C

CARBON-14 DATING

The introduction of the carbon-14 (^{14}C, radiocarbon) dating method in 1947 (for which Willard F. Libby received the Nobel Prize for Chemistry in 1960) transformed many aspects of environmental science by permitting numerical dating of fossils, artifacts and deposits whose age previously had to be estimated. Organisms and events could now be put into chronological order and correlated objectively, and the search for mechanisms of change placed on a sounder footing, leading to a better understanding of such matters as the viscosity of the Earth's mantle, the mechanisms of climatic change, processes of organic evolution and extinction, and climatic history, within the 70 000 or so years spanned by the method.

Three isotopes of carbon are present in the atmosphere in the ratio 100 : 1 : 0.01, of which two, ^{12}C and ^{13}C, are stable. The third, ^{14}C, is radioactive and thus subject to decay, but it is continually replenished by the action of cosmic rays, which interact with ^{14}N atoms in the upper atmosphere to form ^{14}C. The radiocarbon is oxidized to form CO_2, which is then incorporated into plants by photosynthesis or dissolved in the ocean and used to build carbonate structures by mollusks and corals. The current estimate of the half life ($t_{1/2}$) of ^{14}C is 5730 $\pm$30 years, but Libby's original value of 5568$\pm$30 is used in many date lists for consistency. The "Libby" age can be adjusted to the new value by multiplying by 1.03. His use of BP for Before Present (= 1950) also persists.

Originally Libby analyzed his samples as solid carbon. Nowadays in most radiocarbon laboratories ^{14}C content is measured by converting the sample into CO_2, whose radioactivity relative to a modern standard is counted in a gas-proportional counter or by synthesizing benzene (C_6H_6) from the gas and using a liquid scintillation counter. A third method, accelerator mass spectrometry (AMS), allows ^{14}C atoms to be counted directly and thus requires much smaller samples, typically 15 mg as opposed to 1 g of carbon. The $\pm$ value that follows the age is generally a statement of the counting error at 1 s.d., but some laboratories include analytical and other error estimates.

The method can be used for any organic material but some substances have proved less troublesome than others. Wood, charcoal and peat, suitably pretreated, are often favored, but bone collagen and unrecrystallized shell and coral yield reliable ages. Besides exercising great care in field attribution and handling, the collector can check the sample for contamination by old or young carbon by means of microscopy (both optical and SEM), X-ray diffraction and the $^{13}C/^{12}C$ (stable isotope) ratios measured by mass spectrometry. Corrections need to be made for ^{14}C contributed to the atmosphere by thermonuclear weapons testing after 1952 and by dead CO_2 produced from the burning of coal and oil fuels (the Suess effect).

Variations in the ^{14}C reservoir (the de Vries effect) can be allowed for by reference to calibration curves based on tree ring ages for sequoia (*Sequoia gigantea*) and bristlecone pine (*Pinus aristata*), which are available for the last 8000 years. The main sources of variation are the intensity of the cosmic ray flux, the strength of the Earth's magnetic field and changes in the Earth's carbon reservoir stemming from climatic changes. The solution of parochial dating errors is thus proving a source of information on environmental change both on Earth and on the sun.

Further information on carbon-14 dating can be gained from Bradley (1985, Ch. 3), Worsley (1981), and Raaen et al. (1968). Practical applications can be found in Ozer and Vita-Finzi (1986).

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Cross-references

Archeoclimatology
Climatic Variation: Historical
Tree-Ring Analysis

CENTERS OF ACTION

If there were no continents and seasonal variation on the Earth, an ideal atmospheric circulation system would be denoted by: (1) an equatorial belt of low pressure; (2) polar centers of high pressure; and (3) two intermediate belts, one of high pressure (the Horse Latitudes) at about 30 degrees and the other of low pressure in the vicinity of 60 degrees. The low-pressure zones would be centers of convergence and the high-pressure zones centers of divergence. Since the Earth is rotating, wave patterns would develop along the convergences and these would become the great storm-generating belts. However, the critical modification in this ideal pattern is the asymmetric presence of continents; meridional land masses and mountain belts block the 60 degree convergence zone in the northern hemisphere, but in the southern hemisphere this is a clear waterway – the Southern Ocean – all around the Earth.

Due to the different thermal characteristics of continents and oceans, semipermanent centers of high and low pressure tend to build up and remain along the zonal (east–west) pressure belts, with greater contrasts in the northern hemisphere as it contains most of the world's land mass. Over the oceans the sign of the pressure extremes remains constant in each pressure mass, but over continents it tends to alternate in summer (low) and winter (high), the so-called monsoon effect. Since the surface winds diverge from high-pressure areas, the semipermanent highs are known as source regions for airmasses. These semipermanent or regularly recurring pressure centers are known, as Rossby (1945) says, "by the somewhat misleading name" of centers of action. It is important to recognize that these centers are the result and not the cause of global circulation.

The semipermanent centers of action in the northern hemisphere are: (1) low pressure: Aleutian Low, Icelandic Low; and (2) high pressure: Azores or Bermuda High, Hawaiian or North Pacific High (Figure C1).

The continental highs in the northern hemisphere winter are the North American or Canadian High and the Siberian or Asiatic High; in summer these tend to be replaced by lows, but are more complicated by the zonal circulation, especially in the case of the smaller land mass of North America.

With the exception of the South Polar High over Antarctica, comparable centers of action are not as well developed in the southern hemisphere, though well-defined oceanic high-pressure centers occur: the South Atlantic or St Helena High, the Indian Ocean High, and the South Pacific High. In summer, lows develop over the continents and are replaced by highs in the winter; this is similar to, but not so marked as, the Siberian Low and High. These differences occur due to the continuous waterbody of the Southern Ocean symmetrically surrounding Antarctica, which is itself symmetrical about the South Pole, and because the other three southern continents occupy only minor fractions of the total area, the steady circulation is striking. Indeed, Gentilli (1949) found that the maximum zonal index was 6 (i.e. well-marked westerlies) and there was never a negative (easterly) reading between 35°S and 55°S; in contrast, the index varied from +15 to −5 over a 12-month period in North America.

It is known that large-scale phenomena such as the El Niño/Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO) affect the climate over a large part of the Earth on time scales that range from years to centuries. Kirov and Georgieva (2002) compared the century-long variability of ENSO and NAO with variations of solar activity, and suggest that the influence of solar activity on these phenomena is mediated by the atmospheric centers of action which react by changes in intensity and location.

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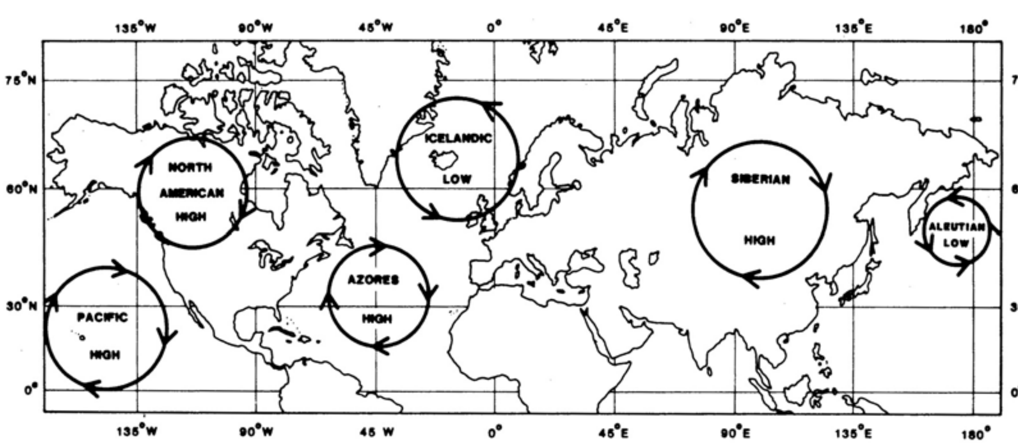

Figure C1 Idealized map showing locations of centers of action in the northern hemisphere.

Cross-references

Airmass Climatology
Aleutian Low
Atmospheric Circulation, Global
Azores (Bermuda) High
Icelandic Low
North American (Canadian) High
North Pacific (Hawaiian) High
Oscillations
Siberian (Asiatic) High
Teleconnections
Zonal Index

CENTRAL AMERICA AND THE CARIBBEAN, CLIMATE OF

Central America and the Caribbean span the deep tropics and subtropics. Because of the tropical maritime location temperature changes throughout the region are generally small, and rainfall is by far the most important meteorological element. In general the climate of the region is controlled by the migration of synoptic features, and the mean climate strongly reflects the annual cycle of these features.

The most dominant synoptic influence is the subtropical high of the north Atlantic. Subsidence associated with the spreading of the subtropical high from the north Atlantic to the north American landmass dominates during boreal winter, as do the strong easterly trades found on its equatorward flank. Coupled with a strong trade inversion, a cold ocean and reduced atmospheric humidity, the region is generally at its driest during the winter. With the onset of boreal spring, however, the subtropical high moves offshore and trade wind intensity decreases, with convergence characterizing their downstream. Especially for Central America, the variation in the strength of the trades is an important determinant of climate throughout the year. There is also a high and weak trade inversion, the ocean warms and atmospheric moisture is abundant. The region is consequently at its wettest in the northern summer half-year.

The contrast in summer and winter rainfall defines most climate classifications of the region. For example, Rudloff (1981) suggests that the climate of Central America and the Caribbean can be classified as dry-winter tropical. The contrast proves an important control for agricultural and tourism activity, water resource allocation, hydrological considerations and fishing – activities which are of utmost importance to the region.

The dry winter/wet summer regime, however, only *broadly* defines the climate of the region as orography and elevation are significant modifiers on the subregional scale. The region is one of complex and diverse topography including continental territories, island chains, and mountain ranges of varying orientations and elevations. The topography interacts with the large-scale circulation to produce local variations in the climate, including significant variations in annual rainfall totals, length of the rainy season, and the timing of maxima and minima. As examples, the windward slopes of the larger mountainous islands of the Greater Antilles have significantly higher rainfall totals than the smaller flatter islands in the eastern Caribbean Sea. Similarly, the Caribbean coastal stations of Honduras, Costa Rica and Panama possess a strikingly homogeneous rainfall regime (i.e. in stark contrast to the dominant dry winter/summer regime) due to strong interactions of low-level winds with the topography. The subregional variations make generalizations about the climate of the region difficult – a fact which must be borne in mind when considering the general overviews presented in the following sections.

Besides the subtropical high, other significant synoptic influences include: (a) the seasonal migration of the Intertropical Convergence Zone (ITCZ) – mainly affecting the Pacific side of southern Central America; (b) the intrusions of polar fronts of midlatitude origin (called "Nortes" in Spanish) which modify the dry winter and early summer climates of the northern Caribbean and north Central American; and (c) westward propagating tropical disturbances – a summer season feature associated with much rainfall especially over the Caribbean region.

Central America

Temperature

Temperature ranges during the course of the year in the Central America region are generally small. They may exceed 4°C to the north, from the Yucatan to Honduras, but drop to less than 2°C in the south and along the coastal zones. Maximum temperatures are dependent on altitude given the mountainous interior of the region. The common usage of the terms *tierra caliente* ("hot land"), *tierra templada* ("temperate land"), *tierra fria* ("cold land"), and *tierra helada* ("frozen land") suggests a longstanding recognition of the effect of altitude on temperatures. The terms loosely denote shifts in the mean temperature regime with increasing altitude, but their usage is relative (varies with location), with no consistent assignment of a temperature to elevation. Minimum temperatures similarly exhibit dependence on altitude, but also show the effect of the winter intrusions of cold air from North America, and can dip as low as 7°C.

Portig (see Schwerdtfeger, 1976) classifies the annual cycle of air surface temperature in Central America as tropical, predominantly maritime, with small annual changes and dependent on cloud cover and altitude. The dominant annual cycle of the Central American region (excepting the Atlantic coasts of Honduras and northern Nicaragua) is monsoonal, with highest temperatures occurring just before the summer rains (Figure C2). Temperatures are at their lowest during January, largely due to the cooling effect of the strong trade winds (Figure C3). Maximum temperatures occur during April and are associated with a decrease in trade wind strength, lower cloud coverture and higher values of solar radiation (Alfaro, 2000a). There is a temperature minimum in July which coincides with the onset of the midsummer drought (or in Spanish the *veranillos* or *caniculas* – see below). During this period the trade winds briefly increase in intensity (see again Figure C3), the subtropical ridge over the Caribbean intensifies, and there is a second minimum and maximum in cloud coverture and radiation respectively.

Precipitation

Mean annual rainfall totals vary over a wide range in Central America in keeping with the diversity in topographic conditions. Annual totals of less than 100 cm are typical of the plains of Guatemala, Honduras and northwestern Nicaragua, and portions of the Pacific coast of El Salvador, Honduras and

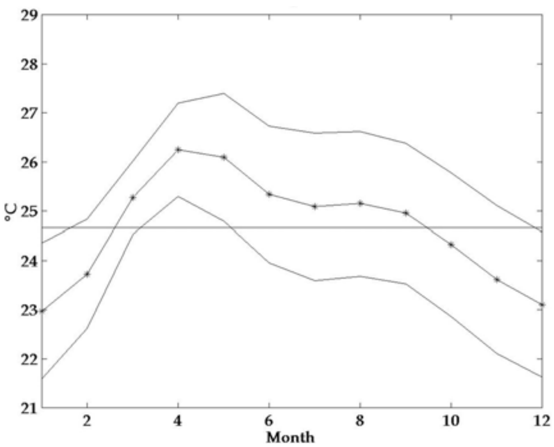

Figure C2 The central line (with asterisks) depicts the mean annual cycle of temperature over Central America as determined from principal component analysis. The upper and lower solid lines represent one standard deviation. The horizontal line is the annual mean of 24.7°C.

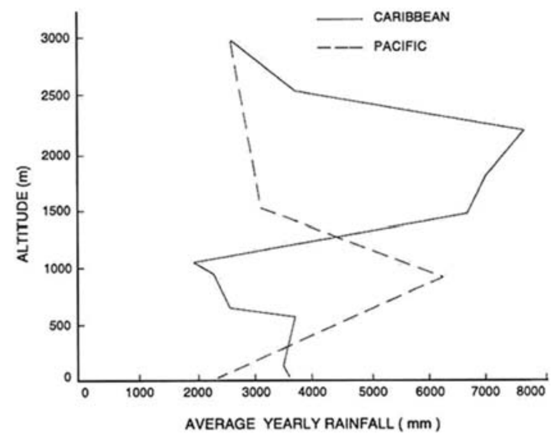

Figure C4 Distribution of average yearly rainfall with altitude on the Caribbean and Pacific sides of a topographic profile in Central America. The periods of analysis are 1966–1984 and 1981–1984 for the Caribbean and the Pacific sides respectively. (After Fernandez et al., 1996.)

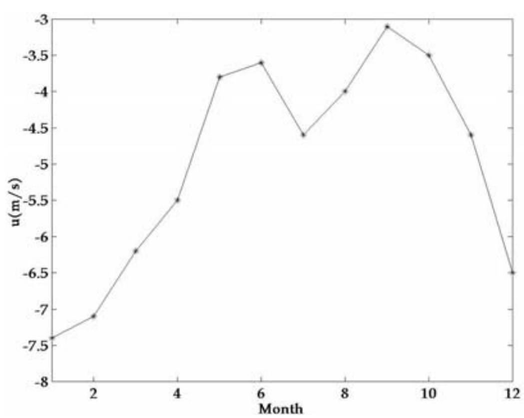

Figure C3 Annual cycle for zonal wind component (u) at the 900 hPa level. Monthly values are from Juan Santamaría synoptic station (10°00′N, 84°12′W, from Alvarado, 1999), period used: 1972–1989). The annual average value is $-5\,\mathrm{m\,s^{-1}}$. Notice that negative values are westward winds.

Panama. In contrast the northern and southern mountain ranges of Guatemala and the low mountains between Costa Rica and Panama receive large amounts of rainfall (in excess of 250 cm), as do the Atlantic coastal regions from Belize through Guatemala, the south coast of Costa Rica and a section of Panama's Atlantic coast. In general the Caribbean seaboard receives more rainfall than the Pacific side, reflecting the influence of tropical disturbances from the Caribbean Sea and the windward interaction of the trade winds with the mountains chains.

Fernandez et al. (1996) explored the variability of rainfall with altitude in Central America by examining the vertical rainfall distribution in Costa Rica along a topographic profile which crossed the country from the Pacific to the Caribbean coasts. The mountain profile, with highest peak of approximately 3000 m, is oriented parallel to the prevailing large-scale northeasterly trade winds. Their analysis of rainfall amounts and the seasonal and diurnal variations at 14 rain-gauge stations located on or close to the topographic profile reveals considerable variation with altitude. Maximum rainfall on both the windward (Caribbean) and leeward (Pacific) sides of the main mountain range occur at intermediate altitudes rather than on the mountain tops. (See also the work of Chacon and Fernandez, 1985.) Average yearly maxima of 7735 mm on the windward side and 6692 mm on the leeward side were observed at about 2000 m and 800 m respectively (see Figure C4).

Analysis of the mean annual cycle of precipitation reveals two dominant modes (Figure C5). The first and more representative mode is characterized by two rainfall maxima in June and September, an extended dry season from November to May, and a shorter dry season in July–August. Figure C5a suggests this regime as characteristic (with a few exceptions) of the entire Central America. This regime is largely explained by the seasonal migration of the subtropical north Atlantic high and the ITCZ.

The dry season of winter and early spring accounts for less than 20% of the annual precipitation total of Central America. It is more intense on the Pacific slopes of the isthmus, possibly due to an additional drying effect caused by the seasonal reversal of the winds on the Pacific side which blow offshore during winter. The ITCZ is also at its maximum southeast position during February and March. The season is further characterized by strong Atlantic trades (Figure C3), high values of total radiation (direct plus diffuse radiation) and sunshine hours in the low troposphere levels, despite the fact that radiation at the top of the atmosphere is at minimum. Cortez (2000) shows that there is no evidence of deep convection during the period as mean OLR values are greater than $240\,\mathrm{W\,m^{-2}}$, and that there is poor humidity convergence over almost all the isthmus excepting the southeast.

The onset of the rainy season is in May. There is a latitudinal variation of onset dates with earlier onset (early May) in the south and late onset (late May) in the north. The latitudinal variation may be partially explained by the northward migration

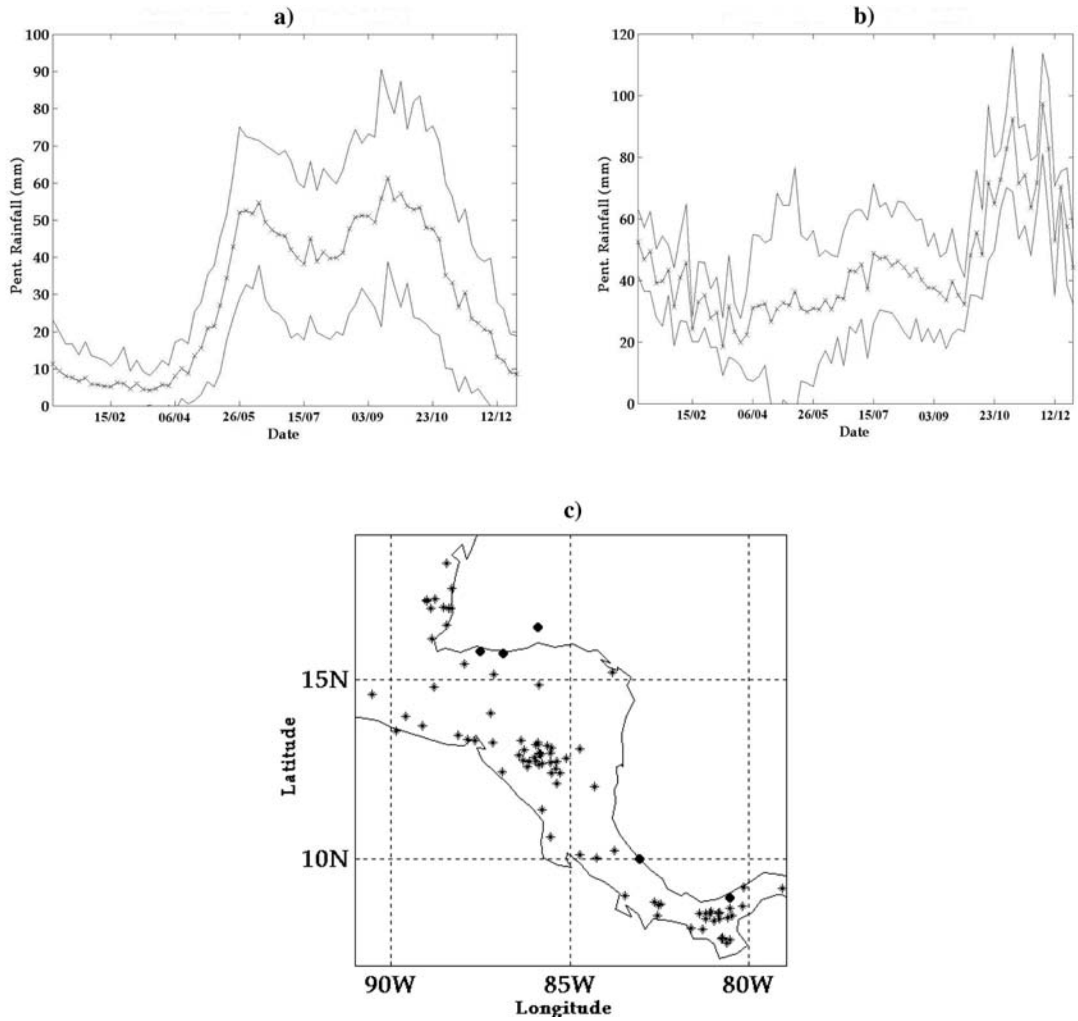

Figure C5 Lines with asterisks are the (**a**) first and (**b**) second pentad rainfall dominant annual cycle patterns, using empirical orthogonal function analysis of 94 gauge stations. Bands are given for one standard deviation (solid lines). (**c**) Asterisks (solid circles) are locations at which the first (second) mode dominates.

of the ITCZ that causes instability and humidity convergence in southern Central America from boreal late spring to early fall. As noted by Alfaro (2000b), however, the migration of the ITCZ cannot explain the generalized deep convection over the entire region during the rainy season as the ITCZ is not normally found at latitudes higher than 10–12°N. The first rainfall peak occurs in mid-June, and during this period (and for the entire rainy season), the trade winds are weak (Figure C3), allowing for the formation of mesoscale systems such as sea breezes, though mostly on the Pacific side.

By the middle of July there is a distinct decrease in rainfall over much of Central America. The brief dry period is known as the *veranillo, canícula,* or the midsummer drought (MSD, see Magaña et al., 1999). Again there is some latitudinal variation in the onset dates of the MSD, with earlier onset in the Southern-Pacific slope and later onset in the Northern-Caribbean slope. During the MSD there is an increase in the trade wind intensity associated with a brief retrogression of the north Atlantic subtropical high.

The second rainfall maximum occurs in mid-September/October, and is wetter than the first. The subsequent demise of the rainfall season also shows latitudinal variation with the tendency for earlier end dates in the north Pacific sector and later end dates in the south. This latitudinal behavior is only partially explained by the southward migration of the ITCZ, since latest end dates are found on the north Caribbean coast. The reason for such late end dates in the north is unclear. It could reflect the influence of cold fronts over the North-Caribbean Central American coast during boreal winter and/or the intensification of the winter trade winds which transport humidity and produce precipitation on the windward side.

The Caribbean coasts of Honduras, Costa Rica and Panama (Figure C5c) exhibit a second rainfall mode, different from the dominant dry-winter/wet-summer regime. It is difficult to define a dry season for this region as rainfall is nearly homogeneous between January and the middle of October, with typical accumulations of between 180 and 300 mm per month. The period accounts for approximately 60% of the annual total in this region. From mid-October onwards there is a marked increase in precipitation accumulation until the end of the year, with a maximum in the beginning of December. This is related to a maximum in humidity convergence at the end of the year in the southeastern region. Significantly, there is no rainfall minimum in July, with instead a relative maximum characterizing the mode 2 regions. This might result from the low-level jet influence and the interaction between topography and the low-level wind (Amador, 1998). Interestingly it is the same mechanism (i.e. the intensification of the trades) which both yields the MSD over most of Central America and makes the Caribbean slope wetter than the Pacific slope during July.

The Caribbean

Temperature

Temperatures on the islands of the Caribbean remain fairly constant throughout the year, with a small annual range of only 2–7°C (see Figure C6). As opposed to the dominant monsoonal type variation of Central America, the annual cycle of temperature for the Caribbean exhibits a summer–winter variation similar to that of more northern latitudes, but with high tropical temperatures and in the context of the small range. In the mean, temperatures steadily increase from May and peak in August–September. The summer months are the warmest with mean temperatures peaking in the upper 20 s Celsius, but rarely exceeding 37°C for any station. For the smaller islands of the Lesser Antilles, the summer heat is tempered by the easterly trades.

Coolest temperatures occur during boreal winter and early spring (December to April) with values generally in the low 20 s for sea-level and coastal stations. Temperatures, however, never generally fall below 18°C. The summer–winter variation becomes more pronounced for territories at increasingly higher latitudes, which also come under the occasional influence of intruding polar fronts in winter. Inland, on the larger islands of the Greater Antilles, the cooling influence of the sea breeze is lost, though in some cases this is compensated for by increasing altitude.

Precipitation

The differences in size, shape, topography, and orientation with respect to the trade winds greatly influence the amount of rainfall received by the island territories of the Caribbean. The larger and more mountainous islands of the Greater Antilles (Cuba, Jamaica, Hispaniola, Puerto Rico) receive rainfall amounts of up to 160 cm per year, with in excess of 500 cm on the highest peaks. There is, however, a distinct rainshadow effect on their southern coasts, which are distinctly arid. To their east the northern Windward Islands (as far as Antigua) receive lower rainfall amounts, tending to be slightly drier. Precipitation increases with southerly latitude and the Windward Islands, Trinidad and Barbados tend to be well watered. Normal yearly rainfall is between 180 and 230 cm for the Windward Islands (from Montserrat through Grenada), 150 cm for Trinidad, and 125 cm for Barbados. The dry belt of the Caribbean is found over the southwestern islands of the Netherland Antilles (Aruba, Bonaire, Curacao), which possess a semiarid climate and receive less than 60 cm per year on average. In general the eastern sides and windward slopes of the Caribbean islands receive more rainfall because of the prevailing northeasterly trade winds, while the lee slopes and interior portions of the mountainous islands are drier.

Except for the Netherland Antilles, the Caribbean islands generally receive most of their yearly rainfall in the summer months (Figure C6). This reflects the influence of the migrating north Atlantic subtropical high, the warming of the Caribbean Sea, a reduction in trade wind strength, and the appearance in mid-June of easterly waves/tropical disturbances which leave the west coast of Africa and traverse through the Caribbean region. These easterly waves are convection centers carried by the trades that frequently mature into tropical storms and hurricanes under conducive atmospheric and oceanic conditions such as warm sea surface temperatures and low vertical shear. They represent the primary source of Caribbean rainfall, and their onset in June and demise in early November roughly coincide with the mean Caribbean rainy season. Unlike Central America, the southernmost Caribbean territories lie too far north (~11°N) to come under the direct influence of the ITCZ.

The northwestern Caribbean territories (Jamaica, eastern Cuba, and Hispaniola) possess a distinct midsummer break in rainfall similar to the MSD of Central America (see Figure C6). Onset, however, tends to be later than in Central America, with the MSD for the Caribbean territories occurring in mid to late July/early August. Consequently the annual cycle of rainfall for the northwest Caribbean is bimodal with an initial peak in May/June and a second greater maximum in October. The islands of the Lesser Antilles generally lack the earlier maximum, with rainfall totals increasing from the onset of the rainy season in May to a late fall rainfall peak. The eastern side of Dominica and Martinique has one long rainfall season with a peak in November, while Barbados has a broad minimum in February and March, and a broad maximum between September and November. This lack of an early rainfall season peak in the eastern Caribbean might be accounted for by the fact that only the western Caribbean Sea is warm enough to support convection (i.e. exceeds the 27°C convection threshold) at the onset of the rainy season (Wang and Enfield, 2001, 2003), and it is only the southwestern Caribbean which is conducive to tropical cyclogenesis at the start of the rainy season.

A slightly different rainfall regime characterizes the flat islands of the Netherland Antilles, which have their main rainy season from October through January, with the "small rains" from June to September. The islands are at their driest between February and May, with an absolute minimum in April.

Outside of the rainy season, rainfall in the Caribbean dry season can occur from the southern tip of intruding cold fronts form North America. The influence of the fronts is particularly noticeable in northern Caribbean territories (especially the Bahamas) but rarely extends below Jamaica or east of Puerto Rico.

Hurricanes

Tropical storms and hurricanes are seasonally common in the northern Caribbean and Gulf of Mexico. They merit brief discussion due to the significant loss of life, the extensive infrastructural damage, and the disruption to the Central American and Caribbean economies and way of life that occur with hurricane landfall or even passage nearby. In the mean eight hurricanes will pass near or through the Caribbean region in a year, but this number can vary significantly from year to year. Global climatic phenomena such as the El Niño Southern

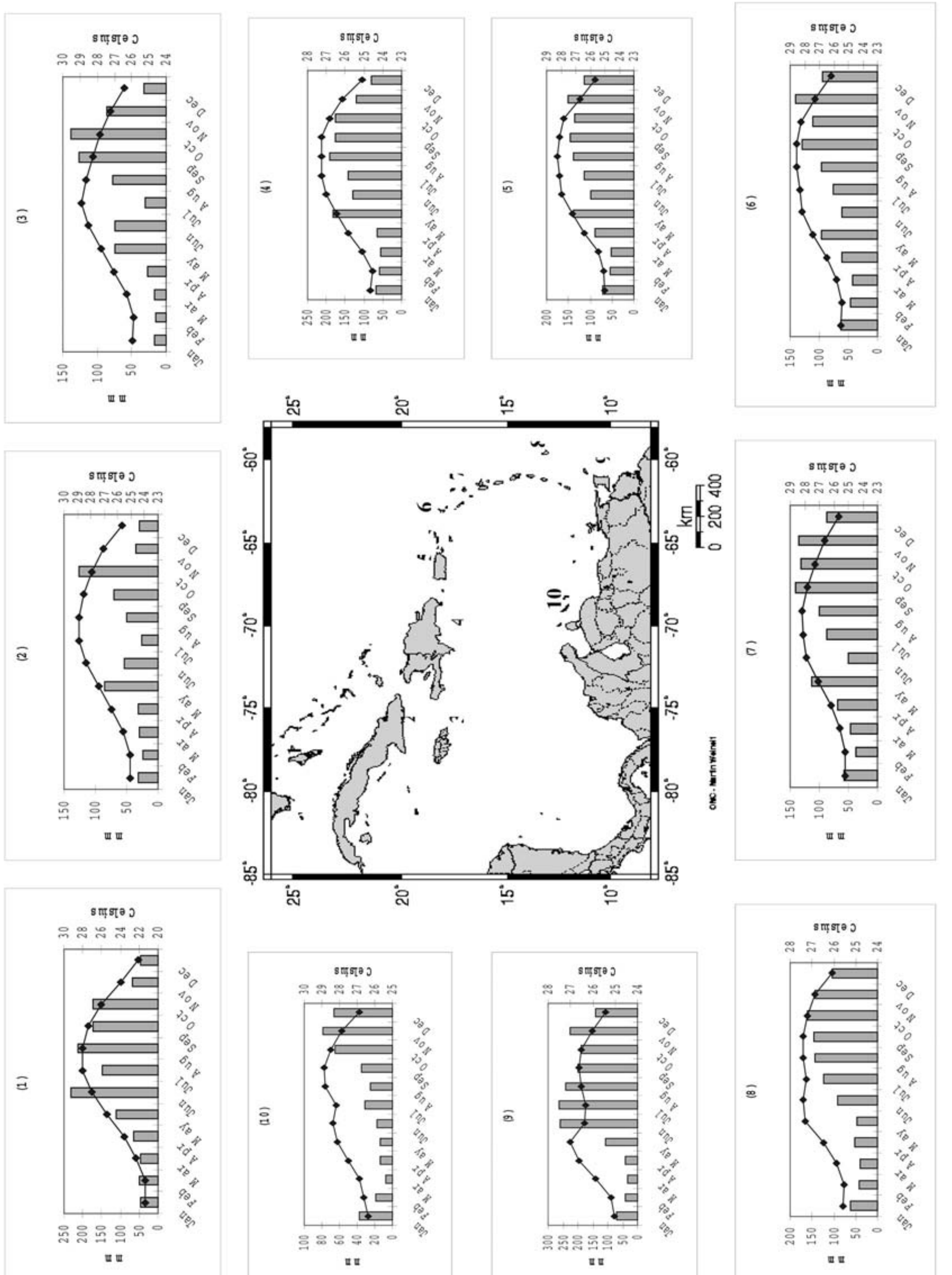

Figure C6 Precipitation and temperature climatology for 10 selected stations in the Caribbean. See Table C1 for station description. Bars denote precipitation in mm. The line graph denotes variation in temperature in °C.

