AN OVERVIEW OF CENTURY TIME-SCALE VARIABILITY IN THE CLIMATE SYSTEM: OBSERVATIONS AND MODELS

THOMAS F. STOCKER
Climate and Environmental Physics
Bern, Switzerland

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Abstract

Estimates of the development of the Earth's climate subject to anthropogenic forcing depend critically on our knowledge of natural climate variability on time scales of decades to centuries. Time scales extracted from high-resolution proxy records and observations indicate that the spectrum of climate variability exhibits significant power in the range of decades to centuries superimposed on a red-noise continuum. The classical view of climate variability is based on the concept that observed fluctuations have their origin in periodic forcings on the same time scale, i.e. that the climate system behaves like a linear system that is externally forced. The present sensitivity of the climate system, however, would require strong positive feedback mechanisms to translate the weak forcing signals (e.g. variability of solar irradiation) into detectable fluctuations in observed and proxy variables. Instead, it is proposed that these fluctuations are linked to interactions within and between the different climate system components. An overview of recent modeling results and the discussion of mechanisms involved show that such interactions internal to the climate system cannot only exhibit the correct time scales but also easily account for the amplitudes observed.

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1 Introduction

Variability is a fundamental property of our climate system. Two decades ago Mitchell (1976) proposed a theoretical framework of climate variability based on a schematic spectrum of climate variations spanning time scales from one hour to some billions of years. He distinguishes two different types of processes in the climate system: (i) internal stochastic mechanisms; and (ii), external forcing mechanisms including their resonant amplification of internal modes. The spectrum in Mitchell's original figure (Fig. 1) thus consists of a background continuum on which is superimposed a series of spectral peaks. The power of the background continuum is stronger for lower frequencies and is the result of a number of red-noise spectra with increasing characteristic time scales. According to Mitchell this is the consequence of the stochastic aspect of the different climate system components for which first-order autoregressive processes are good conceptual models.

The purpose of this paper, which is an extended and updated version of Stocker (1995), is to give an overview on recent results that indicate the importance of these processes on time scales of many decades to centuries. Apart from a general interest this time scale is of particular importance, because detection of anthropogenic climate change depends on our knowledge of the time scales and patterns associated with the natural level of variability on the decadal to century time scale. A second purpose is to summarize those mechanisms of interdecadal-to-century variability that are quantitatively capable of causing detectable fluctuations. Up to now, our knowledge about these processes comes mostly from a hierarchy of dynamical models of the climate system and still only marginally from an observational or proxy network. The development and expansion of the latter is an important task of future research.

Many aspects of Mitchell's schematic picture can be found in records of climatic data both, directly observed and proxy. For instance, the red noise background of the spectrum (i.e. longer time scales exhibit stronger spectral power) can be seen in observed data from the atmosphere and ocean, and model simulations exhibit very similar spectral characteristics (Delworth *et al.* (1993), their Fig. 2, reproduced here as Fig. 5a). The spectra of longer time series (oxygen isotopes on planktonic foraminifera shells found in sea sediments) also confirm the red-noise continuum background and the presence of preferred time scales of variability (see e.g. Imbrie *et al.* (1992,

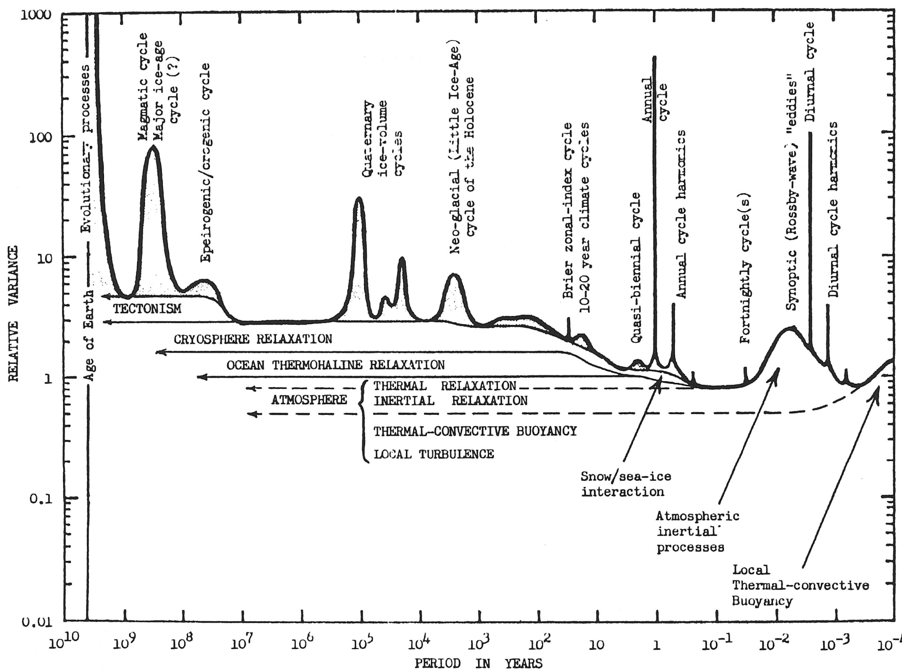

Figure 1: The original Figure of Mitchell (1976) showing a schematic spectrum of climate variability which is a compisite of a red noise background due to stochastic fluctuations and distinct peaks resulting from various external forcings. Note that no peaks are in the El Niño and decadal-to-century band [From Mitchell (1976)].

Imbrie *et al.* (1993)). Of more interest due their predictive potential are the external mechanisms that produce variability on distinct time scales such as the diurnal and seasonal cycles, cyclic processes without a single distinct time scale such as ENSO, the Milankovic cycles and even slower, tectonic processes.

While this concept is a useful starting point, recent high-resolution paleoclimatic archives (e.g. sea sediment cores by Lehman and Keigwin (1992, Bond *et al.* (1993) and the two recent ice cores from Summit, Greenland) have clearly demonstrated that additional aspects of climate variability must be taken into account. Cyclic or periodic fluctuations in a linear, dissipative system are due only to external forcing. While the climate system is dissipative, it is certainly not a linear system.

