# Natural Climate Variability

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*Climate varies on all time scales, from one year to the next, as well as from one decade, century or millennium to the next. The complex nature of this variability is a major obstacle to the reliable identification of global changes brought about – in the past, present or future – by the presence and activities of humanity on this planet.*

*This article discusses observed climate variability on various time scales, paleoclimatic, interdecadal to centennial, seasonal to interannual, and intraseasonal. Key ideas to explain the observed variability are presented, and some of the models used to explore these ideas are mentioned. Finally, we discuss briefly the implications of these ideas for the detection and prediction of climate change.*

Climate variations happen on all time scales, as well as on all spatial scales, from the regional to the global. As an example, the oceanic El Niño (*see* **El Niño and La Niña: Causes and Global Consequences**, Volume 1) phenomenon is most pronounced in the tropical Pacific off the coast of Peru, but the associated Southern Oscillation in the atmosphere has far-reaching, nearly global implications. The combined El Niño/Southern Oscillation (ENSO) (*see* **El Niño/Southern Oscillation (ENSO)**, Volume 1) variability has a seasonal component, hence the name El Niño, due to its preference for appearing first at Christmas time. It also has a quasi-biennial component, with a characteristic recurrence time of 2–2.5 years, and a low-frequency one, with a recurrence of 4–5 years. The interaction of these three distinct modes of variability renders the evolution of sea-surface temperatures across the tropical Pacific fairly irregular. An additional cause of the irregularity observed in these temperatures lies in the frequent weather perturbations that affect the ocean's surface.

Natural climate variability includes in general, as it does in the ENSO example above, three types of phenomena.

1. Variations that are directly driven by a purely periodic external force, like the diurnal or the seasonal cycle of insolation, are the easiest to understand and predict. The diurnal variations are due to the rotation of Earth on its axis, the seasonal ones to its revolution around the Sun. They bring about the temperature and precipitation variations between day and night and between summer and winter, respectively. On much longer time scales, multiply periodic (often called quasi-periodic) variations in Earth's orbit around the Sun affect the intensity of the solar radiation that reaches the Earth's surface.
2. Variations due to the non-linear interplay of feedbacks within the climate system are harder to understand and to predict than the previous ones. For instance, a temperature drop within the system will increase the amount of snow and ice, and thus lead to further cooling; this is the so-called ice-albedo feedback, explored in some detail by energy-balance models (*see* **Energy Balance Climate Models**, Volume 1). On the other hand, the increase of trace-gas concentrations in the atmosphere, such as that of carbon dioxide ($CO_2$), will increase surface temperatures through the greenhouse effect (*see* **Projection of Future Changes in Climate**, Volume 1). This temperature increase will release even greater amounts of trace gases from the upper ocean or, on the contrary, trap them in terrestrial vegetation. Each climate feedback can enhance or countervail the effect of another feedback (*see* **Climate Feedbacks**, Volume 1). The distinct feedback mechanisms identified so far in the climate system are numerous and complex. Their number and complexity contribute significantly to the difficulty of reliably detecting human-induced climate change (*see* **Climate Change, Detection and Attribution**, Volume 1).
3. Variations associated with random fluctuations in physical or chemical factors are hardest to predict in detail for any length of time. These factors can be external to the climate system itself, such as aerosol loading due to volcanic eruptions (*see* **Volcanic Eruptions**, Volume 1). They can also be internal to the system, such as weather fluctuations. The latter are known to be unpredictable on the longer time scales of seasons to millennia. Still their averaged effect over these time scales may result in heat-transport variations between the equator and the poles.

The periodically driven variations – in the absence of any other source of variability – are themselves purely periodic and thus highly predictable. A single sudden change imposed on the system, whether in the solar irradiance received at the top of the atmosphere, in the atmosphere's reflectivity or opacity, or in other forcings or parameters of the system, has a relatively simple effect, conceptually speaking, on such a periodic variation.

A sudden impulse, as described in the previous paragraph, can change, for example, either the mean about which hemispheric temperature oscillates or the amplitude of the oscillation or both. Such a change would thus be easily detectable, speaking again in relative terms, when compared with one that a sudden impulse would cause in the other two types of variations, those resulting from the interplay of non-linear feedbacks (Lorenz, 1963) or those produced by stochastic forcing (Hasselmann, 1976). The characteristics of the latter two types of variations may be affected by an imposed change in diverse and complicated ways, ways that may be hard to distinguish from those of a spontaneous change within the system (*see* **Chaos and Predictability**, Volume 1).

