

ABRUPT CLIMATE CHANGE

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Introduction

High-resolution paleoclimatic records from ice and sediment cores and other sources have revealed a number of dramatic climatic changes that occurred over surprisingly short times – a few decades or in some cases a few years. In Greenland, for example, temperature rose by 5–10°C, snowfall rates doubled, and wind-blown dust decreased by an order of magnitude within 40 years at the end of the last glacial period. In the Sahara, an abrupt transition occurred around 5500 years ago from a relatively green shrubland supporting significant populations of animals and humans to the dry desert we know today.

One could define an abrupt climate change simply as a large and rapid one – occurring faster than in a given time (say 30 years). The change from winter to summer, a very large change (in many places larger than the glacial–interglacial transition) occurring within six months, is, however, not an abrupt change in climate (or weather), it is rather a gradual transition following the solar forcing in its near-sinusoidal path. The term abrupt implies not just rapidity but also reaching a breaking point, a threshold – it implies a change that does not smoothly follow the forcing but is rapid in comparison to it. This physical definition thus equates abrupt climate change with a strongly nonlinear response to the forcing. In this definition, the quaternary transitions from glacial to interglacial conditions and back, taking a few hundred or thousand years, are a prime example of abrupt climate change, as the underlying cause, the earth's orbital variations (Milankovich cycles), have timescales of tens of thousands of years. On the other hand, anthropogenic global warming occurring within a hundred years is not as such an abrupt climate change as long as it smoothly follows the increase in atmospheric carbon dioxide. Only if global warming triggered a nonlinear response, like a rapid ocean circulation change or decay of the West Antarctic Ice Sheet (WAIS), would one speak of an abrupt climate change.

Paleoclimatic Data

A wealth of paleoclimatic data has been recovered from ice cores, sediment cores, corals, tree rings, and other sources, and there have been significant advances in analysis and dating techniques. These advances allow a description of the characteristics of past climatic changes, including many abrupt ones, in terms of geographical patterns, timing and affected climatic variables. For example, the ratio of oxygen isotopes in ice cores yields information about the temperature in the cloud from which the snow fell. Another way to determine temperature is to measure the isotopic composition of the nitrogen gas trapped in the ice, and it is also possible to directly measure the temperature in the borehole with a thermometer. Each method has advantages and drawbacks in terms of time resolution and reliability of the temperature calibration. Dust, carbon dioxide, and methane content of the prehistoric atmosphere can also be determined from ice cores.

On long timescales, climatic variability throughout the past two million years at least has been dominated by the Milankovich cycles in the earth's orbit around the sun – the cycles of precession, obliquity, and eccentricity with periods of roughly 23 000, 41 000, and 100 000 years, respectively. Since the middle Pleistocene transition 1.2 million years ago, the regular glaciations of our planet follow the 100 000-year eccentricity cycle; even though this has only a rather weak influence on the solar radiation reaching the earth, it modulates the much stronger other two cycles. The prevalence of the 100 000-year cycle in climate is thus apparently a highly nonlinear response to the forcing that is still poorly understood. The terminations of glaciations occur rather abruptly (**Figure 1**). Greenland ice cores show that the transition from the last Ice Age to the warm Holocene climate took about 1500 years, with much of the change occurring in only 40 years.

The ice ages were not just generally colder than the present climate but were also punctuated by abrupt climatic transitions. The best evidence for these transitions, known as Dansgaard–Oeschger (D/O) events, comes from the last ice age (**Figure 2**). D/O events typically start with an abrupt warming by around 5°C within a few decades or less, followed by gradual cooling over several hundred or thousand years. The cooling phase often ends with an abrupt final temperature drop back to cold