Therefore, processes additional to those mentioned above can generate vari-

ability. First, a non-linear system can exhibit *self-sustained oscillations* in response to external forcing. Their characteristic time scale is determined by feedback mechanisms internal to the system. Second, external forcing can generate *abrupt reorganisation* of the system, i.e. switches from one equilibrium state to another. While these switches are caused by 'external' perturbations (external for the ocean, i.e. for example atmosphere or ice sheets), the evolution of the system occurs on a time scale determined, again, by internal mechanisms.
The four features of climate change are shown in Fig. 2:

1. *EXTERNALLY FORCED CHANGES,*
2. *SELF-SUSTAINED OSCILLATIONS,*
3. *NON-DETERMINISTIC (CHAOTIC) VARIABILITY,*
4. *ABRUPT REORGANISATIONS.*

Various proxy and direct climatic data from the ice, sea and lake sediment archive illustrate the climatic evolution during the last 40,000 years. The transition from the last glacial to the Holocene is an **externally forced change** due to changes in the distribution of solar radiation and the operation of a number of feedback mechanisms (greenhouse gases, albedo). This is manifested in the gradual increase of $\delta^{18}O$ starting around 20,000 BP and in many proxy records from various archives (Imbrie *et al.* 1992; Imbrie *et al.* 1993) and strong support for the Milankovic theory of climate change.
Self-sustained oscillations of the ocean-atmosphere-terrestrial ice sheet system are likely to be responsible for the climatic swings between milder and colder phases (Dansgaard/Oeschger cycles) during the glacial and it appears that they involve large changes of ice sheet volume which, when melting, influence significantly ocean circulation (Bond *et al.* 1992; Bond *et al.* 1993; Paillard and Labeyrie 1994).
Non-deterministic (chaotic) variability is visible indirectly in the $\delta^{18}O$ record of the last 10,000 years (Holocene) which fluctuates about a well defined mean value; variations are likely due to changes in the hydrological cycle. Classically, this dynamical behaviour is thought to be most relevant for mesoscale processes (and hence comparatively short time scales) in atmosphere and ocean but there are first indications that also on the decadal-to-century time scale the concept of non-deterministic evolution is

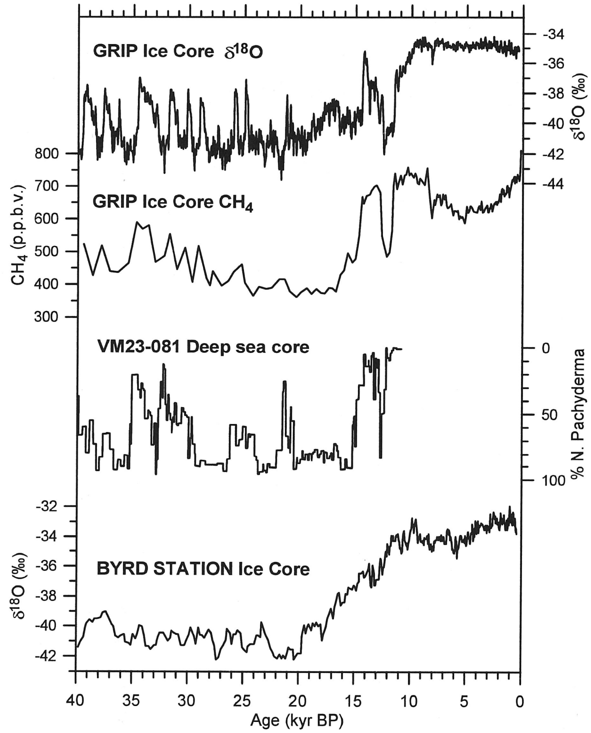

Figure 2: Climatic change over the last 40'000 years as obtained from the measurement of $\delta^{18}O$ on water (from Johnsen et al. (1992), Dansgaard et al. (1993) and Hammer et al. (1994)) and CH_4 on air of bubbles trapped in the ice core (from Chappellaz et al. (1993) and Blunier et al. (1995). Four different features of climate variability are evident in the different time series: (i) slow, astronomically forced transition from the glacial to the interglacial ($\delta^{18}O$ records and CH_4); (ii) self-sustained oscillations during the glacial (foraminifera assemblages in a sea sediment core (from Bond et al. (1993)) and ice cores before 20 kyr BP); (iii) non-deterministic (chaotic) variability during the Holocene ($\delta^{18}O$ records); and (iv) abrupt reorganisations during the Bølling/Allerød/Younger Dryas Period (all but Byrd Station core) [Figure compiled by T. Blunier]

required to describe the system (Saltzman 1983; James and James 1992; Mysak *et al.* 1993; Roebber 1995). Methane, on the other hand, exhibits surprising systematic changes during the Holocene (Blunier *et al.* 1995) which would rather hint at an externally forced cause (receding continental ice sheet give way to changing and evolving biosphere which influences methane production, see Blunier (1995)). This clearly indicates the need for a variety of climate proxies in order to characterize the dynamics and variability during a given period.

Three events of **abrupt reorganisations** recorded in sea sediments and ice cores, are superimposed on the longer term glacial-interglacial transition. The abrupt warming into the Bølling/Allerød, the cooling initiating the Younger Dryas and its termination all occur on time scales of a few decades to a few years (Dansgaard *et al.* 1989; Taylor *et al.* 1993). Model simulations have shown that such changes can be understood in terms of abrupt reorganisations, i.e. switches from one mode of operation to another, initiated by perturbations such as melting terrestrial ice sheets (Wright and Stocker 1993). It is important to distinguish between abrupt reorganisations and chaotic variability where the system resides for some time in one dynamical regime and then switches to another (Lorenz 1963; Lorenz 1990). Abrupt reorganisations are rapid changes between different *equilibrium* states when system parameters are slowly changing (Stocker and Wright 1991; Mikolajewicz and Maier-Reimer 1994; Rahmstorf 1995).