An artist's rendering of climate variability on all time scales is provided in Figure 1(a). It is meant to summarize our knowledge of the relative power $S$, i.e., the amount of variability in a given frequency band, between $f$ and $(f + \delta f)$; here frequency $f$ is the inverse of period $T$, $f = 1/T$, and $\delta f$ indicates a small increment. This power spectrum is not computed directly by spectral analysis from a time series of a given climatic quantity, such as (local or global) temperature; indeed, there is no single time series that is $10^7$ years long and has a sampling interval of hours, as the figure would suggest.

Instead, Figure 1(a) represents a composite of information obtained by analyzing the spectral content of many different climatic series. The figure reflects the three types of variability mentioned earlier: sharp lines that correspond to periodically forced variations, at one day and one year; broader peaks that arise from internal modes of variability; and a continuous portion of the spectrum that reflects stochastically forced variations, as well as deterministic chaos. The latter represents the irregular variations that result from the deterministic interplay of non-linear feedbacks.

Between the two sharp lines at one day and one year lies the synoptic variability of mid-latitude weather systems, concentrated at 3–7 days, as well as intraseasonal variability, i.e., variability that occurs on the time scale of 1–3 months. The latter is also called atmospheric low-frequency variability. This name refers to the fact that the variability in question has lower frequency, or longer period, than the so-called baroclinic instability of large-scale atmospheric flow that gives rise to the development of weather systems. The periods associated with intraseasonal variability exceed even the duration of these weather systems' complete life cycle, from their birth in the storm tracks off the east coasts of the major landmasses to their decay further to the east, across an entire ocean basin. Intraseasonal variability comprises phenomena such as the 40–50-day Madden-Julian oscillation of winds and cloudiness in the tropics, as well as the alternation between episodes of zonal and blocked flow in mid-latitudes. Both of these phenomena involve exchanges of angular momentum between the atmosphere, the oceans and the land: as the winds speed up, the Earth rotation slows down, and vice-versa (Ghil and Childress, 1987; Ghil and Robertson, 2000).

Immediately to the left of the seasonal cycle in Figure 1(a) lies interannual, i.e. year to year, variability. This variability is dominated by ENSO-related phenomena and involves the interaction of the seasonal cycle with internal modes of variability of the ocean-atmosphere system in the tropical

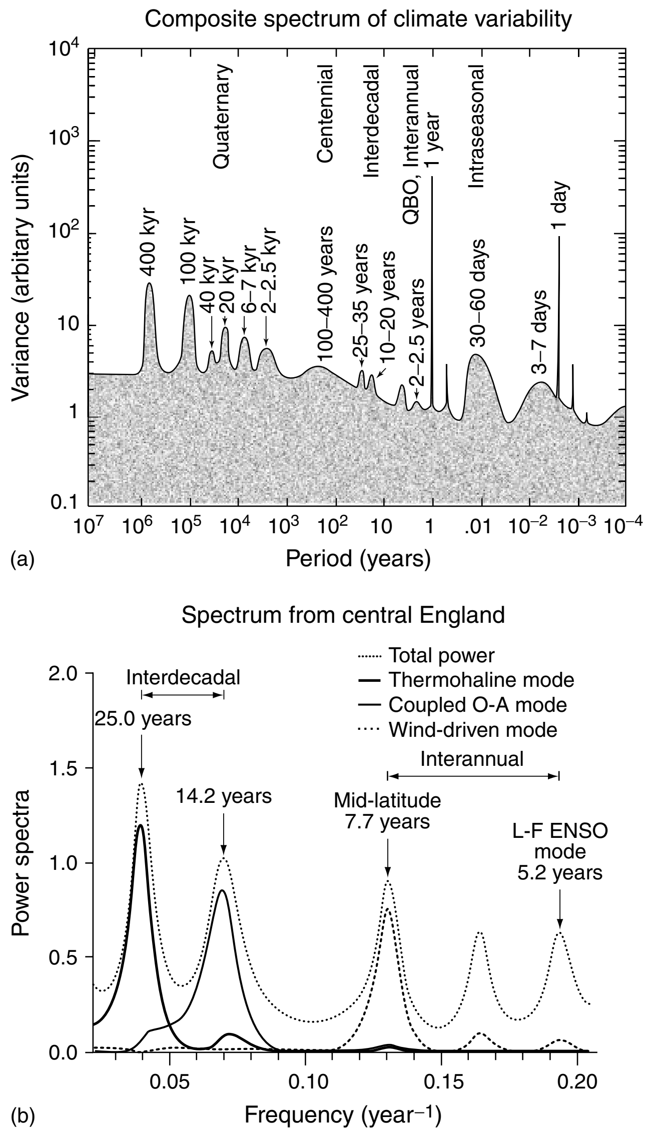


**Figure 1** Power spectra of climate variability. (a) Composite spectrum over the last 10 Myr (1 Myr = $10^6$ year; 1 kyr = $10^3$ year). (b) Spectrum of the Central England record; physical causes of the peaks are tentative (after Plaut *et al.*, 1995)

