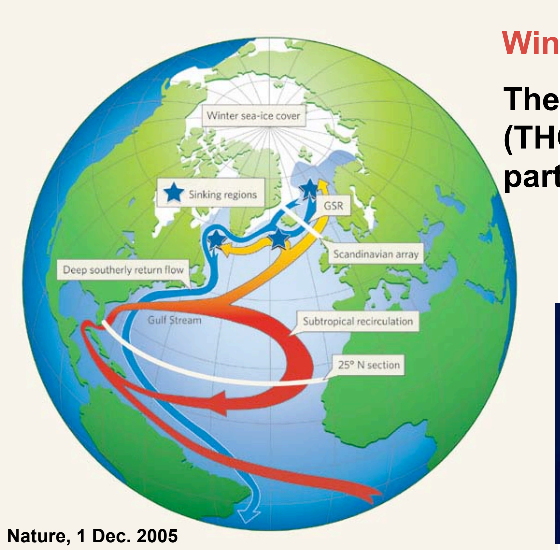
Climate Dynamics (lecture 10)

Stommel model of the thermohaline circulation (THC)

Role of the ocean in climate variability, in particular during last glacial period

http://www.phys.uu.nl/~nvdelden/

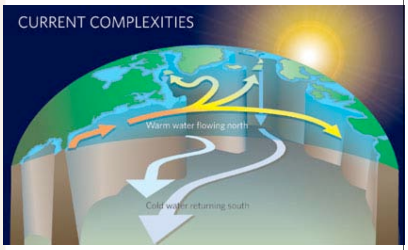
The meridional overturning circulation (MOC)



Nature, 1 Dec. 2005

Wind- and density-driven

The thermohaline circulation (THC) is the density-driven part

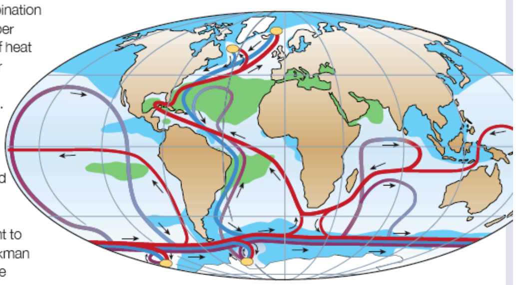


Nature, 19 Jan. 2006

Box 1

Nature 419, 207-214 (12 September 2002)

Some key facts about ocean circulation



The large-scale ocean circulation can be thought of as a combination of currents driven directly by winds (mostly confined to the upper several hundred metres of the sea), currents driven by fluxes of heat and freshwater across the sea surface and subsequent interior mixing of heat and salt (the so-called thermohaline circulation), and tides (driven by the gravitational pull of the Moon and Sun). These driving mechanisms interact in nonlinear ways (since all currents change the heat and salt distribution) so that no unique decomposition exists. Nevertheless the distinction is useful, particularly when changes in wind or in surface heat and freshwater fluxes are considered for their effects on the circulation.

An important way in which wind-driven currents are thought to lead to climatic changes is through their effect on upwelling (Ekman divergence) near coasts and the Equator, changing sea surface temperatures. This mechanism plays a part in the El Niño/Southern Oscillation cycle. The thermohaline circulation is most interesting for its highly nonlinear response to changes in surface freshwater forcing[88], allowing large changes in heat transport to occur (see Box 2). Tides are relevant to the climate system because they form one of the main sources of turbulent energy (in addition to that provided by the wind) to mix the ocean[89].

A highly simplified cartoon of the global thermohaline circulation (sometimes called 'conveyor belt') is shown in the figure above (modified from the original by Broecker). Near-surface waters (red lines) flow towards three main deep-water formation regions (yellow ovals) — in the northern North Atlantic, the Ross Sea and the Weddell Sea — and recirculate at depth (deep currents shown in blue, bottom currents in purple; green shading indicates salinity above 36‰, blue shading indicates salinity below 34‰). A recent estimate of the rate of deep-water formation is 15 ± 2 Sv (1 Sv $= 10^6$ m^3 s^{-1}) in the North Atlantic and 21 ± 6 Sv in the Southern Ocean[90]. Northward heat transport into the northern Atlantic peaks at 1.3 ± 0.1 PW (1 PW $= 10^{15}$ W) in the subtropics[90]; this heat transport warms the northern Atlantic regional air temperatures by up to 10 °C over the ocean with the effect declining inland.