In the search for mechanisms of periodic climatic variability it is often tempting to look for periodic processes in some forcing variables and then postulate enhancing feedback mechanisms which cause a response of sufficient amplitude in the climate variable under consideration. Examples are the various solar cycles such as the sunspot (10–11 yr), Hale (22 yr), and Gleissberg (84 yr) cycles. Although these time scales do occur in spectra of various climate proxies abundantly (see below), it seems unlikely that such cycles are due to direct solar forcing. The global sensitivity of the climate system to changes in the shortwave irradiation is estimated at about $0.14\,K/Wm^{-2}$ (based on an AGCM simulation by Lean and Rind (1994)) with a spatially rather uniform response. The only effect on solar irradiation that has been directly measured is that of the 11-year solar cycle whose peak-to-peak amplitude has been determined at about $2.5\,Wm^{-2}$ (ERBE 1990). This would result in a temperature variation of about 0.35 K which would have to be detected as a globally uniform signal. Significant positive feedback mechanisms would have to operate in the climate system to amplify such a signal to the amplitude seen in many climate records (order of 1 K) and to impose regional patterns such as observed in the climatic reconstruction (e.g. Briffa *et al.* (1992)).

2 Observations and Proxy Indicators

2.1 Direct Observations

Direct observations begin to exhibit large-scale changes in the climate system on the interdecadal time scale. From the comparison of systematic temperature and salinity measurements of the last 40 years a distinct warming in depths between 700 m and 3000 m could be identified in the North Atlantic (Roemmich and Wunsch 1984; Levitus 1989c). Moreover, Parilla *et al.* (1994) find, superimposed on this warming, zonally alternating regions of cooling at 24°N and attribute this to large-scale decadal variability in the North Atlantic. There is also evidence of long-term variability and even step-like changes of properties in the marginal seas of the North Atlantic. Schlosser *et al.* (1991) found a significant reduction of ventilation of the deepwater in the Greenland Sea on the basis of tracers, while Lazier (1996) shows a notable decrease in salinity in the Labrador Sea over the last thirty years. Although these latter two studies do not show cycles they provide potentially important information on mechanisms and focal areas of variability.

Variability on decadal-to-century time scales is also found in atmospheric variables such as air temperature, wind speed and sea level pressure and can be related to concurrent changes in the underlying ocean. Deser and Blackmon (1993) find a cycle of 9–12 years in their analysis of 90 years of winter time means of sea surface temperature (SST), air temperature, wind speed and sea level pressure in the North Atlantic region. They identify two modes in the North Atlantic. The first is a north-south aligned dipole pattern with a negative anomaly of sea level pressure to the north. This results in a westerly wind anomaly enhancing the prevailing wind field and increasing the sea-air fluxes leading to negative SST anomalies. The second pattern is associated with the Gulf Stream where strong positive SST anomalies are found between the period 1939–1968 and 1900–1929. It is probably a response to the global surface warming trend in the 1920s and resulted in a general weakening of the basin-scale atmospheric circulation over the North Atlantic.

Kushnir (1994) has also analysed the above data set with an emphasis to identify the difference patterns between cold (1900–14, and 70–84) and

warm periods (1925–39 and 50–64) during this century. He found that the difference between warm and cold years consists of positive SST anomalies along the east coast of Greenland, the entrance of the Labrador Sea and off-shore of Newfoundland/Nova Scotia and negative SST anomalies along the northern part of the US east coast (Fig. 3). This is associated with a basin-wide cyclonic sea level pressure anomaly very similar to that reported by Deser and Blackmon (1993). The SST anomaly pattern is present in both the warm and the cold season (Fig. 3a, b), whereas the pressure anomaly is primarily a winter time phenomenon. While in the southern region of this SST anomaly the atmospheric circulation anomalies tend to attenuate the SST anomalies, they are maintained by them in the northern region. Hydrographic changes in the shallow and intermediate North Atlantic reported by Levitus (1989a), Levitus (1989b) are consistent with these findings. A similar structure (cooling at the Labrador Sea entrance and east coast of Greenland and warming is again found in the subsurface temperature (125 m) trends from 1966 to 1990 (Levitus *et al.* 1994)).
It is very important to address the question whether the North Atlantic is a region of enhanced variability as suggested by some proxy records or whether this is solely a result of the sparsity of data elsewhere. Schlesinger and Ramankutty (1994) analyze global data sets of surface air temperature covering the period 1858–1992 in 11 different geographical regions. Using singular spectrum analysis, they found a 50–80 year cycle in this variable in various sub-regions of the globe with the largest amplitudes in the North Atlantic and North American regions. This is an important hint towards an underlying mechanism. Similar decadal-to-century cycles at 24 and 100 years have been found in the Central England temperature time series, the longest instrumental record available, covering 318 years (Stocker and Mysak 1992). They are statistically significant at the 99% level.

2.2 Proxy Data

A review of long-term cyclic fluctuations on the century time scale is given in Stocker and Mysak (1992). Cycles of 50 years and longer are abundant in high-resolution proxy records (Fig. 4), but a clear and unequivocal time scale is missing. Stuiver (1980) tested the hypothesis whether such cycles found in proxy data could be due to solar variability for which $\Delta^{14}C$ from tree rings was used as a proxy. A statistically significant relationship could not be established.

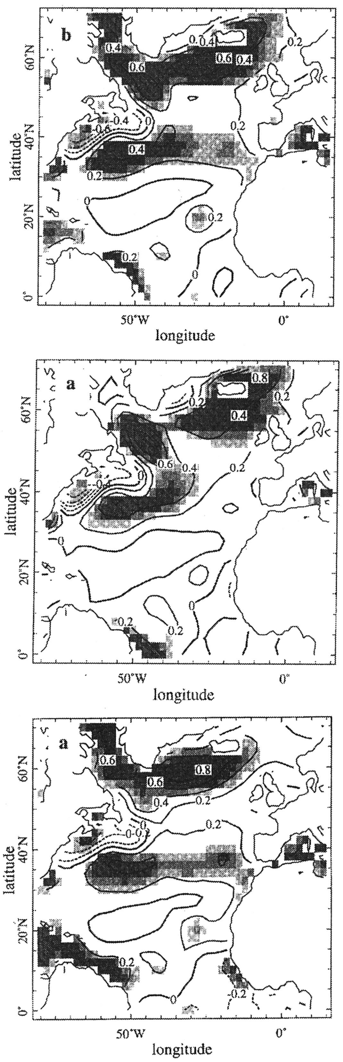

Figure 3: Differences of observed sea surface temperatures during 15 consecutive years of above average (1950–64) and below average (1970–84) temperatures in annual means (top), winter months (centre, Dec-Apr) and summer months (bottom, June-Oct). Large-scale, coherent anomalies can cause anomalous circulation patterns in ocean and atmosphere as independent numerical models also demonstrate. The SST patterns are better developed during the winter season [From Kushnir (1994)].