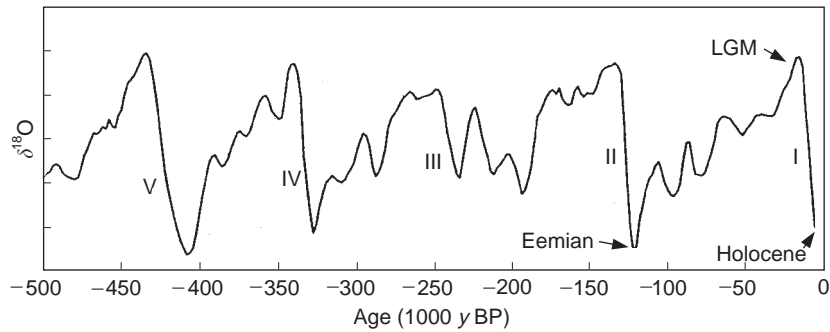


Figure 1 Record of $\delta^{18}\text{O}$ from marine sediments (arbitrary units), reflecting mainly the changes in global ice volume during the past 500 000 years. Note the rapid terminations (labeled with roman numbers) of glacial periods.

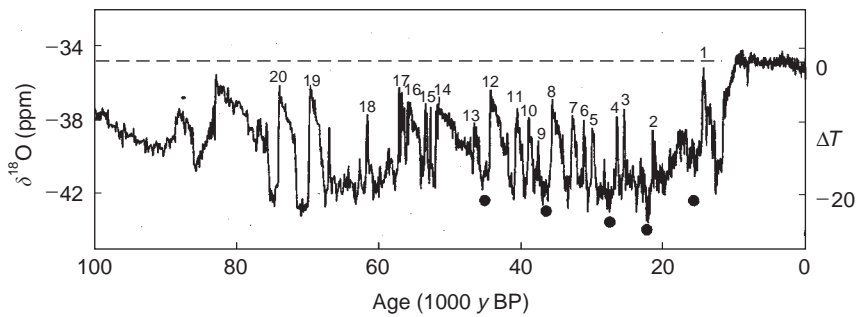


Figure 2 Record $\delta^{18}\text{O}$ from the GRIP ice core, a proxy for atmospheric temperature over Greenland (approximate temperature range ΔT ($^{\circ}\text{C}$) is given on the right). Note the relatively stable Holocene climate during the past 10 000 years and before that the much colder glacial climate punctuated by Dansgaard-Oeschger warm events (numbered). The timing of Heinrich events 1 to 6 is marked by black dots.

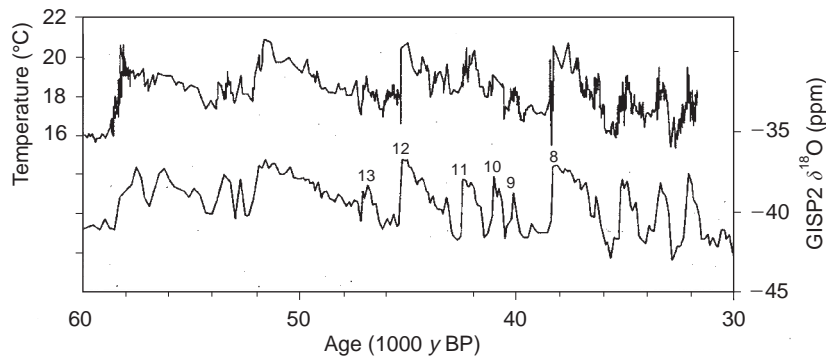


Figure 3 Sea surface temperatures derived from alkenones in marine sediments from the sub-tropical Atlantic (Bermuda Rise, upper curve) compared to $\delta^{18}\text{O}$ values from the GISP2 ice core in Greenland.