Little is currently known about present-day natural variability of this circulation (see ref. 91 for a review), or about the effects of such variability on surface climate. Variations of the Atlantic thermohaline circulation on timescales of several decades are found in many coupled climate models, with a typical amplitude of a few sverdrup; they are probably damped oscillations driven by stochastic variations in surface fluxes (that is, weather variability)[92]. Good observational time series of integral measures of this circulation are lacking, although some data suggest that such decadal variability also exists in nature, and is correlated with the North Atlantic Oscillation (NAO)[93,94]. The NAO also seems to orchestrate the location and intensity of deep convection in the northern Atlantic[95]. Lack of data makes it hard to establish whether a longer-term trend in the circulation exists, although there is intriguing evidence for trends in temperature and salinity[96,97] that may indicate a gradual weakening of the overflow from the Nordic Seas into the Atlantic in recent decades.

A highly simplified but very interesting model of the THC

Stommel model of the THC

(Taylor, 2005)

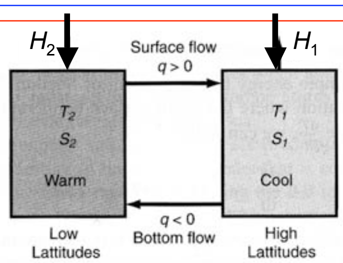


Fig. 11.6
A two-box model of the ocean circulation. The temperatures T_1 and T_2 are fixed so the principal variable is the salinity, S.

Two reservoirs of well-mixed water connected by "pipes" represent the polar and equatorial regions of the ocean at temperatures T_1 and T_2 (fixed for simplicity).The principle variable is the salinity of the water, S, which is affected by a "virtual" flux H of salt from the atmosphere (see also the previous slide). The flow of water q between the boxes is proportional to the density difference. Conservation salt is expressed by

$$\frac{dS_1}{dt} = H_1 + |q|(S_2 - S_1); \quad \frac{dS_2}{dt} = H_2 + |q|(S_1 - S_2)$$

Density of sea-water

The flux is given by

$$q = \frac{k}{\rho_0}(\rho_1 - \rho_2)$$

k is an unknown coefficient with the dimension [s^{-1}]. The equation of state for sea water is (approximately) (see the figure) $\rho = \rho_0(1 - \alpha T + \beta S)$

$\alpha\ (>0)$ is the thermal **expansion** coeff.;
$\beta\ (>0)$ is the haline **contraction** coeff.

$$q = \frac{k}{\rho_0}\Delta\rho = k(\alpha\Delta T - \beta\Delta S)$$

with

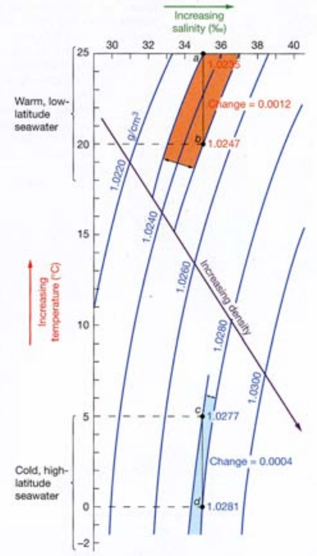


Figure 5.22 Seawater density varies with temperature and salinity. Blue curves show density, which increases with increasing salinity and decreases with increasing temperature. In high-latitude areas of cold water, temperature has less effect on density than in high-temperature, low-latitude areas.

Stommel model

$$\Delta T = T_2 - T_1;\ \Delta S = S_2 - S_1;\ \Delta\rho = \rho_1 - \rho_2$$

$$\frac{dS_1}{dt} = H_1 + |q|(S_2 - S_1);\quad \frac{dS_2}{dt} = H_2 + |q|(S_1 - S_2)$$

The salt flux from the atmosphere is given by H_i.