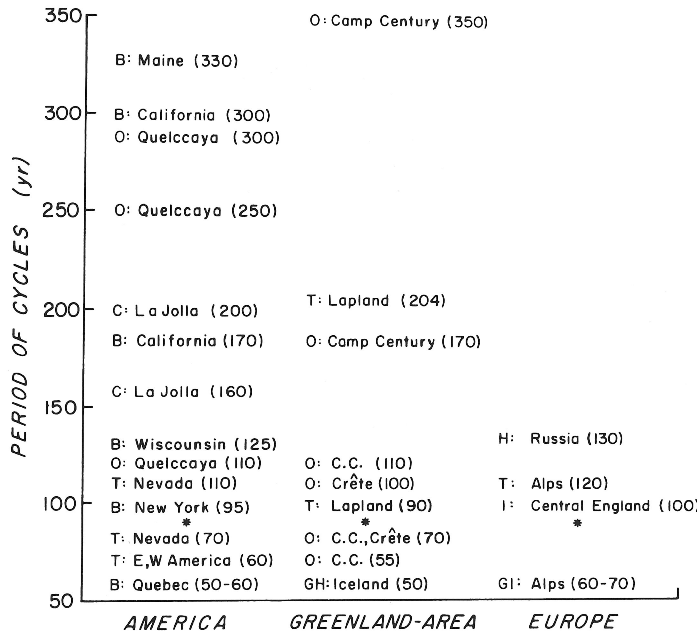

Figure 4: Summary of climatic variations observed on the century time scale. The findings are spatially and temporally ordered. The capital letters denote evidence in biological (B), radio carbon (C), glaciological (G), historical (H), instrumental (I), oxygen isotope and other ice core parameters (O) and tree ring records (T). The observed cycle period is given in brackets. C.C. denotes the Camp Century $\delta^{18}O$ record and $*$ indicates the 83-yr cycle in marine air and sea surface temperatures of Folland et al., (1984). [From Stocker et al. (1992)]

Briffa *et al.* (1992) have spectrally analyzed their 1480-year long tree ring record of Fennoscandia from which summer temperatures are reconstructed. They find significant power in the band of 30–40 years. An interesting feature is the fact that the periods are not stable: they vary by several years depending on the sub-interval considered. Such behavior is not expected if the oscillation is due to external forcing at a fixed period. It rather suggests that an internal mechanism may be responsible for these cycles.

Only recently researchers have begun to synthesize information derived from *different* paleoclimatic archives. The study of Mann *et al.* (1995) combines 35 different proxy records (tree rings, ice and coral cores, lake varves, historical records) distributed mainly in the northern hemisphere

with a few "stations" in the south and covering the last 500 years. This allows them not only to determine time scales of variability but is a first estimate of spatial correlation and patterns of natural variability as recorded in high-resolution proxy data time series of sufficient length. Mann *et al.* (1995) find cycles in the interdecadal (15-35 years) and century (50-150 years) range that are significant. As with the summer temperatures derived from tree ring widths of Fennoscandinavia, these variations are not stable throughout time. Century scale cycles are particularly strong during the mid 17 and 18 century, a time where fluctuations to colder conditions have been reported extensively and are referred to as "Little Ice Age" (Bradley and Jones 1995). The spatial distribution of phase relations indicates that the variability is mostly confined to the North Atlantic region with a clear phase difference of 45°-135° between the eastern and western side of the Atlantic basin. Mann *et al.* (1995) interpret this as the result of in-phase and out-of-phase variability. The former may be due to the basin-wide transport of heat by the meridional overturning circulation and its century scale variability (Mikolajewicz and Maier-Reimer 1990; Mysak *et al.* 1993) while the latter is consistent with a mechanism described by Delworth *et al.* (1993) (see below).
It appears from this that the North Atlantic region is one of the pace makers of climate variability. However, one should note that still only few proxy data come from regions other than Europe or North America, and that therefore our view may be biased. Future emphasis must be given to the retrieval and analysis of high-resolution paleorecords of the tropical regions and the southern hemisphere. In the meantime, models are the only tool that can help us understand mechanisms of variability on the decadal-to-century time scale. Also, they are suitable in pointing to locations of increased variability where proxy records, if made available, should most likely exhibit variability.

3 Models and Mechanisms

As mentioned above, the weakness of the hypothesis of a purely solar origin of decadal-to-century variability lies in the fact that the sensitivity of the climate system to changes in the shortwave irradiance is about a factor of 5 smaller than that for longwave emission changes including water vapour feedback (Cess *et al.* (1989) give a mean sensitivity of $0.68\,\mathrm{K/Wm^{-2}}$). By monitoring the most recent solar cycle it was found that corresponding solar irradiance changes have an amplitude of less than $2\,Wm^{-2}$ (ERBE 1990) which would yield an amplitude of about $0.3\,K$ using the above sensitivity. It is therefore important to look for possible alternative explanations and mechanisms of climate variability.

Modeling has become an important branch of climate research because only with physically based models is it possible to *quantitatively* test and verify hypotheses on climate change. Ocean, atmosphere and coupled models are successfully simulating the large-scale fields of the climate system (Trenberth 1992), and it is now timely to assess these models' capability of simulating also the natural variability.

During the last decade, oceanic circulation models have made significant progress due to the implementation of mixed boundary conditions that take into account feedback mechanisms between atmosphere and ocean. SST anomalies generate local heat flux anomalies that operate to remove the SST anomalies within a few weeks. Sea surface salinity (SSS) anomalies, on the other hand, do not influence the hydrological cycle, i.e. the surface freshwater balance, and hence can have a longer lasting impact on the surface buoyancy distribution. By relaxing SST to a fixed surface air temperature and keeping the surface freshwater fluxes constant, one arrives at a first approximation of the important difference between the feedback character of ocean-to-atmosphere heat and freshwater fluxes (Stommel 1961; Rooth 1982). However, local heat flux anomalies are bound to also change surface air temperature, an effect which is explicitly excluded when using mixed boundary conditions. Recognizing these limitations several studies have proposed improved parameterizations of the surface fluxes by formulating various types of energy balance models coupled to the ocean circulation models (e.g. Stocker *et al.* (1992), Zhang *et al.* (1993), Rahmstorf and Willebrand (1995), Lohmann *et al.* (1996)). Two effects are of importance: (i) SST anomalies cause heat flux anomalies which modify the surface air temperature locally; (ii) due to the possibility of meridional heat flux in the atmospheric part of the coupled model far-field effects can occur.