Table C1 Information for the 10 Caribbean stations used to generate Figure C6. Values in column 1 correspond with a specific subplot in Figure C6. Values in column 6 indicate available monthly data used to calculate the climatology. Asterisked years indicate years with 6 months or more of missing data. Data source for stations 1, 2, 3, 4, 5, 6, 8 and 9 is the Global Historical Climate Network (GHCN). Data source for stations 7 and 10 is the respective meteorological services

Figure No.	Station name	WMO identifier	Coordinates	Elevation	Data range
1	Nassau International Airport (Bahamas)	78073	25.05N, 77.47E	7 m	1960–2000
2	Guantanamo Bay (Cuba)	78367	19.90N, 75.13E	23 m	1960–2000 *1992–1996
3	Kingston, Norman Manley International Airport (Jamaica)	78397	7.90N, 76.80E	14 m	1960–2000
4	Santo Domingo (Dominican Republic)	78486	18.43N, 69.88E	14 m	1960–2000 *91–95
5	San Juan International Airport (Puerto Rico)	78526	18.43N, 66.00E	19 m	1960–2000
6	Juliana Airport (St. Maarten)	78866	18.05N, 63.12E	9 m	1960–2000
7	V.C. Bird International Airport (Antigua)	78862	17.12N, 61.78E	8 m	1960–1995
8	Grantley Adams International Airport (Barbados)	78954	13.07N, 59.48E	56 m	1960–2000 *89–91, 93
9	Piarco International Airport (Trinidad and Tobago)	78970	10.62N, 61.35E	15 m	1960–2000 *81–83, 85–86, 95
10	Queen Beatrix International Airport (Aruba)	78982	12.50N, 70.02E	18 m	1971–2000

Oscillation (ENSO) seem to play a role in determining the number of storms which will develop and pass through the Caribbean. During the warm phase there is an apparent decrease in the frequency of tropical storms due to an increase in wind shear over the Caribbean during the hurricane season (Banichevich and Lizano, 1998). There is also evidence of decadal variation in storm activity with some decades on average being less active (1970s to 1990s) than others (1920s to 1960s) (Goldenberg et al., 2001).

The hurricane season runs from June to November with a peak in activity in September. The peak occurs prior to the rainfall maximum of the Caribbean, suggesting that although hurricanes are significant they are not the primary rainfall mechanism for the Caribbean territories, as is often thought, mainly because their effects over specific locations depend on the cyclone's relative position and their velocity over the Caribbean Sea. The coincidence of hurricane peak activity and maximum tropical wave activity is, however, not surprising given the importance of the waves to cyclogenesis. Most (though not all) of the tropical cyclones that ply the Caribbean Sea originate from tropical waves in the easterlies. A warm ocean and low vertical shear fuel their development. Development during the peak period generally occurs in the 10–20°N latitudinal band (termed the main development region or MDR) just east of the Lesser Antilles in the eastern north Atlantic. Early in the hurricane season, however (June), the development region resides in the Caribbean and Gulf of Mexico.

Hurricane development is suppressed in July leading to a bimodal distribution of the annual cycle of north Atlantic hurricanes. This suppression coincides with the previously noted intensification of the trades in July, which yields upwelling in the southwestern Caribbean and lower sea surface temperatures, high vertical shear and reduced convection.

Sometimes (as with Joan in 1988), hurricanes pass over Central America and continue their activity in the eastern tropical Pacific. The eastern tropical Pacific is itself an important region of hurricane development but these affect mainly the Pacific coast of Mexico, and occasionally the northern part of Central America (Guatemala and El Salvador).

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Cross-references

Azores (Bermuda) High
Humid Climates
Intertropical Convergence Zone
Orographic Precipitation
Trade Winds and the Trade Wind Inversion
Tropical and Equatorial Climates
Tropical Cyclones

CLIMATE AFFAIRS

Climate affairs is a multidisciplinary concept that encompasses climate science, climate impacts on ecosystems and on societies, climate politics, policy, law and politics, climate economics, and climate ethics. While individual and societal concerns about climate have existed for millennia, the notion of climate affairs is new and is designed to meet the growing educational needs related to climate-related issues.

Climate in its various forms – variability, fluctuations, change, extremes – is appearing at the top of governmental lists of concern about global environmental and technological change. Aside from specific weather and climate phenomena that are of concern in their own right, there is a deep concern about what the adverse effects might be of those phenomena on human activities or on the resources on which societies depend: food production and security, water resources, energy, public health, public safety, economy and environment. Running through each of these categories are ethical and equity issues. One could argue that global change represents a version of a new Cold War; only this time the conflict is between societal activities and natural physical and biological processes, instead of between two nuclear superpowers.

Rich and poor societies alike have increasingly come to realize the extent to which their activities (e.g. industrialization processes and land-use practices) can affect the local and global atmosphere as well as be affected by it. In addition, an increasing number of government, individual, and corporate decisions are being made for which a knowledge of climate affairs is required. As a result, there is a growing awareness among educators in many disciplines of the need for a better understanding of just how climate variability, change and extremes affect the environment and the socioeconomic affairs of people, cultures and nations.

Climate issues have become increasingly important to governments, corporations, and the public. One could surmise that this sharp increase in interest in climate issues and weather has been a result of the end of the Cold War in the early 1990s. At the same time various storms and anomalies have been labeled as the "storm of the century", the worst hurricane, the most costly drought, the longest El Niño, and so forth, in history. Some observers have linked "blockbuster" weather and climate episodes to human-induced global warming of the atmosphere, while others have argued that these are random extreme events under normal global climate conditions.

Since the late 1980s, scientific and government interest in human-induced influences on the naturally occurring greenhouse effect has increased steadily and to new heights, leading up to serious international negotiations to control the emissions of greenhouse gases (carbon dioxide, nitrous oxides, CFCs, methane). The perception about the global climate regime is that climate anomalies have become more frequent, more costly and more deadly. The negative impacts of a changing climate on food, water, energy, health and public safety of a human-induced global warming loom large in the 21st century.

For our purposes, climate encompasses *variability* from season to season and from year to year, *fluctuations* on the order of decades, *change* on the order of centuries and beyond, and *extreme meteorological events*.

Climate variability

Although climate varies on several time scales from seasons to millennia, most people are most concerned about climate variability within a season, from one season to the next, and from one year to the next. These are the time scales that are perceived by people and societies to directly, as well as indirectly, affect them and resources such as food, water, energy, ecological and human health. People everywhere are aware of growing seasons, hunger seasons, rainy seasons, fire seasons, hurricane seasons, planting seasons, harvest time, and so on. Forecasters are now focused on producing reliable seasonal forecasts well in advance of the season, as these forecasts are most useful to societies and individuals, affording them time to take strategic evasive action. When the natural flow of the seasons is disrupted, social and economic problems arise for individuals as well as governments. Year-to-year climate variability is also of concern to society. The El Niño–Southern Oscillation (ENSO) cycle of warm–neutral–cold episodes operates on an annual basis with El Niño events recurring on the order of 2–10 years, often interspersed with the cold episodes.

There is a major difference in societal and environmental impacts in the way annual climate is sequenced. For example, if a dry year is followed by a wet year, societal (and individual) responses will differ from a dry year followed by a second and a third dry year. The same applies to wet years. As an example, in an arid area, societal coping strategies are designed to cope with one or two dry years but not with a third.

Climate fluctuations

Fluctuations refers to variability on the order of decades. For example, the frequency of hurricanes in the tropical Atlantic has changed over periods of decades: 1930–1960 was a very active period, whereas 1960–1995 was relatively inactive. Some researchers suggest that we are about to enter a multi-decadal period with an increase in the frequency of tropical storms on the order of the earlier (1930–1960) active hurricane period.

As another example, El Niño events were less frequent than La Niña between 1950 and 1975, but have become more frequent in the last quarter of the twentieth century. While this time scale is of interest to certain specialists dealing with long-range issues, such as engineers who deal with the planning of water resources and structures that have to last for many decades, they are apparently of much less concern to individuals and policy makers. The decadal scale is one about which societies should be more concerned and knowledgeable than they are at present.

Climate change (new global climate state)

Climate change is a euphemism for global warming. Scientific reports strongly suggest that human activities such as the burning of fossil fuels, tropical deforestation and the emission of other greenhouse gases are responsible for enhancing the naturally occurring greenhouse effect.

Political leaders in many countries have become directly involved in and concerned about the global warming issue. It relates to international energy politics as well as domestic energy debates. Governments equate the use of energy with economic progress. The popular source of energy derives from fossil fuel use (coal, oil and natural gas). These gases when emitted alter the radiative balance of the atmosphere, trapping outgoing longwave radiation in the Earth's atmosphere, which leads to global warming.

Furthermore, in order to prepare society to cope with the projected global and regional changes in climate, societies must evaluate how well they cope with variability and extremes today.

Extreme meteorological events (EMEs)

Societies everywhere worry about particular extreme weather, climate and climate-related events. Although each society faces its own subset of such extremes – droughts, floods, cyclones, heat waves, vector-borne infectious disease outbreaks, etc. – it does not have to face the same set of extremes as other societies with different climate regimes. While Floridians are concerned about hurricanes and fires, Californians are concerned about El Niño-related coastal storms and torrential rains in the southern part of the state. States in the Great Plains and the Gulf states are concerned about tornadoes and severe weather to a much greater extent than states in other parts of the US. There are, however, many similarities in the way that societies prepare for or cope with extreme events.

Researchers concerned about climate variability and those concerned about climate change are interested in EMEs, but for different reasons: the former community wants to forecast their onset and prepare society for their impacts, whereas the latter community of researchers wants to identify changes in frequency, intensity, duration or location of such extremes.

In order for a climate affairs activity (education, training, research) to be complete it must incorporate information about climate science, climate impacts, climate politics, policy and law, climate economics and climate ethics and equity.

Climate science

The objective of a climate science section of climate affairs is threefold:

1. To understand the climate system.
2. To understand its components.
3. To recognize society as a component.

It is important that climate affairs students and professionals understand the workings of the climate system. That system has numerous components. The shells of mollusks, for example, retain carbon for millennia. Termites produce methane, as do animals in feedlots, and so forth. The system's major components are sea ice, forests, oceans, deserts, clouds, the sun, topography, rangelands, and wetlands. Human activities have become a notable forcing factor as far as atmospheric processes are concerned and are now an integral part of the climate system along with its physical components. Social scientists are increasingly recognizing the importance of understanding the physical conditions under which societies they study must operate.

Climate impacts

Climate variability, change and extremes have positive and negative impacts on environment and society. Those impacts can vary over time even in the same location, depending on changes in the vulnerability of ecosystems and societies. The level of vulnerability or resilience of a society or an ecosystem is not just dependent on the intensity of a weather or climate anomaly.

Attributing a specific impact to a climate anomaly is not easy, as there are often several factors that have to be taken into consideration in doling out "blame". Often, the way attribution is done by the public is as follows: a major anomaly – drought, flood, frost, fire, severe storm, cyclone – occurs and most adverse impacts that happen to take place are identified; almost immediately they are attributed to it. However, a more considered assessment is necessary, so that appropriate safeguards can be developed to enable a society to cope with similar events in the future. Wrongly attributing climate-related impacts to a specific cause can prove to be wasteful of scarce resources. It can also lead to false expectations about the level of protection that a society might have from climate-related anomalies.

Climate impacts on terrestrial and marine ecosystems

For the most part the media concentrate their attention and coverage on visible climate and climate-related impacts on land-based ecosystems, both managed (irrigated, dryland, rangeland) and unmanaged (forested and other wilderness areas). Droughts and floods capture the most attention. More correctly, premature concern during dry spells about drought often fill the airwaves and the printed media, only to have that concern evaporate as a timely rain saves a crop. For example, in the midst of a devastating El Niño-related drought in 1997 in Australia, timely rains occurred over a few weeks and prevented a devastating agricultural production year.

Climate impacts on marine ecosystems receive much less attention for a variety of reasons: fishing activities take place where most people do not live; the fishing sectors in most countries involve a relatively small number of workers; it is an economic problem for the most part, rather than one of life and death; the relationship between the viability of living marine resources and variability in air–sea interactions in different parts of the world's oceans are much less studied and not well understood. There are notable examples, however, of climate–marine life interactions: the Peruvian anchoveta, Ecuadorian shrimp, Pacific Northwest Salmon, Pacific Sardine, among others.

The impacts of climate anomalies in industrialized countries are economic issues, for the most part, whereas in developing countries the impacts of anomalies on societies tend to create life–death situations for relatively large segments of their populations. In other words, rich countries have the economic (and therefore technological) wherewithal to minimize or mitigate climate impacts on their populations.

The public now knows that human activities can affect atmospheric chemistry and atmospheric processes at all levels from local (i.e. urban heat island effect) to regional (i.e. tropical deforestation, desertification, and SO_2 emissions), to global (i.e. ozone depletion as a result of CFC emissions and global warming as a result of greenhouse gas emissions). Research on human activities that can affect the atmosphere continues, as researchers seek to reduce scientific uncertainties surrounding their impacts on the atmosphere. Improved understanding of this interaction could provide a sound, non-controversial basis for societal actions to minimize human influences on the atmosphere at various spatial levels.

Attempts to identify human impacts on the atmosphere and the impact of atmospheric processes on society depend on both qualitative and quantitative methods of assessment. It is likely that there will always be some level of scientific uncertainty that surrounds society–atmosphere interactions and, as new methods are developed, a refining of our understanding of those interactions will take place. However, there may be useful qualitative measures to identify how societies might be affected by climate variability, change and extremes. Even the use of anecdotal information can be instructive about climate impacts in certain societies. Historical accounts of impacts and responses to climate-related problems can be very instructive.

One qualitative approach has been referred to as "forecasting by analogy". While the future impacts of a climate anomaly will not likely be a mirror image of a recent past anomaly, this approach can help decision makers identify the strengths and weaknesses in societal and especially institutional responses. The strengths can be maintained or enhanced, and the weaknesses can be addressed.

Climate politics, policy and law

Climate politics refers to the process to achieve (or not achieve) certain objectives by way of climate and climate-related policies and laws. The outcome of politics is climate policies and laws. There are many climate-related political issues in need of discussion and resolution. Politics related to climate issues occurs at local, national, regional and global levels.

Climate has become a highly charged (politically and economically) aspect of environmental change at local to global levels. Climate as a scientific concern is not only important to comprehend in its own right, but understanding it and its various aspects helps us to cope more effectively with other issues considered to be climate-sensitive (fisheries, health, history, war, economy, policy, disasters, institutional change, etc.).

Today, the international community is negotiating the development of what could be called a "Law of the Atmosphere", through the IPCC (Intergovernmental Panel on Climate Change) process. Coupling together international agreements related to the atmosphere that have been developed over time, a body of laws related to the chemistry of the atmosphere and atmospheric processes is being accumulated: the Vienna Convention, the Montreal Protocol, the Convention on Long-Range Transboundary Air Pollution, and the UN Framework Convention on Climate Change (UNFCCC) and subsequent Kyoto Protocol and the entire set of Conference of Parties (COP) conferences. In the process of developing what will essentially become a Law of the Atmosphere, negotiators continue to seek assistance from experts in the physical, biological, and social sciences. Thus, expertise on climate-related issues has been emerging within educational, governmental and non-governmental institutions around the world.

Actions have been taken since the mid-1970s to reduce and, later, to eliminate ozone-depleting chemicals from industrial use in both the industrialized and developing countries. The Montreal Protocol in 1987 (and later amended several times) has sharply reduced CFC emissions. Since the late 1980s to the present, there have been several international conferences on global warming. The most noteworthy is the UNFCCC and the various Conferences of Parties to develop a protocol for the effective control of greenhouse gas emissions.

Acid rain has also generated international discussions, conferences and agreements on the science and impacts. Transboundary acid rain concerns have led to a dilemma. The SO_2 output of industrialization processes has been found to counteract the warming of the atmosphere regionally caused by the burning of the suite of fossil fuels. Recently, considerable renewed concern has focused on water issues and how those issues are or can be affected by climate variability, fluctuations, change and extremes (e.g. the Water Forum in 2001 in Japan).

Climate economics

What was the cost of that storm, drought, flood, frost or El Niño event? When people think of climate economics, this is most likely what they will think about – cost or benefit of a climate-related impact. To do this properly, however, researchers have to get a better handle (e.g. understanding) on "attribution" or what impacts can appropriately be blamed on or associated with a weather or climate event. A climate anomaly affects a society that is also affected at the same time by other factors, such as poverty, poor land management, inappropriate land use, conflict, trans-boundary air or water disputes, bureaucratic rivalries, domestic political rivalries, and so forth. A researcher has to take great care in teasing out of this situation, what was the contribution of the climate anomaly. Sometimes the anomaly may not have been the most important factor by itself but became important when it combined with some of these other factors.

Financial loss or gain as the result of climate variability, change or extremes, in fact, is only one aspect of climate economics. Economic researchers are also concerned about the following, when it comes to assessing climate–society–environment interactions: risk analysis, discount rates, externalities – present vs. future generation welfare, forecast value, free market vs. government intervention, prevent, mitigate or adapt, climate variability and economic development.

Climate ethics

Of all the aspects of a climate affairs activity, the most neglected has been that of climate ethics and equity. This aspect encompasses a wide range of ethical and equity issues, such as the following, that are directly or indirectly affected by climate variability, change or extremes.

Inter- vs. intra-generational equity

Governments everywhere face the same question in the face of a climate issue: Who to help in the event of climate-related problems, present generations or future ones? For example, should the Brazilian government preserve Brazil's tropical rainforests, for example, for potential discovery of pharmaceuticals or cut them down to clear the land for cultivation and livestock and sell the timber? There is a desire also on the part of many to pass something of value on to future generations. As a result, they want to manage the Earth's resources (soil, vegetation, water, air) with the needs and welfare of future generations in mind.

Environmental justice

It is usually the poor and elderly in society who are most adversely affected by a drought, a flood or a heat wave. They are the ones whose living space has been relegated to precarious regions or areas or conditions: flood plains, steep slopes, arid lands, swampland, short growing seasons, and so forth. A forecast issued on the Internet gives those who have computers an advantage to receive that timely information before those who do have that access to the information highway. The rich in society have other options that the poor do not have available to avoid climate-related harm. Seldom are the upper classes in a society in harm's way to the same extent as the poor.

North–south cleavages and climate change

Today, there is considerable debate over responsibility for the global warming of the Earth's atmosphere. The developing countries argue forcefully and effectively that the industrialized countries were able to develop rapidly as a result of their burning of fossil fuels during *their* industrial revolution. As a result, the rich countries have saturated the atmosphere with radiatively active greenhouse gases. They argue that it is the responsibility of the industrialized countries to take the first steps – unconditionally – to reduce their greenhouse gas emissions. To counter, the industrialized and the coal- and oil-producing countries argue that the developing countries will be supplying the lion's share of greenhouse gases emissions in future decades.

There are several sayings that relate to the ethical aspects of climate, broadly defined: the "polluter pays principle", the "precautionary principle", and "common but differentiated responsibilities". Each of these notions generates concerns about equity among competing countries, companies and ideologies. Thus, climate ethics and equity is an integral part of climate affairs.

In sum, climate affairs is an integrator of disciplinary interests in issues that are directly or indirectly affected by the behavior (as well as perceptions of the behavior) and impacts of climate on environment and on society. It can also serve to stimulate new ways of thinking about traditional issues of disciplinary scientific research; conflicts over climate-related resources; political, economic, and technological development; famine; and political instability.

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Cross-references

Applied Climatology
Climate Change Impacts: Potential Environmental and Societal Consequences
Climate Hazards
Climate Data Centers
Commerce and Climate

CLIMATIC CHANGE AND ANCIENT CIVILIZATIONS

Climatic change during the postglacial period

The Quaternary period is characterized as a glacial age. The coldest period of the last glaciation was about 18 000 years BP (based on uncalibrate ^{14}C estimate), when margins of ice sheets and glaciers began to retreat in northern Europe, North America, and regions of the Eurasian continent. The end of that glacial age brought great changes in the landscape, due not only to the retreat of ice sheets and glaciers but also to the rise in sea level caused by melting ice. The rise in sea level occurred not only in the postglacial period but also between 17 000 and 14 000 years BP and between 10 000 and 7000 years BP. The latter phase was drastic in northern Europe particularly rapid, and north America because of glacio-isostatic rebound. Worldwide rise in sea level from the last glaciation to the present is estimated to be 140 m in total.

The hypsithermal

The last glacial stage ended about 10 000 years BP, after which the climate became warmer and reached a peak period called the

Climatic Optimum or the Hypsithermal, about that lasted about 6000 to 4500 years BP. Lamb (1982a) reconstructed average latitudes of the lowest and highest air-pressure axes at sea level in the European sector of the northern hemisphere. In winter, during the Hypsithermal, the dry belt under the subtropical high was located at around 40–45°N (around 30°N today) and the wet belt under the subpolar low at around 57–58°N (about 50°N today). On the other hand, in summer the dry belt under the subtropical high was located at around 50°N during the Hypsithermal (around 33–35°N today); the wet belt under the subpolar low has not fluctuated greatly since 8000 years BP. Suzuki (2000) compiled the climatic change curves during the last 10 000 years all over the world, which were based on the continuous data analyzed, as shown in Figure C7. Even though the ranges of fluctuations and the local differences are different from place to place, and for the climatic elements, the general tendency mentioned above can be observed. Circulation patterns over Africa of 8000 years BP were reconstructed by Messerli (1980), as shown in Figure C8. The difference between Hypsithermal and present-day circulation patterns is strikingly clear in this illustration. Climatic fluctuations in the arid belt of the "Old World" since the last glacial maximum, taking examples in Africa, were discussed by Flohn and Nicholson (1980), Nicholson and Flohn (1980), Tyson (1986), and Tyson and Lindberg (1992).

In most parts of the world the climate between 7000 years BP or earlier and by 5000 years BP was warmer by 1–3°C than it is today (Harding, 1982). The middle latitudes enjoyed a warm climate and the Intertropical Convergence Zone (ITCZ) shifted northward in the northern hemisphere (Suzuki, 1975, 2000). One result of this shift was that the arid regions from the Sahel to northwestern India, which today lie in the Subtropical High-Pressure Zone (STHP) area, during the Hypsithermal experienced a moister regime south of the STHP. Europe experienced a warm, moist climate, and extensive beech and oak forests covered England and the Scandinavian peninsula. Overall mean temperatures in Europe and North America seem to have been up to 2°C higher than at present (Lamb et al., 1966; Lamb, 1977). In East Asia it was 5–8°C colder 10 000 years BP but 2–3°C warmer 5000–6000 years BP than today (Yoshino and Urushibara-Yoshino, 1978).

In the middle latitudes, glaciers on the mountains almost disappeared during the Hypsithermal. Ice in the Arctic Sea melted away and the ice sheets and glaciers in Greenland and Antarctica shrank in this period. The minimum reconstructed 1000–500-hPa-layer thickness of 5150–5200 gpm was found over Arctic North America (Lamb, 1974). By 6000 years BP thermal gradients had weakened and the area of steep gradient shifted north of its current position. The great cold trough had broadened, weakened, and shifted east, out over the Atlantic in summer. It appears that the trough was located farther east than where it had been over North America in winter at 8500 years BP (Lamb et al., 1966). In short, the circulation became weaker on the whole, more zonally oriented, with its action centers located farther north. The North American cold trough was displaced eastward and the wavelength was increased, associated with a spread of westerlies at surface level.

Post-Hypsithermal

After the Hypsithermal the climate deteriorated once more. Large forests disappeared from England and Scandinavia during 3500–3000 years BP. (These dates "BP" are all uncalibrated.) This tendency was observed not only in Europe but in all parts of the world. Suzuki (2000) concluded that there is evidence of a sharp decrease in air temperature around 3500 years BP in the Mediterranean region, East Asia, Australia, North America and South America. Of course, the decrease in temperature did not occur at the same time all over the world but it can be summarized that in 3500–3000 years BP temperature dropped by 2–3°C from the maximum of the Hypsithermal period. In the Near East this period was characterized by a mild dryness and in northwestern India a sharp dryness. Such severe desertification at the zone around 25–35°N was caused by the northward shift of the subtropical high-pressure zone in this period. The polar frontal zone also shifted poleward and wet conditions prevailed along both zones. There are other indications that the climate became slightly wetter in the tropical zone during this period, which may have been caused by the intensification of tropical westerlies. In short, in the period 3500–3000 yr BP the temperature decrease and the change in wet and dry conditions caused by the shifting of frontal zones and the subtropical high-pressure zone were striking features.

By 2500 years BP the renewed cooling trend of climates in the north had steepened the thermal gradient again, resulting in stronger circulation, longer wavelength, and more penetration of the westerlies over Europe (Lamb et al., 1966). Around 2850 years BP, glaciers in northern Europe and the Alps advanced suddenly and rivers were frozen in many parts of Europe due to the cold climate. East Asia was also cold during this period.

Suzuki (2000) compiled also the climatic change curves during the last 1100 years all over the world, as shown in Figure C9. The fluctuation ranges and phases are rather different region to region. Roughly speaking, however, the fourth to fifth centuries AD were colder and wetter in most regions. The evidence of climatic changes has been clarified recently (Bradley and Jones, 1992).

Ancient civilizations

After the climax stage of the last glaciations, the climate became warmer and reached its peak of the Hypsithermal as mentioned above. Until 12 000 years BP, human inhabitants in the northern part of the Near East had lived in caves and hunted wild game in mountain areas. With the climatic change to warmer conditions they came down to open living sites in the foothills where the ground was more favorable for cultivation (Wright, 1968). A northward shift in the tree line (tundra boundary) was seen in Europe. In accordance with the invasion of woodland-type vegetation, insects, birds, and fish extended their ranges.

During the warmest postglacial time, between 6500 and 5000 years BP, the Sahara was much more humid than today. Animals and humans could roam about and cross what is now the world's greatest desert. The moist condition of the Sahara is supported by the study of the change in water levels of Lake Chad and other lakes (Messerli, 1980). Although the levels fluctuated severely in the beginning of the postglacial time, they reached their highest levels about 8000 years BP, with a second peak about 5000 years BP.

Cooling first occurred in higher latitudes after about 5500 years BP. Especially after 4800 years BP the moist regime began to decline. The drier conditions confined human settlement and animals to oases and river valleys such as the Nile, Mesopotamia, Indus, and Hwang-ho (Yellow River). In this period, summer temperatures were 1–3°C higher than today's but winter temperatures were variable. The drying tendencies from about 5500 to 4800

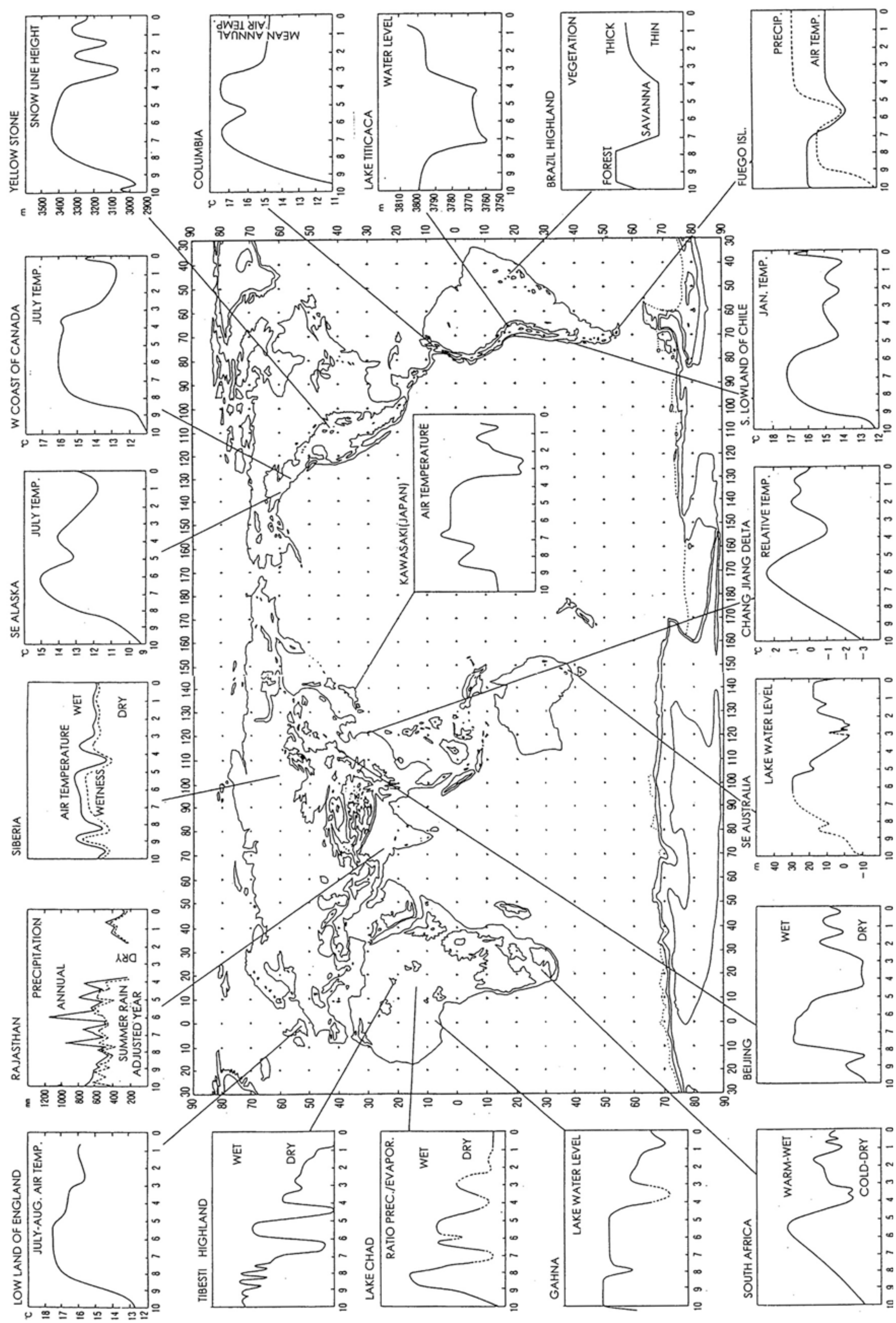

Figure C7 Climatic change curves during the last 10 000 years (after Suzuki, 2000, pp. 8–9).

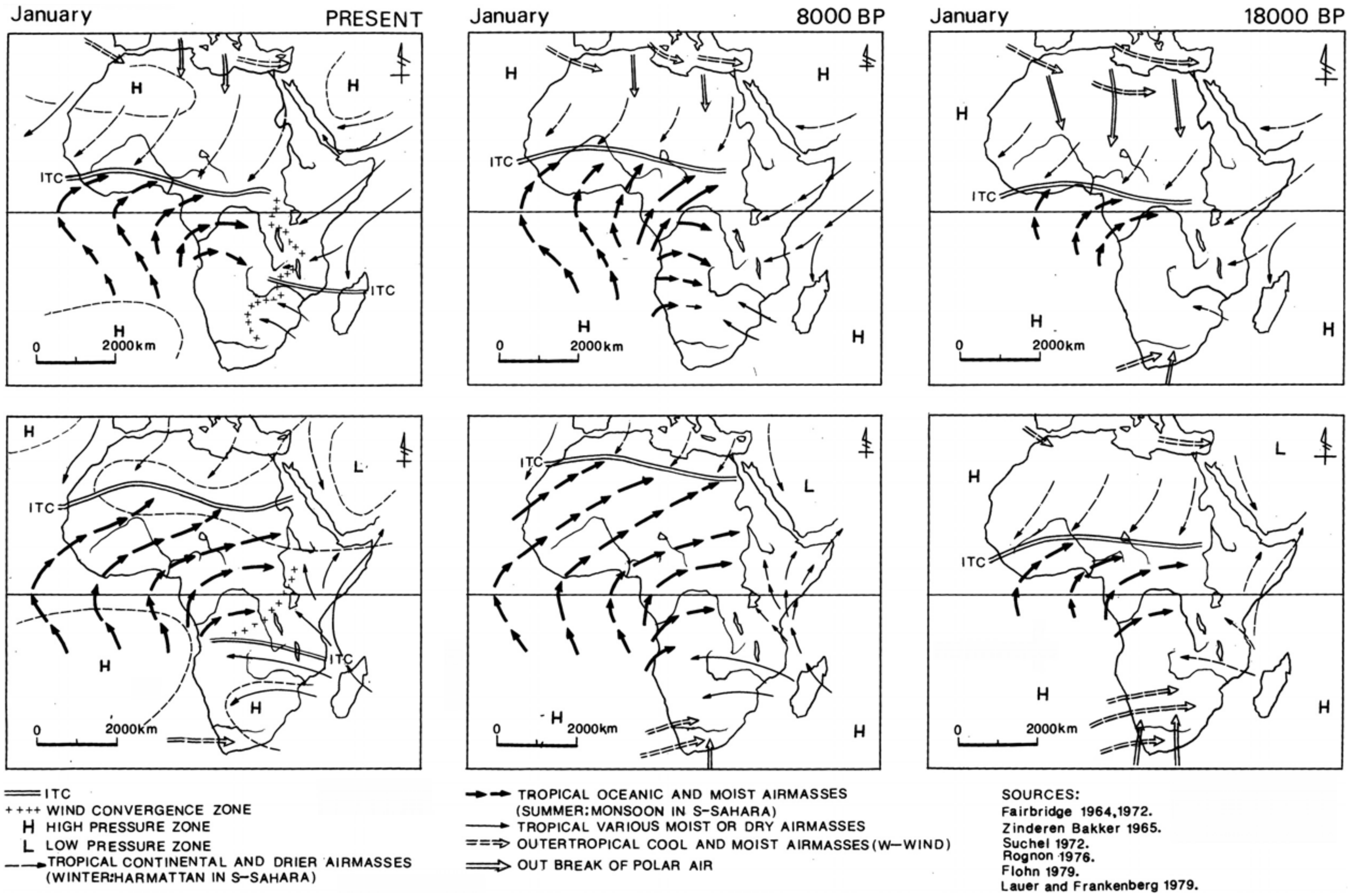

Figure C8 Circulation patterns in Africa at present, 8000 years BP, and 18 000 years BP (after Messerli, 1980, p.84).