H_i is prescribed as follows* $H_i = -\lambda_i(S_i - S_{i0})$

Equilibrium value in absence of meridional transport

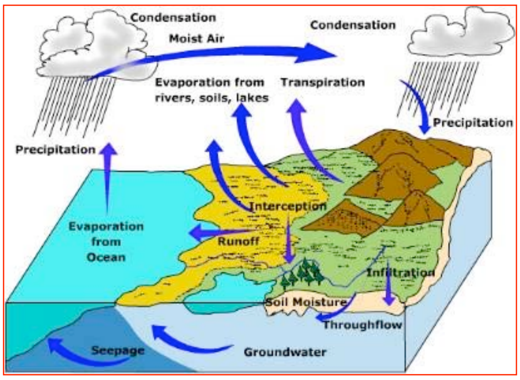
What processes govern these equilibrium values?

*Later we will prescribe H in a different manner

Stommel model

$$\frac{dS_1}{dt} = H_1 + |q|(S_2 - S_1); \quad \frac{dS_2}{dt} = H_2 + |q|(S_1 - S_2)$$

$$H_i = -\lambda_i(S_i - S_{i0}) \qquad q = \frac{k}{\rho_0}\Delta\rho = k(\alpha\Delta T - \beta\Delta S)$$



Two timescales:

$$\tau_1 \equiv \frac{1}{k}; \tau_2 \equiv \frac{1}{\lambda_i}$$

Associated with intensity of the ocean circulation

Associated with freshening of the ocean

Stommel model

$$\frac{dS_1}{dt} = H_1 + |q|(S_2 - S_1); \quad \frac{dS_2}{dt} = H_2 + |q|(S_1 - S_2)$$

$$H_i = -\lambda_i(S_i - S_{i0}) \qquad q = \frac{k}{\rho_0}\Delta\rho = k(\alpha\Delta T - \beta\Delta S)$$

$\Delta S \equiv S_2 - S_1 \qquad \lambda_1 = \lambda_2 = \lambda$

$$\boxed{\frac{d\Delta S}{dt} = -\lambda(\Delta S - \Delta S_0) - 2|k(\alpha\Delta T - \beta\Delta S)|\Delta S}$$

$\Delta S_0 \equiv S_{20} - S_{10}$

Free parameters are ΔT and ΔS_0

Which relaxation parameter do you think is greater, λ or k? Why?

Stommel model

Two timescales:

$$\tau_1 \equiv \frac{1}{k}; \tau_2 \equiv \frac{1}{\lambda_i}$$

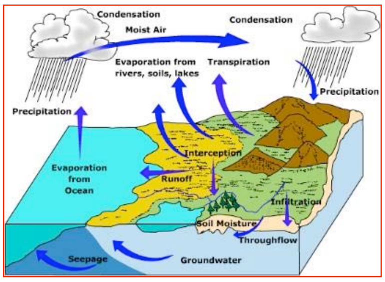
τ_2 associated with freshening of the ocean

Total global fresh water input is 1 Sv=10^6 m^{3}s^{-1}.

Associated timescale of freshening of the upper 100 m of the ocean is

$$\frac{V}{dV/dt} \approx \frac{100 \times 0.7 \times \{\text{area globe}\}}{10^6} \approx 10^3 \text{years}$$

HUGE!



τ_1 associated with intensity of the ocean circulation

Intensity of the Gulfstream is 30-150 Sv.