The basic mechanism for oscillations due to different feedback processes of SST and SSS anomalies was summarized by Welander (1986). He showed that self-sustained oscillations and different equilibrium states can be realized in a circular convection loop in which one side is heated and salted while the other side is cooled and freshened. This is reminiscent of the low and high latitudes where the surface ocean is heated and evaporation causes an increase in salinity whereas the opposite happens in the high

latitudes. He showed that if the time scales characteristic for temperature and salinity anomalies are different, multiple steady states as well as self-sustained oscillations are possible.
In more complex, 2- and 3-dimensional models mechanisms generating natural variability are more difficult to understand, and the range of mechanisms and time scales is quite broad. At present, we do not have a unifying theory explaining natural variability on the century time scale but we are in the stage of collecting evidence for such fluctuations both from the observational and the modeling side. It is hoped that over the coming years the representation of simple atmospheres used in driving ocean models becomes more realistic and that with this improvement, a more consistent picture will emerge. For now, a list of the type and time scales of variability found in numerical models (ocean, atmosphere, coupled) helps us in discussing physically plausible mechanisms of variability.
Table 1 gives a summary of self-sustained oscillations found in a number of ocean, atmosphere and coupled models. We focus only on the most robust cycles in these models. The decadal-to-century time scale does in most cases include the ocean circulation, in particular the thermohaline part of it. Mechanisms are connected with mainly with the thermohaline but also with the wind-driven circulation as well as with the hydrological cycle. Note that long-term variability is also found in atmospheric GCMs suggesting interesting possibilities of interaction between the atmosphere and the ocean also on interdecadal time scales.

Model Type	Geometry	Perturbation to Forcing	Surface Flux Parameterisation
Gyre–Thermohaline Circulation			
B&C 3D OGCM	box, flat bottom, hemispheric	–	mixed
B&C 3D OGCM	1 sector ocean, flat bottom, fixed ACC	var induced by mismatch of surface heat flux and heat storage	no-heat-capacity atm (i.e. effective restoring time of 1-2 yr), salt flux constant
3D B&C OGCM	box, flat bottom, hemispheric	increase of freshwater flux	freshwater flux only, no thermal forcing
B&C 3D OGCM	1 sector ocean, flat bottom, fixed ACC	var induced by mismatch of surface fluxes	mixed, constant heat or freshwater fluxes
3D OGCM	hemispheric box	–	mixed
3D OGCM	hemispheric box	–	mixed
SST Anomalies in the Northwest Atlantic			
3D A/OGCM	global	–	coupled, flux correction
3D OGCM	hemispheric box	–	constant heat flux, mixed or zero heat capacity atm.
Marginal Seas			
B&C 3D OGCM	flat N Atlantic and Labrador Sea	stochastic freshwater flux	mixed
3D LSG OGCM	global, topography	stochastic freshwater flux	mixed
Basin-Scale Thermohaline Circulation			
2D, zonally av. OCM	1 sector ocean	stochastic freshwater flux	mixed
3D LSG OGCM	global, topography	stochastic freshwater flux	mixed
The Southern Ocean-North Atlantic Connection			
3D LSG OGCM	global, topography	modified constant or stochastic freshwater flux	mixed
Other Mechanisms			
2D, zonally averaged OCM, ice	1 sector ocean	–	mixed
3D OGCM	hemispheric box	–	mixed
3D AGCM, coupled to LSG OGCM	global	–	coupled, flux correction
3D AGCM, dry	global, flat surface	–	surface energy balance

Table 1: Internal variability found in various numerical models ordered ...

Period	Mechanism	Reference
Gyre–Thermohaline Circulation		
9	advection of SSS anomaly by subtrop. and subpolar gyres	(Weaver and Sarachik 1991a)
24	advection of SST anomalies by subtropical gyre, overturning influences exposure time of SST anomalies	(Cai and Godfrey 1995)
20–350	advection of SSS anomalies influencing overturning (transition to chaos)	(Huang and Chou 1994)
20–100	advection of SSS or SST anomalies	(Cai 1995)
250–500	succession of haloclines and flushes triggered by gyre salt transport	(Winton 1993)
> 500	advection of SSS anomalies by gyres and vertical diffusion	(Winton 1993)
SST Anomalies in the Northwest Atlantic		
40–60	baroclinic vortex in West Atlantic, meridional heat flux and local heat storiage	(Delworth *et al.* 1993)
50–70	meridional heat transport and local heat storage	(Greatbatch and Zhang 1995)
Marginal Seas		
20	Labrador Sea, zonal and meridional overturning	(Weaver *et al.* 1994)
10–40	Labrador Sea operates as stochastic integrator sending SSS anomalies into N Atlantic	(Weisse *et al.* 1993)
Basin-Scale Thermohaline Circulation		
200–300	large-scale SSS advection	(Mysak *et al.* 1993)
320	large-scale SSS advection in the Atlantic around entire deep circulation loop	(Mikolajewicz and Maier-Reimer 1990)
The Southern Ocean-North Atlantic Connection		
320	halocline in ACC influences NADW, its outflow causes deep heating in ACC and triggers convection	(Pierce *et al.* 1995)
Other Mechanisms		
13.5	brine release–deep water formation–meridional heat flux feedback	(Yang and Neelin 1993)
17	ice cover–thermal insulation feedback	(Zhang *et al.* 1995)
10–20	advection of T anomalies in the Pacific	(Von Storch 1994)
5–40	chaotic nature, subtropical and mid-latitude atmospheric jets	(James and James 1992)

...according to time scale and mechanisms discussed in the text.

The table is grouped into six categories of mechanisms that will be discussed below and ordered according to associated time scales. The latter is not rigorous, of course, since periods depend on model parameters. Nevertheless, it contains information which processes in the climate system might be important on a given time scale. This should contribute to the identification of the most robust and fundamental modes of variability and change and distinguish them from more model-specific fluctuations.