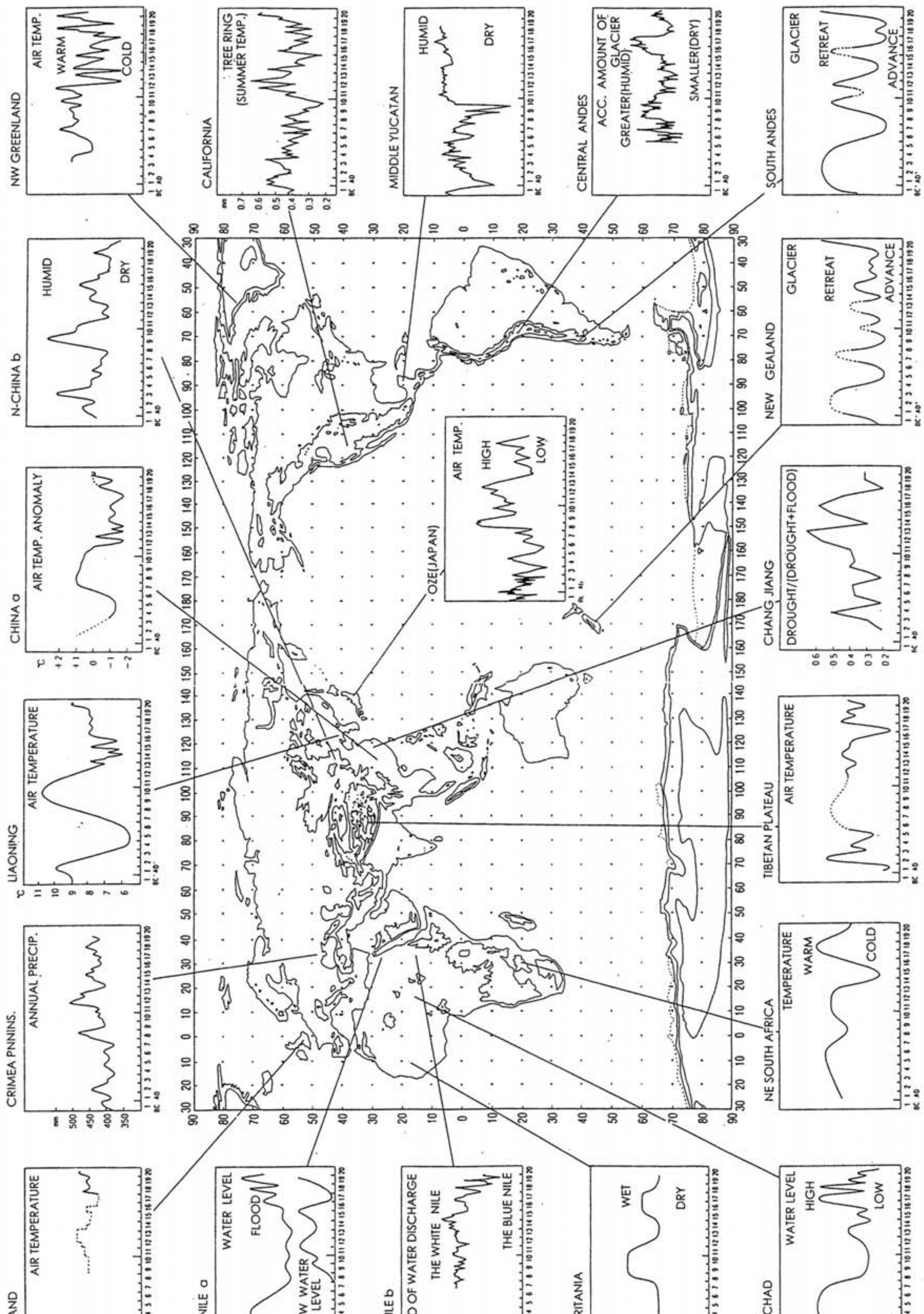

Figure C9 Climatic change curves during the last 2100 years (after Suzuki, 2000, pp. 126–127).

years BP onward seem to be have been related to a climatic development of hemispheric, and probably global, extent (Lamb, 1982a). Butzer (1966) also suggested that the Sahara recorded a moister subpluvial period between about 5500 and 2350 BC (with interruption by one or more dry spells) and a hyperarid oscillation between 2350 and 870 BC. This can also be observed on the curve of ratio for precipitation/evaporation in Lake Chad in Figure C7.

The dry and wet conditions also differed, from region to region during the Hypsithermal, as also shown in Figure C7. The regional change in cultural practices during the postglacial period is closely related to the climatic change, mainly due to the change of hunted animals. Irwin-William and Haynes (1970) reported changes in Paleo-Indian cultural areas in the southwestern United States since 11 500 years BP. After the last glacial period the dryness increased gradually and the Hypsithermal period was the driest in the Paleo-Indian region. Accordingly, their cultural area retreated owing to the decrease in the number of bison. Then, in the period between 5000 and 4500 years BP, a wet condition pervaded this region and a dramatic increase in population took place.

In Egypt, between the first and the second Dynasty (2152–2040 BC), and between the second and the third Dynasty (1552–1069 BC), periods of chaos prevailed. As one of the causes for confusion, Bell (1975) pointed out the anomalous changes of flood levels of the River Nile; which have a close relation to Egypt culture. These periods coincide roughly with the water-level changes of Lakes Faiyum and Moeris (Hassan, 1986), and the sea-level change of the Mediterranean (Fairbridge, 1965) as well as water-level changes of lakes in East Africa (Butzer, 1976) which were of course related to the monsoon rainfall and evaporation changes of the region.

In China, during the Hypsithermal period, the overall average temperature may have been about 2°C warmer and the midwinter temperature about 5°C higher than today. This enabled the cultures of the Yanshao and Yin-Hsu periods to flourish, because there were rich subtropical fauna and flora concentrated around old cities such as Sian (Xi'an) and Anyang (Chu, 1973). It is assumed that, at the beginning of the Chou Dynasty (1066–256 BC), the climate was warm enough to grow bamboo extensively in the Hwang-Ho valley. The cultivation was destroyed, however, and when protracted droughts followed, as well as freezing up of the Yangtze River region.

The first Chinese Neolithic agricultural civilization developed in the northern China plain. The sudden advance to a Bronze-Age culture and the cultivation with irrigation of wheat and millet seem to suggest some contact with the European culture across Central Asia, possibly during a moist period.

In Japan the Hypsithermal corresponded with the early and middle Johmon (Jomon) period (Fukui, 1977). The Johmon culture seems to have been established by the people who came from the west during the warmer Johmon period. After about 4000 years BP the climate tended to deteriorate; it was warm but unstable. In the second half of the latter stage of the Johmon period a cooler climate affected northern Japan: the wetter region retreated to southwestern Japan due to the southward shift of the polar frontal zone during the warm season (Yasuda, 1995). There is evidence that the Johmon culture developed first in the east, then migrated to the southwest. This may be attributed to the shifting of wetter regions in Japan (Yasuda, 1997).

The Yayoi period in Japan began about 2500 years bp, when the climate became cooler. This cooler period coincided with the above-mentioned cool period of the Chou Dynasty in China. There was a warm stage between 800 and 400 bc. Air temperature in the Yayoi period seems to have been 1.0–1.58C lower than today. In northern Japan it was cooler and wetter than today, with its peak in about 2500 years bp (Yasuda, 1978, 1995). Paddy rice cultivation spread from southwestern to northeastern and northern Japan and established the basis of Japanese culture today, even though the climatic conditions were not friendly in this period. It is interesting that even among the different cultures in different periods there are similar phenomena; namely, the coincidence in timing of the rise of a new culture with the decline of the prevailing favorable climate, as in the timing of the rise of the stone age culture in the Old World and its decline during the Hypsithermal.

A question is, why were flourishing civilizations not in the valleys of other great rivers such as the Ganges, the Yangtze, and the Mississippi, as they were in the Nile, Hwang Ho, etc.? At least one reason for this absence can be attributed to the climate. It seems that the drier conditions in the less inhabited valleys during the Hypsithermal did not experience the passage of the ITCZ or polar frontal zones in this period.

Summarizing the relationship between the beginning of ancient civilizations and their climates, we may conclude that the ancient major civilizations began around 5000 years BP – this was the time after the peak of the Hypsithermal. Strictly speaking, civilization began in the valleys along great rivers located in the marginal regions that started to experience cool and dry conditions. The rise of Egypt and the organized cultivation of the Nile valley by the use of early floods for irrigation may have been the necessary response to the increased food demand in a newly habitable region. In other words, civilization came about through the need to organize irrigation systems to produce food for the increased population, while the refugees presumably provided slave labor (Suzuki, 1979; Lamb, 1982a). The disruption of established ways, which the climatic events caused, provided the challenge and stimulus for undertaking deliberate cultivation and invention of new tools.

The decline of ancient civilization took place around 3500 years BP, which was a turning point in postglacial climatic history. It is obvious that the ancient civilizations were not able to continue in the valleys along great rivers because increasingly drier climates caused limited crop production.

The agricultural areas had been greatly reduced by 3500 years BP, and agricultural production was influenced by the decreased rainfall, e.g. several hundred millimeter decreases in rainfall in the Indus valley. Outside the agricultural areas, people had to shift to regions to live in accordance with the retreat and advance of grassland. The climate change was the cause of decline in the ancient civilizations and the migration of peoples.

In Roman times (Rome was founded in 753 BC) the Mediterranean world had a cooler climate with more winter rains (primarily from 600 to 200 BC). This period was one of great fertility for Greece, northern Africa, the Carthaginian, and later Roman croplands. For a few hundred years there was a warming tendency and increasing dry weather until about AD 400.

During the days of early Babylonia and Egypt there were two great inventions. One was the art of fashioning iron into tools with a cutting edge, and the other was the building of seagoing ships. The first invention meant that humans could now live quite comfortably in colder or more humid climates than those of Babylonia or Egypt. The second, shipbuilding, can be considered in relation to forest management. During the greatest days of Greece, the contrast between wet winters and dry summers was less marked than today. Such conditions were favorable for the growth of forests.

Viewpoints

There are several viewpoints on the relationship between climate and civilization (Pittock et al., 1978). In particular, the rise and fall of ancient civilizations seems to have been closely related to the change in climate. The ancient civilizations were built up on the basis of hunting and nomadism, which depended on the distribution of vegetation and animals. The flourishing of the ancient civilizations was supported by agricultural production and a definite form of organized village life.

Huntington (1945) wrote on the role of climate in developing civilizations in the Babylonian and Mesopotamian regions. He pointed out rightly that "the real problem [in developing civilizations] is to determine the exact nature of the [climatic] influence, its magnitude, and the extent to which its favorable and unfavorable aspects have counteracted one another". Because of his general deterministic treatment of the relationship between climate and human activities, including civilization, his descriptions are not accepted by many people today. However, at least as far as his account of the ancient, civilizations is concerned, his description is a sophisticated one. As pointed out by Spate (1952) and Oliver (1973), the historian Toynbee presented many similar ideas, although in a much more literary fashion.

It seems that the second half of the twentieth century was a time to discuss how climatic change may have influenced history (Manley, 1958; Carpenter, 1966; Lamb, 1968; Claiborn, 1970; Le Roy Ladurie, 1971; Singh, 1971; Chu, 1973; Dansgaard et al., 1975; Bryson, 1978; Suzuki, 1979, 2000; Wigley et al., 1981; Lamb, 1982a; Issar, 1998). Ranging from determinism to probablism, and possiblism to voluntarism, there can be various viewpoints on climate–civilization relationships (Oliver, 1973). Civilization is effected by economic determinism, political instability, racial invasion, decreasing activity, disaster by disease, population decreases etc., as well as climatic change (Yoshino et al., 1993). However, at the present stage, it can be concluded that even though the Holocene fluctuations are of relatively small magnitude, they have been sufficient to trigger cultural change in marginal situations. As Bryson (1978) wrote, cultural change includes changes in the economic base, such as agriculture, hunter–gathering, and herding. How humans responded to cultural change, i.e. in situational modification of lifestyle, migration, or literal disappearance of the people, must be determined locally.

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Cross-references

Climatology, History of
Determinism, Climatic
Climatic Variation: Historical
Climatic Variation: Instrumental Record
Desertification
Tree-Ring Analysis

CLIMATE CHANGE AND GLOBAL WARMING[1]

Coal, oil, and natural gas that were formed from the fossilized remains of ancient plants and animals (leading to the term *fossil fuels*) provide 80–85% of the world's energy. This energy contributes to sustaining the world's standard of living and provides much of the power for transportation, generation of electricity, home heating, and food production. Compared to other sources of energy, fossil fuels are relatively inexpensive, transportable, safe, and abundant. At the same time their use contributes to environmental problems such as urban air pollution, acid rain, and elevated levels of mercury (particularly in the Arctic) that arise from trace elements in the fuels, and to long-term climate change that arises because combustion of these fuels releases carbon to the atmosphere that has been geologically sequestered for hundreds of millions of years. The health and environmental problems from the trace element emissions can be reduced by conventional emission control strategies. However, while some carbon can likely be sequestered in the deep ocean or in soils, sharply cutting overall carbon emissions will require substantial reductions in the combustion of fossil fuels for generating energy. The governments of the world began this process by negotiating the UN Framework Convention on Climate Change in 1992 and many of the developed nations are taking the first implementation step by adopting the Kyoto Protocol that was negotiated in 1997.

Concerns about the potential climatic influences of the burning of fossil fuels are not new. During the nineteenth century, initial interest focused on determining where the carbon went when coal was burned; analyses sought to determine if the atmospheric concentration of carbon dioxide (CO_2) would increase or if new carbonaceous fuels were being formed as rapidly as they were being combusted. In 1897 Swedish scientist Svante Arrhenius suggested that combustion of fossil fuels would cause the atmospheric CO_2 concentration to build up, and he calculated that this would lead to an increase in global temperatures because this gas absorbs and re-radiates infrared radiation. Just before World War II, British scientist G. S. Callendar reported observations indicating both a rising atmospheric CO_2 concentration and rising temperatures over much of the northern hemisphere. However, because analyses of ocean chemistry suggested that the oceans could take up the excess CO_2, and because northern hemisphere temperatures seemed to be cooling during the late 1940s and early 1950s, Callendar's results were only slowly accepted. In 1957 American scientists Roger Revelle and Hans Suess recognized that, because of ocean chemistry and the slow mixing of surface and deep ocean waters, the oceans would only be able to slowly moderate the excess CO_2 concentration that was being created by the combustion of fossil fuels. They coined the phrase "great geophysical experiment" to describe what was occurring. To provide the observations needed to confirm their findings, C. David Keeling established a monitoring station atop Mauna Loa in Hawaii to sample the CO_2 concentration in relatively clean and well-mixed air that was expected to represent the average conditions for the northern hemisphere. Over the next several years his observations of the rising CO_2 concentration, along with new calculations with atmospheric radiation models carried out by Syukuro Manabe and his colleagues at the Geophysical Fluid Dynamics Laboratory in Princeton, NJ, led to increasing acceptance of the suggestions of Arrhenius and Callendar. In a 1965 report of the US Government's Council on Environmental Quality, a panel of prominent scientists, led by Roger Revelle, reported to President Lyndon Johnson on the potential for global climate change and consequent impacts.

In the nearly four decades since this first report to government policy makers was submitted, much more information has become available on all aspects of this issue, which has come to be known as human-induced "global warming", even though this term greatly oversimplifies the complexity and diversity of the response of the climate system. Of particular importance have been the international scientific assessments prepared by the Intergovernmental Panel on Climate Change (IPCC), which was established in 1990 under the auspices of the United Nations (see listings of IPCC assessments in the Bibliography; references to the full scientific literature are provided in the technical chapters of these assessments). These assessment reports periodically summarize and evaluate the state of scientific understanding, and the IPCC findings have been unanimously accepted by representatives of the roughly 150 member nations, and endorsed by the leading national academies of science in the world.

Although uncertainties remain about many aspects of the chain of events from emissions to changes in concentration to climatic effects to societal and environmental impacts and options for addressing and responding to the issue, the broad outlines of the global warming situation are quite well established. This article summarizes the state of current understanding about changes in atmospheric composition and climate that are likely to occur as a result of past and future combustion of fossil fuels. An accompanying article provides an overview of the potential impacts of climate change and the challenge of adapting to the

[1] This paper is updated from MacCracken, M. C., 2001, Global warming: a science overview, in *Global Warming and Energy Policy*. New York: Kluwer/Plenum, Kluwer, pp. 151–159.

changes, and indicates the level of effort required and underway to limit long-term climate change.

Changes in atmospheric composition

That the composition of the atmosphere is changing is evident from many types of observations. For example, over the past several decades, dozens of observing stations have been established in pristine areas from the South Pole to Point Barrow, Alaska. Observations at all of these stations indicate that the global background concentrations of a number of gases present in the atmosphere at trace levels are increasing. In particular, the atmospheric concentrations of carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), and of various halocarbons – including hydrofluorocarbons (HCFC) and, until very recently, chlorofluorocarbons (CFC) – are all increasing. Attention is focused on these gases, which are collectively referred to as "greenhouse" gases, because these gases can exert a warming influence that enhances the Earth's natural greenhouse effect through their absorption and downward re-emission of the infrared (heat) radiation emitted from the surface and lower atmosphere (see further explanation in next section).

Changes in greenhouse gas concentration

Atmospheric concentrations of greenhouse gases have been measured since 1957 and reconstructed for periods prior to when the instrumental record began. For example, by measuring the composition of air trapped in the bubbles in ice cores drilled through the Antarctic and Greenland ice sheets, the history of the CO_2 concentration now reaches back over 400 000 years. These records indicate that, except for a number of the halocarbons that were created in the early twentieth century for use as refrigerants, these gases were present in the natural (or preindustrial) atmosphere, but at much lower concentrations than at present. By looking at such factors as the history of emissions versus concentrations, analyses of carbon isotopes, tree rings, and other scientific measures, it is clear that human activities have caused the increase from the preindustrial level that has persisted since the end of the last glacial.

Taken together, these results indicate that, as of late 2004, the CO_2 concentration of about 380 parts per million by volume (ppmv) is about 35% above its preindustrial level of about 280 ppmv. In addition, the long record indicates that the present value has not been exceeded back through at least the past few glacial cycles (and likely as far back as several million years or more). The analyses also indicate that the increase in the CO_2 concentration since the mid-nineteenth century has been due primarily to the combustion of fossil fuels and secondarily to the release of carbon occurring in the clearing of forested land and the plowing of soils for agriculture (see Figure C10).

Similarly, the CH_4 concentration is up over 150% compared to its preindustrial value. The analyses indicate that its increase is due primarily to human-induced emissions, including particularly from rice agriculture, ruminant livestock, biomass burning, landfills, and fossil fuel development, transmission, and combustion. Analyses of past concentrations and emissions of halocarbons, however, indicate that many of these compounds were not present in the preindustrial atmosphere and are solely a result of human activities.

By comparing emissions and concentrations it is evident that the persistence (or lifetimes) of the excess contributions of these gases in the atmosphere range from decades (for CH_4) to centuries (for CO_2 and some halocarbons) to thousands of years (for some perfluorocarbons). For this reason, even in the absence of future emissions, the excesses of their concentrations above natural levels are expected to persist for many centuries. Because of their long lifetimes, the concentrations of these gases tend to be quite uniform around the world, with local variations occurring indicating strong source regions (and for CO_2, possibly strong seasonal uptake or release by vegetation).

Aerosols

Observations of atmospheric composition also indicate that human activities are contributing to an increase in the atmospheric concentrations of small particles (called aerosols), primarily as a result of emission of sulfur dioxide (SO_2), soot, and various organic compounds. A large fraction of the emissions of these human-induced aerosols results from the combustion of fossil fuels (primarily from coal combustion and diesel and two-stroke engines), with a somewhat smaller fraction resulting from biomass burning. Once in the atmosphere these compounds are transformed and combined in various ways. For example, SO_2 emissions are transformed into sulfate aerosols that create the whitish haze common over and downwind of many industrialized areas. Changes in land cover, especially where this leads to desertification, can also lead to increased lofting of dust particles into the atmosphere.

Of critical importance is that the typical lifetime of aerosols in the atmosphere is less than 10 days. Sulfate and nitrate compounds, for example, are often removed when they first encounter rain systems, causing the acidification of precipitation known popularly as acid rain. Because of their relatively short lifetime in the atmosphere compared to greenhouse gases, the radiative influence of aerosols is most often regional. As a result, large and sustained emissions must occur for aerosols to have a climatic influence that is comparable to the warming influences of the long-lived greenhouse gases.

Because aerosols have direct effects on human health and on visibility, control measures have tended to limit aerosol buildup in most developed nations; however, aerosol concentrations have become quite high in particular regions in many developing nations and in regions where biomass burning is extensive. Recent research is indicating that, in addition to the pollutant effects, the aerosols are causing regional disturbances of the climate. For example, aerosols lofted in southern Asia, primarily from two-stroke engines and inefficient combustion in homes and factories, may be contributing to the diminishment of the monsoon (Lelieveld et al., 2001). In addition, aerosols can have hemispheric effects when they happen to be carried across oceans by the large-scale atmospheric circulation and do not encounter precipitation systems.

Summary

Although natural processes can also affect the atmospheric concentrations of gases and aerosols, observations indicate that natural factors have not been an important cause of significant changes in atmospheric composition over the past 10 000 years. Combined with the evidence of changes in source emissions, this preindustrial stability in atmospheric composition makes it clear that human activities have been the major cause of the dramatic increases in greenhouse gas concentrations since the start of the Industrial Revolution.

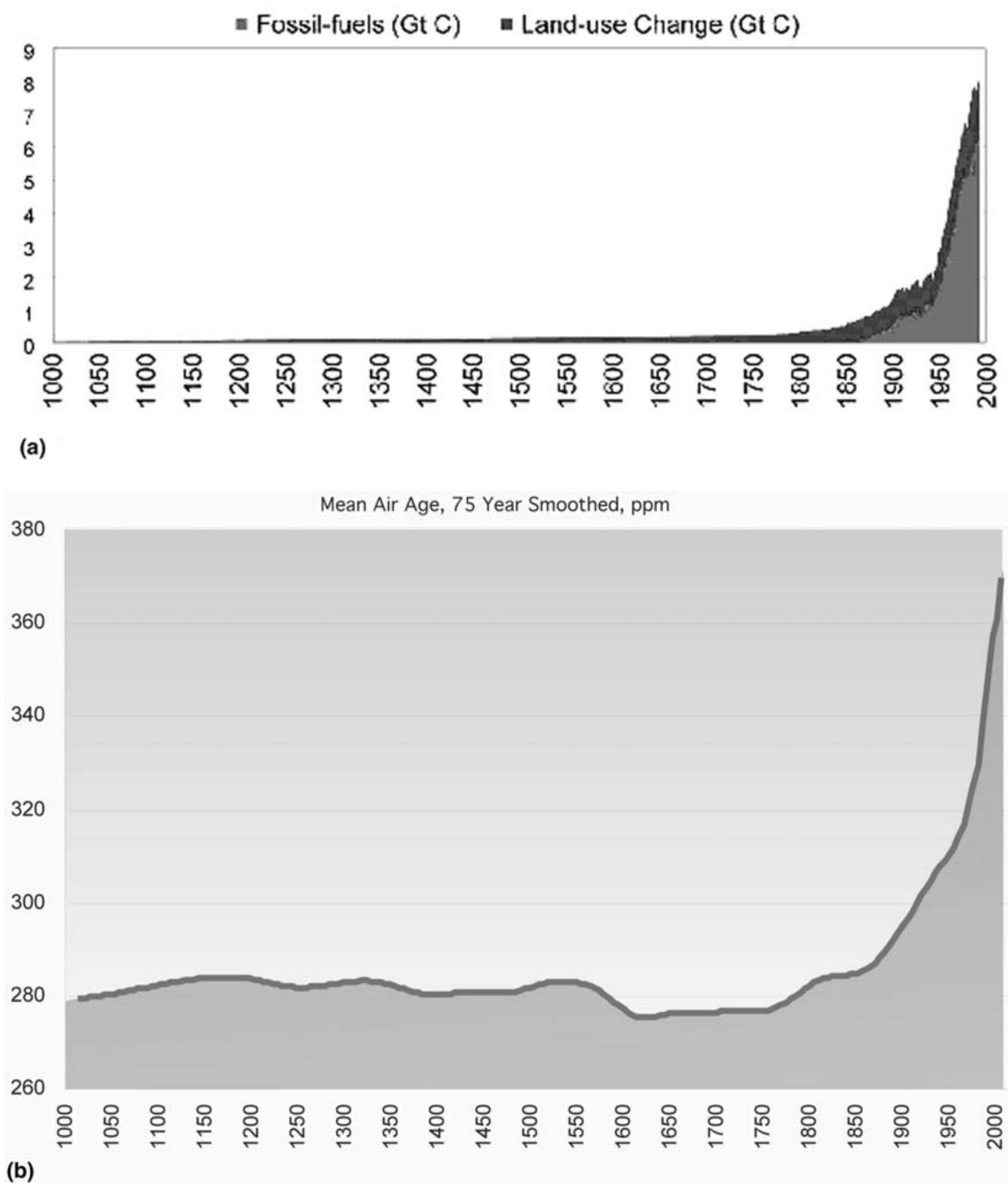

Figure C10 (**a**) Reconstructions of CO_2 emissions for the last 1000 years. Emissions of CO_2 since 1750 are based on compliations for land clearing and for combustion available from the Carbon Dioxide Information and Analysis Center (CDIAC, http//cdiac.esd.ornl.gov). (**b**) Reconstruction of the CO_2 concentration for the last 1000 years. This record is based on the concentration of CO_2 in ice core bubbles up to the early twentieth century (Etheridge et al., 1998) and instrumental records from Mauna Loa since 1957 (Keeling, 1999).

Intensification of the greenhouse effect

From laboratory experiments, from study of the atmospheres of Mars and Venus, from observations and study of energy fluxes in the atmosphere and from space, and from reconstructions of past climatic changes and their likely causes, it is very clear that the atmospheric concentrations and distributions of radiatively active gases play a very important role in determining the surface temperature of the Earth and other planets. Figure C11 provides a schematic diagram of the energy processes and fluxes that determine the Earth's temperature.

Of the solar radiation reaching the top of the atmosphere, about 30% is reflected back to space by the atmosphere (primarily by clouds) and the surface; about 20% is absorbed in the atmosphere (primarily by water vapor, clouds, and aerosols); and about 50% is absorbed at the surface. For each part of a system to come to a steady-state temperature the energy absorbed in each component of the system must be balanced by the energy lost. For the planet as a whole, the radiation that is emitted away as infrared (or heat) radiation must balance the amount of solar energy absorbed. Were the Earth's

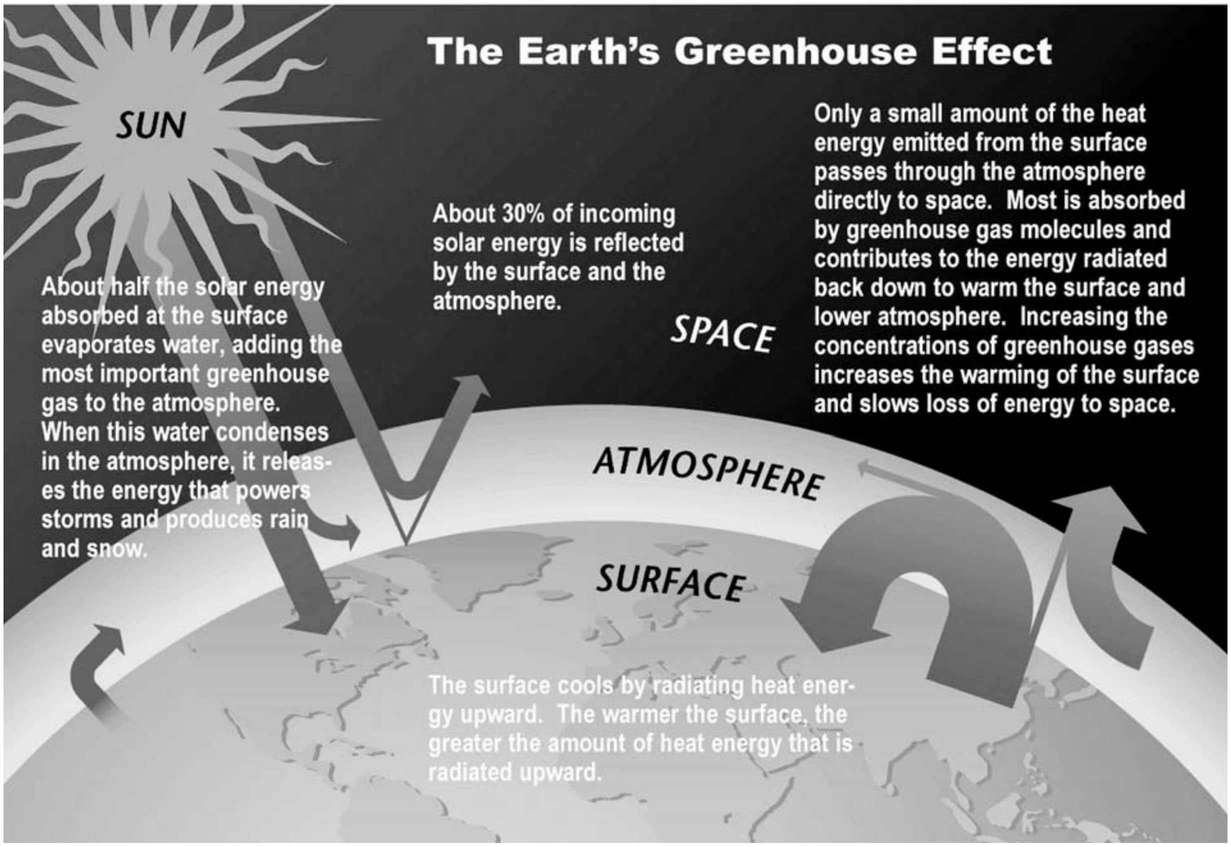

Figure C11 Schematic diagram of the energy fluxes that create the Earth's greenhouse effect. The width of the arrows is proportional to the amount of energy. Total incoming energy from the Sun averages 342 Wm^{-2} (NAST, 2000).

atmosphere transparent, and its surface a simple radiator of energy to space, the Stefan–Boltzmann equation could be used to calculate that the Earth's average surface temperature would equilibrate at close to 0°F (−18°C), given the current reflectivity of the Earth–atmosphere system. Such a temperature would be much too cold to sustain life as we know it.

However, the Earth's atmosphere is not transparent to infrared radiation; instead, a number of gases are able to absorb and emit infrared energy (to be radiatively active, molecules need to have at least three atoms so that various rotational and vibrational bands can be excited by the radiation). As a result of this property of particular atmospheric gases, a warming influence on the climate is created. This warming effect occurs because a large fraction of the infrared radiation emitted by the surface and by the greenhouse gases and low clouds in the atmosphere is absorbed by these radiatively active atmospheric gases before the energy is lost to space. For example, less than 10% of the infrared radiation emitted by the surface is at wavelengths that pass through the atmosphere directly to space without being absorbed. The energy that is absorbed in the atmosphere tends to warm it, in turn causing the radiatively active gases in the atmosphere and the clouds to emit infrared radiation. A significant fraction of this energy is radiated back toward the surface, providing additional energy to warm the surface. This radiation in turn causes the surface to warm, which raises its temperature and causes more radiation to be emitted upward, where much of it is again absorbed, providing more energy to be radiated back to the surface.

This emission–absorption–re-emission process that recycles energy back to the surface is popularly called the greenhouse effect, even though the processes are radiative and so are different from the limitations on loss of energy by convection and to freely flowing air that keep a greenhouse warm and humid. The Earth's natural greenhouse effect raises the average surface temperature from about −18°C (the temperature at which the Earth would radiate away as much energy as it absorbs from the Sun given its current albedo) to the observed value of about 15°C.

Radiative effects of greenhouse gases

The amount of this greenhouse warming is determined by the concentrations of greenhouse gases in the atmosphere. If the concentrations of greenhouse gases are increased, this causes an intensification of the natural greenhouse effect, inducing some warming. An additional warming influence results because the atmospheric temperature decreases with altitude up to the tropopause (about 10–12 km above the surface) before temperatures start to rise again in the stratosphere, which is warmed primarily as a result of solar absorption by ozone (O_3) molecules. When the concentrations of greenhouse gases are increased and the atmosphere becomes more opaque to infrared radiation, the atmosphere's vertical temperature structure (i.e.

its lapse rate) causes the absorption and reemission of infrared radiation to the surface to occur from lower and warmer layers in the atmosphere. Because the emission of infrared energy is proportional to the fourth power of temperature, this has the effect of increasing the downward-emitted radiation, tending to further enhance the natural greenhouse effect. Similarly, because of the resulting changes in the atmosphere's infrared opacity, increasing the concentrations of greenhouse gases tends to cause the emission of infrared radiation outward to space to occur from higher and colder layers; this leads to less radiation being emitted to space for a given average temperature of the lower atmosphere. With less infrared energy being emitted from the surface–atmosphere system to space, the excess of absorbed energy will cause the system to warm in order to re-establish the planetary energy balance.