Therefore τ_1 is about a factor 100 smaller than τ_2

Therefore:

$$\lambda_i \approx 3\times10^{-11}\text{s}^{-1}; k \approx 3\times10^{-9}\text{s}^{-1}$$

$$(\lambda_1 = \lambda_2)$$

Steady states of Stommel model

$$\frac{d\Delta S}{dt} = -\lambda(\Delta S - \Delta S_0) - 2|k(\alpha\Delta T - \beta\Delta S)|\Delta S = 0$$

if $\alpha\Delta T > \beta\Delta S$ $\quad -\lambda(\Delta S - \Delta S_0) - 2k(\alpha\Delta T - \beta\Delta S)\Delta S = 0$ (q>0)

$$\beta\Delta S = +\frac{1}{2}\left(\frac{\lambda}{2k} + \alpha\Delta T\right) \pm \frac{1}{2}\left(\left(\frac{\lambda}{2k} + \alpha\Delta T\right)^2 - \frac{2\lambda\beta\Delta S_0}{k}\right)^{1/2}$$

$\alpha\Delta T > \beta\Delta S$!!!

two solutions possible (thermally driven) **(solution 1&2)**

if $\alpha\Delta T < \beta\Delta S$ $\quad -\lambda(\Delta S - \Delta S_0) - 2k(\beta\Delta S - \alpha\Delta T)\Delta S = 0$ (q<0)

$$\beta\Delta S = \frac{1}{2}\left(\frac{-\lambda}{2k} + \alpha\Delta T\right) \pm \frac{1}{2}\left(\left(\frac{-\lambda}{2k} + \alpha\Delta T\right)^2 + \frac{2\lambda\beta\Delta S_0}{k}\right)^{1/2}$$

$\alpha\Delta T < \beta\Delta S$!!!

(solution 3)

Minus-sign discarded because $\beta\Delta S > 0$

So, this gives one solution (salt driven circulation)

Multiple equilibria

two solutions for thermally driven circulation

if $\left(\frac{\lambda}{2k}+\alpha\Delta T\right)^2 > \frac{2\lambda\beta\Delta S_0}{k}$

Exercise

Plot the steady state solutions in a graph with *q* along the vertical axis and the meridional temperature difference along the horizontal axis and determine the stability of these solutions. Discuss the implications of the result (see also , the following slide).

Multiple equilibria

The thermohaline circulation is responsible for a large part of the heat transport

The thermohaline circulation is sensitive to freshwater"forcing"

Stefan Rahmstorf, 2002: Nature, 419, 207-214

Box 2

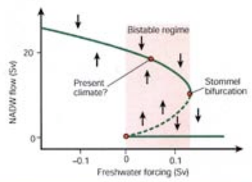
Stability and nonlinearity of the thermohaline circulation

The thermohaline circulation is thermally driven: highest surface densities in the world ocean are reached where water is coldest, in spite of the lower salt content there compared with the warmer tropical and subtropical regions. Nevertheless, the influence of salinity is important and is the main cause of the nonlinearity of the system. Salinity is involved in a positive feedback. Higher salinity in the deep-water formation area enhances the circulation, and the circulation in turn transports higher salinity waters into the deep-water formation regions (which tend to be regions of net precipitation, that is, freshwater would accumulate and surface salinity would drop if the circulation stopped). This leads to two possible equilibrium states, with and without North Atlantic Deep Water (NADW) formation[4]. This was first described in a classic paper by Stommel[37] with the help of a simple box model.

The stability properties are illustrated in the diagram below, which plots the strength of the thermohaline circulation as a function of the freshwater input into the North Atlantic. The simple presentation shows the bistable regime and a saddle-node bifurcation point where the circulation breaks down (for a more detailed discussion, see ref. 88).

This salt-transport feedback is not the only feedback rendering the system nonlinear. The convective mixing process at high latitudes is itself a highly nonlinear, self-sustaining process, which at least in models can lead to multiple stable convection patterns[3,38]. Together, these two positive feedback mechanisms allow two types of transitions between distinct circulation modes: on/off switches of NADW formation, and shifts in the location of convection. These two mechanisms are crucial in attempts to explain glacial climate changes.

The thermohaline circulation takes several thousand years to reach full equilibrium. The transient response to a change in forcing can therefore deviate substantially from the equilibrium solutions and is in many cases more linear[89].