3.1 Gyre–Thermohaline Circulation

Weaver and Sarachik (1991a) report self-sustained oscillations in a hemispheric 3-D OGCM under mixed boundary conditions. Formed in the western boundary current, warm and saline anomalies travel eastward and are picked up by the sub-polar gyre which transports them into the region of deep water formation; this journey takes about 8 to 9 years. There, they influence the basin-scale overturning and so feedback to the surface advection of these anomalies. The role of the latitudinal structure of the freshwater forcing was also studied (Weaver *et al.* 1993). When precipitation in the high latitude is increased internal variability on the decadal (as before) and the interdecadal (15–20 yr) time scales is generated. For sufficiently strong forcing sequences of violent overturnings and little deep water formation could be excited. As these are connected to diffusive processes in the ocean interior during the time of reduced ventilation, time scales are on the order of 500 years and are decreasing with increasing amplitude of the stochasting forcing.
Similar oscillations were found by Winton (1993) using a frictional-geostrophic model. Important in these models is the fact that anomalies are advected by the gyre circulation near the surface which determines the decadal time scale. These oscillations affect the entire water column in the high latitudes by turning on and off deep water formation, and the amplitude of the changes of the meridional heat flux are of order $0.2 \times 10^{15}\,W$. While these models allow us to isolate and investigate various mechanisms of internal variability, immediate application to the real world is limited because of the simple parameterization of ocean-atmosphere interaction, simplified geometry, and their still fairly coarse resolution.

3.2 SST Anomalies in the Northwest Atlantic

Coupled climate models are also beginning to exhibit natural variability. Delworth *et al.* (1993) integrate the GFDL climate model for 600 years and find natural variability whose spectral properties are remarkably similar to observations (Fig. 5a). Superimposed on a red-noise spectrum a number of spectral peaks are visible. Interdecadal oscillations of 40–60 yr are evident in the maximum meridional overturning in the North Atlantic reaching amplitudes of about 2 Sv (Fig. 5b). When the thermohaline circulation is weak, decreased advection of lower-latitude warm and saline waters into the central regions of the North Atlantic generates a pool of anomalously cold and fresh water. The thermal anomaly dominates and hence generates a geostrophically controlled cyclonic circulation at the surface (baroclinic vortex). The western half of this anomalous circulation enhances the mean northward current of warm and saline waters located in the center of the Atlantic which is part of the large-scale conveyor belt circulation. The strengthened conveyor then carries more saline and warm low-latitude waters into this region. Again, the thermal contribution is stronger and creates an anomalous warm pool which is associated with anti-cyclonic circulation. The latter weakens the mean flow again, and the cycle begins anew.

The oscillation is distinctly irregular, a common feature of nonlinear dynamical systems. Although the first 200 years of the integration show a quasi-periodic cycle of 40–50 years, the periods are longer in the following 400 years. This is indicative of some preconditioning of the ocean independent of the feedback mechanism. One plausible possibility is the volume, i.e. the heat and salt content, of the anomalous pool. The bigger the pool, the longer it takes a certain mass transport anomaly to erode the SST anomaly. While the amplitude of the anomaly governs the strength of this anomalous mass flux via the pressure gradients, the spatial extent itself does not influence it but determines the time scale on which the anomaly can be removed. For an estimate of orders of magnitude we assume a typical extent of this pool (65°W–45°W, 35°N–50°N, (Delworth *et al.* 1993)) and a depth of the anomaly of roughly 300m (Greatbatch and Zhang 1995). The corresponding volume of $8.2 \cdot 10^{14} m^3$ is renewed once in 26 years by a flux anomaly of 1 Sv. Changing pool sizes are thus likely to be responsible for the changes of the period lengths during different segments of the 600-year run of Delworth *et al.* (1993). It is intriguing that a similar in-

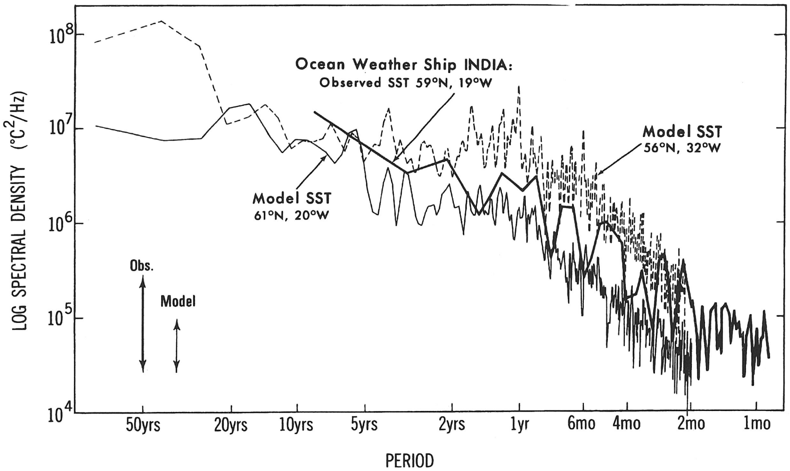

Figure 5a: Comparison of the spectral properties of a coupled A-OGCM with observations. The total amount of variability in the displayed frequency band is remarkably similar to observations [From Delworth et al. (1993)]

stationarity of interdecadal cycles is clearly present in proxy data (Briffa *et al.* 1992; Mann *et al.* 1995).

There are still limitations of such models. In order to achieve a stable climate consistent with the observations spatially dependent flux corrections have to be applied. Although these corrections do not impose a time scale on the model, they act as an additional forcing which, in a nonlinear system, could generate additional variability. Also, the region where the oscillation is observed (northwest Atlantic) is one of the areas of large flux correction (Manabe and Stouffer 1988). On the other hand, the model does suggest a deterministic mechanism with time scales and amplitudes similar to those of proxy records and also indicates the focal regions of oscillatory activity. Moreover, the spatial patterns and depth structures of temperature and salinity in the ocean and sea level pressure that are suggested by the model can be directly compared to long-term observations. The basic mechanism is also present in an ocean-only model driven by *constant* surface heat flux only, suggesting that the phenomenon is primarily thermally driven (Greatbatch and Zhang 1995). Occurring in two entirely different models is good evidence that this mechanism is a very robust feature. Greatbatch and Zhang (1995) have also found that the inclusion