Pacific (Philander, 1990; Neelin *et al.*, 1998). Additional forms of interannual and interdecadal variability in higher latitudes are associated with the North Atlantic Oscillation, the Pacific Decadal Oscillation, and the Arctic Oscillation (*see* **Arctic Oscillation**, Volume 1; **North Atlantic Oscillation**, Volume 1; **Pacific–Decadal Oscillation**, Volume 1).

The emphasis of climate research in the second half of the 20th century has shifted farther and farther to the left in the composite spectrum of Figure 1(a). It has proceeded from the study of weather systems in the 1950s and 1960s to that of intraseasonal variability in the 1970s and 1980s and on to interannual variability in the 1980s and

1990s. The greatest excitement – among scientists as well as the public – is currently being generated by interdecadal variability, namely climate variability on the time scale of a few decades, i.e., the time scale of an individual human's life cycle.

This progression of interest towards longer and longer time scales is due to two complementary causes. First, the non-linear feedbacks between different components of the climate system that affect the longer-term interactions pose a greater challenge to the scientific community. Second, the greater scientific challenge goes hand in hand with humanity's increasing desire to predict and eventually control its environment farther and farther into the future (National Research Council, 1995).

Clearly, controlling our environment is predicated upon the ability to understand and predict it first. Thus, weather prediction can be performed with the same accuracy now for 3–5 days as it was for 12–24 hours in the 1960s, due to the very substantial progress in understanding and modeling weather phenomena over the last 40 years. We have learned to carry out predictions of certain climate variables over certain areas, with some useful skill, for up to six months in advance. This skill is a result of considerable advances in understanding ENSO and modeling coupled ocean-atmosphere interactions over the last 20 years. It is reasonable, therefore, to attack now the even more challenging problems of understanding and predicting climate variability for the next decade and century.

Prediction is greatly facilitated by regularity. Thus, the prediction of sea-surface temperatures on seasonal to interannual time scales is helped by the near-repetition of warmer temperatures in the eastern tropical Pacific every boreal winter, as well as roughly every second year, and every fourth or fifth year. The positive interference of the quasi-biennial mode (2–2.5 years) with the low-frequency mode (4–5 years) yields major warm and cold events, but the episodes of mutual reinforcement between modes still occur rather irregularly. This irregularity precludes, at present, greater certainty in predictions for 6 months or longer (Neelin *et al.*, 1998).

The power spectrum shown in Figure 1(b) is based on the longest instrumentally measured record of any climatic variable, the 335 year long record of Central England temperatures. It represents an up-to-date blow-up of the interannual to interdecadal portion of Figure 1(a). The broad peaks are due to the climate system's internal processes: each spectral component can be associated, at least tentatively, with a mode of interannual or interdecadal variability. Thus the rightmost peak, with its period of 5.2 years, can be confidently attributed to the remote effect of ENSO's low-frequency mode. The 7.7-year peak is currently believed to capture a North Atlantic mode of variability that arises from the Gulf Stream's interannual cycle of meandering and intensification (Moron *et al.*, 1998; Speich *et al.*, 1998). The two interdecadal peaks, near 14 and 25 years, are also present in global records, instrumental as well as paleoclimatic. They seem to be associated with oscillatory modes in the global oceans' thermohaline circulation (*see* **Ocean Circulation**, Volume 1) and its coupling to the atmosphere above (Ghil and Robertson, 2000).

Finally, the leftmost part of Figure 1(a) represents paleoclimatic variability. The information summarized here comes exclusively from proxy indicators of climate. These indicators include coral records and tree rings for the historic past, and marine-sediment and ice-core records for the last two million years of Earth history, known as the Quaternary. During the Quaternary era, large ice sheets were present on the planet, especially in its Northern Hemisphere. This era is therefore termed an ice age, when compared to such ice-free epochs of the past as the Cretaceous.