(‘stadial’) conditions. Although first seen in the Greenland ice cores, the D/O events are not a local feature of Greenland climate. **Figure 3** shows that subtropical sea surface temperatures in the Atlantic closely mirror the sequence of events in Greenland. Similar records have been found near Santa Barbara, California, in the Cariaco Basin off Venezuela, and

off the coast of India. D/O climate change is centered on the North Atlantic and regions with strong atmospheric response to changes in the North Atlantic, with little response in the Southern Ocean or Antarctica. The ‘waiting time’ between successive D/O events is most often around 1500 years or, with decreasing probability, 3000 or 4500

years. This suggests the existence of an as yet unexplained 1500-year cycle that often (but not always) triggers a D/O event. The second major type of climatic event in glacial times is the Heinrich (H) event. Heinrich events involve surging of the Laurentide Ice Sheet through Hudson Strait, occurring in the cold stadial phase of some D/O cycles. They have a variable spacing of several thousand years. The icebergs released to the North Atlantic during H events leave tell-tale drop-stones in the ocean sediments when they melt. Sediment data further suggest that H events shut down or at least drastically reduce the formation of North Atlantic Deep Water (NADW). Records from the South Atlantic and parts of Antarctica show that the cold Heinrich events in the North Atlantic were associated with unusual warming there (a fact sometimes referred to as ‘bipolar see-saw’).

At the end of the last glacial, a particularly interesting abrupt climatic change took place, the so-called Younger Dryas event (12 800–11 500 years ago). Conditions had already warmed to near-interglacial conditions and continental ice sheets were retreating, when within decades the climate in the North Atlantic region switched back to glacial conditions for more than a thousand years. It has been speculated that the cooling resulted from a sudden influx of fresh water into the North Atlantic through St. Lawrence River, when an ice barrier holding back a huge meltwater lake on the North American continent broke. This could have shut down the Atlantic thermohaline circulation (i.e., the circulation driven by temperature and salinity differences), but evidence is controversial. Alternatively, the Younger Dryas may simply have been the last cold stadial period of the glacial following a temporary D/O warming event.

Does abrupt climate change occur only during glacial times? Evidence for the last interglacial, the Eemian, is so far inconclusive. During the present interglacial, the Holocene, climate was much more stable than during the last glacial. However, two abrupt events stand out. One is the 8200-year event that shows up as a cold spike in arctic ice cores and affected the North Atlantic region. The second major change is the abrupt desertification of the Sahara 5500 years ago. There is much evidence from cave paintings, fire remains, bones, ancient lake sediments, and the like that the Sahara was a partly swampy savannah before this time. The best evidence for the abrupt ending of this benign climate comes from Atlantic sediments off north-eastern Africa, which show a sudden and dramatic step-function increase in wind-blown dust, witnessing a drying of the adjacent continent.

Mechanisms of Abrupt Climate Change

The increased spatial coverage, quality, and time resolution of paleoclimatic data as well as advances in computer modeling have led to a greater understanding of the mechanisms of abrupt climate change, although many aspects are still in dispute and not fully understood.

The simplest concept for a mechanism causing abrupt climatic change is that of a threshold. A gradual change in external forcing (e.g., the change in insolation due to the Milankovich cycles) or in an internal climatic parameter (e.g., the slow build-up or melting of continental ice) continues until a specific threshold value is reached where some qualitative change in climate is triggered. Various such critical thresholds are known to exist in the climate system. Continental ice sheets may have a stability threshold where they start to surge; the thermohaline ocean circulation has thresholds where deep water formation shuts down or shifts location; methane hydrates in the seafloor have a temperature threshold where they change into the gas phase and bubble up into the atmosphere; and the atmosphere itself may have thresholds where large-scale circulation regimes switch.

For the D/O events, Heinrich events, and the Younger Dryas event discussed above, the paleoclimatic data clearly point to a crucial role of Atlantic ocean circulation changes. Modeling and analytical studies of the Atlantic thermohaline circulation (sometimes called the ‘conveyor belt’) show that there are two positive feedback mechanisms leading to threshold behavior. The first, called advective feedback, is caused by the large-scale northward transport of salt by the Atlantic currents, which in turn strengthens the circulation by increasing density in the northern latitudes. The second, called convective feedback, is caused by the fact that oceanic convection creates conditions favorable for further convection. These (interconnected) feedbacks make convection and the large-scale thermohaline circulation self-sustaining within certain limits, with well-defined thresholds where the circulation changes to a qualitatively different mode.