Numerical integration of the Stommel model

$$\frac{dS_1}{dt} = H_1 + |q|(S_2 - S_1); \quad \frac{dS_2}{dt} = H_2 + |q|(S_1 - S_2) \qquad q = \frac{k}{\rho_0}\Delta\rho = k(\alpha\Delta T - \beta\Delta S)$$

$$H_i = -\lambda_i(S_i - S_{i0})$$

$$\frac{d\Delta S}{dt} = -\lambda(\Delta S - \Delta S_0) - 2|k(\alpha\Delta T - \beta\Delta S)|\Delta S$$

$\Delta S_0 = 10$ i.e. it is 10 parts per thousands saltier in the south than in the north

$\lambda_i = 3\times10^{-11}\mathrm{s}^{-1}; k = 3\times10^{-9}\mathrm{s}^{-1}$

$\alpha = 0.0002\ \mathrm{K}^{-1};\ \beta = 0.001$

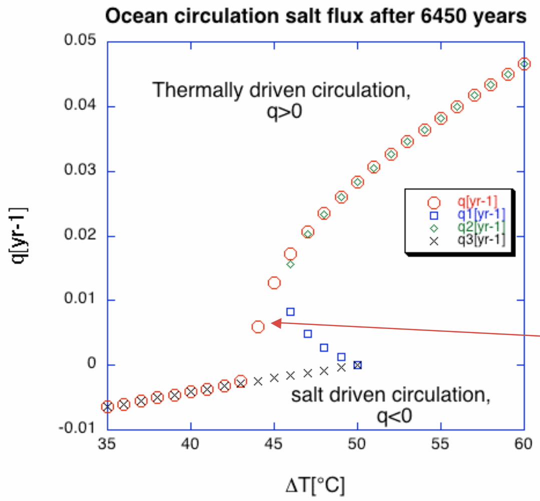
At $t = 0$ $\Delta S = 10$ parts per thousand

The Runge-Kutta scheme is used to approximate the time derivative

Result of an integration lasting 6450 years

Circulation intensity

See:
http://www.phys.uu.nl/~nvdelden/Stommel.htm



q: flux at t=6450 years
q1: solution 1
q2: solution 2
q3: solution 3

$\Delta S_0 = 10$

not in a steady state

Different formulation salt flux

Stommel-Taylor model

$$\Delta T = T_2 - T_1;\ \Delta S = S_2 - S_1;\ \Delta\rho = \rho_1 - \rho_2$$

$$\frac{dS_1}{dt} = H_1 + |q|(S_2 - S_1);\ \frac{dS_2}{dt} = H_2 + |q|(S_1 - S_2)$$

The salt flux from the atmosphere is given by H_i.

H_i is prescribed as follows $H_2 = -H_1 \equiv H > 0$

$$\frac{d\Delta S}{dt} = 2H - 2|k(\alpha\Delta T - \beta\Delta S)|\Delta S$$

$$Y \equiv \beta\Delta S;\ X \equiv \alpha\Delta T$$

$$\frac{dY}{dt} = 2\beta H - 2|k(X - Y)|Y$$

Stommel-Taylor model

$$\frac{dY}{dt} = 2\beta H - 2|k(X - Y)|Y$$

Steady states: $0 = 2\beta H - 2|k(X - Y)|Y$

if $\alpha\Delta T > \beta\Delta S$ $\quad Y \equiv Y_0 = \frac{X}{2} \pm \frac{1}{2}\left(X^2 - \frac{4\beta H}{k}\right)^{1/2}$ **(solution 1&2)**

if $\alpha\Delta T < \beta\Delta S$ $\quad Y \equiv Y_0 = \frac{X}{2} \pm \frac{1}{2}\left(X^2 + \frac{4\beta H}{k}\right)^{1/2}$ **(solution 3)**

Minus-sign discarded because $\beta\Delta S > 0$

Stability analysis

(solution 3) $Y_0 = \frac{X}{2} + \frac{1}{2}\left(X^2 + \frac{4\beta H}{k}\right)^{1/2}$ **(1)** $\alpha\Delta T < \beta\Delta S$

Salt driven

$$\frac{dY}{dt} = 2\beta H - 2|k(X-Y)|Y \quad \textbf{(2)}$$

Suppose $Y \equiv Y_0 + Y'$ $\quad Y' << Y_0$

(small perturbation to the steady state)