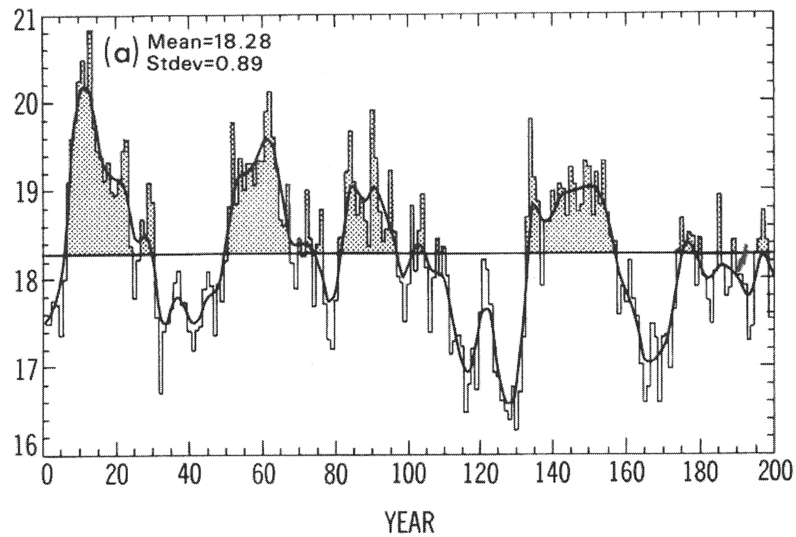

Figure 5b: Time series of the maximum overturning streamfunction and running average in the North Atlantic for the first 200 years of their 600-year integration. [From Delworth et al. (1993)]

of freshwater flux forcing increases the period and decreases the amplitude of the oscillation. A changing hydrological cycle in the atmosphere thus represents a second mechanism for instationarity of periods.

3.3 Marginal Seas

Recent modeling results indicate that marginal seas are potentially very important pacemakers for self-sustained variability of the ocean circulation. Weaver *et al.* (1994) found oscillations of about 20 years in the Labrador Sea of their model. The mechanism is linked to an interaction between the meridional (and zonal) pressure gradients and the geostrophically controlled zonal (and meridional) overturning in the Labrador Sea. In contrast to the mechanism discussed in the previous section, this variability is neither dependent on the surface freshwater budget nor on the wind-driven gyral circulation. While the resolution of this model in the most important area is admittedly very low (only 3 tracer grid points), it is encouraging to note that SST anomalies are very similar to those reported by Kushnir (1994) based on observations.

A complementary study with a global model was done by Weisse *et al.* (1993). A stochastic freshwater flux perturbation excites century (320 yr) and decadal scale oscillations. The latter are localized, again, in the Labrador Sea area, have periods in the range of 10–40 years, but the mechanism is distinctly different from that above. The relatively isolated marginal sea integrates the stochastic freshwater flux perturbations and sends salinity anomalies into the North Atlantic on a time scale of about 10–40 years. This time scale is determined by the flushing time of the upper 250 m of the basin, since the stratification is quite stable. Once the perturbations arrive in the North Atlantic, where the stratification between 50°N and 60°N is weak, they strongly influence the deep water formation rates and create the variability observed.
These two mechanisms differ distinctly from each other in that in the first study, the Labrador Sea itself generates the variability by changing rates of local deep water formation, whereas in the other case the same region merely appears as a storage of perturbations which, once accumulated, act outside the basin. An increased resolution and, with it, a better representation of the water masses in the Labrador Sea will refine our understanding of its role in controlling the natural variability in the North Atlantic region.

3.4 Basin-Scale Thermohaline Circulation

Basin-scale variations of the Atlantic meridional overturning and hence meridional heat flux are possible mechanisms for century time scale variability. The Hamburg global OGCM was run under mixed boundary conditions including a stochastic freshwater flux perturbation (Mikolajewicz and Maier-Reimer 1990). Large fluctuations with a time scale of 320 years are found in the mass transport through Drake Passage and the meridional heat flux and overturning in the North Atlantic. Amplitudes would be large enough to be detected in paleoclimatic archives (the heat budget in the Southern Ocean shows peak-to-peak amplitudes of up to $3 \times 10^{15}\,W$); events are not regular cycles but appear in the time series as distinct events. The mechanism is associated with the long residence time of SSS anomalies. The random freshwater flux perturbations create local salinity anomalies that are advected northward by the near-surface circulation in the Atlantic. Depending on the spatial structure of the mean freshwater fluxes they are enhanced or removed before they reach the deep water formation area in the high-latitude Atlantic. If they are enhanced, they tend to accelerate

meridional overturning. The authors trace these anomalies along the entire loop of overturning so that they eventually surface again in the South Atlantic/Southern Ocean region. Although much diffused and diminished by that time, they are the seed of new anomalies that start their northbound travel at the sea surface.
Mysak *et al.* (1993) forced a zonally averaged one-basin ocean circulation model with random freshwater fluxes and found oscillations with time scales between 200 and 300 years over a wide range of parameters (Fig. 6). In contrast to the previous model, salinity anomalies could not be traced around a complete loop of the circulation but became well mixed once they were in the deep ocean. This indicates that these cycles are primarily due to an interaction between the detailed spatial structure of the surface freshwater flux and the strength of the thermohaline overturning which determines the preferred time scale.

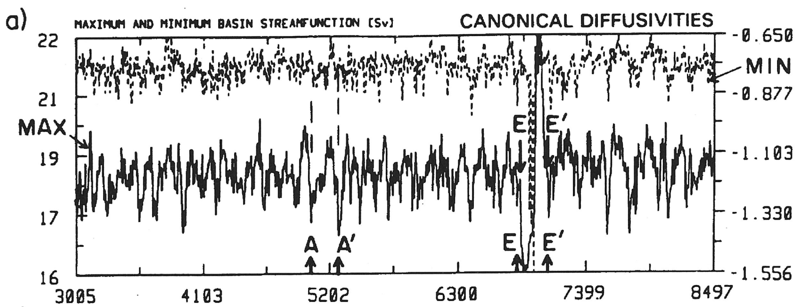

Figure 6a: Time series of the minimum (dashed) and maximum (solid) overturning streamfunction in a one-basin 2-d ocean model. [From Mysak et al. (1993)]