Ice ages occupy on the whole no more than about one tenth of documented Earth history. Their main interest lies in the fact that natural climatic variability is higher during an Ice Age than during more benign geological times. The Quaternary represents an alternation of warmer and colder episodes, called glaciation cycles. This cyclicity is manifest in the broad peaks present in Figure 1(a) between roughly 1 kyr and 1 Myr. Of these, the three peaks near 20 kyr, 40 kyr and 400 kyr reflect quasi-periodic variations in Earth's orbit. These orbital variations (*see* **Orbital Variations**, Volume 1) are due to the perturbing forces exerted by Jupiter and the other planets on Earth's otherwise purely periodic, Keplerian motion around the Sun. They represent, respectively, variations in precession, obliquity and eccentricity, three parameters that are used to describe Earth's secularly varying orbit.

The large 100-kyr peak continues to puzzle paleoclimatologists. It has been attributed variously to an internal mode of the climate system, to the system's resonant response to small eccentricity variations or to a combination tone between orbital frequencies (Ghil and Childress, 1987). In the latter theory, the dominant 100-kyr peak corresponds to the non-linearly resonant amplification of a difference tone between the two distinct precessional lines at 19 kyr and 23 kyr. Certain climate simulations from the geological past attribute a critical role to natural changes in $CO_2$ levels, both in the initiation of the Quaternary as a whole and in generating the 100-kyr peak during the Late Pleistocene, i.e., the last one million years (*see* **Climate Model Simulations of the Geological Past**, Volume 1).

The peak at 6–7 kyr was predicted theoretically in the early 1980s as an internal oscillatory mode due to the interplay between the positive ice-albedo feedback and the negative precipitation-temperature feedback. In this theory, when coupling an energy-balance model with an ice-sheet

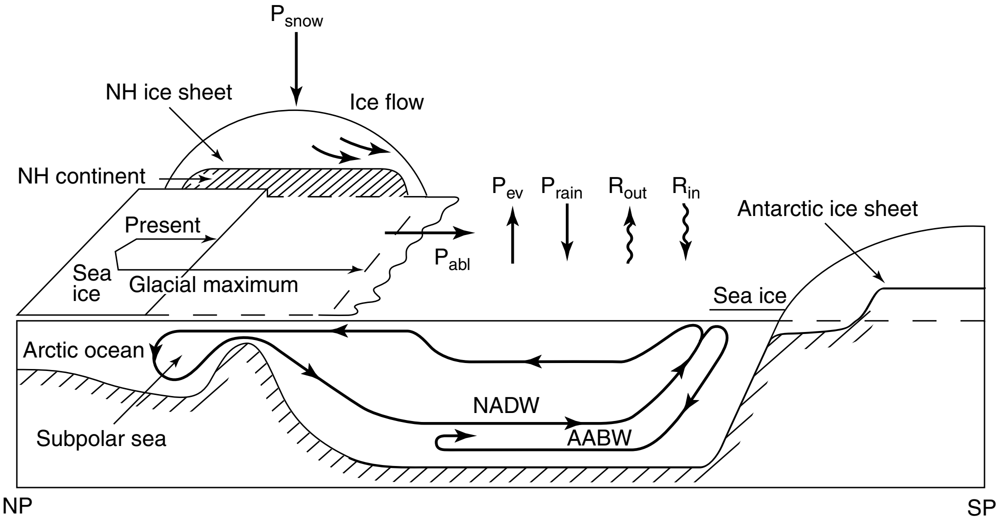


**Figure 2** Schematic diagram of the Atlantic Ocean's thermohaline circulation and of the radiative, hydrologic and cryospheric processes likely to affect it (after Ghil *et al.*, 1987)

model, global temperature drops as ice mass increases. The temperature drop diminishes the evaporation in low latitudes and hence the ice accumulation in high latitudes; this in turn decreases the ice extent and allows temperatures to rise again. Heinrich events (*see* **Heinrich (H-) Events**, Volume 1) were discovered in the late 1980s and found in the 1990s to have a near-periodicity of 6–7 kyr, as predicted by this theory (Ghil, 1994).

The peak at 2–2.5 kyr is associated with the names of W Dansgaard and H Oeschger and was discovered in Greenland ice cores (*see* **Dansgaard–Oescheger Cycles**, Volume 1). It seems to arise from the alternative intensification and weakening of the Atlantic's thermohaline circulation, as represented in Figure 2. Both the region of formation of North Atlantic Deep Water (NADW) and its flux (i.e., its mass transport per unit time) vary in a fairly irregular manner, but with a millennial-scale mean periodicity. They are associated with variations in sea-ice cover, as well as in temperature and precipitation over the adjacent land areas.