The most important radiatively active (or greenhouse) gas is water vapor. Not only does water vapor absorb infrared radiation emitted from the Earth's surface and lower atmosphere, but it also absorbs infrared radiation from the Sun. In addition, under appropriate conditions, water vapor can condense and form clouds that absorb and emit infrared radiation as well as absorbing and scattering solar radiation. The amount of water vapor in the atmosphere is largely determined by the large-scale surface temperature of the Earth, with atmospheric circulation and precipitation processes causing its concentration to be highly variable throughout the atmosphere. Although most of the water vapor is found within about 2 km of the surface, the small amounts at higher levels in the troposphere have very strong influences on the fluxes of infrared radiation. Because of this, changes in atmospheric convection that lead to small changes in the amount of upper tropospheric water vapor could have important climatic effects.

Although human activities loft significant amounts of water vapor into the atmosphere, these additions tend to reduce the surface–atmosphere gradient of water vapor pressure. This tends to reduce the natural rate of evaporation of water vapor, thereby moderating the radiative effects of the injected water vapor. In addition, although there is a lot of water vapor in the atmosphere, the average lifetime of a water vapor molecule in the atmosphere is only about 8 days. This short lifetime makes it very difficult for the direct effects of human activities (e.g. by water vapor injection from cooling towers) to sustain a significant increase in water vapor's natural concentration. The only exception is for the case of water vapor injected by airplane exhausts at altitudes where water vapor molecule lifetimes are relatively long. Present levels of injection in this manner exert a relatively small warming influence that is balanced somewhat by the additional reflection of solar radiation from contrails.

Other than water vapor, the primary radiatively active gases are CO_2, CH_4, N_2O, and a number of HCFCs and CFCs. Because the concentrations of these gases are being directly affected by human activities, these greenhouse gases are usually referred to as the anthropogenic greenhouse gases (strictly speaking, their concentrations are being anthropogenically modified). In addition, the tropospheric and stratospheric concentrations of O_3, which is also a greenhouse gas, are being indirectly affected through chemical reactions caused by the emissions of gases such as non-methane hydrocarbons and nitrogen oxides, many of which are also tied to use of fossil fuels for energy.

Changes in the radiation balance

Changes in the concentrations of aerosols can also affect the radiation balance. Sulfate aerosols tend to be quite bright, and can exert a cooling influence on the climate by reflecting away a higher fraction of solar radiation than darker ground and ocean surfaces that typically lie below. Dust lofted by the winds also generally has a cooling influence on the climate. Because they are so dark and absorbing, soot aerosols by themselves tend to enhance absorption of solar radiation, thereby creating a warming influence, especially over very light surfaces. The various types of aerosols can also combine to form mixed-composition aerosols that can exert either warming or cooling influences, creating significant uncertainty in the overall effect of changes in aerosol loading.

Changes in land cover can also alter the energy balance of the Earth system. Although changes in vegetation from one year to the next can seem quite small, over time they can become quite extensive. For example, most of the northeastern US was forested in pre-settlement times, then mostly cleared by the early twentieth century due to the demands for farmland, wood for ships and buildings, and charcoal to make steel. This change tended to brighten the surface, exerting a cooling influence. Over the last 100 years, however, as farming shifted to the Great Plains and the Southwest, the forests have regrown and now cover the region, tending to darken it and exert a warming influence. Some studies indicate that such changes can have not only regional effects on the climate but, by altering the atmosphere's heat balance, these changes can also cause shifts in the atmospheric circulation that result in climatic changes elsewhere. With changes of land cover occurring in many regions around the globe (e.g. deforestation in Amazonia and in southern and western Asia), significant regional perturbations in climate are likely to be already occurring along with the growing increment in the global average temperature due to the rising concentration of greenhouse gases.

In addition to changes in the radiation balance occurring due to human activities, changes can result from natural processes. Most noticeable are the influences of volcanic eruptions, which can loft millions of tonnes of SO_2 and dust into the stratosphere, where it can spread out over much of the globe and persist for up to a few years. Major eruptions, such as Mount Pinatubo in the Philippines in 1991, can reduce incoming solar radiation by of order of 1% and cause cooling of order of 0.5°C for a year or more. Major volcanic activity has tended to be infrequent, and no long-term trend in the effects of volcanoes on the radiation budget has been found, although there have been some relatively active and inactive decades.

Variations in the Sun's output are a second natural source of changes in the Earth's radiation balance. Satellite observations have confirmed indications from the surface that the Sun's output varies with the cycle of sunspots, and other indications suggest that there may be multidecadal to multicentury variations. These variations are estimated to amount to perhaps a few tenths of 1% of the solar flux, and may have caused variations in global average temperature of roughly 0.5°C over the last few thousand years.

Over periods of many thousands of years, other natural forcings can also come into play. For example, the Yugoslavian scientist Milutin Milankovitch was the first to calculate the effect on the Earth's radiation balance resulting from its orbit about the Sun being elliptical rather than circular. Over the course of tens of thousands of years, changes in eccentricity (i.e. departure from circularity), tilt of the Earth's axis, and precession of the seasons (time of year of nearest approach to the Sun) cause a redistribution of the incoming solar radiation in latitude and season. That the timing of these orbital changes and the cycling of the glaciers coincide suggests that these changes have been the drivers of glacial cycling, especially over the last million years. The ice-core records of changes in atmospheric composition suggest that variations in greenhouse gas concentrations are likely one of several additional processes that have contributed to amplifying the climatic response to the orbital changes identified by Milankovitch.

The relative climatic influences of natural forcings and of changes in the atmospheric amounts of gases and aerosols are typically compared by using atmospheric radiation models to calculate their ability to change the fluxes of solar and terrestrial radiation at the tropopause; that is, to alter the amount of energy entering or leaving the surface–troposphere system. The change in these fluxes as a result of changes in the amounts of gases and aerosols is referred to as the "radiative forcing" that they exert. Based on the latest international assessment report of the IPCC (2001), Table C2 summarizes the best estimates of the changes in abundances of various gases and aerosols and the radiative forcings that are estimated to have resulted since

Table C2 Greenhouse gases and aerosols contributing to changes in climate (based on information in IPCC, 2001)

Greenhouse gas or aerosol or other forcing mechanism	Approximate abundance in 1750	Approximate abundance in 2000	Estimated radiative forcing from 1750–2000 ($W\,m^{-2}$)	Projected abundance in 2100 based on SRES emissions scenarios (IPCC, 2000)	Projected radiative forcing from 1750–2100 ($W\,m^{-2}$)
Carbon dioxide (CO_2)	278 ppmv	369 ppmv	1.46 (plus or minus 10 per cent for all well-mixed greenhouse gases)	549–970 ppmv (reference model)	3.64–6.69
Methane (CH_4)	700 ppbv	1745 ppbv	0.48	1574–3413 ppbv	0.42–1.09
Nitrous oxide (N_2O)	270 ppbv	314 ppbv	0.15	354–460 ppbv	0.27–0.57
Perfluoromethane (CF_4)	40 pptv	80 pptv	0.003	208–397 pptv	0.013–0.029
CFC-11 ($CFCl_3$)	0	268 pptv	0.07	45 pptv per Montreal Protocol	0.01
CFC-12 (CF_2Cl_2)	0	533 pptv	0.17	222 pptv per Montreal Protocol	0.07
Other anthropogenic halocarbons (generally contain C, H, and Cl, Br, or F)	0	Various	~0.1	Various	~0.06–0.33
Tropospheric ozone (O_3)	Increase of ~8 Dobson units		0.35 (0.2–0.5)	−18% to +5%	0.21–1.27
Stratospheric ozone (O_3)	Depletion by several percent since about 1970		−0.15 (−0.05 to −0.25) (1979–1997)	Recovery from halocarbon injections, but some depletion due to stratospheric cooling	Uncertain, but possibly less negative
Sulfate aerosol (SO_4) – direct effect		0.52 TgS	−0.4 (−0.2 to 0.8)	0.15–0.45 TgS	−0.12 to −0.35
Black carbon (BC)		0.26 Tg	0.2 (0.1–0.4)	0.13–0.68 Tg	0.2–1.05
Organic carbon (OC)		1.52 Tg	−0.1 (−0.03 to 0.3)	0.77–4.00 Tg	−0.88 to −1.32
Biomass burning aerosols			−0.2 (−0.07 to 0.6)		
Aerosol – indirect effects due to change in cloud reflectivity			0 to −2.0		Uncertain
Mineral dust			−0.6 to 0.4		
Contrails due to jet aircraft			0.02 (0.00 to 0.07)		
Aviation-induced cirrus			0 to 0.04		
Change in surface vegetation			−0.20 (0 to −0.4		
Solar irradiance			0.30 (0.1–0.5)		

[a] From 1993 to 2002, on assignment from the Lawrence Livemore National Laboratory to the Office of the US Global Change Research Program, serving as Executive Director of the Office from 1993 to 1997 and as Executive Director of the National Assessment Coordination Office from 1997 to 2001. Presently serving as Chief Scientist for Climate Change Programs at the Climate Institute, Washington DC, and president (2003–07) of the International Association of Meteorology and Atmospheric Sciences.

the beginning of the Industrial Revolution in about 1750. Although there is considerable uncertainty in the estimates of the radiative forcing of the various types of aerosols on the radiation balance, the net effect of all of the changes that have occurred is most likely to have been significantly positive, thereby exerting a warming influence on the global climate. Indeed, observations from space-based instruments since the 1970s clearly indicate that the rising concentrations of the anthropogenic greenhouse gases are indeed tending to enhance the natural greenhouse effect.

In addition to the direct effects of changes in greenhouse gas concentrations and aerosol loading on the radiation balance, there are several types of indirect influences on the radiative forcing of a gas or aerosol. For example, changes in the amounts and types of aerosols can affect the amounts and reflectivities of clouds that in turn alter the fluxes of solar and terrestrial radiation. There can also be indirect influences through various effects on atmospheric chemistry. For example, the chemical decomposition of methane in the stratosphere creates water vapor that has a powerful greenhouse effect, and the release of various halocarbons has depleted the stratospheric ozone layer, affecting its role in the infrared radiation balance. Estimates of these indirect influences are also included in Table C2.

Feedback mechanisms

The intensity of the Earth's natural greenhouse effect is also altered through various feedback mechanisms that can amplify or moderate the system's response to the changes in radiative forcing. For example, the warming influence caused by increases in the concentrations of CO_2, CH_4, and other anthropogenic greenhouse gases can be significantly amplified by a positive water-vapor feedback mechanism. This positive feedback occurs because more water vapor can be present in a warmer atmosphere, and an increase in atmospheric water vapor enhances the greenhouse effect and causes a further warming that in turn allows more water vapor to be present in the atmosphere. At the same time, however, changes in the amount of atmospheric water vapor (more specifically, in conditions affecting water vapor removal when the relative humidity exceeds saturation) and in atmospheric circulation can alter the extent and distribution of clouds. These changes in cloud cover and distribution can in turn affect the extent of the absorption and scattering of solar radiation and the absorption and re-emission of infrared radiation, thereby creating both amplifying and moderating feedbacks.

There are many other feedbacks and interactions. For example, warming tends to reduce the amount of snow and sea ice, which in turn reduces the reflectivity of the surface (i.e. the albedo), allowing increased absorption of solar radiation, and reinforces the original warming. Over the longer term, melting of glaciers and ice sheets can similarly enhance the warming influence. Changes in the climate can also affect the type of vegetation and its extent, affecting not only the surface albedo, but also its roughness, the rate of evapotranspiration from the surface, and the lofting of dust. The ocean circulation can also be affected by the climate, with wind changes affecting surface current and changes in ocean temperature and salinity causing global-scale changes in the thermohaline circulation that ventilates the deep ocean. Over thousands of years and longer, glaciological and isostatic processes can affect the height of the land and even the shapes of coastlines, which in turn can affect atmospheric and oceanic conditions. There are thus a great many processes and feedbacks that can potentially come into play once changes in the climate are initiated.

Climate sensitivity

Estimates of the climate's sensitivity to changes in radiative forcing have been derived from theoretical analysis, from reconstructions of climate changes in the geological past, and from simulations using global climate models. These approaches indicate that, at steady state, each increase in the radiative forcing by $1\,W\,m^{-2}$ seems to be associated with an increase in the global average temperature of about 0.4–1.2°C. This degree of sensitivity is also often expressed in terms of the warming that would result from the change in radiative forcing associated with a doubling of the atmospheric CO_2 concentration. When expressed in this manner, the radiative forcing from the doubling the CO_2 concentration (i.e. almost $4\ Wm^{-2}$) would lead to an increase in the global average temperature of 1.5–4.5°C. It is important to emphasize that these estimates are after a new steady-state climate is established, and that, while much of the readjustment would occur over a few decades, the full response would take many centuries because of the time it takes for the deep ocean temperature and feedback mechanisms involving glacial ice and ecological systems to come to a new equilibrium.

Overall, there is broad scientific agreement that increases in the atmospheric concentrations of the anthropogenic greenhouse gases will enhance the Earth's natural greenhouse effect and tend to raise the global average surface temperature by of order of a few degrees Celsius. The remaining uncertainties relate to pinning down by how much and how rapidly warming would occur.

Detecting and attributing past changes in the climate

With the evidence being clear-cut that the concentrations of greenhouse gases have risen significantly since the start of the Industrial Revolution, and that increasing the concentrations of greenhouse gases will induce a warming influence on the Earth's climate, then a key test of scientific understanding of the climatic influence of these changes is to determine if the time history and magnitude of historic changes in the climate match those expected to be occurring as a result of past changes in radiative forcing. Complications in this analysis arise for several reasons. First, there are shortcomings in the climate record, including a near-absence of observations at times before human-induced forcing began to have an influence on the global climate. As a result, *detecting* whether changes have actually occurred requires care to distinguish actual change from changes arising due to limitations in the data (e.g. due to the changing spatial coverage over time) and from the natural variability of the climate caused mainly by varying atmosphere–ocean interactions. Second, there are multiple influences contributing to changes in the radiative forcing; most are not definitively quantified and it is not even clear if all are yet recognized. As a result, *attributing* particular changes in the climate to particular factors requires very careful forensic analysis.

To have the best chance of both detecting and attributing the human influence, it is most useful to look at the longest records of the changing climatic state. Instrumental records of temperature go back only a few hundred years, and a reasonable network of observation stations was only established around the middle of the nineteenth century. Accounting carefully for changes in measurement technique and the local environment, records of how the monthly and annual average temperatures at each of thousands of locations have changed from their local baseline value can be used to construct a space and time history of such changes. These results can then be used to calculate an area-weighted global average of the change in local near-surface air temperature. This quantity is often referred to simply as the change in the global average temperature, even though this terminology really obscures how its value is determined and unfortunately implies that the global average temperature is actually measured in some direct way by some single instrument.

Detecting changes in climate

Spatially representative estimates of the change in global average temperature go back to about 1860. These records indicate a warming of over 0.6°C, mostly during the early and late twentieth century. Extensive proxy records (e.g. records derived from tree rings, ice cores, coral growth, etc.) have been used to attempt to reconstruct the amount and pattern of changes in surface temperature further back into the past. For the northern hemisphere, proxy records going back about 1000 years indicate that a slow cooling was under way for most of this period. These records also clearly indicate that the twentieth century was much warmer than earlier centuries and that the decade of the 1990s was almost certainly the warmest decade of the last 1000 years. This temporal history is clearly evident in Figure C12, which indicates that a sharp rise in the average temperature began during the late nineteenth century and has continued throughout the twentieth century. It is also clear that the twentieth-century warming has been much more persistent than earlier fluctuations, which were likely caused by the inherent natural variability of the ocean–atmosphere system (i.e. internally created variability) and natural variations in solar radiation and the occasional eruption of volcanoes (i.e. externally caused variability).

The global warming evident at the surface is also consistent with observations of increasing temperatures measured in boreholes (i.e. dry wells), retreating mountain glaciers, diminishing sea ice in the Arctic, increasing concentrations of atmospheric water vapor, rising sea level due to melting of mountain glaciers and thermal expansion in response to recent warming (augmenting the natural rise due to the long-term melting of parts of Antarctica), and related changes in other variables. As a whole, most records indicate that long-term changes in the climate have been occurring, even though some analyses of some shorter records indicate very little change (e.g. one analysis of changes in tropospheric radiance measured using microwave instruments on a succession of satellites indicates virtually no increase in the implied tropospheric average temperature since 1979, although another analysis and balloon observations of changes in tropopause height do indicate that tropospheric warming has been occurring).

Attributing climate change to human activities

A key question is whether these changes are mainly a natural fluctuation or whether human activities are playing a significant role. The twentieth-century warming is largely being attributed to

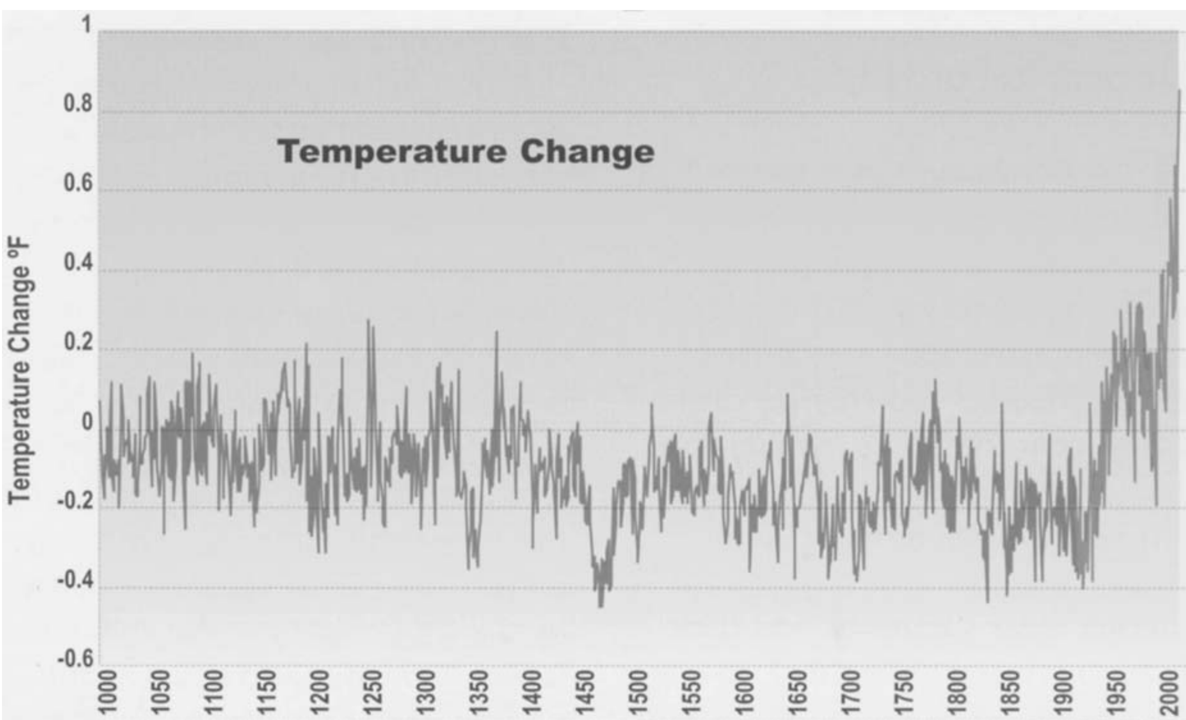

Figure C12 Reconstruction of the northern hemisphere average surface temperature for the last 1000 years. Up to the mid-nineteenth century the record is based on proxy records (e.g. tree rings analyzed by Mann et al., 1999). The estimated uncertainty for an individual year is several tenths of a degree. Since the mid-nineteenth century the record is based on instrumental data, for which the comparable uncertainty is about 0.2°F.

human activities because of the coincidence in timing with the warming influences exerted by the changes in greenhouse gas concentrations, the very large and unusual magnitude of the changes compared to past natural fluctuations, the warming of the lower atmosphere and cooling of the upper atmosphere (a sign of a change due to changes in greenhouse gas concentrations rather than changes in solar radiation), and the global pattern of warming (with larger changes in high than low latitudes). Some uncertainty is introduced because some of the warming occurred before the sharpest rise in greenhouse gas concentrations in the latter half of the twentieth century. For example, some analyses indicate that as much as 20–40% of the warming may be due to a coincidental increase in solar radiation, although other factors, such as changes in land cover or in soot emissions, may also have had an influence. In addition, some uncertainty has been introduced because the rise in tropospheric temperatures over the past two decades may have been a bit slower than the rise in surface temperature. Whether this difference is real or arises from, for example, calibration issues with the satellite instrumentation, natural variations in Earth-surface temperatures, or the confounding influences of ozone depletion, volcanic eruptions, and atmosphere–ocean interactions, is not yet clear.

Taking all of the scientific results into consideration, the IPCC concluded in its Second Assessment Report (IPCC, 1996) that "The balance of evidence suggests a discernible human influence on the global climate". This conclusion, in essence, is equivalent to the criterion for a civil rather than a criminal conviction in a court of law. The IPCC's Third Assessment Report (IPCC, 2001) documented even more clearly that the magnitude and timing of the warming during the twentieth century quite closely matches the changes that would be expected from the combined influences of human and known natural influences, concluding that "there is new and stronger evidence that most of the warming over the past 50 years is attributable to human activities".

Projections of future changes in emissions and atmospheric composition

The continued reliance on fossil fuels for generation of energy and the release of other greenhouse gases are sure to cause further changes in atmospheric composition. Along with the continuing adjustment of the climate system to past changes in atmospheric composition, these changes will far exceed changes in other contributions to radiative forcing, making them the driving influences for future climate change. While it is not possible to provide definitive predictions of how the world will evolve out for decades to centuries into the future, the IPCC (2000) reviewed projections of various trends for the future and, using a number of storylines about how population, energy use and energy technologies, economic development, and international cooperation might evolve, developed a set of plausible scenarios of future emissions of greenhouse gases and some types of aerosols and aerosol precursors. These scenarios indicate, for example, that annual emissions of CO_2 in the year 2100 could range from roughly 5 to over 30 GtC/yr (billions of tonnes of carbon per year) compared to present emissions of about 6 GtC/yr.

A rough indication of what the emissions might be can be derived by simply extending trends from current values. With the global population of 6 billion people and current use of fossil fuels, each person on Earth is presently responsible, on average, for emission of about 1 tC/yr (tonne of carbon per year). Per-capita use, however, varies widely across the world, reaching nearly 6 tC/yr in North America and about 3 tC/yr in Western Europe, but amounting to only about 0.5–1.0 tC/yr in developing countries such as China and India.

Central projections for the year 2100 foresee a global population of 8–10 billion. With the rising standard of living and the energy required to sustain it, and with continued reliance of fossil fuels for most of the world's energy, global average per-capita emissions of CO_2 seem likely to at least double, with most growth resulting from increased emissions in the highly populated, but currently underdeveloped, emerging economies. If this occurs, total annual emissions could more than triple, reaching roughly 20 GtC/yr or more. IPCC (2000) also considered the potential for variations of various kinds around an estimate such as this, with the possibility that intensified use of coal, which is among the least expensive and most available sources of energy in the developing world, could cause emissions to be even greater, whereas a drop in the rate of population growth and a greatly accelerated switch to non-carbon-based fuels could return emissions in 2100 to about current levels after higher levels through most of the twenty-first century.

If global emissions of CO_2 do gradually increase to about 20 GtC/yr by 2100, models of the Earth's carbon cycle that are verified against observations of past increases in concentration, project that the atmospheric CO_2 concentration would rise to just over 700 ppmv. This would be almost double its present concentration, and over 250% above its preindustrial value.

For the broader set of emission scenarios that are based on consideration of ranges in global population, energy technologies, economic development, and other factors, projections indicate that the atmospheric CO_2 concentration is likely to reach at least 500 ppmv and may even exceed over 900 ppmv in 2100. The projections also indicate that the CO_2 concentration would continue to increase during the twenty-second century unless global emissions have been reduced to 1–2 GtC/yr by 2100.

Depending on various control measures, the rise in the CO_2 concentration would be expected to be accompanied by increases in the concentrations of CH_4 and N_2O, as well as of sulfate and soot aerosols, although it would take an unrealistically large (and unhealthy) amount of aerosols to affect radiative forcing to an extent that would counterbalance the radiative forcing created by the increases in the concentrations of the anthropogenic greenhouse gases that are projected. Based on IPCC scenarios and analyses (IPCC, 2000, 2001), Table C2 provides estimates of the future abundances of the various gases and aerosols as well as an estimate of the total anthropogenic contribution to radiative forcing expected by the year 2100.

Because the historical temperature record includes natural fluctuations, and because the relative contributions to radiative forcing of the various greenhouse gases and aerosols are changing, extrapolation of historic trends in temperature into the future is not a particularly reliable means for projecting future trends. Similarly, although the geological record provides evidence that both the atmospheric composition and climate have changed in the past, only a rough indication can be provided of how the climate will change over the twenty-first century because the change in atmospheric composition is occurring so rapidly. Because of these limitations, and in order to get a more complete sense of how the many climatic variables would be expected to change, projections of the climatic effects of the various scenarios of changes in greenhouse gas and aerosol emissions must necessarily be based on the use of computer-based climate models.

Global climate models

Global climate models (also known as Earth system models) are assembled by coupling general circulation models of the atmosphere and oceans to land surface and cryospheric models. Each of these component models is constructed by applying the fundamental conservation laws of physics (i.e. for mass, momentum, and energy) and chemistry (i.e. for water and various other species, including chemical transformations) to many thousands of grid cells that are used to cover (or tile) the Earth. Although atmospheric and oceanic models with resolutions of order 10 km in horizontal resolution are starting to become available, developing reliable projections of climatic change requires running multiple simulations for hundreds to thousands of years. As a result of limits imposed by present computer speed and resources, a significantly coarser resolution (e.g. 100–200 km) is being used for most model simulations of climate change. Because this resolution is not adequate to represent the details of, for example, convective cloud development and wind drag caused by variable terrain, the effects of these processes are represented in the models in terms of larger-scale variables using parameterizations that are typically derived from the results of large-scale observations or extensive field investigations of how atmospheric or oceanic processes work. In that these parameterizations are only approximate, and because it is not possible to ensure that all processes have been represented, it is essential to test the models and to carefully evaluate their results.

To develop a sense of confidence in the models a wide range of tests is often performed. The most intensive efforts have been conducted under the auspices of the Program for Climate Model Diagnosis and Intercomparison, which has programs to compare the results of the various model components with observations and with the performance of other models (e.g. Gates et al., 1999). Comparisons of model simulations to the observed behavior of the climate typically cover recent decades, recent centuries, and even geological periods in the past. In general, the models do reasonably well at representing the latitudinal and seasonal distributions of the climate, but are still somewhat limited in their representations of climatic variability on scales from the El Niño/Southern Oscillation to the North Atlantic Oscillation. As indicated above, when driven by both natural and human-induced forcings, the models reasonably represent the changes in global average temperature observed over the twentieth century (see IPCC, 2001, chapter 8).

Projections of changes in climate and sea level

To generate projections of changes in the future climate, the models are typically initiated at conditions that have been reconstructed for the mid-nineteenth century before human activities were contributing substantially to radiative forcing. A control simulation with no change in forcing is then run to establish an estimate of the baseline climate in the absence of human activities. These results are then compared to the results of perturbation simulations that are driven by the scenarios of future emissions and/or changes in atmospheric composition, depending on the model and the type of forcing.

Based on the model simulations, and supported by theoretical analyses and extrapolations of recent trends, the set of IPCC emissions scenarios (IPCC, 2000) for the twenty-first century is projected to cause the global average temperature to increase by about 1.4–5.8°C over the period 1990–2100 (IPCC, 2001). A warming of this amount would be roughly 2–9 times as large as the warming during the twentieth century. Such a warming would mean that human activities would have increased global average temperature over its preindustrial average by from about 2°C to over 6°C. Model simulations are increasingly providing consistent indications of the warming to be expected in various regions of the Earth, indicating larger changes occurring in higher rather than lower latitudes and over land rather than the ocean (IPCC, 2001, chapter 10). Increases in the average temperatures projected for the end of the twenty-first century, even were they to be at the lower bound, would significantly exceed the temperatures that occurred over the last 1000 years, and most likely since the emergence from the last glacial maximum.

Many types of changes in the weather and climate would result. For example, shifts in storm tracks and an intensification of evaporation and precipitation cycles would be expected to alter the frequency and intensity of floods and droughts. IPCC (2001) also projects that the intensity of various other extreme weather and climate events would increase.

In addition, the projections indicate that the melting of mountain glaciers and thermal expansion of warming seawater would cause the rate of sea level rise to increase from the observed rate of 0.1 to 0.2 m/century during the twentieth century. Because of uncertainties relating to the fates of the Greenland and West Antarctic ice sheets, the estimated rate of rise is quite broad, ranging from about 0.1 to 0.9 m/century during the twenty-first century. The lower estimate of projected sea level rise reflects the possibility that, for perhaps the next century, increased snowfall could cause ice to accumulate over major areas of Greenland and Antarctica more rapidly than it would be lost at the edges. Higher rates of rise could occur during the latter part of the twenty-first century and are projected for the following centuries as the warming over Greenland (and perhaps West Antarctica) initiates a millennium-long melting that could add 0.5–1 m/century to the rate of future sea level rise.

As was the case for the sudden appearance of the Antarctic ozone hole, there is also a real possibility for surprising changes to occur, especially given the potential for thresholds and nonlinearities. One of the possibilities is the disruption of the Gulf Stream and the larger-scale deep ocean circulation of which it is a part. The weakening of the world-girdling thermohaline circulation apparently occurred during the Younger Dryas period about 11 000 years ago, which was an interruption in the recovery from the glacial conditions that had prevailed for the preceding 100 000 years. The weakening of the global ocean circulation, in that case likely caused by a sudden release into the ocean of glacial meltwaters, led to a strong cooling centered over Europe. Were a similar weakening of the thermohaline circulation to occur during the twenty-first century, the cooling temperatures would likely only be regionally significant, even though it might only moderate the influence of global warming. However, because such a change would reduce the rate of cold-water transport into the deep ocean, the rate of sea level rise would tend to increase, creating problems for coastal regions around the world.

An associated item in this volume (Climate Change Impacts) summarizes the expected types of impacts resulting from changes in climate of the magnitude discussed here. That item also indicates the magnitude of the emissions reduction that would be required to stabilize the atmospheric concentrations of greenhouse gases and eventually stabilize the climate, as has been set as an international objective in the Framework Convention on Climate Change that was negotiated in 1992, and has since been agreed to by most of the nations of the world.

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Cross-references

Aerosols
Climate Change Impacts: Potential Environmental and Societal Consequences
Climate Variations
Energy Budget Climatology
Greenhouse Effect and Greenhouse Gases
Models, Climatic

CLIMATE CHANGE AND HUMAN HEALTH

Global climate change is one of a larger set of large-scale environmental changes occurring in today's world. These changes reflect, in various ways, the increasing human domination of the ecosphere, as human numbers continue their unprecedented expansion and as human economic activities intensify (Vitousek et al., 1997). All of these changes – climate change, stratospheric ozone depletion, biodiversity loss, worldwide land degradation, freshwater depletion, disruption of the elemental cycles of nitrogen and sulfur, and the global dissemination of persistent organic pollutants – have great consequences for the sustainability of ecological systems, for food production, for human economic activities and, via those and other pathways, for human population health.

There is a growing realization that the sustainability of human health should be a central consideration in the public debate on sustainable development (McMichael et al., 2000). After all, society's ultimate, if unstated, objective is to maximize the wellbeing, health and survival of its population. The advent of climate change and other global environmental changes, as an indicator that our current developmental trajectory is unsustainable, is obliging us to take a more ecological view of population health. That is, population health is an index of the extent of our success in the longer-term management of social and natural environments.

A change in world climate will influence the structure and functioning of many ecosystems and the biological health of many plants and creatures. Indeed, it has emerged, in recent decades, that many non-human physical and biological systems have undergone alterations that are reasonably attributable to climate change. This includes the retreat of glaciers and sea ice, and the earlier occurrence of bird-nesting and insect migrations (IPCC, 2001; Root et al., 2003). Likewise, there would be various health impacts in human populations. These would differ by location since climate change will induce a range of environmental changes specific to the particular geographical setting. Besides, the vulnerability of each human population will vary as a function of locality, level of material resources, technological assets and type of governance.

Some of the anticipated health impacts would be beneficial. For example, milder winters would reduce the seasonal wintertime peak in deaths that currently occurs in temperate countries, whereas in currently hot regions a further increase in temperatures might reduce the viability of disease-transmitting mosquito populations. Overall, however, scientists assess that most of the health impacts of climate change would be adverse. The health impacts of climate change do not entail novel processes and unfamiliar disease outcomes (in contrast to the recent surprise appearances of HIV/AIDS and human "mad cow disease"). Rather, they entail climate-induced changes in the frequency or severity of familiar health risks – such as floods, storms and fires; the mortality toll of heatwaves; the range and seasonality of infectious diseases; the productivity of local agro-ecosystems; the health consequences of altered freshwater supplies; and the many repercussions of economic dislocation and population displacement.

The human species, via its social organization and cultural practices, is much better – and often intentionally – buffered against environmental stressors than are all other plant and animal species. Hence, *Homo sapiens* is likely to be affected less soon and less sensitively than most other species. Not surprisingly, therefore, at this early stage in the process of global climate change there is little empirical evidence that climate change has already begun to affect human health.