Three main circulation modes have been identified both in sediment data and in models (Figure 4): (A) a warm or interglacial mode with deep water forming in the Nordic Seas and large oceanic heat transport to northern high latitudes; (B) a cold or stadial mode with deep water forming south of the shallow sill between Greenland, Iceland, and Scotland and with greatly reduced heat transport to high latitudes; and (C) a ‘switched off’ or ‘Heinrich’ mode

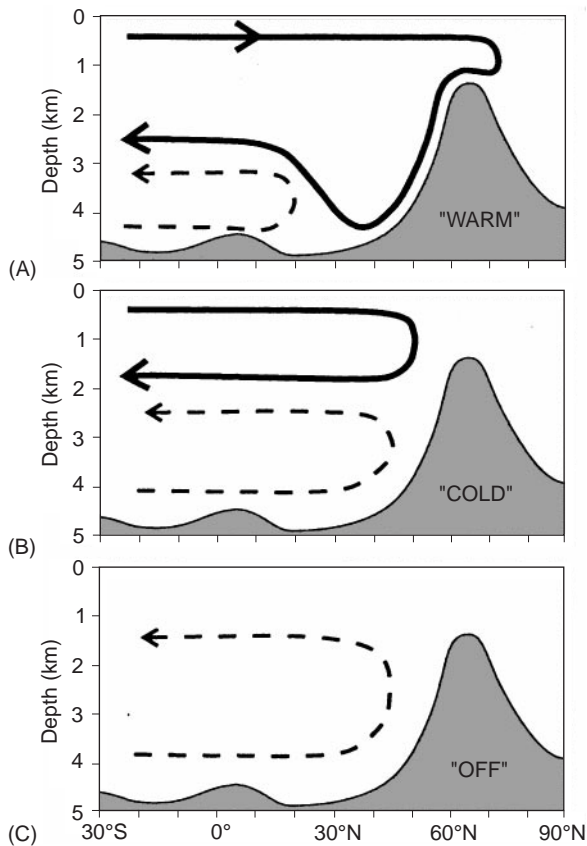


Figure 4 Schematic of three major modes of Atlantic ocean circulation. (A) Warm or interglacial mode; (B) cold or stadial mode; (C) 'off' or Heinrich mode. In the warm mode the Atlantic thermohaline circulation reaches north over the Greenland–Iceland–Scotland ridge into the Nordic Seas, while in the cold mode it stops south of Iceland. Switches between circulation modes at certain thresholds can pace and amplify climatic changes.

with practically no deep water formation in the North Atlantic. In the last mode, the Atlantic deep circulation is dominated by inflow of Antarctic Bottom Water (AABW) from the south.

Many features of abrupt glacial climate can be explained by switches between these three circulation modes. Model simulations suggest that the cold stadial mode is the only stable mode in a glacial climate; it prevails during the cold stadial periods of the last glacial. D/O events can be interpreted as temporary incursions of warm Atlantic waters into the Nordic Seas and deep water formation there, i.e., a switch to the warm mode causing abrupt climatic warming in the North Atlantic region. As this mode is not stable in glacial conditions, the circulation starts to gradually weaken and temperatures start to decline again immediately after the incursion, until the threshold is reached where con-

vection in the Nordic Seas stops and the system reverts to the stable stadial mode. Heinrich events can be interpreted as a switch from the stadial mode to the Heinrich mode, i.e., a shutdown of North Atlantic deep water formation. As this mode is probably also unstable in glacial conditions, the system spontaneously reverts to the stadial or to the warm mode after a waiting time of centuries, the timescale being determined partly by slow oceanic mixing processes.