Substitute in (2) using (1):

$$\frac{dY'}{dt} = -2k\left(X^2 + \frac{4\beta H}{k}\right)^{1/2} Y'$$

$Y' = A\exp(\sigma t)$ **growthrate:** $\sigma = -2k\left(X^2 + \frac{4\beta H}{k}\right)^{1/2} < 0$

Therefore perturbation dies out: solution 3 is always stable to small perturbations

Stability analysis

$\alpha\Delta T > \beta\Delta S$ temperature driven

Same analysis as on previous slide:

solution 1: $Y_0 = \frac{X}{2} - \frac{1}{2}\left(X^2 - \frac{4\beta H}{k}\right)^{1/2}$ **is stable** $(\sigma < 0)$

if $X^2 > \frac{4\beta H}{k}$

solution 2: $Y_0 = \frac{X}{2} + \frac{1}{2}\left(X^2 - \frac{4\beta H}{k}\right)^{1/2}$ **is unstable** $(\sigma > 0)$

if $X^2 > \frac{4\beta H}{k}$

System can "jump" from one stable steady state to another stable steady state

Stability analysis

Condition for existence of temperature driven circulation:

$$X^2 > \frac{4\beta H}{k}$$

corresponds to

$$E = \frac{\beta H}{kX^2} = \frac{\beta H}{k(\alpha\Delta T)^2} < \frac{1}{4}$$

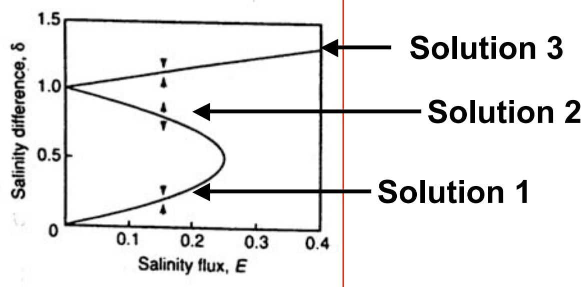
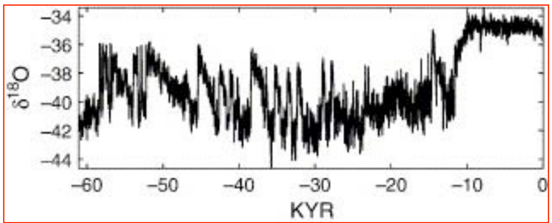


Fig. 11.7
Dimensionless salinity difference $\delta = \beta\Delta S/\alpha\Delta T$ plotted against dimensionless surface salinity flux $E = \beta H/k(\alpha\Delta T)^2$ showing equilibrium solutions (lines) and tendencies (arrows). The latter indicate that the middle branch of the curve consists of solutions that are unstable.

Taylor (2005)

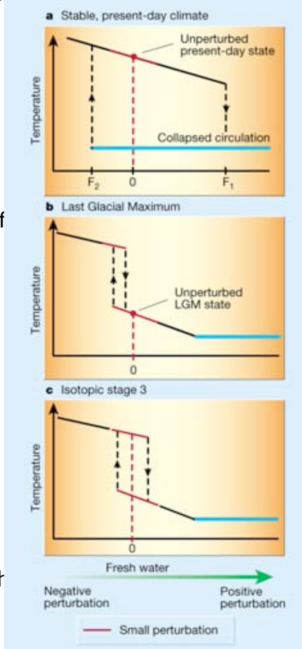
Does this kind of non-linear behaviour have something to do with…

Strong climate variations during last glacial period (?)