Climate and human health: an ancient struggle

> Whoever wishes to investigate medicine properly, should proceed thus: in the first place to consider the seasons of the year, and what effects each of them produces, for they are not all alike, but differ much from themselves in regard to their changes. [Hippocrates, in *Airs, Waters, and Places* (Hippocrates, 1978)]

The idea that states of human health and disease are linked with climatic conditions almost certainly predates written history. The Greek physician Hippocrates, around 400 BC, related epidemics to seasonal weather change. He wrote that physicians should have “due regard to the seasons of the year, and the diseases which they produce, and to the states of the wind peculiar to each country and the qualities of its waters”. He exhorted them to observe “the waters which people use, whether they be marshy and soft, or hard and running from elevated and rocky situations, and then if saltish and unfit for cooking” and to note also “the localities of towns, and of the surrounding country, whether they are low or high, hot or cold, wet or dry . . . and of the diet and regimen of the inhabitants”.

Two thousand years later, Robert Plot, Secretary to the newly-founded Royal Society in England, collated weather observations in 1683–84 and noted that if the same observations were made “in many foreign and remote parts at the same time” we would “probably in time thereby learn to be forewarned certainly of divers emergencies (such as heats, colds, dearths, plagues, and other epidemical distempers)”.

In those intervening 2000 years many climatic disasters have affected communities and populations around the world, causing starvation, infectious disease, social collapse and, sometimes, the disappearance of whole populations. One such example is the mysterious demise of the two Viking settlements in Greenland in the fourteenth and fifteenth centuries, as temperatures in and around Europe began to fall (Pringle, 1997). These culturally conservative livestock-dependent settlements were established during the tenth century ad early in the Medieval Warm Period. They could not cope, however, with the progressive deterioration in climate that ensued from the late Middle Ages. Food production declined and food importation became more difficult as sea ice persisted. The Viking settlements eventually died out or were abandoned.

There are numerous historical accounts of acute famines occurring in response to climatic fluctuations (Bryson and Murray, 1977). Throughout pre-industrial Europe, food supplies were marginal, and the mass of people survived on monotonous diets of vegetables, grain gruel and bread. A particularly dramatic example in Europe was the great medieval famine of 1315–17. Climatic conditions were deteriorating, and the cold and soggy conditions led to widespread crop failures, food price rises, hunger and death. Social unrest increased, robberies multiplied, and bands of desperate peasants swarmed over the countryside. Animal diseases proliferated, contributing to the die-off of over half the sheep and oxen in Europe. This tumultuous event, and the Black Death that followed 30 years later, may have contributed to the weakening and dissolution of feudalism in Europe.

Over these and the ensuing centuries, average daily food intakes were less than 2000 calories, falling to around 1800 calories in the poorer regions of Europe. This caused widespread malnutrition, susceptibility to infectious disease, and low life expectancy. Recurrent famines culled the populations, often drastically. In Tuscany, for example, there were over 100 years of recorded famine between the fourteenth and eighteenth centuries. Meanwhile in China, where the rural diet of vegetables and rice accounted for an estimated 98% of caloric intake, in the 2000 years between around 100 BC and AD 1900 there were famines that involved at least one province in over 90% of years (Bryson and Murray, 1977).

The health impacts of climate change

It is important to stress, again, that there is yet little direct empirical evidence of climate change affecting human health over recent decades. In part this is because epidemiologists have had little interest in studying the relationship of climatic variations to human health. The low premium accorded to this topic has presumably reflected a general assumption that there are few opportunities for direct intervention to reduce adverse health impacts from climatic influences. Nevertheless, the position has changed markedly in recent years in the wake of newly recognized “anthropogenic” global climate change.

This upsurge in research activity falls into three categories: (a) studies seeking to broaden the empirical information base by examining how recent variations or trends in climatic conditions have related to changes in health outcomes, (b) studies seeking evidence of early impacts of global climate change on human health, and (c) studies that use existing empirical and theoretical knowledge to model the future impacts of given scenarios of climate change.

It is usual to distinguish several generic categories of health impacts due to climate change – first, those that arise from relatively direct impacts of alterations in temperature, precipitation and extreme weather events; second, those that occur in response to climate-induced changes in ecological processes and systems; and, third, those that result from the economic and demographic dislocation of human communities.

Direct effects

The more direct and immediate impacts include those due to changes in exposure to very cold and very hot weather extremes and those due to increases in extreme weather events (floods, tropical cyclones, storm-surges and droughts). Climate change would also directly increase the production of certain air pollutants, such as tropospheric ozone (the formation of which is affected by both temperature and level of sunlight), and various aeroallergens (spores and molds) involved in the causation of asthma, hay fever and other allergic disorders.

Populations display a characteristic pattern of daily deaths (and hospitalizations) in relation to daily temperature (Wilkinson et al., 2002). Typically, death rates increase both with greater heat and greater cold. In cool temperate countries, climate change would result in a decrease in winter mortality because of less severe winters, and this may offset the increases in summer heat-related mortality (Langford and Bentham, 1995; Rooney et al., 1998), at least initially. In warm temperate and tropical countries the overall number of temperature-related deaths is likely to increase. The net balance of future changes in hot and cold effects remains contentious and, anyway, varies substantially between different geographic regions, states of economic development, and between urban and rural populations.

The extent to which the regional frequency of extreme weather events will alter because of climate change remains somewhat uncertain. Climatologists, however, are becoming increasingly confident of their modeled regional predictions, as the downscaling of their climate models improves. Further, extreme weather events appear to have increased in tempo during the 1990s (e.g. McMichael et al., 2003; Milly et al., 2002), with commensurate risks to human life and limb. Nevertheless, despite the potentially great impact of such events on deaths, injuries and consequent diseases (infections, malnutrition and mental health disorders), estimating the future profile of health impacts is at best an indicative exercise. Hurricane Mitch, centered in Honduras in 1998, had a death toll of over 11 000 people, and was followed by tens of thousands of new cases of malaria, cholera and dengue fever.

Indirect effects (especially infectious diseases and food yields)

The indirect, second, category of health impact refers particularly to changes in the transmission patterns of infectious diseases and changes in regional food-producing systems. These impacts are, in essence, a result of ecologically mediated mechanisms. In the longer term it is probable that these indirect impacts on human health would have greater magnitude than the more direct impacts (McMichael et al., 1996, 2003).

For vector-borne infectious diseases the geographic range and abundance of vector organisms (and, in some cases, their intermediate hosts) are affected by various meteorological factors (temperature, precipitation, humidity, surface water and wind), biotic factors (vegetation, host species, predators, competitors, and parasites) and human interventions. The rate of maturation of the pathogen within the vector is typically sensitive to temperature, in a curvilinear fashion. This means, for example, that an increase of 1°C across different parts of the temperature range would yield very different increases in transmissibility.

There is some evidence that tick-borne encephalitis, a viral disease of humans, has increased its geographic range within Sweden over the past two decades, as winter temperatures have increased (Lindgren and Gustafson, 2001). Other studies have found some evidence that malaria has extended its range in parts of highland Eastern Africa in association with local warming (Patz et al., 2002) and that, in northeastern Australia, Ross River virus disease has increased its geographic distribution since 1990 (Tong et al., 2001).

Considerable recent effort has sought to develop better mathematical models for making scenario-based health impact projections (McMichael et al., 2003; Epstein, 1999; Martens, 1998). The models in current use have well-recognized limitations. For example, the transmission cycles of some vector-borne diseases are highly complex, involving multiple vector and host species. This can have the effect of limiting the geographical scale of predictive modeling (from the global to, say, the subnational level), as well as increasing the range of uncertainty around estimates of climate change impact on disease transmission (Woodruff et al., 2002). Nonetheless, mathematical modeling studies have provided useful indicative forecasts of the likely direction and magnitude of impacts, and are an important stimulus to further research.

Mathematical models have forecast that an increase in worldwide average temperature and associated changes in precipitation patterns would cause a net increase in the geographic range of malaria-transmitting mosquito species (Martens et al., 1999) – although some localized decreases may also occur in regions that become too hot or dry, such as in the Sahel region of Sub-Saharan Africa (McMichael et al., 1996; Martens, 1998). Further, temperature-related changes in the life-cycle dynamics of both the vector species and the pathogens (protozoa, bacteria and viruses) would increase the potential transmission of many vector-borne infectious diseases such as malaria (mosquito), dengue fever (another of the world's great vector-borne viral infectious diseases, transmitted by mosquito) and leishmaniasis (sand-fly) (Patz et al., 1996; Martens, 1998). Models indicate that dengue fever would extend its range and seasonality (Hales et al., 2002). In Australia, a combination of heavy rainfall and high temperatures have been used to predict epidemics of Ross River virus disease (transmitted by mosquitoes) with a likely increase in future transmission expected in some regions (Woodruff et al., 2002).

Climate change would also affect the transmission of water-borne infectious diseases via several mechanisms (McMichael et al., 2003), including intensification of rainfall episodes, leading to flooding that causes contamination of drinking water supplies. Bacterial water-borne illnesses such as gastroenteritis due to coliform bacteria, giardiasis, and cholera may thus be affected; so too would various protozoal water-borne infectious diseases such as cryptosporidiosis. Similarly, warmer temperatures would tend to increase the summer seasonal peaks of food-borne bacterial enteric infections, such as those due to *Salmonella* and *Campylobacter*. In low-income countries, with poor hygiene and resources, rates of serious child diarrheal disease would rise. The sensitivity of child diarrhea to variations in climatic conditions has been well demonstrated in studies in Lima, Peru and in Fiji (Singh et al., 2001; Checkley et al., 2000).

Another fundamental potential impact on human health would arise from any downturn in food availability. Climate change is likely to affect crop yields, livestock health and resultant food products, and fisheries. However, the processes are generally complex – and often entail seemingly stochastic events (e.g. the occurrence of new infectious diseases in plants and animals). Cereal grains account for around 70% of world food energy – both via direct consumption and via the feeding of grains to livestock for meat production. Various modeling studies have made estimates of the impact of climate change upon cereal grain yields – allowing for the expected "fertilization" effect of increasing concentration of atmospheric carbon dioxide. Globally, a slight downturn in grain production appears likely, especially in the latter half of the twenty-first century (Parry and Carter, 1998; McMichael et al., 2003). This downturn would be greater in the already food-insecure regions in South Asia, parts of Africa and Central America. Such downturns would increase the number of malnourished people in the world (currently an estimated 830 million) by at least several percent overall – with higher percentage decreases in Sub-Saharan Africa and South Asia. World cereal grain yields have become a little more unstable during the 1990s, displaying increased inter-annual variability: could this be (partly) due to changing climatic conditions?

Effects of economic, social and demographic disruption

The third category of health impact is somewhat speculative and difficult to model quantitatively. The situation of several small island states in the Pacific region typifies issues that are likely to be experienced, at some level, in other parts of the world. These islands are registering growing concern about sea-level rise and disturbance of natural and managed food systems, due to the depletion of freshwater stocks, arable land, and damage to coastal industries. The economic downturns, unemployment, and civil strife typically associated with such large-scale changes, and the resulting population displacement, are also conducive of a range of risks to health. Refugee populations typically experience mental health problems, malnutrition, infectious diseases, and the physical hazards of new and improvised living environments.

The anticipated health consequences of economic and social disruptions due to climate change may not become evident for several decades. A usual difficulty for researchers is in deciding whether an event, such as climate change, has caused such outcomes, given the many, often interrelated, causal influences involved. It will therefore be difficult to detect early climate impact "signals" against the considerable background "noise".

Population vulnerability, and adaptive responses

The magnitude of the impact of climate change upon health depends both on the extent of that change and on characteristics of the target population. Although changing climate conditions may make the environment more favorable for disease transmission and other negative health impacts, this does not necessarily mean these will occur, as long as there is the capacity to adapt to the changed conditions. For example, warmer temperatures are predicted to extend the distribution of many disease-carrying vectors into new areas. This increased risk does not have to mean a large increase in cases of disease, provided there is a continuing expansion of the public health response to the disease, additional quarantine efforts, etc.

A population's "vulnerability" is a function both of the sensitivity of the exposure–response function in that population and of the population's adaptive (i.e. impact-lessening) capacity (Parry and Carter, 1998). That adaptive capacity depends on factors such as population density, level of economic development, food availability, local environmental conditions, preexisting health status, and the quality and availability of public health care. It also depends on various structural and politically determined characteristics, including social-cultural rigidity, international connectedness and political flexibility (Woodward et al., 2000). The level of risk to the population's health is therefore a function of that vulnerability and the amount of exposure to climatic or associated environmental factors.

The reduction of socioeconomic vulnerability remains a high priority. Poor populations will be at greatest health risk because of their lack of access to material and information resources and because of their typically lower average levels of health and resilience (nutritional and otherwise). The long-term improvement in the health of impoverished populations will require income redistribution, improved employment opportunities, better housing and stronger public health infrastructure – including primary health care, disease control, sanitation and disaster preparedness and relief.

Adaptation will be either reactive, in response to climate impacts, or anticipatory, in order to reduce vulnerability. These adaptive actions can be categorized as: (a) administrative or legislative, (b) engineering, or (c) personal (behavioral) (Patz, 1996). Legislative or regulatory action can be taken by government, requiring compliance by all (or by designated classes of) people. Alternatively, voluntary adaptive action may be encouraged via advocacy, education or economic incentives. The former type of action would normally be taken at a supranational, national or community level; the latter would range from supranational to individual levels. Adaptation can be undertaken at the international/national level, the community level and the individual level.

It would be shortsighted to expect that adaptation will provide a complete antidote to the problems of climate change. Reduction of greenhouse gas emissions remains a primary preventive health strategy, with appreciable 'collateral' health benefits in the short term. Transport policies to reduce dependence on the motorcar could lead, for example, to improved air quality and increased physical activity (linked to a reduction in obesity, diabetes and heart disease at the population level). Even so, we are already committed to some degree of global climate change over coming decades despite any mitigation actions we might take now. It is important to begin assessing, for any particular population, both its vulnerability and its adaptation options.

Conclusion

Although it is now generally agreed that human-induced climate change is under way, and will continue to occur, the magnitude and rate of change remain uncertain. Future levels of greenhouse gas emissions will be affected by population and economic growth, as well as by technological change, energy preferences, and social behavior. Our ability to explain the chaotic variations within the climate system is improving, but there will always be inaccuracies involved in modeling natural systems.

In addition to climatic change and increased variability, numerous other factors (i.e. social, behavioral, environmental) influence the risk of the diseases discussed in this item. Teasing out the climatic contribution is a major challenge to health researchers. Given the unavoidable uncertainties pertaining to the projected health impacts of climate change, social policy should be developed within the framework of the "precautionary principle" – that is, the principle of proceeding on the basis of the incomplete evidence that is available in relation to future processes that are never fully knowable in advance (McMichael et al., 2003).

Anthony J. McMichael and Rosalie E. Woodruff

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Cross-references

Air Pollution Climatology
Applied Climatology
Climate Change Impacts: Potential Environmental and Societal Consequences
Climate Comfort Indices
Climatology, History of
Determinism, Climatic
Human Health and Climate
Seasonal Affective Disorder

CLIMATE CHANGE IMPACTS: POTENTIAL ENVIRONMENTAL AND SOCIETAL CONSEQUENCES[1]

With some changes in climate having already occurred, with the climate continuing to adjust to past changes in atmospheric composition, and with continuing greenhouse gas emissions certain to cause further climate change, there is no doubt that the world will be undergoing relatively large changes in climate that will be ongoing for centuries. Such changes will inevitably cause consequences for society and the environment.

[1] This item is updated from MacCracken, M.C., 2001, Global warming: a science overview, pp. 151–159 in *Global Warming and Energy Policy*. New York: Kluwer Plenum.

In developing estimates for these consequences, it is important to consider not only the impacts that would be expected were the projected changes in climate to affect systems as they exist at present, but also to carry out the analyses in the context of how these systems are changing (naturally and due to other human-induced stresses) and in recognition of the possible adaptation of strategies for responding and adapting to the projected combination of stresses. Because vulnerability is really the difference between the sensitivity to the impact and the ability to adapt, all considered in the context of other stresses on society and the environment and resources available to help respond, the vulnerability of each nation will depend particularly on its own situation and the local ability and commitment to respond.

The Second and Third Assessment Reports of the IPCC summarize present understanding of the types of consequences for continental-scale regions around the world (IPCC, 1996a,b, 2001b). Figure C13, taken from IPCC (2001b), provides a summary diagram of how the relative importance of the potential impacts is likely to increase as the amount of warming increases. This figure also makes clear that, because climate change will occur with differing intensities at different latitudes, and because different emissions scenarios will lead to different overall rates of change, different countries are likely to be in very different situations.

Recognizing this, many nations have begun to analyze the potential consequences that they anticipate having to deal with. Information for many nations on potential impacts and adaptive measures can be found in their national assessment reports (e.g. for the US, see NAST, 2000 and 2001; for Canada, see Environment Canada, 1998). Many of the nations of the world summarize their findings in reports called for by the United Nations Framework Convention on Climate Change (UNFCCC); for example, for the US, see Department of State (2002) and for other nations see "National Communications" section at http://unfccc.int/). Generally, it has been found that the impacts are likely to be relatively more important for developing than developed nations because developing nations must devote their more limited resources to more immediate challenges and so do not have the flexibility and resources to adapt that developed nation economies may have available. Because it is impossible to be comprehensive in a brief review, both the IPCC and the national assessment reports should be consulted to gain a fuller understanding.

The regional and temporal variations in projected consequences create significant difficulties for negotiators in determining criteria for meeting the central objective of the UNFCCC that was negotiated and internationally accepted in 1992. This objective is to achieve "stabilization of the greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system" while at the same time doing this in a way that would "allow ecosystems to adapt naturally to climate change, ... ensure that food production is not threatened, and ... enable economic development to proceed in a sustainable manner".

In thinking about the importance of potential impacts it is also important to remember that fossil fuels provide wide-ranging benefits to society, providing roughly 80% of the world's energy and so sustaining the world's economy and enabling the present standard-of-living of peoples around the world. Recognizing this, some argue that, if the use of the energy source that supports the global economy is to be sharply reduced, such impacts will have to be relatively certain and substantial. In contrast, others point out that we have only one "spaceship Earth", so

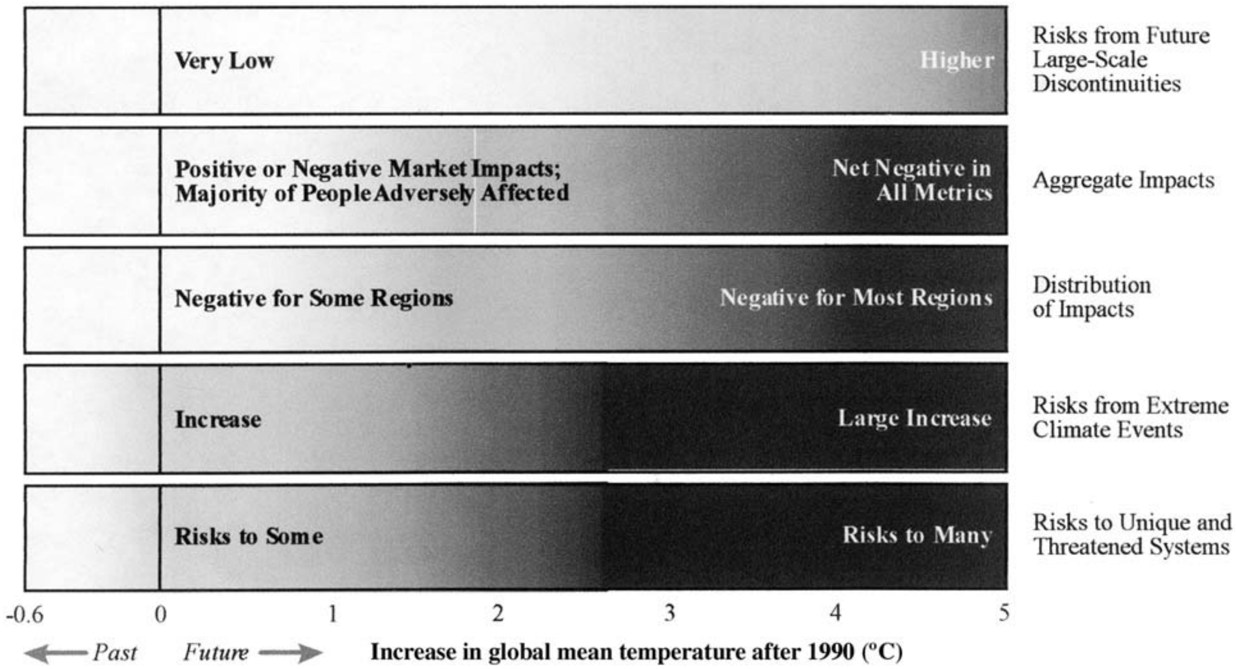

Figure C13 Estimated level of risk as a function of the change in global mean temperature from its value 1990 (in °C), with the areas to the left of the vertical line referring to the risk created by warming prior to 1990. Each horizontal bar refers to a different reason for being concerned. The more intense the shading, the greater the risk. Figure developed as part of IPCC's Third Assessment (IPCC, 2001b).

precaution suggests that the chance of significant or irreversible changes should be avoided. This article provides an overview of some of the important categories of potential impacts that have been identified, providing a sense of the types of impacts (for further details, see the various IPCC assessments).

Human health

Given the large number of people that could be affected, even a modest health impact could create very significant societal and economic consequences. For many regions, significant increases in summertime temperature (and in the absolute humidity that is often associated with such high temperatures) could lead to much more frequent and intense occurrence of dangerously high heat index conditions. While more extensive availability of air-conditioning and warning of dangerous events could limit an increase in heat-related mortality in developed nations, the threat for developing nations where such remedies might not be available could be quite significant.

With the poleward movement of wintertime frosts, mosquitoes and other disease vectors are likely to spread poleward; this has the potential to increase the incidence of infectious diseases if these influences cannot be offset by more attention to appropriate public health measures, enhanced building and community design, and new medicines. Again, those in developed nations are likely to be much better off than those in developing nations; for example, the 17-year total of cases of dengue fever, which is mosquito-borne, is about a thousand times as high in northeastern Mexico as in Texas, mainly due to the relative level of development. The increased intensity of extreme events such as hurricanes and typhoons is also likely to injure or kill more people (and disrupt communities) unless steps are taken to enhance risk-averse planning and construction and to provide better warning and recovery capabilities, and again, developing countries are much more vulnerable.

Food production

The UNFCCC objective specifically mentions the need to ensure that food production is not threatened, for if there is not enough food to survive into the future, the amount of climate change will not really matter. Fortunately, laboratory and field studies indicate that CO_2 concentrations are very likely to enhance the growth of many types of crops and to improve their water use efficiency. If this happens widely (i.e. if other constraints on agriculture do not arise), changes in climate would be expected to contribute to an increase in agricultural productivity (i.e. yield per hectare), thus working in the same positive direction as improvements in agricultural practices, seed strains, and technology. If these positive influences exceed the negative influences caused by deteriorating soil fertility and altered climatic conditions, the potential exists for increasing overall agricultural production while using less land for agriculture. Such an outcome would have the potential for reducing food costs for the public, and there are some studies indicating that the past increase in the CO_2 concentration has contributed to overall food production.

Given the varying soil, climatic, and agricultural capabilities around the world, however, the benefits and impacts are unlikely to be evenly spread. For example, the altered climatic conditions facing farmers in some low-latitude developing nations might well exceed the tolerances for present crops, or, for farmers used to growing only a few types of crops, could at least require a disruptive change in what is grown. For farmers in developed nations, many are facing economic hardships now because of the low prices that result, at least in part, from overproduction. This overproduction presently causes many governments to have to provide subsidies or to impose trade barriers of various kinds. Thus, if changes in climate lead to increases in productivity that cause additional downward pressure on commodity prices or to an increase in costs of production (e.g. to pay for additional irrigation), income for farmers in marginal growing areas is likely to decrease. Even though these farmers might benefit from some gain in overall crop productivity, it is likely to be less than the increase for farmers in the most productive areas, meaning that those in marginal areas are unlikely to remain competitive. This could cause economic and environmental problems in nearby rural communities unless other profitable crops are identified and farmers have the incentive and resources to tend and conserve the land resources.

Ecosystems and ecosystem services

As is clearly evident from a hike up a mountain or a vacation excursion across a continent, the prevailing vegetation is closely tied to the climate that a location experiences. Although the world's vegetation shifted dramatically as the climate warmed at the end of the last glacial maximum, the shifts took place over many millennia and were possible in part because the world's wildlife and plant species had been squeezed equatorward and so had a base from which to spread. For the future, the amount and rate of climatic change are projected to be much, much greater than has been experienced over the past 20 000 years. In addition, the ultimate climatic conditions are projected to reach levels that are well beyond existing conditions, thereby rapidly recreating climatic conditions that have not been experienced in the world for many millions, even tens of millions, of years. In that the ecosystems that make up our landscape provide such a range of services, including water and air purification, supply of food and fiber products, and recreational and esthetic enjoyment, the likelihood of significant changes in ecosystems prompted the provision in the UNFCCC objective that necessitates understanding how ecosystems are likely to change and adapt.

Ecosystem simulation models, which are calibrated based on past and present climate–vegetation relationships, project that changes are likely to be quite extensive. In general, temperature changes are likely to increase with latitude and to be larger over land areas than over ocean areas. Precipitation in low and high latitudes is expected to increase, but with an expanded subtropics where increased evaporation likely further reduces soil moisture. Wintertime precipitation in middle- and high latitudes is projected to increase (proportionately) more than summertime precipitation. Evaporation rates are likely to increase, especially in summer, causing reductions in soil moisture that could affect pest and fire disturbance frequencies for forests and grasslands. As the seasonal patterns of temperature and soil moisture change, the competition among various species will be changed, leading to changes in prevailing tree and grass types and then in the types, abundance, and migration of birds and other wildlife. Some, but not all, of these effects may be offset by the increased CO_2 concentration, which is expected to help many types of plants grow better if other factors, such as nutrients, are not limiting. To the extent that regions accumulate carbon in vegetation and dry out during persistent warm episodes, fire risk is likely to increase even though plant growth may overall be more vigorous. At the same time, increased precipitation over deserts caused by shifting atmospheric circulation patterns, even though perhaps seasonally concentrated, could lead to more vegetation in some regions.

Climatic changes will also lead to significant impacts on freshwater and marine ecosystems. Already, bleaching of coral reef ecosystems is being attributed, at least in part, to the warming oceanic conditions. The increasing CO_2 concentration also affects ocean chemistry, and for coral reefs the impact is to dramatically shrink the regions of the world where temperatures and ocean chemistry are conducive to generation of strong reef structures. Fish populations are also very sensitive to the temperatures of streams and ocean waters, and significant shifts are projected as the climate changes. Changes in ocean productivity also cause effects all the way up the food chain, ultimately affecting the resources, and so the populations, of species as varied as whales, seals, and penguins.

What is most important to understand is that the notion of ecosystem migration is a misconception – as climate changes, the dominant species in particular locations will change, and so the local ecosystem will change as species now dominant become, if they can, dominant elsewhere. However, because each species is distinct, this process is likely to lead to the deterioration of existing ecosystems and the creation of new ones. During this century the changes are likely to be relatively rapid, and the complexity and resilience of current systems is likely to be disrupted because there will not be sufficient time for adjustment and evolution to take place. As a result, the ecological services that society depends on seem likely to be diminished.

Water supplies

Water is absolutely vital to society, and water resource limitations are currently affecting many nations. By affecting the hydrological cycle in a variety of ways, climate change is likely to exacerbate the availability of water for peoples around the world. Changes in the location and timing of storms will alter the timing and amount of precipitation, soil moisture, and runoff. For example, the intensity of convective rainfall is projected to increase, creating the potential for enhanced flooding in watersheds that experience frequent rainfall. Conversely, increased summertime evaporation will reduce soil moisture and may also reduce groundwater recharge in areas such as the Great Plains, and cause lower levels in the lake and river systems, reducing water supplies for communities and agriculture as well as opportunities for shipping and recreation.

Warmer conditions are also likely to increase the rate of evaporation and to change precipitation in cold regions from snow to rain. These and other types of changes will thus affect the overall amounts and availability of water resources, as well as potentially intensifying both floods and droughts. Adapting to such changes will require changes in how river and reservoir systems are managed in order to satisfy the demands of water for communities, industry, hydroelectric power, irrigation, water levels, groundwater recharge, and preservation of fish migrations and aquatic ecosystems. The challenges will be especially severe in mountainous regions where the snowpack (and even mountain glaciers) presently serve as virtual water reservoirs, typically providing water during the spring and summer. Because wintertime precipitation is projected, on a relative basis, to increase the most, avoiding floods is likely to require an even greater lowering of reservoir levels in winter, even though this would reduce water availability in summer when demand is likely to be highest. In addition, because water resources are under increasing stress due to many non-climatic factors, the impacts of climate change are likely to be a significant exacerbation of these other influences.

Coastal endangerment

After the rapid rise of sea level at the end of the last interglacial, reconstructions of beach height, taking into account isostatic adjustments of the deformed land surface, suggest that sea level was relatively constant through most of the Holocene. Indications are that global sea level started rising during the nineteenth century, rose by about 0.1–0.2 m during the twentieth century, and will rise more rapidly in the future. If the rate of sea level rise triples (using mid-range estimates from IPCC, 2001a), this would put many coastal regions at significant risk of inundation. For regions currently experiencing subsidence of their coastlines (e.g. river deltas as sediments compact and are not renewed, regions just equatorward of ice sheet edges), there could be an even more significant acceleration in the rate of rise. Because there are few resources to protect sensitive natural areas such as wetlands and fish- and bird-breeding grounds, accelerating loss is very likely. The rate of loss will be greatest during coastal storms when storm surges (and therefore wind-whipped waves) will reach further inland and further up rivers and estuaries.

There are also many cities and significant infrastructure along coastlines. For highly developed areas, retreat would be very costly, and so strengthening of coastal protection is essential, not just to protect against sea-level rise, but also to reduce current vulnerability to the storm surge from coastal storms and hurricanes. For smaller communities and residential structures, protective structures such as levees are likely to be too expensive and so retreat will be required, or forced by events. Already, some coastal villages in Arctic regions are at risk because the wintertime retreat of sea ice is increasingly exposing these communities to high waves whipped up by winter storms.

With the likelihood that the rate of sea-level rise will continue at significant levels for many centuries, determining when near-term protective measures are best replaced by long-term relocation will be very difficult for many nations, especially with the increasing desire of affluent populations to live near the coastline and enjoy its scenic and recreational opportunities. However, it is essential that such analyses be done as the choice is important in determining how best to deal with rising sea level and its impacts over time (e.g. if the coastline is to be preserved, roads need to be parallel to the coast; if retreat is to occur, roads need to be perpendicular to the coast to allow phased movement).

Societal infrastructure

Although societal infrastructure is generally designed to withstand a wide variety of weather extremes, the changing climate is projected to create changes in the intensity and frequency of extreme events. For example, the climate model of the Canadian Climate Centre projects that the return period of 100-year storms could decrease by half or more over the next 100 years. Given the economic costs and psychological impacts of recovering from such intense and damaging storms and conditions, a phased response that leads to infrastructure designs that provide more resilience has the potential of limiting potential consequences. Although the incremental cost of providing more resilience is not likely to be high for any single project in developed nations, there is an extensive amount of infrastructure in these nations and so total costs could be quite significant, although mostly hidden within large budgets. For developing countries the cost of designing resilient infrastructure is likely to be a greater increment, but, in that old infrastructure is not having to be replaced, may be feasible. What is most important is that such precautionary investments (e.g. elevating new sewer plants to withstand projected changes in sea level) be made, starting as early as possible, rather than further investing in facilities that will soon have to be reconstructed.

Transportation systems represent a very large societal investment. Present intensities of severe weather and floods typically cause disruptive economic impacts and inconvenience that can become quite important for particular regions during particular extreme events. While information is only starting to emerge about how climate change might lead to changes in weather extremes, a range of possible types of impacts seem possible, including some that are location-dependent and some that are event-specific. Location-dependent consequences could include: lower levels in rivers and lakes used for transportation and as sources of water for communities and industry, relocation of coastal channels as a result of shifting sediments and barrier islands, and opening of Arctic shipping routes as sea ice melts back. Event-specific consequences could include: more frequent occurrence of heavy and extreme rains (a trend already evident during the twentieth century); reduced or shifted occurrence of winter snow cover that might reduce winter trucking and air traffic delays; altered frequency, location, or intensity of hurricanes and typhoons accompanied by an increase in flooding rains; and warmer summertime temperatures that raise the heat index and may increase the need for air pollution controls. By reducing the density of the air, warmer temperatures will also cause reductions in combustion efficiency, which would both increase costs and require longer runways or a lower load for aircraft, just as now occurs for airports in mountainous regions. Starting to consider climate variability and change now in the design of various types of infrastructure could well be a very cost-effective means of enhancing the short- and long-term resilience of society to weather extremes and climate change.

Air quality

Warmer temperatures generally tend to accelerate the formation of photochemical smog. The rising temperature and rising absolute (although perhaps not relative) humidity will raise the urban heat index significantly, potentially contributing to greater air pollution. Meeting air quality standards in the future is therefore likely to require reductions in pollutant emissions beyond those currently in force (although, of course, a move away from the combustion engine might make this change much easier). Increasing amounts of photochemical pollution could also increase impacts on stressed ecosystems, although the increasing concentration of CO_2 may help to alleviate some types of impacts by causing the leaf's stomata to close somewhat. In some regions, summertime dryness would be expected to exacerbate the potential for fire, creating the potential for increased amounts of smoke, whereas in other regions the climatic changes may make dust more of a problem.