This interpretation is consistent with the observed patterns of surface temperature change. The warming during D/O events is centered on the North Atlantic because this is where the change in oceanic heat transport occurs; the warm mode delivers heat to much higher latitudes than does the cold mode. A switch to the Heinrich mode, on the other hand, strongly reduces the interhemispheric heat transport from the South Atlantic to the North Atlantic. This cools the northern hemisphere while warming the southern hemisphere, explaining the 'bipolar seesaw' response in climate. It should also be noted that the initial transient response can differ from the equilibrium response as the oceanic heat storage capacity is large. The patterns of these abrupt changes differ from the longer-timescale (many thousands of years) response to the Milankovich cycles because for the latter the slow changes in atmospheric greenhouse gases (e.g., CO_2) and continental ice cover act to globally synchronize and amplify climatic change.

While the threshold behavior of the Atlantic ocean can dramatically shape and amplify climatic change, the question remains what triggers the mode switches. As mentioned above, D/O switches appear to be paced by an underlying 1500-year cyclicity that is as yet unexplained. This could either be an external (astronomical or solar) cycle or an internal oscillation of the climate system, perhaps also involving the Atlantic thermohaline circulation. The irregularity in D/O event timing is probably the result of the presence of stochastic variability in the climate system as well as the presence of longer-term trends such as the slow build-up of large continental ice sheets.

The ocean circulation change during Heinrich events, on the other hand, can be explained by the large amounts of fresh water entering the North Atlantic at these times in the form of icebergs. Simulations show that the observed amounts of fresh water are sufficient to shut down deep water formation in the North Atlantic. The nonlinear dynamics of ice sheets provide a plausible trigger mechanism. Ice sheets may grow for many thousands of years until their base melts owing to

geothermal heating, when the ice sheet becomes unstable and surges.

Thresholds in ocean circulation and continental or sea ice dynamics are not the only mechanisms that can cause abrupt climatic changes. In the desertification of the Sahara in the mid-Holocene, probably neither of these mechanisms was involved. Rather, an unstable positive feedback between vegetation cover (affecting albedo and evapotranspiration) and monsoon circulation in the atmosphere appears to have been responsible.

It is almost certain that there are further nonlinearities in the climate system that could have caused abrupt climatic changes in the past or may do so in the future. We are only beginning to understand abrupt climate change, and the interpretations presented here – while consistent with data and model results – are not the only possible interpretations. Reflecting the state of this science, they are current working hypotheses rather than established and well-tested theory.

Risk of Future Abrupt Changes

The prevalence of abrupt nonlinear (rather than smooth and gradual) climatic change in the past naturally leads to the question whether such changes can be expected in the future, either by natural causes or by human interference. The main outside driving forces of past climatic changes are the Milankovich cycles. Close inspection of these cycles as well as modelling results indicates that we are presently enjoying an unusually quiet period in the climatic effect of these cycles, owing to the present minimum in eccentricity of the earth's orbit. The next large change in solar radiation that could trigger a new ice age is probably tens of thousands of years away. If this is correct, it makes the Holocene an unusually long interglacial, comparable to the Holstein interglacial that occurred around 400 000 years ago when the earth's orbit went through a similar pattern. This stable orbital situation leaves unpredictable events (such as meteorite impacts or a series of extremely large volcanic eruptions) and anthropogenic interference as possible causes for abrupt climatic changes in the lifetime of the next few generations of humans.

Significant anthropogenic warming of the lower atmosphere and ocean surface will almost certainly occur in this century, raising concerns that nonlinear thresholds in the climate system could be exceeded and abrupt changes could be triggered at some point. Processes that have been (rather speculatively) mentioned in this context include a collapse of the West Antarctic Ice Sheet, a strongly enhanced

greenhouse effect due to melting of permafrost or triggering of methane hydrate deposits at the seafloor, a large-scale wilting of forests when drought tolerance thresholds are exceeded, nonlinear changes in monsoon regimes, and abrupt changes in ocean circulation.

