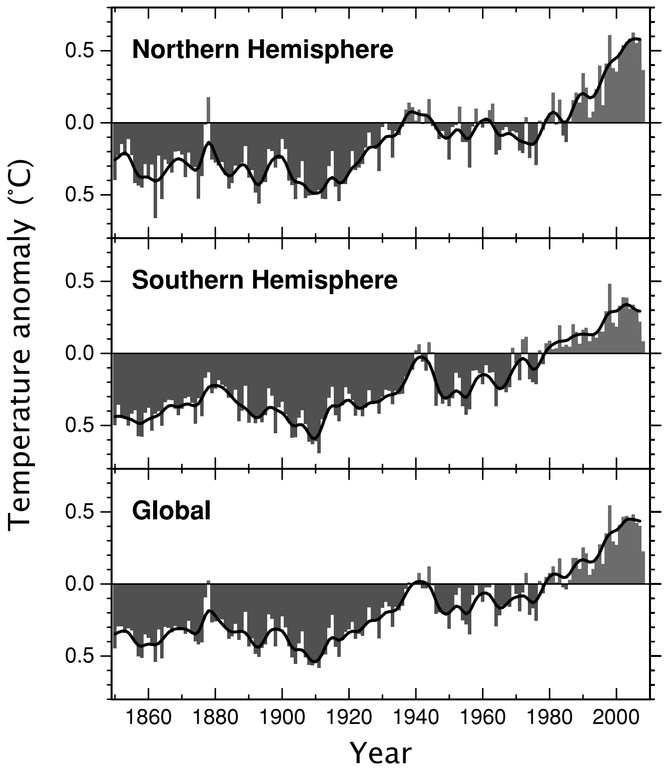


Figure 1.1 The global average surface temperature (anomaly from 1961–1990 average) from the HadCRUT3 data, from Brohan *et al.* (2006). In addition to the general warming trend some decadal variability seems apparent, presumably due to natural variability in the system.

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 realisation 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. 1.2). 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 time series are not qualitatively different from the original (lower right panel of Fig. 1.2). In this figure we show the power spectra of the original time series (thick solid line), a sampling of the spectra computed from shuffled time series (thin solid lines), the mean of the these (thick dashed line) and the mean plus or minus one standard deviation (thin dashed 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