3.5 The Southern Ocean-North Atlantic Connection

Very recently two studies have re-examined natural variability on the century time scale found in the Hamburg ocean model (Pierce *et al.* 1995; Osborn 1995). Although the cycles have the same period they identify and demonstrate a mechanism that is significantly different from that proposed by Mikolajewicz and Maier-Reimer (1990). Also, variability need not be generated by a stochastic perturbation of the freshwater flux but can be initiated by a slight change (±10%) of the magnitude of the freshwater flux, i.e. an imposed incompatibility of the fluxes with the steady state (a

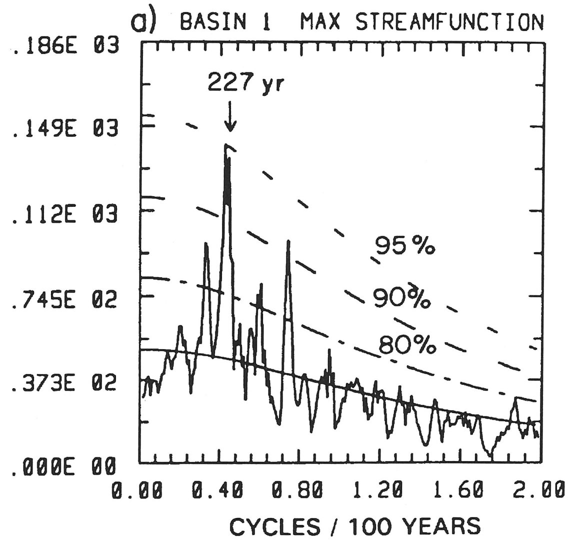

Figure 6b: Spectrum of the time series in Fig. 6a. Significant power is found on the century time scale. [From Mysak et al. (1993)]

similar way of inducing variability was presented by Cai (1994)). This is an indication that models which use the classical mixed boundary conditions, where the salt flux is fully consistent with the steady-state circulation field, probably underestimate natural variability.

The mechanism of variability is based on a newly found coupling between the hemispheres by the Atlantic thermohaline circulation. The model oscillates between two extreme states during one cycle: (i) strong deep water formation in the Southern Ocean with significant influx of Antarctic Bottom Water into the Atlantic where the thermohaline circulation is slightly weakened; (ii) halocline around Antarctica, no AABW in the Atlantic with slightly increased Atlantic overturning. The development and destruction of the Circum-Antarctic halocline has – via the JEBAR effect (see Cai (1994) for an example in a 3D model) – a strong impact on barotropic transport through the Drake Passage which changes between 60 and 160 Sv!

A cycle evolves as follows. When a strong halocline is present in the Southern Ocean, surface heat exchange is strongly reduced (due to stable stratification and ice cover). The warmer waters exiting from the Atlantic ocean

at a depth of 2–3 km manage to slowly heat up the subsurface Southern Ocean until it is destabilised. Convection then establishes fairly quickly all around Antarctica and reaches full strength within less than 100 years. During the following 200 years convection slowly decreases again due to the action of net precipitation in that area. The result is an asymmetric oscillation pattern. Pierce *et al.* (1995) show that this oscillation is only possible when a nonlinear equation of state is used.
The above mechanism is similar to the flushes found in many other models (Marotzke 1989; Wright and Stocker 1991; Weaver and Sarachik 1991b; Winton and Sarachik 1993) whose characteristic feature is a decoupling of the lower ocean from the surface. In contrast to these earlier studies where an entire basin had to be destabilized by *diffusion* and hence evolved on much longer time scales of $o(10^3 - 10^4\text{yrs})$, destabilisation here is due to *advection* of warmer NADW, i.e. an "efficient" process with a time scale of a few hundred years.

3.6 Other Mechanisms

Internal variability on the interdecadal time scale is found in simple ocean models when a thermodynamic sea ice component is included. Formation of a sea ice cover releases salt into the water column and induces convection. By altering density gradients this increases the thermohaline circulation advecting more heat northward which then melts the sea ice cover. Melting increases the surface density which tends to decrease the thermohaline circulation leading to a net cooling. This feedback loop was found in a 2-dimensional ocean circulation model coupled to a thermodynamic sea ice model (Yang and Neelin 1993). Zhang *et al.* (1995) showed that this process is also present in a 3-dimensional OGCM coupled to an ice model and found the destabilizing influence of brine rejection less important than that of slow heating below the ice cover due to thermal insulation.
In the previous section mainly oceanic processes localized in or related to the Atlantic have been discussed. However, there are also recent model examples of interdecadal cycles in the Pacific and in the atmosphere alone. James and James (1992) find in their atmospheric circulation model variability on time scales of 10 to 40 years and associate them with changes in the structure of the subtropical and mid-latitude jets.
Von Storch (1994), on the other hand, uses a coupled A/OGCM and identifies two types of low-frequency variability. The atmospheric fluctuations

are essentially red noise, and there appears an out-of-phase relationship between the stratosphere and the troposphere. In the ocean there is an irregular cycle of about 17 years located in the Pacific Ocean. Note that no distinct variability is found in the Atlantic basin neither in marginal seas nor basins-wide.

4 Conclusions

The study and quantitative understanding of internal variability in the climate system is still in its infancy. Focal regions where internal variability is enhanced or generated have been identified and preliminary results have shed light on a palette of mechanisms. However, we are still at the stage where each model tends to produce its characteristic set of fluctuations with the associated time scales. Important results have been found regarding the role of the ocean, the influence of atmosphere-ocean exchange fluxes and the dynamics in marginal seas, but a consistent *quantitative theory* on climatic cycles on the decadal-to-century time scale is still missing. Further work is urgently needed both with respect of models and observations. We need to improve the formulation of surface exchange processes (boundary conditions) especially in the ocean-only models, the parameterisation of convection and deep water formation; the resolution must be increased in order to better represent topography and marginal but important regions of the ocean basins must be included.

More important, however, are efforts to obtain and analyze high-quality terrestrial and oceanic proxy data which allow annual or seasonal resolution. Special attention must be focused on the transfer function, i.e. how is the climate signal transferred into the archive and how is the possible temporal evolution of such a transfer function. Ongoing European and US efforts to retrieve two high-resolution ice cores from different locations in Antarctica will provide vital information about natural variability centered in the Southern Ocean. Finally, spatial networks of homogenized paleoclimatic data will certainly allow us to find some of the keys to better understand natural variability in the climate system.

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