International interactions

While it is natural to look most intently at consequences within a particular nation, each nation is intimately coupled to the world in many ways. For example, economic markets and investments are now coupled internationally and a nation's foods come from both within and outside its borders. Because of international travel for business and pleasure, health-related impacts are also international. Many resources, from water- and hydropower-derived electricity to fisheries and migrating species, are shared across borders, or move and are transferred internationally. Finally, people continue to relocate from one nation to another, becoming citizens of one area while tied by family and history to other areas, especially when disasters strike. Increasingly, all countries are connected to what happens outside their borders, and those in one country will be affected by the repercussions of the societal and environmental consequences experienced by others.

Summary of impacts

Given the variety of potential impacts and the varying capabilities and resources for responding to them, international assessments (e.g., IPCC, 1996b, 2001c) provide primarily a qualitative indication of the potential consequences. In addition to the scientific challenges of developing the estimates, issues of equity and cultural values introduce many more complications. For example, IPCC (1996b) struggled with the issue of whether the imputed value of a human life should be considered the same in a developed and a developing nation even though the lifetime earnings and economic evaluations of a life lost are quite different across the world.

Present assessments indicate that there are very likely to be important consequences, some negative and some positive, and that impacts are likely to be greater for developing than for developed nations. However, most analyses currently treat rather large regions and so average out the most costly and most beneficial outcomes, thereby perhaps obscuring what could be rather large implications for localized groups.

The mitigating effects of emissions reductions

In recognition of the potential for significant climatic and environmental change and consequent impacts, the nations of the world in 1992 agreed to the United Nations Framework Convention on Climate Change (UNFCCC). As indicated in the preceding section, this agreement set as its objective the "stabilization of the greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system". Because of difficulties in deciding upon what might be considered dangerous, and because uncertainties in understanding of the climate system create only a loose coupling to the changes in atmospheric emissions and composition and consequences that might occur, discussions about potential stabilization levels have so far been based mainly on a sense of the level of change in atmospheric composition that seems likely to have significant consequences, considered generally in the context of what might be technologically and economically possible. For example, considerable attention has been given to attempting to stabilize the atmospheric CO_2 concentration at double its preindustrial level (i.e. at about 550 ppmv). Although not discussed below, similar limitations would need to be imposed on the other greenhouse gases and aerosols if atmospheric composition is to be stabilized.

Based on current scientific understanding of the carbon cycle, limiting the atmospheric CO_2 concentration to 550 ppmv would require *the world as a whole* to remain, on average, at the present per-capita CO_2 emission level of about 1 ton of carbon (tC) per year throughout the twenty-first century. This contrasts with the expectation that per-capita emissions are likely to double over this period in the absence of agreements to control emissions. As context for considering the challenge that this would impose, the typical American is now responsible for emission of 5–6 tC/year and the typical European is responsible for emission of about 3 tC/year, ignoring any imputed contribution that might arise from considering the energy used to create imported products on which they depend. Maintaining the CO_2 concentration at 550 ppmv beyond the twenty-first century would require, for example, that global emissions for the twenty-second century drop to more than a factor of 2 below current global emissions levels (i.e. to less than half of the 6 billion tonnes of carbon now emitted each year). Such an emissions level would require that global per-capita emissions of carbon be about one-third of the emission level of developing countries today, or one-twentieth of the current per-capita level in North America. Such low levels would not mean that per-capita *energy* use would need to be reduced by this amount, only that net per-capita use of fossil fuels (so emissions minus sequestration) would need to be this low. The IPCC (2001c) suggests that the necessary transition in sources of energy could be accomplished by meeting most of the world's energy needs using renewables and nuclear, if there is significant effort to improve the efficiency of energy end uses, although this is controversial (Hoffert et al., 2002).

As perhaps a start toward this objective, the Kyoto Protocol has been negotiated. Even though its goal is relatively modest in a scientific sense (i.e. by 2008–2012, developed nation emissions of a set of greenhouse gases would be reduced on average to several percent below their 1990 emissions level), it is proving to be a political challenge to get approved and implemented. Even if the US were to participate in its implementation, the emissions limitation would only begin to limit the rate of increase in the global CO_2 concentration because emissions from developing nations would not be similarly controlled and are, not surprisingly, projected to increase significantly along with their economic growth, even though many of these countries are making voluntary efforts to limit growth in emissions. With full implementation of the Kyoto Protocol through the twenty-first century, the CO_2 concentration would be expected to rise to about 660 ppmv as compared to an expectation of a rise to 710 ppmv in the absence of emissions controls. Such analyses make it clear that reducing CO_2 emissions to achieve stabilization at 550 ppmv would thus require significant further steps, and these would need to involve all nations. Accomplishing such an emissions reduction would require much more extensive introduction of non-fossil energy technologies, improvement in energy generation and end-use efficiencies, and switching away from coal to natural gas (unless extensive carbon sequestration of carbon in forests, oceans, or underground can be accomplished).

Absent such efforts on a global basis, the CO_2 concentration is projected to rise to about 800–1100 ppmv (about 3–4 times the preindustrial level) during the twenty-second century. Were this to occur, climate simulations indicate that the resulting warming would be likely to induce such potentially dangerous long-term, global-scale impacts as the initiation of the eventual melting of the Greenland and the West Antarctic ice sheets (each capable of inducing a sea-level rise of up to about 5–7 m over the next several centuries), the loss of coral reef ecosystems due to warming and rapid sea-level rise, the disruption of the global oceanic circulation (which would disrupt the nutrient cycle sustaining ocean ecosystems), and extensive loss or displacement of critical terrestrial ecosystems on which societies depend for many ecological services (e.g. see O'Neill and Oppenheimer, 2002).

What seems most clear from present-day energy analyses is that there is no "silver bullet" that could easily accomplish the major emissions reduction that is needed to achieve the UNFCCC objective of stabilization of atmospheric composition, even in the face of significant potential impacts. Rather, moving toward the objective will require a multifaceted international approach involving a much more aggressive (although not unprecedented) rate of improvement in energy efficiency, broad-based use of non-fossil technologies (often selecting energy sources based on local resources and climatic conditions), and an accelerated rate of technology development and implementation.

Conclusion

A major reason for controversy about dealing with the climate change issue results from differing perspectives among and within nations about how to weight a wide range of influential factors. These factors include, among other aspects:

1. the need for scientific certainty versus making decisions in the face of uncertainty;
2. the flexibility and adaptability of the energy system that provides for the national and global standard of living;
3. the potential for improving efficiency and developing new technologies;
4. the potential risk to "Spaceship Earth" that is being imposed by this inadvertent and virtually irreversible geophysical experiment;
5. the economic and environmental costs and benefits of taking early actions to reduce emissions (including what factors to consider in the analysis and how to weight the relative importance of long-term potential impacts versus better defined near-term costs); and
6. the weight to give matters of equity involving relative impacts for rich versus poor within a nation, for developed versus developing nations, and for current generations versus future generations.

Moving toward an international consensus on these issues sufficient to negotiate international agreements will require that the publics and governmental officials of all nations become better informed about the science of climate change, about potential impacts and their implications, and about potential options for and costs of reducing emissions. Moving toward collective action will require finding approaches that tend to balance and reconcile these (and additional) diverse, yet simultaneously legitimate, concerns about how best to proceed.

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Cross-references

Agroclimatology
Air Pollution Climatology
Bioclimatology
Coastal Climate
Cultural Climatology
Climate Change and Global Warming
Human Health and Climate
Vegetation and Climate

CLIMATE CLASSIFICATION

The philosophical base of climate classification follows the same logic as all classifications and can be dealt with accordingly. Classification is variously defined but is well described as the systematic grouping of objects or events into classes on the basis of properties or relationships they have in common. The function of classification is also given different interpretations but essentially concerns the needs:

(a) to bring structure, order and simplicity to a complex system,
(b) to provide an intellectual shorthand,
(c) to identify spatial limits and boundaries, and
(d) to provide practical as well as theoretical uses.

Each of these cited functional goals must be considered in the formulation of a useful climate classification scheme. However, the identification of spatial limits and boundaries has proved of special interest to climatologists. Much of the literature dealing with climatic classification deals with limits of a climate group rather than the climate itself.

Of the many ways in which to classify components of the world's climate, the simplest is through classification using a single variable. Each distributional map of any climatic variable is, by definition, a classification scheme. The long history of climatic classification has its roots in the orderly presentation of the distribution of a single variable.

Single-variable classification

The earliest writings providing a consideration of the different climatic environments over the Earth can be traced to the contributions of thinkers in the sixth century BC and to such scholars as Aristotle and Plutarch. The ancient Greeks had postulated a spherical Earth and, through knowledge of the travel of the sun, identified the five-zone system – a torrid zone, two temperate zones and two frigid zones – still identified today. Aristotle provided the first quantitative boundaries of the system by identifying the Tropics, a division based upon astronomy and geography. Following the earlier Greeks, Ptolemy (AD 90–168) used day length as the classification variable and identified seven climates based upon duration of the longest day.

The ideas of Greek and Egyptian scholars passed to the Arab schools and such writers as Ibn Hauqual (ninth century), al Biruni (tenth century) and Idrisi (twelfth century) reproduced and expanded upon the work of the ancients. This work carried over to the geographical writings of the Renaissance and, in 1650,Varenius presented a table, the climata, giving the day lengths at the solstices for zones of the Earth. From the findings of these early theorists it is apparent that day length zones rather than temperature zones provided the quantitative expression of climate.

Distributions other than those based upon day-length appeared soon after thermometric data were assembled in the early nineteenth century. In 1817 von Humboldt produced the first isothermal map; while in 1846 Dove prepared monthly temperature distributions and went on to compute mean temperature of latitudes and hence the concept of temperature anomalies. Thereafter contributions became more numerous. Supan (1879) developed the first map of climatic belts based upon mean annual temperature and temperature of the warmest month. Later, in 1884, he produced a system in which zones were identified by regional names.

Single-variable classifications also used precipitation data and it was here that Koppen, the best known of climatic classifiers, made his first major climatological contribution. Both Schimper (1898) and Herbertson (1905) drew upon Koppen's work to produce regionalized world precipitation maps. As more data became available, world maps and regions were constructed for most climatic variables, and classification passed to the use of multiple variables

Vegetation distribution and climatic classification

The early development of climatic classification is related closely to vegetation studies. Of significance in these early studies of vegetation was the approach used. Systematic botanists arranged plants of the world into classes, genera, etc. using what might be termed a floristic approach; plant geographers used a physiognomy of vegetation in which plant communities expressed in such terms as forests, meadows, moors, etc., were identified. Table C3 provides an example showing the relationship between climate and vegetation.

In 1874, De Candolle proposed a classification that was designed for tracing the development of plants through geologic time. Six subdivisions were initially used but one, the megistotherms (mean annual temperature above 30°C), while prevalent in earlier geologic times, are found today only in the vicinity of hot springs. The remaining five, megatherms, mesotherms, microtherms, hekistotherms and xerophiles, were designated with letters A through E, with B referring to the xerophiles. Given that De Candolle was identifying groups as parallel zones, the insertion of moisture-based unit (B) in a temperature-based grouping is not really illogical. It has led, however, to many misunderstandings in its transfer to the Köppen system.

The Köppen system

From the pioneer work on vegetation classification came climate organizational schemes based upon vegetation distribution. This realm of classification is dominated by the work of Wladimir Köppen, who published his first paper in 1868 and continued to be a productive scholar up to his death

Table C3 Correspondence between climatic and vegetation types

Climate	Vegetation type name	Vegetation
Rainy tropical	Malayan	Evergreen rain forest
Subhumid tropical	Nicaraguan	Deciduous or monsoon forest
	Timoran	Savanna forest or woodland
	Visayan	Tropic grassland
Warm semiarid	Tampicoan	Thorn forest, thorn scrub
	Tamaulipan	Desert savanna, wetter parts
Warm arid	Tamaulipan	Desert savanna, drier parts
	Sonoran	Subtropic desert
	Tripolian	Short grass; desert grass
Hyperarid	Atacaman	"Barren" desert
Rainy subtropical	Kyushun	Warm temperate rain forest
	Argentinean	Prairie
Summer-dry subtropical	Mediterranean	Sclerophyll woodland and scrub
Rainy marine	Tasman	Subantarctic forest
Wet-winter temperate	Oregonian	Conifer forest
Rainy temperate	Virginian	Mixed deciduous and conifer forest
Cool semiarid	Patagonian	Cold desert, wetter parts
Cool arid	Patagonian	Cold desert, drier parts
Subpolar	Alaskan	Taiga forest
Polar	Aleutian	Tundra and polar barrens

After Putnam et al. (1960).

in 1940. Köppen's work in classification may be viewed as a climatic determinism of vegetation types, while his greatest contribution was to stress the fundamental unity of pattern in the location of climatic regions throughout the world.

Table C4 provides the outline of Köppen's 1900 classification. It is seen that the idea of providing letters (A – E) is introduced but not emphasized. Instead, Köppen retained the ninteenth-century approach of using plant and animal names for identified regions. Some of these, for example the Penguin Climates, are certainly interestingly informative. The classification also contained some quantitative boundaries, but the well-known characteristics of the classification known to us did not appear until 1918.

The 1918 Köppen classification provides the basis for that used today (Table C5, Figure C14). Each climatic type is given a distinctive upper- or lower-case letter, each of which has a specific meaning. Local regional and biological names are dropped. Over time, changes to the initial system have been both suggested and implemented. The main changes introduced

Table C4 Climatic regions identified in Köppen's (1900) classification

A. Megathermal or tropical Lowland climates; coldest months >18°C

1. Liana
2. Baobab

B. Xerophytic climates – arid and semiarid. Continuous scarcity of precipitation

I. Coastal deserts of low latitudes

3. Garua (Fog)

II. Lowland desert and steppe experiencing great summer heat

4. Date palm
5. Mesquite
6. Tragacanth
7. East Patagonian

III. Lowland desert and steppe with cold winters and short hot summers

8. Buran
9. Prairie

C. Mesothermal or temperate climates. Coldest month below 18°C, warmest month over 22°C

I. Subtropical climates with moist, hot summers

10. Camellia
11. Hickory
12. Maize

II. Subtropical climates with mild, wet winters and dry summers

13. Olive
14. Heather

III. Tropical mountain climates and maritime climates of middle latitudes

15. Fuchsia
16. Upland savanna

D. Microthermal or cool climates. Warmest month between 10°C and 22°C

17. Oak
18. Spruce
19. Southern beech

E. Hekistotherm or cold climates. Warmest month between 0°C and 10°C

20. Tundra
21. Penguin or Antarctic
22. Yak (Pamir)
23. Rhododendron
24. Ice cap

Quantitative values were given for each identified type.

Table C5 The Köppen classification of climate

A. *Temperature of the coolest month above 18°C (64.4°F)*

f: rainfall of driest month at least 6 cm
m: rainfall of driest month > (10 – annual rainfall/25) but < 6 cm
w: rainfall of driest month, 6 cm and dry in low-sun season (but does not meet *m* criteria)
s: rainfall in driest month, 6 cm and dry in high-sun season (but does not meet *m* criteria)

Also recognized:

w′: maximum rainfall in summer
w′: two rainfall maxima
i: annual temperature range < 5°C
g: warmest month precedes summer solstice

B. *Evaporation exceeds precipitation for the year*

BW (desert) and *BS* (steppe) determined as follows:
Substitute *r* (annual precipitation in cm) and *t* (annual average tempature °C) in
$r = 2(t + 14)$ when 70% rain is in summer 6 months
$r = 2t$ when 70% rain is in winter 6 months
$r = 2(t + 7)$ when evenly distributed or neither of above
If $r >$ right-hand side of equation then climate is not a desert
If $r <$ right-hand side of equation then climate is a desert (*BS* or *BW*)
If right-hand side of equation is $< r/2$ then *BW*, else it is *BS*

h: average annual temperature > 18°C
k: average annual temperature < 18°C

Also recognized:

k′: average temperature of warmest month < 18°C
n: high frequency of fog
s: 70% of rainfall in winter 6 months (summer dry)
w: 70% of rainfall in summer 6 months (winter dry)

C. *Coolest month average temperature < 18°C and above −3°C; warmest month average temperature >10°C*

f: at least 3 cm precipitation each month
w: at least 3 times as much precipitation in a summer month as in driest winter month
s: at least 3 times as much precipitation in a winter month as in the driest summer month, with 1 month < 3 cm

a: warmest month > 22°C
b: warmest month < 22°C but at least 4 months > 10°C
c: only 1–3 months > 10°C

Also recognized:

x: rainfall maximum in late spring or early summer: dry late summer
n: high frequency of fog
i: annual temperature range < 5°C
g: warmest month occurs before summer solstice
t′: warmest month occurs in fall
s′: maximum rainfall in fall

D. *Coolest month average temperature < −3°C; Warmest month >10°C*

d: coldest month < −38°C
Other categories same as *C* climates

E. *Warmest month average temperature <10°C*

ET: average temperature of warmest month between 0°C and 10°C
EF: average temperature of warmest month < 0°C

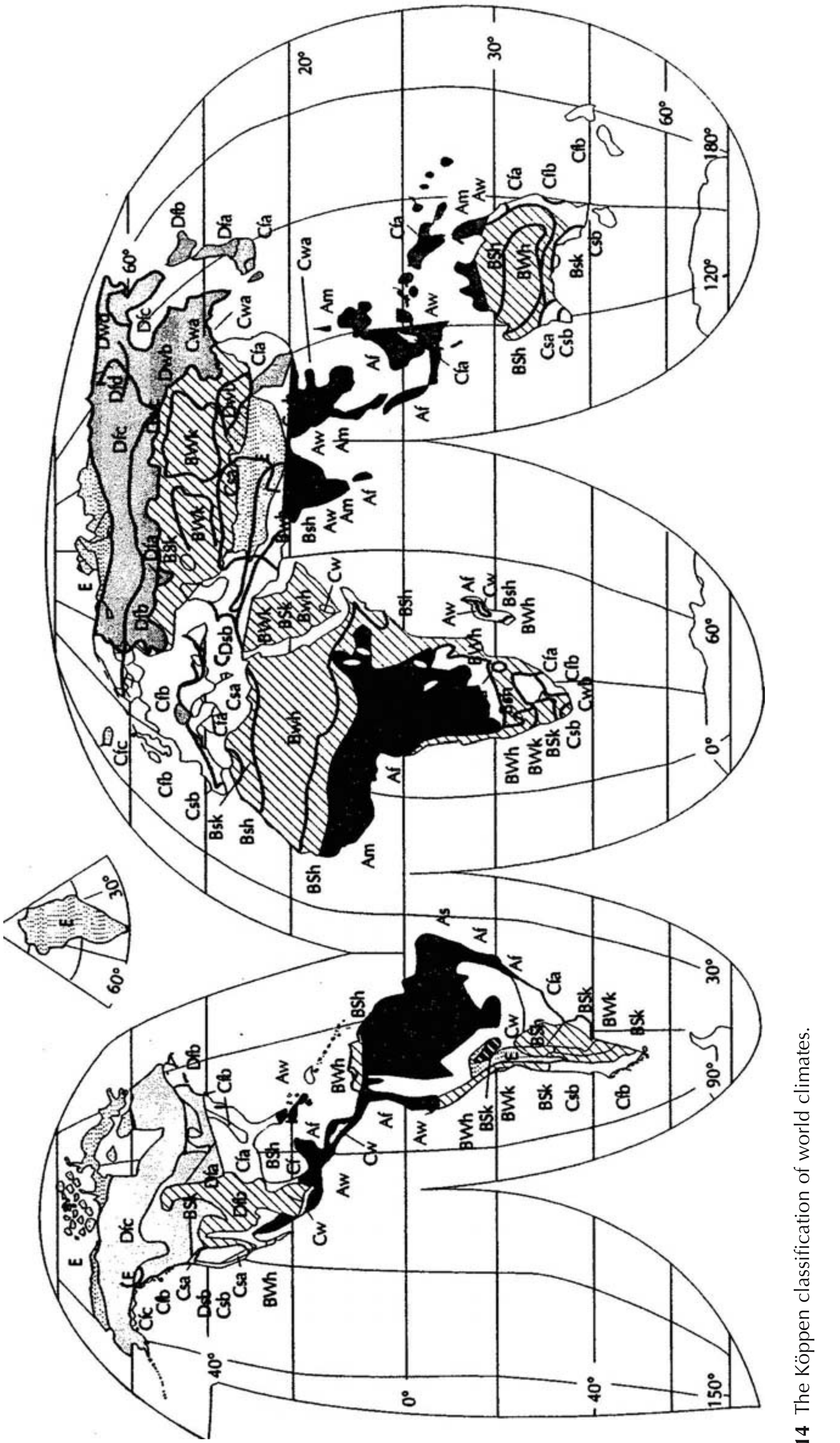

Figure C14 The Köppen classification of world climates.

by Köppen himself largely concerned the definition of the B climates. It remained for others to expound upon the work, and of these there were many. In a survey completed in 1952, Knoch and Schultz identify 70 works concerning the use of the Köppen system. Excellent reviews of the Köppen systems have been given by Thornthwaite (1943) and Wilcock (1968).

So dominant was the Köppen system that many fine alternate vegetation-based systems did not gain the widespread recognition. A system proposed by Vahl (1919), for example, overcomes the reliance upon average annual temperatures of the Köppen approach. Vahl based his classification upon four zonal climates that are subdivided by temperature limits that use both the warmest and coldest months. Another interesting approach was given by Troll (1958) who, by noting that "The life of plants, animals and humans . . . is subject to rhythm of climatic phenomena" produced a classification of the seasonal climates of the Earth. Graphical representation also plays an important role in the classification introduced by Peguy (1961).

A classification of considerable interest was proposed by Holdridge (1947), who explained its formulation. "While attempting to understand relationships between the mountain vegetation of an area of Haiti and other vegetation units of the island and surrounding regions, the literature was searched unsuccessfully for a comprehensive system which presented formation or vegetation units on a relatively equal or comparable basis." To meet this end he produced what is now called the Holdridge Model, which related temperature and precipitation to produce a series of nested hexagons. The ecological value of the system was ably demonstrated by Tosi (1964).

Table C6 Some empirical approaches to climate classification

Author(s)	Purpose	Base of System
Bagnouls and Gaussen (1957)	Biological climates	Duration of the dry season based upon the derived xerothermal index (*X*). Twelve major regions identified according to *X* values, temperature of coldest month, and frost/snow data.
Blair (1942)	An orderly description of world climates	Five main zonal climates distinguished: tropical (*T*), subtropical (*ST*), intermediate (*I*), subpolar (*SP*), and polar (*P*) Fourteen types and six subtypes distinguished using letter notation. Based upon precipitation and temperature data and related to vegetation types.
Brazol (1954)	Human comfort zones	Use of wet and dry bulb temperatures to establish comfort months. Twelve ranked classes ranked from no. 12 – lethal heat – to No. 1 – glacial cold.
Budyko (1958)	Distribution of energy in relation to water budget	Use of rational index of dryness to relate ratio of net radiation to energy required for vaporizing moisture. Moisture regions form basic unit.
Carter and Mather (1966)	Environmental biology	A modification of the 1948 Thornthwaite system.
Creutzberg (1950)	Climate–vegetation relationships	The annual rhythm of climate based on identification of *Isohygromen* (lines of equal duration of humid months) and *Tag-Isochione* (lines of equal daily snow cover duration). Four major zones differentiated, subdivided by monthly moisture values.
de Martonne (1909 and following years)	World regional (land-form) studies	Nine first-order divisions based upon temperature and precipitation criteria. Numerous subdivisions named for local areas in Europe. Considerable attention given to desert limits, but most boundaries derived nonquantitatively.
Emberger (1955)	Biologic (ecologic)–climate relationships	Two main climates differentiated: deserts and non-deserts. Differentiated and subdivided in terms of annual range of temperature and duration of light periods.
Federov	The "complex" method, utilization of day-to-day weather observations	An incomplete system that relies upon codification of daily weather events. For example, the first letter indicates character of prevailing wind; second letter character of temperature; third letter character of precipitation, cloudiness, humidity; fourth letter character various phenomena of atmosphere and state-of-ground surface.
Gorczynski (1945)	The decimal system	Ten "decimal" types associated with five main zonal climates. Emphasis on continental versus marine climates and definition of aridity.
Geiger-Pohl (1953)	World map of climate types	Modification of the Köppen system.
Köppen (1918 and following years)	See description in text	
Malmstrom (1969)	Precipitation effectiveness as a teaching scheme	Retention of the basic concepts of the Thornthwaite (1948) system but with arbitrary threshold values to express water needs of plants. Warmth index ($N-38\,m/100$) also used.
Miller (1931 and following years)	Temperature and precipitation used s main variables	A basic system for location of general world patterns.

Table C6 (*Continued*)

Author(s)	Purpose	Base of System
Papadakis (1996)	Agricultural potential of climatic regions	Use of "crop-ecologic" characteristics of a climate based upon empirically derived threshold values. Ten main climate groups recognized, each divided into subgroups which are themselves divided.
Passarge (1924)	Climate–vegetation relationships	Recognition of five climatic zones and their subdivision into 10 regions emphasizing vegetation distribution.
Penck (1910)	World climates in relation to studies in physical geography and physiography	Recognition of three main types of climates significant in determining weathering and erosion. Humid, Arid, and Nival. Each subdivided into two.
Peguy (1961)	See description in text	
Philippson (1933)	Climatic regionalization on the world, continental and regional levels	Based on temperature of warmest and coldest months and upon precipitation characteristics. Five climatic zones with 21 climatic types and 63 climatic provinces.
Putnam et al. (1960)	Coastal environments of the world, climate–vegetation characteristics	Fourteen types recognized. Climatic characteristics expressed as those occurring between the 25th and 75th percentile of the frequency distribution appropriate to each climatic type. Variables include mean maximum and minimum temperatures, mean annual precipitation and monthly precipitation frequency.
Terjung (1966)	Bioclimatic, based on humans	Use of comfort index and wind effect to identify physiological climates.
Thornthwaite (1948)	See description in text	
Trewartha (1954)	An orderly description of world climates	Modification of the Köppen system.
Vahl (1919)	World climates related to vegetation	Five zones appraised by temperature limits which are a function of the data for warmest and coldest months. Subdivision by precipitation expressed as a percentage of number of wet days in a given humid month.
von Wissman (1948)	World distribution of climate related to vegetation	Related to the Köppen approach. Five temperature zones subdivided by precipitation distribution and temperature regimes.

Table C6 provide a listing of some of empiric approaches to the classification of climate.

Applied empiric classifications

While the vegetation-based classifications, typified by the Köppen system, received most attention, many other authors proposed empiric schemes that might be classed as special-purpose. These applied empiric classifications were as diverse as the things upon which climate has an impact, but their nature can be demonstrated using selected examples.

In early published works using the applied empiric approach, climate was considered a component of the proposed classification rather than the base of the system itself. In the De Martonne (1909) classification the principal goal was to provide a basis for physiographic studies. This was accomplished by identifying nine main regions grouped upon a variety of parameters ranging from temperature thresholds, annual temperature range and annual average precipitation. De Martonne refined his approach in successive published books, many of which were translated into English (De Martonne, 1927). Another physiographer, A. Penck (1910), used an already-existing formula (rainfall/evaporation ratio) to suggest that humid, dry and transitional climates could form the bias of a world classification of utility in physical geographic studies.

The more recent applied empiric systems relate climate to a wide range of applied areas. Terjung (1968) formulated a bioclimatic classification, while in his text *World Climates,* Rudloff (1981) applies a bioclimatic aspect to the Köppen system. Of systems relating climate to crops, the classification outlined by Papadakis (1966) is of interest. In some ways such an approach is an extension of the crop analog studies completed by Nuttenson (1947).

The moisture balance

In a paper published in 1931, C. W. Thornthwaite proposed a climatic classification that appeared as a marked departure from preexisting systems. Unlike most classifications available at the time Thornthwaite based his system on the concept of precipitation and thermal efficiency. Of more significance at present, having superseded the earlier system, is Thornthwaite's 1948 classification (Figure C15). This shows a radical departure from the 1931 system because it makes use of the important concept of evapotranspiration. While the earlier system had been concerned with the loss of moisture through evaporation, the new approach considers loss through the combined process of evaporation and transpiration. Plants are considered as physical mechanisms by which moisture is returned to the air. The combined loss is termed evapotranspiration, and when the amount of moisture available is nonlimiting, the term potential evapotranspiration is used. As with any widely used system, the 1948 classification has been subject to criticism. Many of the criticisms relate to the empiric formula used to express evapotranspiration, and to the

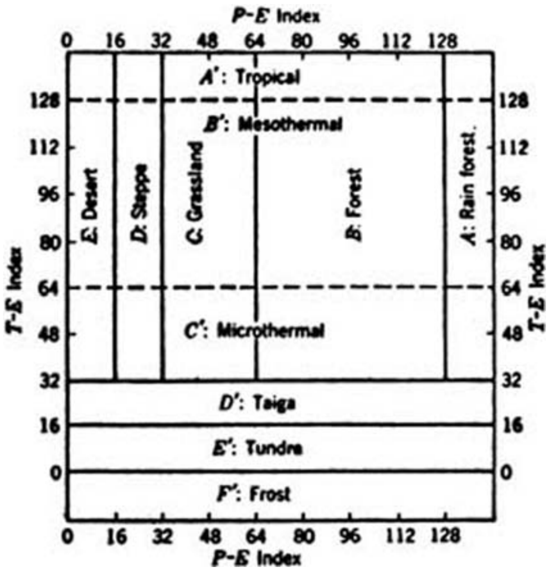

Figure C15 Temperature and humidity provinces of the Thornthwaite classification.

way in which the water budget of a station is manipulated. Its present status is well described by Muller and Oberlander (1984), who note that because its computations are time-consuming, the Thornthwaite system is not often used to define climatic regions on a global continental scale. At the same time it is an invaluable tool in water budget analyses.

Airmass and synoptic climatology

Just as climates may be classified according to the effect of climate upon environmental systems, so may they be grouped according to factors that contribute toward their cause. These causative effects relate to the origin (or genesis) of the climates concerned and resulting systems may be termed the genetic classifications. The development of synoptic meteorology and the concept of airmasses led to enormous improvements in comprehending weather. It seemed that a similar methodology could provide a suitable base for the study of genetic climatology for, as Thornthwaite (1943) suggested "Climatic types are to climate what airmasses are to weather". Subsequent to the basic delineation of airmass types, a number of significant climatological studies based upon airmasses were published. Willett (1933) described the properties of airmasses of the United States, while Showalter (1939) produced airmass identification tables. Despite these advances no comprehensive climatic classification based upon airmasses was formulated and, even at present, such systems are still only first approximations of the climatic complex of the Earth. Valuable maps of seasonal airmasses were presented by Berry and Bollay (1945), while Oliver (1970) used an airmass identification chart to classify world climates using set theory. But of the airmass groupings available perhaps the most utilitarian is that given by Strahler (1951 and following years) in his popular text *Physical Geography* (Figures C16 and C17). Essentially, three main groups of climates are differentiated:

1. Group I. Climates dominated by equatorial and tropical airmasses all the year.
2. Group II. Climates that occur between groups I and III and that are influenced by the interaction between tropical airmasses (group I) and polar airmasses (group III).
3. Group III. Climates controlled by polar airmasses.

Dynamic/synoptic classification

Today, some the most significant classification schemes are developed as part of dynamic/synoptic climatology. The forerunners to the most recent systems are those that were based upon identified physical determinants. One of the earliest workable systems of classifying climate by cause was introduced by Hettner (1930), with that by Flohn (1950) exemplifying a later approach. The nature of more recent work in synoptic classification, and the concepts of grosswetterlagen (large-scale weather patterns) is well described by Barry and Perry (1973). These authors note the various approaches that can be used in synoptic classification, and differentiate the static pattern (the features of circulation including pressure patterns) and the kinematic, in which streamline analysis is of importance. Of note, however, is that many of the studies using such approaches are purpose-oriented classifications and differ in objective and character from the classic world schemes. Barry (in Oliver et al., 1989) has described the significant advances in classification procedures in synoptic climatology made in recent years. He cites the progress in evaluating existing objective procedures (e.g. Yarnal and White, 1984), and the value of clustering techniques. Many of the techniques used are an integral part of the methods used in numerical classification.