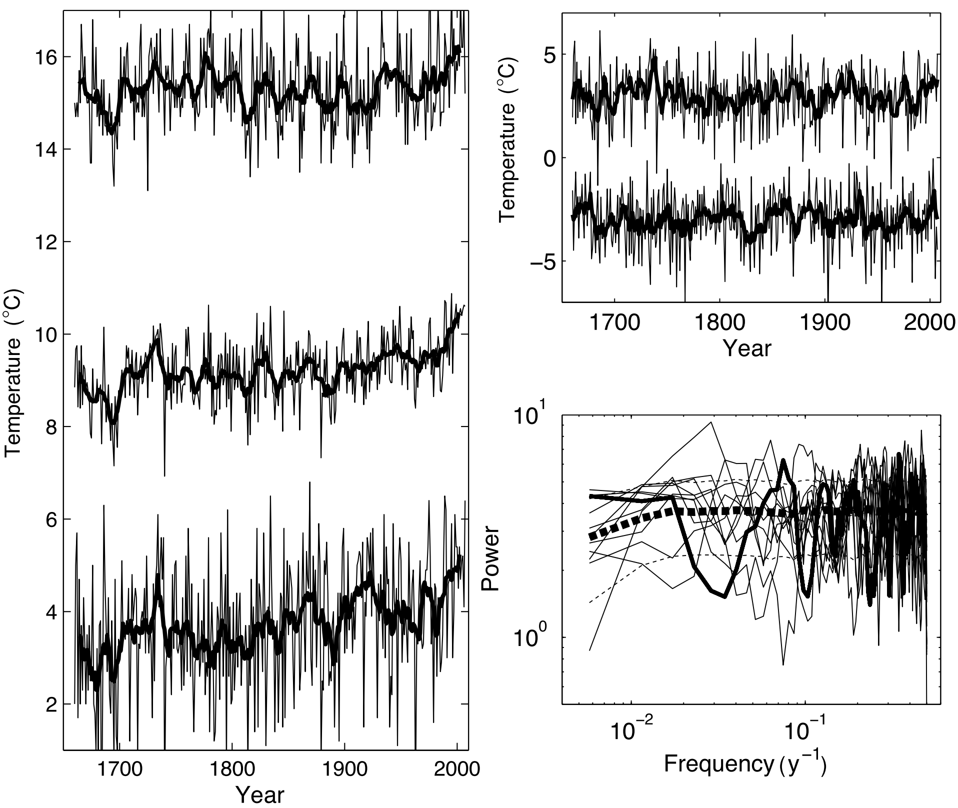


Figure 1.2 Left: central England temperature from 1650 to 2007, using the HadCET data (Parker *et al.* 1992). The three sets of curves show, from the top, summer (JJA), annual average, and winter (DJF) temperatures. The thin curve shows the average over the season or year, and the thicker curve shows the ten-year running mean. Right top: detrended winter temperature anomalies; the thin line is the seasonal temperature, and the thicker line is the temperature after a six-year running mean. The 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. The thick solid line is the power spectrum of the actual winter (DJF) temperatures, the thick dashed line is the mean power spectrum of 1000 shuffled series, with plus and minus one standard deviation marked by thin dashed lines to either side. The multiple thin lines are the power spectra of 10 of the shuffled time series.

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; Tourre *et al.* 1999). Similarly, Biondi *et al.* (2001) conclude the climate in and around the Pacific region has undergone real 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.

### *1.2.2 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 time scales 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 time scales 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 essentially 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 EI Niño/Southern Oscillation (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.

In the next few sections I will discuss these mechanisms, excluding the last item which is well documented elsewhere. The El Niño phenomenon is not discussed 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.

## 1.3 Atmospheric variability

In the extra-tropical atmosphere the primary mechanism of variability on large scales is baroclinic instability, the basic lifecycle 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,

$$\sigma \equiv \frac{0.3\Lambda H}{L_d} = \frac{0.3U}{L_d} \tag{1.3.1}$$

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, $U = 10\ \mathrm{m\,s^{-1}}$ and $L_d = 1000$ km we obtain $\sigma \approx 1/4\ \mathrm{d}^{-1}$. (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. Here $\beta$ is the rate of change of Coriolis parameter with latitude.) 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 time scale, 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 $\mathrm{cm\,s^{-1}}$ 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 that the atmospheric ones.)

Of course baroclinic eddies are nonlinear, 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 intraseasonal variability. The zonal index will also vary on intraseasonal time scales 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 time scales 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 time scales longer than a few weeks, or months at most. Nevertheless, we can also expect some interannual variability essentially as a residual of the intraseasonal variability (e.g. Feldstein 2000), but such interannual variability will be weak and unpredictable.

However, for the atmosphere to produce variability on time scales significantly longer than a few weeks – for example with some peak in the power spectrum at interannual time scales – would likely require there to be some kind of regime behaviour, in which the gross atmospheric behaviour changes on time scales 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 time scale not closely associated with baroclinic waves. Even more strikingly, zonal jets, once formed, can have an extremely long time scale (Panetta 1993; Vallis & Maltrud 1993). 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 intraseasonal variability or, perhaps, mechanisms associated with the slow variability of persistent jets or the quasi-biennial oscillation, whose climate relevance is unclear.
(ii) Suppose that the atmosphere *were* able to produce regime-like behaviour when steadily forced. The difference between any two realistic regimes would still 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 general circulation models (GCMs) do not produce significant variability on interannual time scales (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 idealised 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. 1.3. James and James call this 'ultra-low-frequency variability'.