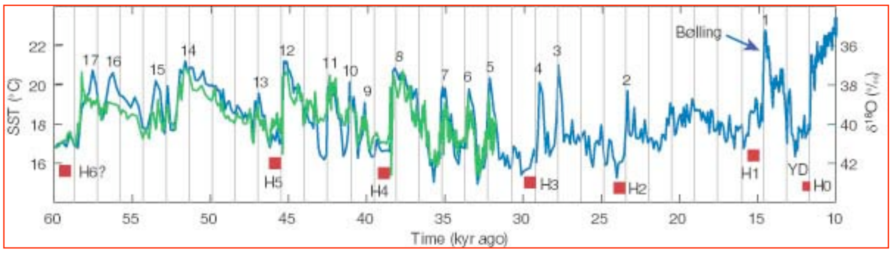
$\delta^{18}O$ from the **GISP2 ice core**. Time runs from left to right. This normalized ratio of $^{18}0$ to $^{16}0$ concentrations is believed to track local atmospheric temperatures in central Greenland to within an approximate factor of two. Large positive spikes are called **Dansgaard-Oeschger (D-O) events** and are correlated with abrupt warming. Note in particular the **quiescence of the Holocene interval** (approximately the last 10,000 yr) **relative to the preceding glacial period**. The Holocene coincides with the removal of the Laurentide and Fennoscandian ice sheets. The range of excursion corresponds to about 15°C. Time control degrades with increasing age of the record.

During glacial times the ocean meridional overturning circulation switched abruptly between cold and warm modes, with the temperature in Greenland changing by up to 10 °C in a matter of decades (see figure).
a, Under present-day conditions, North Atlantic climate has essentially two possible equilibria. When freshwater input exceeds a threshold value F_1, thermohaline circulation jumps from the upper (warm) equilibrium branch to the lower (colder) one, which corresponds to thermohaline collapse (blue line). It can return to the upper branch only if fresh water is removed (by, say, evaporation) and decreases below the threshold value F_2. The hysteresis width F_1–F_2 is large. So present climate is not destabilized by weak freshwater perturbations. **b**, Under the conditions of the Last Glacial Maximum, the hysteresis is much narrower and so the system is much more sensitive to the input or removal of fresh water. Even a slight reduction can induce abrupt warmings, and such **Dansgaard–Oeschger warming events** are evident in the palaeoclimate record. Large inputs of fresh water, as during **Heinrich events (ice-sheet melting)**, will induce a relatively small cooling through thermohaline collapse. **c**, A guess at an intermediate situation, as pertained during isotopic stage 3, around 50,000–30,000 years ago. The warm (upper) branch is more stable than it is under LGM conditions, corresponding to the longer Dansgaard–Oeschger events that occurred at this time.
Nature **409**, 147-148 (11 January 2001)



Looking closer at the last glacial period

D/O events and Heinrich events



Temperature reconstructions from ocean sediments and Greenland ice. Proxy data from the subtropical Atlantic[86] (green) and from the Greenland ice core GISP2 (ref. 87; blue) show several Dansgaard–Oeschger (D/O) warm events (numbered). The timing of Heinrich events is marked in red. Grey lines at intervals of 1,470 years illustrate the tendency of D/O events to occur with this spacing, or multiples thereof.

Nature 419, 207-214 (12 September 2002)

Heinrich events

In the 1980's, studies of rapidly deposited sediments in the North Atlantic detected relatively short climate variations. **Hartmut Heinrich** first connected these variations to major episodes of ice rafting separated by 500-15000 years.

Heinrich events are **massive episodic iceberg discharges** from the Laurentide ice sheet through Hudson Strait, with up to 10% of the ice sheet sliding into the oceans. A highly plausible explanation is that the ice **sheet grew to a critical height where it became unstable**, and a major surge could then start spontaneously or be triggered by a small perturbation

Nature 419, 207-214 (12 September 2002)