What are the implications of the natural variability described so far on our understanding of global climate change? If no variability whatsoever were present, i.e., if the climate system were in equilibrium, external changes would result in a simple shift of this equilibrium (see Figure 3a). The ratio of the amount of this shift, in global mean temperature say, to the equivalent change in irradiance at the top of the atmosphere say, is usually defined as climate sensitivity (*see* **Climate Sensitivity**, Volume 1). The actual forcing change under consideration might be in carbon dioxide concentration or in atmospheric opacity due to aerosol loading or in the mean distance between Earth and Sun (see Box 1 for details).

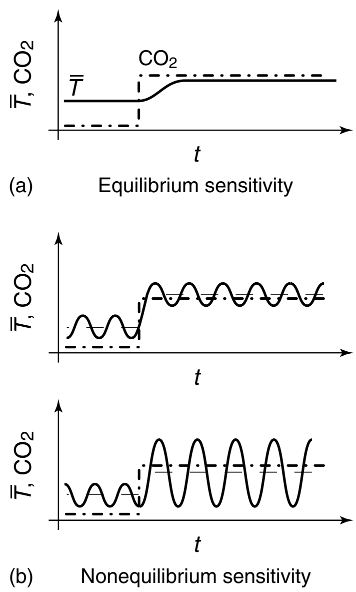


**Figure 3** Climate sensitivity for (a) an equilibrium model and (b) a nonequilibrium, oscillatory model: as a forcing parameter (atmospheric $CO_2$ concentration, dash-dotted line) changes suddenly, global temperature (heavy solid) undergoes a transition. In (a) only the mean temperature (still heavy solid) changes; in (b) the oscillation's amplitude can also decrease (upper panel), increase (lower panel) or stay the same, while the mean (light dashed) adjusts as it does in panel (a)

Figure 3(b) shows how the response of the system would be modified if the system were undergoing periodic oscillations, either as the result of purely periodic forcing or due to an internal mode that is purely oscillatory. In this case, a shift in the mean would be accompanied by a change in the

**Box 1** Computations of the climate's equilibrium sensitivity are often based on the linear, scalar ordinary differential equation

$$c\dot{\theta} = \lambda\theta + Q \quad (1)$$

where $\theta$ is global temperature and

$$\dot{\theta} \equiv \frac{d\theta}{dt}$$

is its rate of change with time $t$.

Here

$$\lambda = \sum \lambda_i \quad (2)$$

is the sum of the (linear) feedbacks $\lambda_i$, positive or negative, and

$$Q = \sum Q_j \quad (3)$$

is the sum of the (positive or negative) forcings $Q_j$, also called source (or sink) terms (see Ghil and Robertson, 2000, and references therein). Typically, $Q_j$ might include absorption by greenhouse gases (positive forcing) and reflection by aerosols (negative forcing).

If climate change during the industrial era were truly governed by simple linear mechanisms, according to Equations (1–3) above, it would be fairly smooth. Provided that only gradual increases in aerosols and greenhouse gases were at work, it would look schematically like the light solid line in Figure 4. In fact, the observed temperature variations in the Northern Hemisphere looked, again schematically, like the heavy solid line in the figure. The difference between the two curves, light and heavy, is due to the presence of natural variability, as well as to that of variable natural forcings, like volcanic eruptions and solar-irradiance variability.

This difference implies that the actual climate system's evolution is more complex than that dictated by a linear response to simple forcings. To simulate and predict this complex evolution requires the use of a set of non-linear (ordinary or partial) differential equations, depending on time $t$ and on a set of parameters $\mu$:

$$\dot{X} = F(X, t, \mu) \quad (4)$$

Here $X$ is a vector of state variables that describes the atmosphere, ocean and other climate subsystems. The components of the right-hand side vector $F$ give the rates of change of the components of $X$. The vector $F$ depends on the state $X$ itself, time $t$ (in the form of time-dependent forcing), and the parameter set $\mu$. This set includes but is not limited to those present in the former scalar equation; typical parameters are those that determine the strength of various (non-linear) feedbacks and the net insolation at the top of the atmosphere.

Gradual changes in $\mu$ can lead to sudden changes in system behavior, from steady state to purely and multiply periodic, and on to chaotic and fully turbulent; these changes are called bifurcations. As a result, the system's predictability can change dramatically.

oscillations' amplitude, increasing or decreasing it. In fact, for a non-linear system, a change in an external parameter can lead from the system's being in equilibrium to its undergoing self-sustained oscillations, i.e., as a result of global change the system might be destabilized and become oscillatory.