The structure of the variability, as represented by the first empirical orthogonal function (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 time scale of about 1 year, which is considerably longer than what might be expected to occur as a result of frictional spin-down effects; these would typically produce reddening on time scales of a few weeks or less. In the James–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

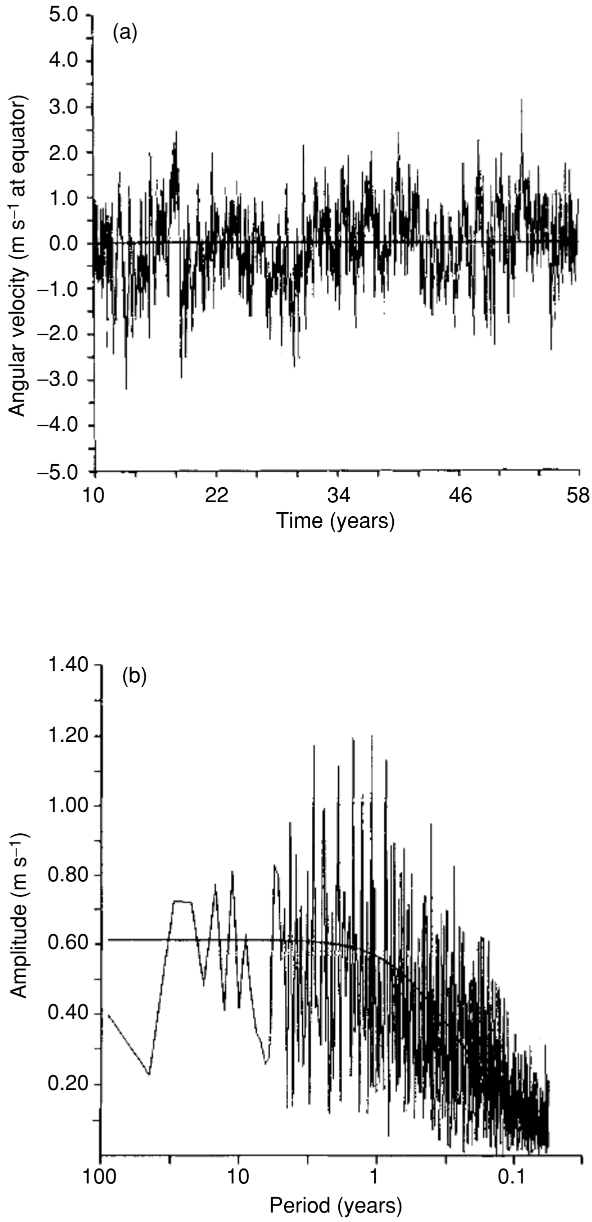


Figure 1.3 (*a*) Time series of atmospheric angular velocity in an integration of a low resolution (T31) primitive equation model with idealised, time-independent forcing. Note the variability at decadal time scales. (*b*) The power spectra of the first principal component of the zonal mean zonal wind, along with a first-order Markov ('red noise') process (thick smooth line). From James & James (1992).

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 & 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.

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 time scales, the question becomes whether such variability is important compared to other mechanisms that can also produce such variability, and we discuss these mechanisms next.

## 1.4 The null hypothesis: reddening of the atmospheric variability by the ocean

The most unequivocal mechanism for producing climate variability, if not predictability, on time scales 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 follows Schopf (1985) and, especially, Barsugli & Battisti (1998).

### *1.4.1 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. 1.4. 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 parameterisations suffice for these.

The physical parameterisations of the model are as follows:

absorption of solar energy at surface: $S(1-\alpha)$ (1.4.1)

sensible, latent and radiative flux from surface to atmosphere: $A_s + B_s T_s$ (1.4.2)

downwards infrared radiation from atmosphere to surface: $A_d + B_d T_a$ (1.4.3)

upwards infrared radiation from atmosphere to space: $A_u + B_u T_a$ (1.4.4)

upwards infrared radiation from surface escaping to space: $CT_s$. (1.4.5)