Numerical classification

Of singular importance in the developmental history of climatic classification is the introduction of numerical classification or numerical taxonomy. Classifications may be derived either by the logical subdivision of a population or by agglomeration of similar individuals within that population. Most of the classification systems described thus far use logical subdivision as the methodological choice. By contrast, the use of any one of the discriminatory techniques now available, permit classification by agglomeration. The advances in electronic computing permit univariate and multivariate systems of climate to be generated from enormous databases and sophisticated techniques have been incorporated in existing classification methodologies. In reality the numerical taxonomic method supplies a rational mathematical approach to the basic needs of a classification. Willmott (1977) points out that decision making in deriving any classification must concern (a) the number of regions to be identified, (b) the identification of boundary criteria, (c) the variables selected to represent the climate, and (d) the methods in which the selected variables may be summarized. Prior to the utilization of taxonomic theory, the availability of extensive databases and electronic computing devices, decisions concerning these were highly arbitrary. Using the newer methodology the decisions obviate arbitrary choices. McBoyle (1971), following Steiner (1965), produced one of the earliest numerical classifications of climate, and his methodology illustrates the nature of the approach. Using 20 variables for 26 stations in Australia, McBoyle completed a factor analysis to produce three factors that together accounted

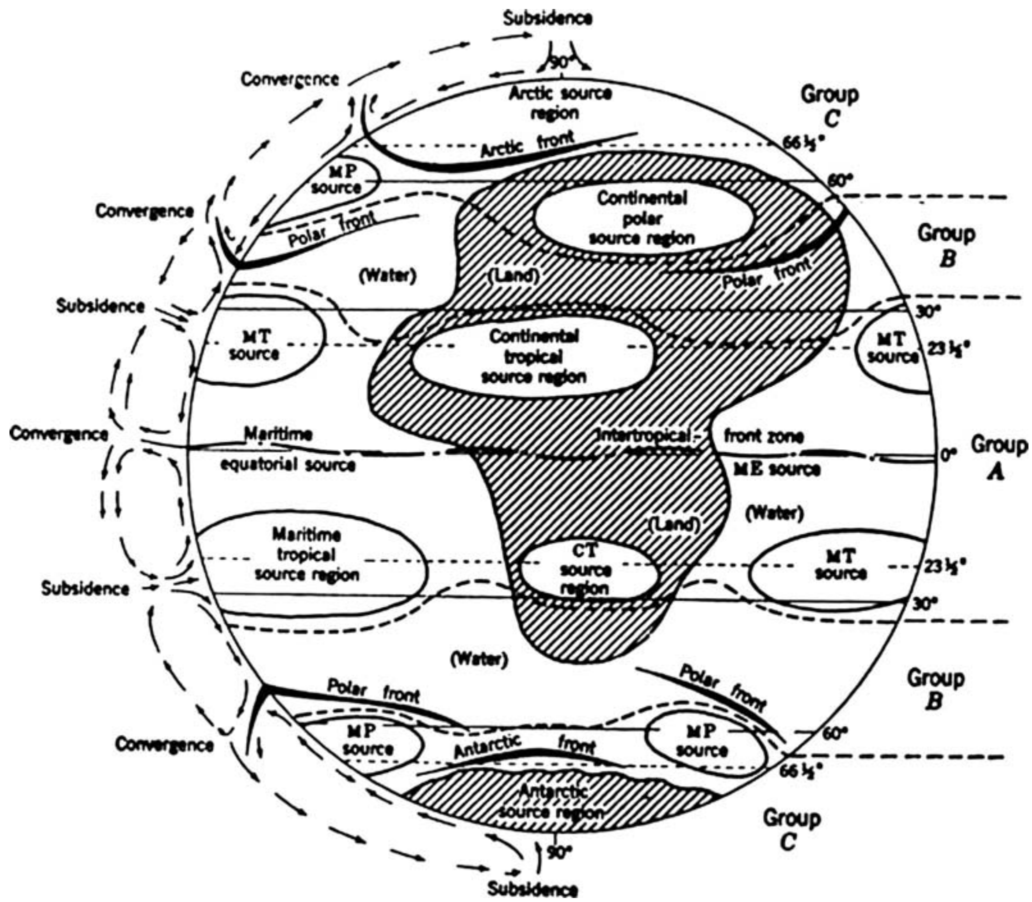

Figure C16 Global diagram showing the basis for three major climate divisions in the Strahler classification (from Strahler, 1951, by permission of John Wiley & Sons).

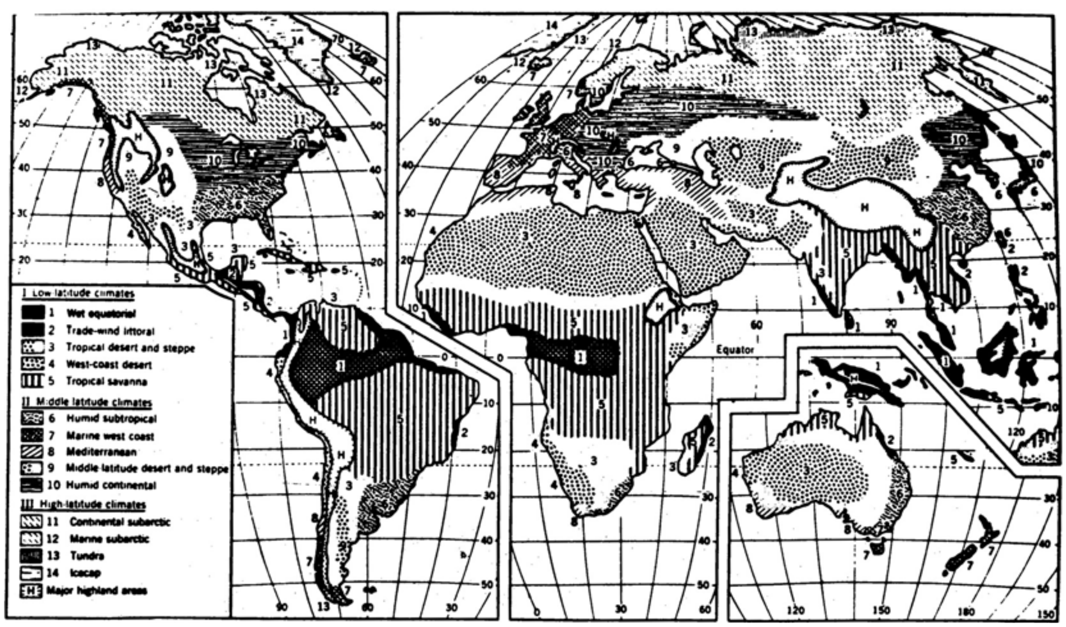

Figure C17 Generalized and simplified map showing distribution of climates according to the Strahler system (from Strahler, 1951, by permission of John Wiley & Sons).

Table C7 Numerical procedures applicable to climatic classification[a]

Similarity matrices
Correlation coefficients
Distance measures
Principal component transformation
Grouping
Taxonomic structure evaluation cophenetic correlation
Cluster selection and testing canonical correlation

[a] Derived from Balling (1984), who provides details of procedures.

for almost 90% of the total variance. Then, through linkage analysis, he showed how a grouping of Australian climates could produce a number of regions most applicable to the needs for which the classification was completed. This latter point is of considerable importance for, ultimately, classification rests with the goals of the user. This example is but one of many numerical classifications now available. In reviewing available grouping methods, Balling (1984) provides a most useful inventory of both the techniques and appropriate examples of their use. Table C7 outlines the approaches he identifies

The introduction of numerical taxonomy has opened new vistas for the regionalization of climate, and perhaps it is this realm that represents the future of climatic classification. In effect, it seems that, for general pedagogic purposes, users appear satisfied with world climatic systems that are available. It is possible that other general-purpose world classification systems may be proposed, but given the history of non-use of some excellent classifications, the likelihood of widespread use and acceptance is minimal. In contrast, the increasing significance of climate impact studies, applied climatic research and scenario analysis suggests that special-purpose classifications will be used, and that these will largely be based upon numerical classification.

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Cross-references

Airmass Climatology
Antarctic Climates
Arctic Climates
Arid Climates
Atmospheric Circulation, Global
Climatology
Evaporation
Evapotranspiration
Humid Climates
Maritime Climate
Middle Latitude Climates

CLIMATE COMFORT INDICES

Climate has profound influence on human comfort and health. Weather elements can act singly or in combination to bring about effects on human bodies. Assessment of the influence of atmospheric environment on human comfort becomes an area of great concern in the context of human health and diseases, recreation, migration, tourism, heating and ventilating industry and architecture.

To obtain relative comfort the human body has to achieve a thermal equilibrium with the surrounding environment. The basis of human thermal response is the energy balance equation, which is expressed by

$$\mathrm{M} \pm \mathrm{R} \pm \mathrm{C} - \mathrm{E} = \pm \mathrm{S} \qquad (1)$$

where M is metabolic rate, R and C are heat exchange through radiation and convection, E is heat loss through evaporation, and S is heat storage in the body. Positive S indicates an energy gain for the body whereas negative S shows an energy loss. Physiologically, least thermal stress is experienced when S equals zero. To maintain a constant core body temperature near 37°C, the human body acts as an energy exchange surface and a balance is conserved through heat exchange with the environment and heat gain from metabolism.

Human thermal comfort is the result of the combined effect of several atmospheric variables. To quantify comfort or discomfort it is necessary to devise biometeorological indices to predict the human responses to weather stress and to assess the physiological strain. These indices incorporate a combination of atmospheric parameters depending upon their applications. Also, the terms in the energy balance equation given earlier may have to be simplified or detailed. In cold environments, evaporative–expired air energy loss is assumed to be constant, and it is presumed that metabolic rates need to increase to make up the losses in the energy balance. On the contrary, in warm environments, sweating, that leads to elevated metabolic rate and inefficiency of cooling due to sweat covering the body surface, makes calculation extremely complicated.

Empirical indices

Table C8 presents representative comfort indices, largely from Driscoll (1985), Beshir and Ramsey (1988), Auliciems (1997) and Pepi (1999). The listing is not inclusive.

The Effective Temperature (ET) combines temperature and humidity into a single index; and is defined as the temperature of still-saturated air that produced the same thermal sensation as the actual atmosphere. ET is criticized for overestimation of the effect of humidity at low air temperature and underestimation at higher air temperature. ET can be modified to include air movement. Its derivative corrected effective temperature (CET) is simple and easy to use, and is considered as a useful index for engineers. Gagge et al. (1971) introduced the new effective temperature (ET*). This revised ET* employs reference conditions of 50% relative humidity and 0.6 clo of clothing insulation instead of 100% humidity and 1.0 clo. ET* can be applied in environments in high altitudes and underground mines.

The Discomfort Index (DI) was developed by the US Weather Bureau (now the National Weather Service). It gives the equivalent temperature at 100% humidity. It was renamed the Temperature–Humidity Index (THI). The index does not include radiation and wind speed.

The major criticism of the THI is the use of 100% saturation in the computation. A new index called the Summer Simmer Index (SSI), which is calculated as the THI in the usual manner, but 10% humidity is employed, was developed. Ten percent humidity is selected because it is a value experienced at typical temperature extremes in dry climates of the United States. So the SSI is an index expressing how hot one feels relative to a dry climate. A new SSI was proposed in 1999 using dewpoint of 2°C (35°F) as the dry equivalent base. This new index is confirmed by subjects testing at Kansas State University.

Humiture (H) incorporates temperature and vapor pressure to characterize warm and hot environment. The Atmospheric Environment Service of Canada modified and changed the units in the index from degree Fahrenheit to degree Celsius and named the new index Humidex (Masterton and Richardson, 1979), that is used to inform the public as to the heat stress.

The Wet-bulb Globe Temperature Index (WBGT) is an index of heat stress and incorporates air temperature, wet bulb temperature and black globe temperature. The WBGT is a better index than the old ET as it shows a good correlation with sweat rate. The index is commonly used by the US Marine Corps to control drill activity outdoors, and by industries to estimate heat stress potential in industrial environments.

The Relative Strain Index (RSI) includes temperature, humidity, air movement, the insulating effect of clothing and net radiation of heat of the body. It assumes that a person, dressed in a light business suit, walking at a moderate pace in a very light air motion, has a metabolic rate of 3.2 km/h.

The Predicted 4-hourly Sweat Rate Index (P_4SR) adds the rate of heat produced from physical work to the WBGT. The index assumes to assess the amount of sweat perspired by a physically fit young man in the conditions under review for a period of 4 hours. It is commonly utilized in hot climates to estimate sweat loss and the required water intake.

The Belding–Hatch Heat Index also involves the heat rate produced by physical work. It compares the amount of sweat

Table C8 Representative comfort indices

Name of index	Description or comment
Effective Temperature (ET)	Physiological principles are not considered. Overestimates the effect of humidity at low temperature.
Corrected Effective Temperature (CET)	A modification of ET. Combines temperature, humidity, air velocity and radiation. Underestimates the effect of humidity and low air movement. Used in the British armed forces.
New Effective Temperature (ET*)	Used by ASHRAE for indoor comfort. Good indicator of physiological strain and warmth discomfort. Difficult to apply and complicated instruments are needed.
Temperature–Humidity Index (THI); formerly called the Discomfort Index (DI)	Gives the equivalent temperature at 100% humidity.
Summer Simmer Index (SSI)	Same as THI but 10% humidity is used. To indicate how hot one feels to a dry climate. New SSI proposed in 1999, also related to a dry environment and a dewpoint base of 2°C (35°F) is used.
Humiture (H)	Incorporates air temperature and vapor pressure. The Atmospheric Environment Service of Canada modified the index and named it the Humidex, that is more sensitive to humidity change.
Wet-Bulb Globe Temperature Index (WBGT)	Good correlation with sweat rate. Practical for industrial purposes. Estimation gets poorer under low humidity conditions. Metabolic workload is not considered.
Relative Strain Index (RSI)	Includes temperature, humidity, air movement, clothing and radiation of heat of the body. Applicable to assess heat stress of manual workers under shelter at various metabolic rates.
Predicted 4-hourly Sweat Rate Index (P_4SR)	Same as WBGT with addition of heat rate produced by physical work. Estimates sweat loss and required water intake.
Belding–Hatch Heat Index	Same as WBGT with addition of heat rate produced by physical work. Evaluates hot working environment. Underestimates the adverse effect of low wind speed and hot humid environment.
Apparent Temperature (AT)	To quantify sultriness.
Heat Index (HI)	A multiple regression model converted from AT. Used by the US NWS to alert the public of heat stress.
Wind Chill	Most widely used cold stress index. A new formula was implemented by the US NWS on 1 November 2001.

needed to maintain thermal equilibrium to the maximum amount of sweat that can be evaporated. It incorporates air temperature, globe temperature and the wet bulb temperature.

Steadman (1979a,b) studied sultriness that led to the development of the Apparent Temperature (AT). AT was an attempt to quantify sultriness. The index incorporated temperature, humidity, clothing and human physiology. The algorithms were simplified and regression equations were provided for indoor, shaded and sunny conditions (Steadman, 1984). The US National Wealth Service (NWS) converted AT into a multiple regression model called Heat Index (HI), which is used to alert the public to heat stress. Table C9 presents the apparent temperatures of HI, and the heat stress categories and their associated symptoms based on the severity of impact on humans.

Wind Chill (WCT), a measure of the combined effects of low temperature and wind, was devised by Siple and Passel (1945) on the basis of experiments examining the freezing rate of water in a cylinder at various temperatures and wind speeds in Antarctica. The index and derived wind chill equivalent temperature are widely used, especially among the media; meanwhile they have also received criticisms over the past five decades. The recent reviews of wind chill are given by Dixon and Prior (1987) and Brauner and Shacham (1995). Steadman (1971) introduced a new wind chill formula that included the effect of clothing and all forms of heat loss. The model was justified in human conditions. However, it is less popular due to its complex equation. Because of the dissatisfaction of the WCT, the Office of the Federal Coordinator for Meteorological Services and Supporting Research (OFCM) formed a special group called the Joint Action Group for Temperature Indices (JAG/TI), that included the Meteorological Services of Canada (MSC), the US NWS and several members of the academic community, to develop a new WCT formula. Besides mathematical modeling, human experiments were also involved. Volunteers, with temperature sensors on their faces, were exposed to various thermal conditions in a wind tunnel. The

Table C9 The Heat Index in terms of air temperature and relative humidity

Air temperature (°C)	Relative humidity (%) 0	10	20	30	40	50	60	70	80	90	100
20	16	17	17	18	19	19	20	20	21	21	21
21	18	18	19	19	20	20	21	21	22	22	23
22	19	19	20	20	21	21	22	22	23	23	24
23	20	20	21	22	22	23	23	24	24	24	25
24	21	22	22	23	23	24	24	25	25	26	26
25	22	23	24	24	24	25	25	26	27	27	28
26	24	24	25	25	26	26	27	27	28	29	30
27	25	25	26	26	27	27	28	29	30	31	33
28	26	26	27	27	28	29	29	31	32	34	36
29	26	27	27	28	29	30	31	33	35	37	40
30	27	28	28	29	30	31	33	35	37	40	45
31	28	29	29	30	31	33	35	37	40	45	
32	29	29	30	31	33	35	37	40	44	51	
33	29	30	31	33	34	36	39	43	49		
34	30	31	32	34	36	38	42	47			
35	31	32	33	35	37	40	45	51			
36	32	33	35	37	39	43	49				
37	32	34	36	38	41	46					
38	33	35	37	40	44	49					
39	34	36	38	41	46						
40	35	37	40	43	49						
41	35	38	41	45							
42	36	39	42	47							
43	37	40	44	49							
44	38	41	45	52							
45	38	42	47								
46	39	43	49								
47	40	44	51								
48	41	45	53								
49	42	47									
50	42	48									

The US National Weather Service has guidelines for the following ranges of heat index:

32–40°C Heatstroke, heat cramps, or heat exhaustion are possible with prolonged exposure and/or physical activity.
41–53°C Heat cramps or heat exhaustion are likely, and heatstroke possible with continued exposure.
≥54°C Heatstroke is highly likely with continued exposure.

NWS implemented the new WCT on 1 November 2001. The new formula is shown in equation (2).

$$\mathrm{WCT} = 13.13 + 0.62T - 13.95V^{0.16} + 0.486TV^{0.16} \quad (2)$$

where T is the air temperature (°C) and V is wind speed (m/s).Using the new WCT formula, frostbite may occur in 30 min or less according to Table C10.

Energy balance models

A large variety of models have been developed to include all energy exchange processes and outdoor environments. These models are extremely complex. The major models are presented in Table C11.

Fanger (1970) devised a model that comprises all forms of heat exchange. The model was based on the results of laboratory experiments involving nearly 1400 subjects wearing light clothing, and the relationship between the computed energy exchange and the predicted mean vote (PMV) of comfort sensation was derived. The PMV of comfort sensation is rated on a seven-point scale ranging from +3 (hot) to −3 (cold). The PMV equation represents the human physiological responses to particular environments. The model was designed for indoor conditions and has been modified to incorporate the complex outdoor radiation conditions, being known as the Klima Michel model (Jendritzky et al., 1979).

Höppe (1984) proposed the MEMI model (Munich Energy-balance Model for Individuals) that took into account the basic thermoregulatory processes, such as constriction or dilation of peripheral blood vessels and sweat rate. The MEMI consists of three equations. They are the energy balance equation of the total body, the equation of heat flux from the body core to the skin, and the equation of heat flux from the skin through clothing layers to the surface of the clothing. The model is the basis for the calculation of the physiological equivalent temperature (PET), that is equivalent to the air temperature at which, in a typical room indoors, the heat budget of the human body is balanced with the same core and skin temperatures as under the outdoor conditions being assessed.

Blazejczyk (1994) proposed the MENEX model (Man–environment Heat Exchange Model), that is the evolution of the model proposed by Budyko (1974). The model also takes into account human thermo-physiological considerations. Its applications include forecasting human thermal conditions outdoors, and evaluations of bioclimates and heat load at work.

De Freitas (1985) developed the STEBIDEX (Skin Temperature Energy Balance Index) and the HEBIDEX (Heat

Table C10 The new Wind Chill Temperature (WCT) chart

	T_{air}											
V_{10}	5	0	−5	−10	−15	−20	−25	−30	−35	−40	−45	−50
5	4	−2	−7	−13	−19	−24	−30	−36	−41	−47	−53	−58
10	3	−3	−9	−15	−21	−27	−33	−39	−45	−51	−57	**−63**
15	2	−4	−11	−17	−23	−29	−35	−41	−48	−54	**−60**	**−66**
20	1	−5	−12	−18	−24	−31	−37	−43	−49	−56	**−62**	**−68**
25	1	−6	−12	−19	−25	−32	−38	−45	−51	−57	**−64**	**−70**
30	0	−7	−13	−20	−26	−33	−39	−46	−52	−59	**−65**	**−72**
35	0	−7	−14	−20	−27	−33	−40	−47	−53	**−60**	**−66**	**−73**
40	−1	−7	−14	−21	−27	−34	−41	−48	−54	**−61**	**−68**	**−74**
45	−1	−8	−15	−21	−28	−35	−42	−48	−55	**−62**	**−69**	**−75**
50	−1	−8	−15	−22	−29	−35	−42	−49	−56	**−63**	**−70**	**−76**
55	−2	−9	−15	−22	−29	−36	−43	−50	−57	**−63**	**−70**	**−77**
60	−2	−9	−16	−23	−30	−37	−43	−50	−57	**−64**	**−71**	**−78**
65	−2	−9	−16	−23	−30	−37	−44	−51	−58	**−65**	**−72**	**−79**
70	−2	−9	−16	−23	−30	−37	−44	−51	−59	**−66**	**−73**	**−80**
75	−3	−10	−17	−24	−31	−38	−45	−52	−59	**−66**	**−73**	**−80**
80	−3	−10	−17	−24	−31	−38	−45	−52	**−60**	**−67**	**−74**	**−81**

Approximate thresholds:
Risk of frostbite in prolonged exposure: wind chill below **−25**
Frostbite possible in 10 minutes at **−35** Warm skin, suddenly exposed. Shorter time if skin is cool at the start.
Frostbite possible in less than 2 minutes at **−60** Warm skin, suddenly exposed. Shorter time if skin is cool at the start.

T_{air} = air temperature in °C and V_{10} = observed wind speed at 10 m elevation, in km/h; courtesy of the Meterological Service of Canada.

Table C11 Major energy balance models

Name of model	Description or comment
Predicted Mean Vote (PMV)	First model including all forms of heat exchange. Mean skin temperature and sweat rate are quantified as comfort values, being only dependent on activities.
MEMI	Thermo-physiological heat balance model. Basis for the calculation of PET.
MENEX	Thermo-physiological heat balance model.
Klima Michel Model	Modification of Fanger's PMV to incorporate complex outdoor radiation.
HEBIDEX and STEBIDEX	To identify the relationship between the environmental stress being experienced and the condition of mind, being expressed in thermal sensation.
STOEC	Based on the fall of body core temperature. Applied in cold environment.

Budget Index). The adequacy of both indices was tested by empirical study of human thermal response outdoors on the beach. The results indicated that the STEBIDEX model could provide a more reliable estimate of thermal sensation.

De Freitas and Symon (1987) developed the STOEC (Survival Time Outdoors in Extreme Cold). The STOEC incorporates temperature, wind, solar radiation, clothing and the human energy budget, which is based on the rate of fall of core temperature form 37°C to 27°C. It provides the calculations for the shortest, longest, mean and indefinite survival times. It is useful in estimating the duration of research and rescue operations.

Conclusion

The trends in the development of comfort indices reflect the move from the conventional static approach to environmental assessment. This change has been toward the use of dynamic energy exchange models, that synthesize the interaction between the human thermo-physiological processes and the atmospheric environments. However, it is important for an index to provide realistic meaning and simple interpretation of abstract index value to the lay public.

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Cross-references

Architecture and Climate
Heat Index
Seasonal Affective Disorder
Human Health and Climate
Wind Chill

CLIMATE DATA CENTERS

With recent advances in computer storage capability, and the proliferation of the Internet, climate data for virtually every country in the world are becoming easily obtainable for researchers and the general public alike. Many countries have set up, within the organizational structure of the various departments of meteorology and climatology, systems for gathering, editing and archiving the volumes of climate information that are available. While the following is not a comprehensive list of the climate centers around the world, it represents places of international renown that service the research and public domain communities by providing raw, edited, or aggregated datasets and have important links to modeling and research centers.

Japan

Tokyo Climate Center

The Tokyo Climate Center (TCC) was established in April 2002 to provide climate data and services to the Asian-Pacific environment and to assist in the mission of the Japan Meteorological Agency and the National Meteorological and Hydrological Service of Japan. The activities of the center include monthly reporting of climate events of a global nature and investigations in the climate systems that may affect the weather of the Asian-Pacific realm, including, and especially reporting on, the El Niño phenomenon and to produce ensemble monthly forecast models and verification of them.

The TCC is involved in assisting with the technical expertise needed to design the delivery of climate services throughout the Asia-Pacific region and to provide climatology data to the region. It also works in concert with other organizations in nations within the region to help facilitate the delivery of climate data, services and forecasts by the meteorological and climatological agencies in other Asia-Pacific nations.

United Kingdom

British Atmospheric Data Center (BADC)

The BADC is the main archive for atmospheric data in the UK. It is one of seven data centers designated by the National Environmental Research Center and is housed within the Space Science and Technology Department of the Rutherford Appleton Laboratory in Oxfordshire. The BADC produces datasets derived from NERC-funded projects, and serves as an efficient link to data provided by such services as the European Center for Medium-Range Weather Forecasting Office (the ECMWF) and the UK Meteorology Office.

The BADC was established in 1994 building upon the work of the former Geophysical Data Facility (GDF). The BADC presently holds up to 70 datasets of a variety of atmospheric variables and is the archive for the ECMWF reanalysis dataset. In addition, research work by BADC staff includes projects delegated to three main groups: atmospheric dynamics, cloud physics and data assimilation. BADC staff are also engaged in research efforts with the RAL SSTD Atmospheric Modeling and Data Interpretation Group (AMDI).

United States

Carbon Dioxide Information Analysis Center, CDIAC, Oak Ridge, Tennessee

The CDIAC, operating out of the Oak Ridge Laboratory, compile data and information related to greenhouse gases, their rates of emission, and the cycle of these gases between the oceans, the biosphere and the atmosphere. They also have surface climate data, edited for selected stations that are part of the United States Historical Climate Network and historic mid-tropospheric height data. The CDIAC also houses surface data for the People's Republic of China. The World Data Center for Atmospheric Trace Gases is housed in the CDIAC.

The Climate Diagnostics Center (NOAA – CIRES, CDC), Boulder, Colorado

The CDC is a research and archival unit designed to analyze climate events. It archives the National Center for Environmental Protections Re-Analysis Dataset, as well as time series of sea-surface temperatures, various El Niño and Southern Oscillation indices, as well as the northern hemisphere teleconnection indices.

US National Snow and Ice Data Center (NSIDC), Boulder, Colorado

The NSIDC, under a joint agreement with the University of Colorado, operates houses the World Data Center for Glaciology. The center compiles, archives and maintains data related to snow cover, snow depth, snow pack, sea-ice extent and depth, freshwater ice and glacier extent.

US National Climate Data Center (NCDC), Asheville, North Carolina

The NCDC has the largest climate and meteorology database in the world. It also houses the World Data Center for Meteorology. Designed originally to house weather records obtained by National Weather Service offices and cooperative stations, it now archives 99% of the data obtained by departments of the National Oceanic and Atmospheric Administration (NOAA). The center also maintains and archives radar data, satellite imagery from 1960, radiosonde,

Table C12 Selected members of the World Data Center network directly related to research in climatology

World Data Centers	Affiliation and location	Data emphasis
Atmospheric Trace Gases	Carbon Dioxide Information Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee	Data related to atmospheric trace gases that affect and contribute to the Earth's energy budget.
Glaciology – USA	National Snow and Ice Data Center, Boulder, Colorado	Snow cover, snow pack, sea-ice extent, sea-ice thickness, images and pictures of historic glacial extent.
Glaciology – Geocryology, China	Lanzhou Institute of Glaciology and Geocryology, Chinese Academey of Sciences. Lanzhou, China	Glacial atlas of China, snow cover, glacial extent and variation, periglacial data and hydrologic data.
Glaciology, UK	The Royal Society and the Scott Polar Research Institute, University of Cambridge	Data related to glaciers, periglacial processes, satellite imagery, snow and ice chemistry.
Airglow – Japan	National Astronomical Observatory, Tokyo, Japan	Airglow data, solar radiation data.
Meteorology – China	Climate Data and Applications Office, Beijing, China	Real-time synoptic data, historical surface climate data, dendrochronology data, glacial data and atmospheric chemistry data.
Meteorology – Russia	Federal Service of Russia for Hydrometeorology and Monitoring of the Environment, Obninsk, Russia	Both surface observations and gridded surface and upper air meteorology data, marine ship observation data, and aerology data.
Meteorology – USA	National Climate Data Center, Ashville, North Carolina	Archives of data from many national and international research projects and experiments including data from the IGY 1957–1958 and International Quiet Sun Year, 1964–1965, among many others.
Paleo-climatology – USA	National Geophysical Data Center (NGDC), Boulder, Colorado	Dendrochronology data, ice-core data, sea-floor sediment cores, Coral data, proxy data on climatic forcing, including volcanic aerosol data, ice volume, atmospheric composition, etc. Numerical model simulation experiments data, climate reconstructions and maps.

rawinsonde, rocketsonde and other advanced technology sources. The center also houses weather data from diaries, notes and other sources. The center also services a network of regional and state climatology data centers designed to deliver climate data of regional and local interest.

International Organizations

The Intergovernmental Panel on Climate Change Data Distribution Center (DCC)

The DCC acts as a gateway to many datasets used to assess changes in the global climate. The DDC acts in concert with the Deutsches Klimareschenzentrum (DKRZ) in Germany and the Climate Research Unit (CRU) in the United Kingdom to archive and distribute much of the data that the IPCC provides. The DDC maintains links to locations that have data in the public domain. Output from Global Circulation Model (GCM) runs are available at the DKRZ distribution site. The DDC also archives the CRU Global Climate Dataset that has monthly surface variables from 1905 to 1995 and is arranged on a 0.5° by 0.5° latitude–longitude grid. This includes all land grid cells except for Antarctica. The DDC also maintains another dataset arranged by country and containing both averages (1961–1990) and month-by-month variations (1901–1995).

World Climate Data Centers

The World Climate Data Center system is a network of data centers that was originally designed to archive the data derived during the International Geophysical Year (IGY) but has subsequently expanded to include climatological, meteorological, astronomical, oceanograhical as well as geophysical datasets. The centers are usually housed within other organizations. Each of the following regions, United States, Europe, Australia, China, Russia, Japan and India, have several centers specializing in one of the areas of geophysical data and research. Table C12 lists those centers that have data directly related to research in climatology in particular.

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Cross-references

Climate Modeling and Research Centers
Models, Climatic
Oscillations
Teleconnections

CLIMATE HAZARDS

This review of climate hazards outlines the great variety of such events and their impacts. It illustrates some of the present concerns of and future challenges for the climatologist.

Nature and variety of the threats

Potentially hazardous atmospheric phenomena include tropical cyclones, thunderstorms, tornadoes, drought, rain, hail, snow, lightning, fog, wind, temperature extremes, air pollution, and climatic change. Hazards may arise from single-element extremes, such as excessively high temperatures causing physiological heat stress; or from various combinations of elements, such as tropical cyclones with high wind, torrential rain and storm surge, all posing threats to people and their property. Estimates of annual global economic losses due to meteorological disasters showed a fourfold increase from the 1960s to the early 1990s to nearly \$90 billion, and insured losses increased about tenfold to over \$50 billion (Bruce, 1994).

Hazards producing disasters with economic, political, and social repercussions have always been newsworthy. The public view of climatic hazards has been nurtured by frequent media reports of events such as the tropical cyclone that killed about 300 000 people in the former East Pakistan in 1970; the severe frost damage to coffee crops in Brazil in 1975, followed by huge increases in world coffee prices; the storm that devastated the Fastnet yacht race in 1980; the exceptional storm in 1988 which left the most severe damage for many generations in southern England; another tropical cyclone in Bangladesh in 1991 which killed about 139 000 people; Yangtze floods which destroyed over four million dwellings in 1991; Hurricane Andrew which produced damage estimated at over \$25 billion in Florida and Louisiana in 1992; ice and snow storms in eastern North America in 1993, 1994 and 1998, with total damages of several billions of dollars; the heatwave in the Midwest and eastern USA in 1995; El Niño-related weather events in 1997–1998 around the world; Hurricane Mitch in Central America in 1998; and in 2002 Typhoon Rusa, the deadliest storm in South Korea in over 40 years, as well as widespread flooding in Europe.

Perhaps Leonardo da Vinci was the first to recognize the downburst associated with severe thunderstorms (Gedzelman, 1990), but there is little evidence of research on climatic hazards before the nineteenth century, with early emphasis mainly on the cataloguing of events, such as tropical cyclones in a particular area (e.g. Knipping, 1893). The work of Barrows (1923) on human structuring of, and adjustment to, environment provided an identifiable start for a more balanced research tradition, the development of which may be traced through studies such as those by Visher (1925) on tropical cyclones in the Pacific Ocean and by Foley (1957) on drought in Australia. Interest and emphasis have shifted to a much greater concern with the nature and alleviation of impacts, as reflected in the wide range of natural, physical, and behavioral scientists, and economists working on frequencies, magnitudes, causes, behavior, and impacts of climatic hazards, as well as with studies of perception, preparedness, planning, mitigation and control. This diversity of interests, and the directions in which research has been heading, is exemplified in works such as those by White (1974), Heathcote and Thom (1979), Hewitt (1983, 1997), Burton et al. (1993) and Smith (2001).

Identification of hazardous events is not always easy, although certain criteria are usually present. They include property damage; economic loss, such as loss of income or a halt in production; major disruption of social services, communications and transportation; excessive strain on essential services such as police, fire, hospitals, and public utilities; and psychological stress, injuries, and fatalities (e.g. Changnon, 1989, 1999; Barker and Miller, 1990; Morison and Butterfield, 1990; Hoque et al., 1993; Brugge, 1994a, b; Kalkstein, 1995; Fink et al., 1996; Curran et al., 2000; Dupigny-Giroux, 2000; Palecki et al., 2001; Ulbrich et al., 2001; Pielke and Carbone, 2002).