Of those possibilities, the risk of a change in ocean circulation is probably the best understood and perhaps also the least unlikely. Two factors could weaken the circulation and bring it closer to a threshold: the warming of the surface and a dilution of high-latitude waters with fresh water. The latter could result from an enhanced atmospheric water cycle and precipitation as well as meltwater runoff from Greenland and other glaciers. Both warming and fresh water input reduce surface density and thereby inhibit deep water formation. Model simulations of global warming scenarios so far suggest three possible responses: a shutdown of convection in the Labrador Sea, one of the two main NADW formation sites; a complete shutdown of NADW formation (i.e., similar to a switch to the Heinrich mode); and a shutdown of AABW formation. A transition to the stadial circulation mode has so far not been simulated, perhaps because convection in the Nordic Seas is strongly wind-driven and is more effectively switched off by increased sea ice cover than by warming.

A shutdown of Labrador Sea convection would be a significant qualitative change in the Atlantic Ocean circulation, but would probably affect only the surface climate of a smaller region surrounding the Labrador Sea. Effects on ecosystems and fisheries have not been investigated but could be severe. A complete shutdown of NADW formation would have wider climatic repercussions. Temperatures in north-western Europe could initially rise several degrees in step with global warming, then abruptly drop back to preindustrial values (the competing effects of raised atmospheric CO₂ and reduced oceanic heat transport almost balancing). If CO₂ levels decline again in future centuries as expected, European temperatures could remain several degrees below present as the Atlantic thermohaline circulation is not expected to recover. Most models indicate that for the present climate (in contrast to the glacial), both the warm mode and the 'switched off' mode of Atlantic circulation are stable. Global effects of a shut-down of deep water renewal include reduced oceanic uptake of CO₂ (enhancing the greenhouse effect) and accelerated sea level rise (due to a faster warming of the deep oceans).

The probability of major climatic thresholds being crossed in the coming centuries is difficult to establish and largely unknown. Currently, this possibility

lies within the (still rather large) uncertainty range for future climate projections, so the risk cannot be ruled out.

See also

Deep-sea Drilling Results. Past Climate From Corals. Thermohaline Circulation. Water Types and Water Masses.

Further Reading

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ABSORBANCE SPECTROSCOPY FOR CHEMICAL SENSORS

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Introduction

Optical methods have played an important role and continue to do so in various fields of chemical analysis. Optical methods that are used in chemical analysis rely on the fact that there is an interaction of electromagnetic radiation with matter. Thus, these methods can provide a rapid and nondestructive tool for analysis. In chemical sensors, a range of wavelengths covered by optical techniques are used that can be divided into the ultraviolet (200–400 nm), the visible (400–780 nm), the near infrared (780–3000 nm), and the infrared (3–50 μm) regions. When electromagnetic radiation in the optical region interacts with matter, dispersion, absorption, diffraction, refraction, and reflection of the light can occur.

Absorbance spectroscopy has been used in conjunction with dry reagent chemistries in the development of optical chemical and biochemical sensors for a variety of applications including environmental and biomedical analyses. These sensors incorporate transducers which consist of active chemicals/biochemicals immobilized on to suitable polymeric materials and interfaced to optical fibres. The

developments of transducers, which are the heart of these sensor devices, determine much of the progress in this field of chemical sensors and require a multidisciplinary team of expertise.

This article will review the practical procedures as applied to absorbance based chemical sensors and present some of the developments indicating their potential application in oceanography.

Optical Fibers

Almost all optical sensors employ optical fibers to transmit light to and from the chemical transducer. Optical fibers consist of a core of certain refractive index (n_1), surrounded by a cladding material of lower refractive index (n_2). Most optical fibers are then covered with a protective jacket that has no influence on the wave-guiding properties of the optical fiber (**Figure 1**). Incident light is transmitted through the fiber when it interacts the cladding interface at an angle greater than the critical angle, so that it is totally internally reflected at the core-cladding interface. The optical fiber light transmission characteristics are described by its numerical aperture (NA), which is directly proportional to the sine of the half angle (α) of the acceptance cone of light entering it which, in turn, is related to n_1 and n_2 , as in eqn [1].

$$\text{NA} = n_0 \sin \alpha, \quad \text{and} \quad \frac{(n_1^2 - n_2^2)^{1/2}}{n_0} \quad [1]$$