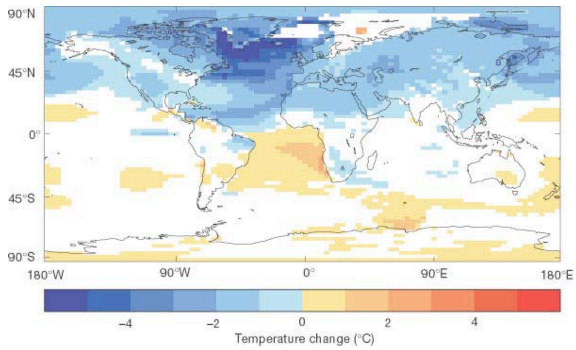
Dansgaard-Oescher event

Dansgaard–Oeschger (D/O) events are perhaps the most pronounced climate changes that have occurred during the past 120 kyr. They are not only large in amplitude, but also abrupt. In the Greenland ice cores, D/O events start with a **rapid warming by 5–10°C within at most a few decades**, followed by a plateau phase with slow cooling lasting several centuries, then a more rapid drop back to cold stadial conditions. The events are not local to Greenland. Amplitudes are largest in the North Atlantic region, and many Southern Hemisphere sites, especially those in the South Atlantic, reveal a hemispheric **'see-saw' effect** (**cooling while the north is warming**). D/O events have curious statistical properties: the waiting time between two consecutive events is often around 1,500 years, with further preferences around 3,000 and 4,500 years (Fig. 3), which suggests a stochastic resonance[35] process at work.
Several ideas have been advanced to explain D/O events, most of which **involve the thermohaline circulation** of the Atlantic. The first of these was probably the idea of **thermohaline circulation bistability**, much like what is seen in the Stommel model. NADW formation is active during the warm phases (interstadials), whereas it is shut off during cold phases (stadials), and some outside trigger causes mode switches between these two stable states.

Younger Dryas cold event

The end of D/O 1 marked the beginning of the last cold event before the Holocene. This event is called the Younger Dryas (YD) cold event. It seems to be special in a number of ways. Because of the high **meltwater influx** at this time, **NADW formation probably stopped**, as during Heinrich events. Nevertheless, it seems **hard to reconcile** the fact that the Younger Dryas event is almost as cold as previous Heinrich events during glacial-maximum conditions with the already elevated CO_2 level in the atmosphere (over 240 p.p.m.) and reduced inland ice volume. Furthermore, there is increasing evidence from New Zealand and South America that the Younger Dryas event was accompanied by a global re-advance of ice, which is also reflected in a temporary halt of sea-level rise. The Younger Dryas event may thus be **more than a change in ocean circulation**; a global forcing causing cooling could be involved, possibly of **solar origin**. A final northern cooling in the history of deglaciation is a short event occurring 8,200 years ago, which has also been linked to a meltwater-induced weakening of the thermohaline circulation.

Climate Model Calculations



Changes in surface air temperature caused by a shutdown of North Atlantic Deep Water (NADW) formation in an ocean–atmosphere circulation model. Note the hemispheric *see-saw* (Northern Hemisphere cools while the Southern Hemisphere warms) and the maximum cooling over the northern Atlantic. In this particular model (HadCM3), the surface cooling resulting from switching off NADW formation is up to 6°C.

Nature 419, 207-214 (12 September 2002)

Conclusion

The study of climate variations over the past 120,000 years has reached a state where palaeoclimatic data provide increasingly reliable information on the driving forces and the responses of the climate system, and where distinct climatic events such as glaciation, deglaciation, D/O events or Heinrich events can be characterized in terms of their spatial patterns and evolution over time. **Understanding the mechanisms behind these climatic changes has moved beyond speculation to specific, testable hypotheses backed up by quantitative simulations.**

It has become clear that the climate system is sensitive to forcing and responds with large and often abrupt changes in surface conditions. **The role of the ocean circulation is that of a highly nonlinear amplifier of climatic changes.** Many issues are still controversial and unresolved, both in terms of the data (for example, whether the late-glacial glacier advance in New Zealand and South America is synchronous with the Younger Dryas cold event in the north) and in terms of the mechanisms (for example, whether Younger Dryas cooling is caused by a meltwater-induced shutdown of NADW formation). But progress has been rapid, and the potential exists to resolve many of these issues in the coming decade or so by collecting more data, refining the analysis methods and improving models.

A better understanding of the carbon cycle remains one of the main challenges; the ocean has a crucial role in this cycle, one that has not been discussed. Reconstructions and modelling of carbon cycle changes can provide useful constraints on ocean circulation changes, and understanding the glacial–interglacial changes in atmospheric CO_2 concentration remains an elusive central piece in the climate puzzle.

Nature 419, 207-214 (12 September 2002)

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