The climate system's behavior, however, is much more complicated than being in equilibrium or in a state of purely periodic oscillations. Thus, the effects of natural or anthropogenic changes in the system's forcing or parameters, including but not restricted to net insolation, cannot be measured by a single quantity like climate sensitivity. Resonances may lead to the amplification of certain oscillatory modes and the entire behavior of the system may change, becoming more or less predictable. The situation we are confronted with is illustrated schematically in Figure 4. The figure highlights the discrepancy between a simple response to human-induced forcings and the temperature variability recorded during the last century-and-a-half (see Box 1 for explanations).

A simple example of the difficulties in assessing human-induced climate change in the past, and hence predicting it for the future, is given in Figure 5. Valley glaciers in France's Chamonix Valley, and elsewhere in the Alps, have been retreating recently. This retreat, however, is far from unprecedented, as the figure shows. Is the present retreat reversible – as it was in the latter part of the 18th century or the early part of the 20th – or is it not?

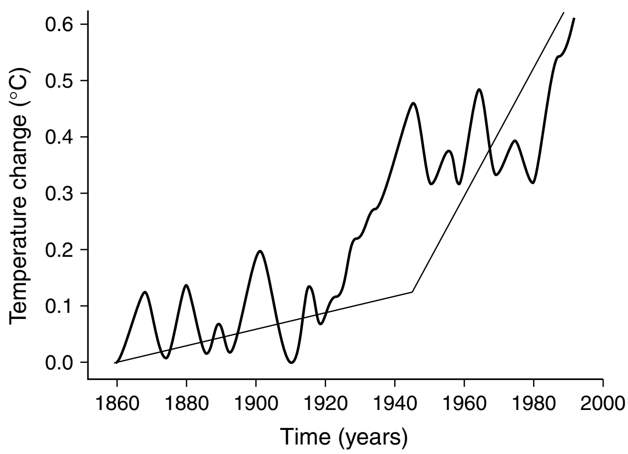


**Figure 4** The role of natural variability in climate-change detection and attribution: equilibrium response of global temperatures to changes in aerosols and trace gases (light solid line) vs. observed temperature variations (heavy solid)

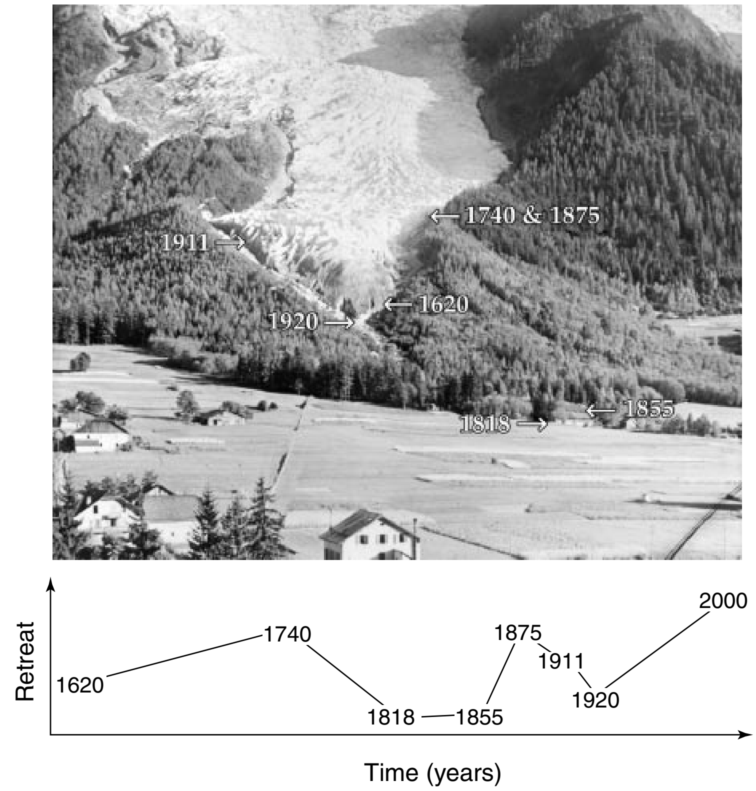


**Figure 5** Retreats and advances of an alpine glacier over the last 300 years (upper panel courtesy of Musée Alpin–Les Amis du Vieux Chamonix)

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