It can be difficult to distinguish between the atmospheric and nonatmospheric factors producing climatic hazards. Hot, dry winds may promote fire disasters but do not cause them. An avalanche depends on the quality and quantity of snow and on the timing of a thaw, but it is unlikely to happen without certain slope characteristics. Increasing losses over time point to the significance of socioeconomic factors in exacerbating the vulnerability of communities to hazard events. More than half of the world's population lives within 60 km of the ocean. Known flood plains, hurricane- and drought-prone regions have experienced development pressures and increased volumes and values of property at risk (e.g. Riebsame et al., 1986; Pielke and Pielke, 1997; Kunkel et al., 1999; Changnon et al., 2000; Easterling et al., 2000). Climate change is also seen by some as creating potential new threats (e.g. Obasi, 1994; Woodhouse and Overpeck, 1998; Yarnal et al., 1999; Parry et al., 2001). There are already very large disparities in the nature and magnitude of hazard losses in different parts of the world; differences mainly attributable to socioeconomic factors (Degg, 1992).

Atmospheric, and socioeconomic, factors fulfill a variety of roles in the development of a hazardous situation. A broad distinction can be made between phenomena such as tropical cyclones or severe local storms, and their associated weather extremes, which involve the sudden impact of very large amounts of energy discharged over relatively short periods; and those features that become hazards only if they exceed tolerable magnitudes within or beyond certain limits (Gentilli, 1979). In the latter category can be included heat waves; cold spells; flood-producing rains; frosts; fogs; droughts; high winds, snow and ice associated with extratropical low-pressure systems; and the effects of climatic change. Some climatic hazards result from human activity. Under the broad umbrella of air pollution, these include hazards to human health, the possible dangers of inadvertent modification of climatic patterns, and the effects of acid rain on natural ecosystems.

The treatment of climatic hazards that follows is not exhaustive, nor is it the only approach possible, but it illustrates various types of threats and their impacts. Examples of alternative approaches include Changnon and Changnon (1992), who group 11 causes of storms into four classes; and Smith (1997, 2001), who classifies climatic hazards into four groups: single-element extremes from common and less common hazards, and compound-element events from primary or secondary hazards.

Sudden-impact hazards

Tropical cyclones

Tropical cyclones can be the most dangerous and deadly storms on Earth. They are usually very mobile and relatively unpredictable. The main dangers to people and property arise from three distinct hazards: violent winds, storm waves and surges, and torrential rain. Winds in Hurricane Gilbert in the Caribbean in 1988 gusted to over 320 km/h (Eden, 1988), and in Hurricane Linda in the East Pacific in 1997 reached about 350 km/h (Brugge, 1998). Sustained winds approaching 280 km/h were reported for Hurricane Gilbert, but most tropical cyclones do not reach such intensity. Death and injury may result from structural collapse or from the impact of flying objects. Rainfall associated with tropical cyclones may total over 1000 mm in 24 h. Cyclone Hyacinthe dumped 6433 mm of rain on Reunion in 14 days in January 1980, with more than 1000 mm falling on each of two successive days (Smithson, 1993). During Hurricane Mitch in October 1998 parts of Honduras received 698 mm in 41 h, leading to widespread flooding and landslides (Hellin and Haigh, 1999). Flooding can be severe when such heavy rain falls in restricted catchments, or when run-off combines with storm wave and storm surge effects. The northern Bay of Bengal suffers a particularly serious storm surge problem because of a combination of large astronomical tides, a funneling coastal configuration, low and flat terrain, and frequent severe tropical storms. The November 1970 storm in this region may have been the deadliest ever to devastate a coastal area in recent times, with a storm surge of over 6 m and about 300 000 people killed (Frank and Husain, 1971), although tropical cyclones in the same region in 1737 and 1876 may have rivaled this (Sensarma, 1994).

Most of the 80–100 tropical cyclones each year form between latitudes 5° and 25°, 60–70% in the northern hemisphere. Tropical cyclone tracks sometimes reach beyond 401 latitude. The area immediately affected by the full force of a storm is typically about 1° by 1° latitude, corresponding to approximately 100 km of coastline. On this basis there are about 800 separate prime target areas around the globe, each with peculiar local conditions, especially with respect to storm surges.

Tropical cyclones are responsible for an annual average of about 20 000 deaths and over $6 billion in damage globally (Obasi, 1994). In some cases, like the 1991 Bangladesh storm which destroyed over 500 000 dwellings, millions of residents can be directly affected (Haque and Blair, 1993). Three types of most vulnerable areas can be identified: densely populated, fertile coastal plains and deltas (e.g. the Ganges); island groups dependent on agricultural economies (e.g. Oceania, the Caribbean, the Philippines); and highly populated coastal regions developed as residential resorts (e.g. Florida, the

Queensland Gold Coast) or for industry (e.g. the Texas Gulf Coast, Japan) (Stevens, 1991).

Regional variation in the socioeconomic impact of tropical cyclones is related to factors such as the geographic vulnerability of communities, their experience with severe storms, population density, coastal and inland topography, land use, social organization, property and infrastructure at risk, and warning and response capabilities. Extensive delta regions in developing countries present the greatest potential for loss of life.

Evacuation is one form of emergency response in developed countries. However, as population densities increase, evacuation plans may prove inadequate, so that evacuation times may exceed what is feasible given the number of people to be moved and likely lead times from existing forecast capabilities (American Meteorological Society, 1993, 2000b). Costs of evacuation, estimated at about \$0.6 million per kilometer of coastline (Pielke and Carbone, 2002), could in some cases at least match those resulting from storm damage. Research after cyclone Tracy hit Darwin has suggested that mass evacuation may pose more problems than it solves, and that greater use of the victims' resources might speed up reconstruction and reduce psychological stress (Western and Milne, 1979). Research on the impact of tropical cyclones in such areas, however, cannot adequately represent impacts in countries with labor-intensive rural economies. Hurricane Gilbert produced damage in Jamaica estimated at \$800–1000 million, and estimates of reinsurance inflows were \$650 million; even the lower value exceeded the country's annual foreign exchange earnings from exports (Barker and Miller, 1990). Shortfalls in domestic food production had to be met by imports, virtually the whole banana crop and most of the coffee crop were wiped out. In Bangladesh in 1991 an estimated 51 000 ha of crops were completely destroyed and over 150 000 ha were partially damaged (Haque and Blair, 1993).

An industrialized society may plan for a low rate of building failures. This requires a high degree of engineering attention to housing design and construction, together with stringent building regulations. However, a developing country may elect to accept a higher level of building failures, adopting a strategy of temporary housing during and after a tropical cyclone. A variety of public buildings would then be strengthened and designed as refugee or reception centers. These contrasting approaches are exemplified by the decisions made by authorities in Darwin after cyclone Tracy in 1974 and by those in Sri Lanka after a rare tropical cyclone in 1978.

Most building damage is caused by the effects of wind on buildings that are not properly engineered, such as domestic houses and small, low-rise industrial and commercial structures. The effects of extreme winds on buildings are well understood, and wind engineering technology in theory can be applied anywhere. The impact of tropical cyclones is greatest when the population is rendered homeless, so housing design in vulnerable areas should include consideration of the effects of wind and water. Areas of infrequent tropical cyclone occurrence (e.g. the Queensland Gold Coast, the Atlantic Coast of the United States north of Cape Hatteras, parts of Mexico, India, and Japan) tend to have greater building vulnerability; whereas communities in areas battered relatively frequently by tropical cyclones have learned to cope with the hazard and tend to be less visible as disaster areas (e.g. Mauritius, Reunion Island, Guam, Northwest Australia, some South Pacific islands).

Improvements in warnings and in community preparedness have decreased the death toll from tropical cyclones, but the average annual inflation-adjusted hurricane losses in the United States have grown from about \$5 billion in the 1940s to more than \$40 billion in the 1990s (Easterling et al., 2000). In developing countries, where some of the expected increase in world population will be in regions subject to tropical cyclone impact, a growing death toll and damage may be unavoidable, partly due to limitations on national resources to apply effective mitigation measures and partly due to a lack of understanding of the vulnerability of certain areas to the effects of tropical cyclones. In all parts of the world there is a continuing need for better community education about the potentially damaging hazards and about long-term mitigation planning (American Meteorological Society, 2000b).

The warning system is a primary feature of organization for disaster preparedness. Key elements in a warning system, apart from early detection and accurate forecasting, are the efficiency of the dissemination process and the reaction of the community. Education is vital to ensure that warnings are understood and that people are aware of the necessity to heed such warnings. Hazard warning involves a sensitive balance between overwarning, resulting in unnecessary and sometimes expensive preparations and leading ultimately to complacency and apathy, and underwarning, which gives insufficient time for adequate protective measures to be taken. The benefits of improvements in tropical cyclone detection technology and in behavior prediction are lost without corresponding progress in ability to utilize the information in planning, organizing, and acting for protection and convenience (Jagger et al., 2002). McAdie and Lawrence (2000) note that tropical cyclone track forecasts for the Atlantic basin improved from 1970 to 1998, but Powell and Aberson (2001) suggest that no statistically significant change is apparent for landfall position forecasts during the last 25 years. In most other parts of the world the situation is no better. In Taiwan, for example, hit by an average of nearly four typhoons per year, forecasting, and hence warning, is difficult because of lack of data over the North Pacific (Wu and Kuo, 1999). Despite some recent possible reductions in track forecast errors there has been little improvement in forecasts of storm intensity or structure, including overall size; and, while there have been advances in basic understanding, accurate predictions of tropical cyclone genesis are still some way off (American Meteorological Society, 2000b).

There were many false warnings in East Pakistan prior to the storm of November 1970, mainly because of lack of facilities to distinguish between killer and nonkiller cyclones. An estimated 90+% of the people in the disaster area knew about the storm, yet less than 1% sought refuge in substantial buildings. Most residents had never experienced a storm surge like that predicted, and thus felt no urgency to leave their homes. In addition, few had the means to move, or anywhere to go in the time available. A comparable situation occurred in Darwin with cyclone Tracy when the potential benefits of technically good predictions were not realized. Records show several near-misses for Darwin, including cyclone Selma only 3 weeks before Tracy, and at least six tropical cyclones which had seriously affected the community. However, the last severe event had been in 1937 and was remembered by few residents in 1974; Christmas Eve as Tracy approached, and Christmas Day when it struck, were not ideal times to muster enthusiasm for effective action.

Severe local storms

Severe local storm hazards are widespread, relatively unpredictable, seemingly impossible to prevent, and often costly in lives and property damage (Atlas, 1976). At any instant there may be about 2000 active thunderstorms around the world (Dudhia, 1996). As separate cells, or as organized line squalls, thunderstorms develop cold downdraughts with high velocities that are capable of causing severe localized damage. The squall of a thunderstorm can gust to 185 km/h, and its effects are often compounded by intense rainfall, large hail, or lightning. Individual storms usually affect only small areas, but there may be many such storms at any one time in a particular region. Their association with flash floods, downbursts, strong winds, tornadoes and lightning makes accurate forecasting vital.

High-intensity, localized thunderstorms may produce flash floods. A storm in August 1988 at Khartoum, with daily rainfall of about 200 mm being more than twice the previous fall on record, left tens of thousands of homes destroyed and vast areas around the city inundated (Hulme and Trilsbach, 1989). An estimated 1.5 million were made homeless, and there were over 100 deaths and hundreds of injuries. Most of the homeless had nowhere to go; even areas which had not suffered damage were under water and there were few areas dry enough for temporary camps to be erected. A flash flood, produced by rainfall approaching a 500-year return period event, at Fort Collins in Colorado in July 1997, caused five deaths, 62 injuries and more than $250 million in property damage (Weaver et al., 2000). Lessons from that event, with a systematic effort to improve awareness, communication and warning, meant that another flash flood in April 1999 caused significantly less damage.

Hailstones are usually less than 10 mm in diameter and cause little damage, but they are occasionally 100 mm or more in diameter and can cause serious damage to crops, buildings, and motor vehicles. About 20 000 severe thunderstorms occur annually in the USA, with damage to property and agriculture of up to $3 billion (Bentley et al., 2002). Hailstones up to softball size were reported in a storm causing about $350 million damage in Denver in June 1984 (Blanchard and Howard, 1986), and there have been at least two reports from South Africa of coconut-sized stones (Perry, 1995). In the period 1982–1989, 250 people were reported killed by hail in India (Nizamuddin, 1993). Giant hailstones weighing 2–3 kg were reported in February 1988 in parts of Orissa state.

Thunderstorms are the most common cause of air traffic delays and play a major role in weather-related aircraft accidents (Bromley, 1977). Turbulence, hail, and wind shear within storm clouds have damaged many aircraft, with wind shear recognized as a major aviation hazard, particularly in the airport environment. The gust front ahead of the cold outflow from thunderstorm downdraughts is particularly hazardous because of associated large surface wind shears and because of its very localized nature. The strongest downdraughts, those most likely to be hazardous to aircraft during takeoff and landing, are called downbursts or microbursts (Fujita and Caracena, 1977; McCarthy and Serafin, 1984; Dudhia, 1997). Very high-resolution modeling of a developing thunderstorm has confirmed reports from aircraft of relatively narrow regions of turbulence in layers more than 1 km deep above the cloud (Lane et al., 2002). Thunderstorm downdraughts may also be a critical factor in driving some bushfire fronts; sometimes the fires themselves having been started by lightning strikes.

Lightning is one of the major causes of fatalities in the USA and is a significant hazard to outdoor activity during the summer months (Watson and Holle, 1996; Holle et al., 1999; Curran et al., 2000). It is probable that lightning deaths and injuries are considerably underestimated (Lopez et al., 1993), with most people displaying inappropriate behavior during thunderstorms and not realizing the range of possible medical implications from a lightning strike, including paralysis, external burns, severe headaches, hearing and memory loss, and many others (Shearman and Ojala, 1999).

Some thunderstorms and the peripheral circulation of some tropical cyclones are accompanied by tornadoes, which are among the smallest but most destructive features of atmospheric circulation. Tornado wind speeds can exceed 350 km/h. The Tri-State tornado in March 1925 traveled about 350 km from Missouri to Indiana at speeds of 91 km/h to 109 km/h. About 80 km^2 of land were totally devastated, 689 people were killed, nearly 2000 people were injured, and over 11 000 people were left homeless (Flora, 1953). Over 2 days in April 1974 an outbreak of 148 tornadoes in the Midwest and southeast USA left 350 dead (Brugge, 1994c). On a single day in November 1981 an outbreak of 102 tornadoes struck parts of England and Wales (Rowe and Meaden, 1985). Just one tornado in May 1996 is reported to have killed at least 400 people in Bangladesh (Snow and Wyatt, 1997).

The global distribution of tornadoes is difficult to determine accurately (Perry and Reynolds, 1993). Most reports of tornadoes come from the United States, but they occur in many parts of the world including most of Europe, northern India and Bangladesh, Australia and New Zealand, Japan, Uruguay, and southern Africa (Fujita, 1973). Frequency data for tornadoes are unreliable for most parts of the world. During the 1950s about 200 tornadoes were reported annually in the United States, but by the late 1990s around 1200 were noted each year (American Meteorological Society, 2000a). Most of the increase may be accounted for by more frequent reports of weak tornadoes, probably because of growing public awareness, rather than any meteorological factors. Numbers of tornadoes reported from other parts of the world are substantially less, perhaps related to lack of awareness and sparse weather-observing networks. A study of tornadoes on the Indian subcontinent identified only 51 events between 1835 and 1977, but the path lengths and widths were larger than those characteristic of the United States, so many smaller tornadoes may have passed unreported (Peterson and Mehta, 1981). Elsewhere, for example, 191 tornadoes were reported in Argentina from 1930 to 1979; 42 in Taiwan from 1951 to 1978; 87 in Japan in 5 years from 1968; and 273 in France from 1680 to 1998 (Snow and Wyatt, 1997; Paul, 1999).

Tornadoes exhibit a considerable range of intensity, size, and duration. Typical New Zealand tornadoes have been reported to have damage paths only 10–15 m wide. The typical or median path of tornadoes in the United States has been given as 3.2 km long by just under 50 m wide (National Severe Storm Forecast Center Staff, 1980), but there have also been reports of tornadoes with paths ranging from less than 50 m to over 400 km. Damage associated with tornadoes is very localized, including severe structural damage to buildings (roofs lifted, walls and windows collapsed) and mobile homes, and tops screwed or snapped off trees, which may also be uprooted.

Tornado-generated missiles, ranging from gravel to semi-trailer trucks, present engineers with major design problems. In tornado-prone areas of the United States, where protection of people in buildings is important, the tornado missile can be the controlling design factor. Buildings such as hospitals, fire stations, and emergency operating centers, where critical functions must be maintained, are particularly susceptible to missile damage (McDonald, 1976). Small outdoor equipment and larger objects such as utility poles can become tornado missiles. A tornado at Lubbock, Texas, in 1970 was responsible for moving a cylindrical tank (3.35 × 12.5 m) weighing over 11 tonnes about 1.21 km; and three 40-passenger school buses apparently became airborne in a tornado at McComb, Mississippi, in 1974.

Increasing public awareness, emphasis on tornado preparedness, improved warning systems, better understanding of tornado formation, the use of tornado drills, spotter groups and other measures at local and state levels have contributed to the decreasing death toll from tornadoes in the United States. However, a single tornado in Bangladesh may kill several hundred people because none of these conditions or measures exists, and because of the very dense population.

The broad environmental conditions leading to the development of severe thunderstorms and tornadoes are reasonably well recognized and can be forecast with some skill, but there is still no reliable method for predicting the development of a specific severe storm (Hoium et al., 1997; Snow and Wyatt, 1998). Technological advances such as Doppler radar, high-resolution satellite imagery, and acoustic sounders have improved prospects for forecasting severe local weather. In the USA there have been marked improvements in the remote detection of severe local storms and flash floods, accompanied by better warnings of such events (Polger et al., 1994; Vasiloff, 2001); but in most regions of the world the observing network and the available technology are inadequate for detection and measurement of most small-scale events.

Some local storms occur in areas where warnings, if they are given, will not help very much. A severe dust storm followed by heavy rain, for example, hit Karachi in May 1986 causing extensive property damage and dislocation of services (Middleton and Chaudhary, 1988). Thousands of bamboo huts and improvised houses with tin or asbestos roofs were blown away; telephone and electricity wires were snapped, giant trees uprooted and vehicles overturned; and more than half the city was plunged into darkness. There is no doubt that improved short-term forecasts can greatly diminish the impact of some local storms, in areas where such forecasts are available and can be readily disseminated. Warnings during the May 1999 tornado outbreak in Oklahoma and Kansas are credited with saving hundreds of lives, and Doppler radar provided significant warning lead time during the tornado outbreak in Georgia in February 2000 (Vasiloff, 2001).

Cumulative hazards

Many atmospheric disasters result from an accumulation of events, that singly would not be hazardous. One dry day or even one dry year does not necessarily constitute a drought, but a succession of abnormally dry years can have disastrous effects on the environment and its inhabitants. Similarly, hot days are common in many parts of the world, but a succession of many very hot days can prove lethal, especially in areas not normally accustomed to heat waves combined with high humidities.

Drought

In the period from 1967 to 1991 droughts were estimated to have affected 50% of the 2.8 billion people who suffered from weather-related disasters (Kogan, 1997). It is difficult to find a generally accepted definition of drought. Drought clearly involves a shortage of water, but realistically can be defined only in terms of a particular need (Linsley, 1982). Drought is not just a physical phenomenon; it results from interplay between a natural event and demands placed on water supply by human use systems (American Meteorological Society, 1997). The absence of a precise definition adds to confusion about whether or not a drought exists and, if it does, its severity. The effects of a drought accumulate slowly over long periods and may linger for years after termination of the event. The reporting of drought occurrence may be overestimated or underestimated, because the material well-being of the reporter may be affected (Heathcote, 1979).

Four types of drought are usually recognized: meteorological or climatological, agricultural, hydrological and socioeconomic. Droughts impact both surface and groundwater resources. They can lead to reductions in water supply, diminished water quality, crop failure, reduced power generation, disturbed riparian habitats, suspended or curtailed recreation activities and a variety of other associated economic and social activities (Riebsame et al., 1991; Woodhouse and Overpeck, 1998). Long-term impacts on plant and animal life have received relatively little attention, and many questions are unanswered concerning the role of drought in ecosystems.

The impact of drought on human activities is usually described in terms of reduced water supplies and economic losses throughout the community. The evaluation of such impacts is complicated, with many factors needing to be taken into consideration. Several uncertainties can be involved, such as shortfalls in expected yields tending to inflate the value of actual yields on which the value of lost production is calculated, yet there being no guarantee that lost production could have been sold anyway.

Frequent droughts around the world, and interest in their possible links with phenomena such as El Niño, keep the hazard in evidence even for the casual observer. The 1975–1976 drought in Western Europe had widespread effects on agriculture, domestic and industrial water supplies, and on river and canal traffic. The drought in Britain, particularly southern England, was the worst for about 250 years. In some parts of the country, water supplies to domestic consumers were cut for up to 17 hours per day, and production of root and vegetable crops was down by as much as 40%. Droughts of the early to mid-1980s in Africa affected more than 40 million people. The Canadian Prairie Provinces are particularly sensitive to rainfall shortages and associated soil losses due to wind erosion (Wheaton and Chakravarti, 1990). In 1984 conditions rivaled those of the "Dust Bowl" years of 1936–1937, with farmers losing up to half of their grain crop to the value of about $2.5 billion (Sweeney, 1985). The 1988 drought in the USA was rated as one of the worst in 100 years, with an estimated impact on the economy of $40 billion (Kogan, 1997). An unexpected impact in this instance was on barge traffic on the lower Mississippi river, the industry suffering an income loss of about 20% (Changnon, 1989). The El Niño event of 1997–1998 was linked to drought in Central America and southeast Asia, with major impacts on vegetable quantity and quality in the former and on coffee and palm oil in the latter, both leading to reduced exports and increased prices on global markets (Changnon, 1999). It has been suggested that worldwide disasters triggered by droughts are twice as frequent during year two of ENSO warm events as during other years, particularly in southern Africa and southeast Asia (Dilley and Heyman, 1995).

Several factors may be implicated as potential causes of drought: ENSO, abnormal sea surface temperature patterns in areas other than the equatorial eastern Pacific, soil moisture desiccation, and nonlinear behavior of the climate system (Orville, 1990). It is tempting also to suggest that climate is changing and that droughts are becoming more frequent and/or more severe. However, there have always been droughts, and records show that events such as those mentioned are within the realm of statistical expectations (Landsberg, 1982). Examination of the paleoclimatic record for the Great Plains suggests that the droughts of the 1930s, 1950s and 1980s were eclipsed several times by droughts within the last 2000 years, and that more severe droughts could occur in the future (Woodhouse and Overpeck, 1998).

The massive Australian drought of 1895–1903, which followed rapid growth of rural enterprises in the 1870s and 1880s, and the great American drought of the 1930s, are good examples of lack of understanding of the environment leading to unwise land use. Farmers in southeastern Australia before 1893 believed that the climate was on their side, and in South Australia, the leading wheat producer, there was a belief that rain followed the plow. Notions that plowing and tree planting could bring rain were widespread, enticing farmers into marginal areas. The number of sheep in Australia fell from an estimated 106 million in 1891 to 54 million in 1902, cattle numbers were almost halved to about 7 million, and dust storms were common as a result of the vast expansion of land plowed for wheat. In areas such as the Sahel, where nomadism and intermittent grazing have been prevalent and more or less in balance with environmental conditions, more intensive exploitation has had disastrous results for social systems and ecosystems when drought has struck.

Drought is a common feature in many countries but is often regarded as an unfortunate and irregular abnormality of the environment. It would be more appropriate to consider drought as part of the normal sequence of events. Society must be prepared to cope with the effects of drought at any time. Impacts in the past have been exacerbated by absence of coping mechanisms, with too little preparation during non-drought periods.

Heat and cold

Excessive heat possibly contributes to more illness and mortality than any other direct, weather-related cause, certainly in regions better equipped to cope with the more violent hazards (Kalkstein, 1995). Most heat-related deaths occur in midlatitude cities, in northern India and China, eastern and Midwestern USA and Western Europe, with infrequent but extreme heat waves. Subtropical and tropical cities with higher mean summer temperatures seem less vulnerable, partly due to acclimatization of residents and perhaps due to more efficient cooling of houses. Heat-related deaths, averaging about 1000 per year, appear to be on the increase in the USA, exceeding those caused by other hazardous weather conditions (Changnon et al., 1996). Heat waves may also be associated with increases in the incidence of rioting, violence and homicide.

Heat-related death rates are usually higher in urban areas than in rural areas. This is almost certainly the result of climatic modification and heat retention due to urbanization, the heat island effect, plus the pollution trapping and concentrating effects of stagnant atmospheric conditions of heat waves, adding to those of heat stress. Death rates are also higher among the aged, especially as a result of aggravating effects on pre-existing conditions such as heart disease or cancer.

An exceptionally severe heat wave in parts of the southwestern USA in June 1994 produced the highest temperatures ever recorded in four states (Brugge, 1995). At the same time authorities in southern Ontario advised residents with heart or respiratory problems to rest, and also requested people to avoid driving, and avoid the use of aerosols that could worsen ground-level ozone. A short but intense heat wave in mid-July 1995 caused over 800 deaths in the USA, over 500 in Chicago alone (Kunkel et al., 1996; Karl and Knight, 1997). About 70 daily maximum temperature records were set during this heat wave at locations from the central and northern Great Plains to the Atlantic coast (Livezey and Tinker, 1996).

Many places in England and Wales experienced record high temperatures during a heat wave in August 1990. Apart from the usual impacts upon those unaccustomed to temperatures over 37°C, tar on roads melted, leading to closures, one runway at Heathrow airport was closed when newly laid tar failed to set, the entire stock of a Liverpool chocolate factory melted, a life-sized waxwork knight at a castle in Essex melted into a puddle, and there was a spate of drownings as people tried to keep cool by swimming (Brugge, 1991).

Increased demands for air-conditioning and refrigeration can produce overloading of power supply systems during heat waves, leading to power restrictions and breakdowns, tending to aggravate the heat-stress situation. Assessment of the causes of death in the 1995 Chicago heat wave included factors such as inadequate warning systems, insufficient time to acclimatize, the heat island effect, an aging population, an inadequate ambulance service, and the inability of many residents to properly ventilate homes due to fear of crime or lack of resources for fans or air-conditioning (Changnon et al., 1996). During a subsequent heat wave in Chicago in July 1999 the death toll was much lower, because lessons had been learned from the previous event. Heat wave plans included timely warnings, activation of cooling shelters, frequent broadcasting of information and the ready availability of helplines (Palecki et al., 2001). In any such event, indeed in almost any hazard event, there are winners. In this case air-conditioner sales increased, sales of ice creams and ice set records, private ambulance operators were busier than normal, utilities not experiencing major equipment failures made record profits, people went to shopping malls and movie theaters to escape the heat in air-conditioned facilities, and merchants at lakeside outlets benefited from record attendances at the height of the heat wave. 'Properly spaced green areas are the most effective and aesthetically pleasing means of controlling urban temperature excess by improving ventilation and circulation and reducing heat storage capacity. Building materials with lower heat conductivity and storage properties, and water bodies within or close to cities help to keep maximum temperatures down and encourage mixing and ventilation resulting from enhanced temperature differentials. Adequate surveillance systems are needed, to alert the public and authorities that potentially dangerous weather is imminent. A promising approach is based on the identification of high risk air masses historically associated with increased mortality (Kalkstein, 1995; Kalkstein et al., 1996). A similar approach, using a weather type classification scheme developed for North America, has been used in several heat stress warning systems worldwide. It has been incorporated in systems developed for Rome, Shanghai, Toronto, Phoenix, New Orleans and Cincinnati (Sheridan, 2002).'

Excessive cold directly causes death through the effects of exposure and is indirectly responsible for deaths from causes such as fatal heart attacks while clearing snow and asphyxiation in stranded vehicles. The annual average mortality rate attributable to winter storms and cold in the USA up to the mid-1990s was between 130 and 200 (Changnon et al., 1996). The number of cold-related fatalities may be increasing, perhaps because an aging population becomes more sensitive to the effects of cold. Over 600 people died from cold-related accidents in each of the North American winters of 1977–1978 and 1978–1979, with temperatures averaging 6°C below normal from December 1978 to February 1979.

The impact of a severe winter extends beyond increased fatalities. Production losses in industry, crop losses, transportation losses in revenue and damage to roads and bridges, losses in retail sales, and losses resulting from increased energy consumption in the winter of 1976–1977 in the United States were at least $40 billion. Unemployment rose by 2 million, the consumer price index had its largest 2-month jump in 3 years, the inflation rate rose, the number of business failures increased, and the balance of payments dropped by $1.4 billion (Hughes, 1982). Record low temperatures were registered in January 1994 across central and eastern USA and eastern Canada, down to −31°C at Akron, Ohio, −30°C at Pittsburgh, −36°C in parts of Minnesota, −41°C in New York State and below −45°C in parts of Ontario (Brugge, 1994b). A wind-chill figure of −70°C was reported from the Midwest. Impacts of the severe cold included school closures; interrupted train services; closed airports; closed public services; record power consumption and electricity blackouts; water mains frozen for up to a week; ice-making machines frozen; canceled garbage collections because of frozen equipment; many weather-related injuries, including a man who set himself on fire trying to light a fire; a halt to brewery production in Milwaukee; and a crack in a gasoline pipeline leaking over 450 000 L, much into the Mississippi. To impacts like these should be added elements such as personal inconvenience, worry and stress, extra work, injuries, higher taxes to cover costs of repairs, and decreased tax income through lost work-days.

Frost presents a hazard to crops around the world. Freezing temperatures in southern Brazil, in 1979 and 1985 at around 20°S, damaged coffee trees and severely reduced coffee production. Severe frost, although infrequent, is also a major climatic hazard in the near equatorial latitudes of the highlands of Papua New Guinea (Brown and Powell, 1974). Risk of ground frost in Papua New Guinea starts at about 1 500 m, increasing with elevation and in valleys and basins. The worst frost in memory occurred in the early 1940s, accompanied by a prolonged drought. Gardens as low as 2000 m were wiped out by weeks of frosts, with sleet and snow at higher levels destroying gardens, food-bearing trees, domestic pigs, and wild animals. Food shortages forced extensive migrations to lower areas, leading to severe social conflict and occasional fighting. In March–April 1990 much of the UK was affected by unusually severe ground frosts and prolonged air frosts, leading to widespread damage to tree fruit crops in flower at the time and to young nursery stock (Morison and Butterfield, 1990). There were also reports of substantial damage to winter-sown cereal crops, with from 10% to 90% of ears in barley and wheat crops being killed. The damage was exacerbated because of previously mild conditions throughout the preceding winter and the corresponding advanced development of many of the plants.

There is growing recognition of the potential value of meteorological information in decision making. Deciduous fruit trees are very susceptible to frost damage in spring, with loss of fruit yield and even permanent tree damage. Several protective devices are available, such as wind machines, sprinklers, and heaters, but these are expensive. Accurate minimum temperature forecasts can help the grower in making decisions whether or not to protect on a particular occasion, and so may help to save money (Katz et al., 1982).

Snow and ice

Snow hardly worries an alpine village; indeed it may be welcomed as improving prospects for the tourism industry, but the same depth of snow may bring panic and total disruption to a lowland city unaccustomed to it. The nature of the snow hazard is influenced by other weather conditions, particularly wind, by topography and by road surface materials. On crowded roads even a shallow snow cover of 20 mm may be sufficient to halt or hinder traffic flow, leading to increased costs in work losses, late delivery of goods, losses of consumer sales and receipt losses by public transportation, apart from increasing accident and casualty rates (Perry, 1981). Snow has widespread impacts on construction, merchandising, manufacturing, agriculture, power supply, communications, recreation, and public health and safety services and can induce heavy financial losses either in terms of direct damage and disruption or in costs of mitigation (Rooney, 1967). However, snow can be beneficial if it covers crops or vegetation and protects them from air frost.

There are numerous examples of the impacts of major snow storms on large urban areas (e.g. Speakman, 1994; Brugge, 1944a; Wild et al., 1996). The severe winter storm of March 1993 along the east coast of North America illustrates the possible impacts of snow on communities. Property damage was estimated at $1 billion; snow clearance costs were estimated at $100 million; and total damage was put at about $4 billion. Structural damage was widespread from Florida to Canada; hundreds of roofs collapsed under snow loads; and over 3 million customers were without electrical power due to fallen trees and high winds, which reached 210 km/h on Cape Breton Island. Over 200 people died from weather-related causes and 48 were lost at sea. Many deaths were attributed to heart attacks clearing snow, others died in fires, drowned or suffered carbon monoxide poisoning when trapped in cars or buildings (Brugge, 1994a).

A record early snow storm which hit parts of the High Plains and Midwest of the USA in October 1997 and an extremely severe ice storm in the northeast and in eastern Canada in January 1998 were both attributed to the El Niño event of 1997–1998, as were many other climatic hazards at that time (Changnon, 1999).

Ice storms can cause major economic and social disruption (Assel et al., 1996). Aircraft icing, through the accretion of supercooled water, has been identified as a cause of many aircraft accidents during winter storms (Rasmussen et al., 1992). The 1998 ice storm was remarkable for its spatial extent and duration, as well as in terms of the severity of its impacts on the region (DeGaetano, 2000; Dupigny-Giroux, 2000). Total storm damages in the USA and Canada exceeded $2 billion, over $800 million to the hydroelectric installations in Quebec alone, where well over 1 million people were at times without power. In Montreal the entire commercial sector was closed down for a week. Deaths occurred from falling ice, house fires, hypothermia and carbon monoxide poisoning. Thousands of farmers in the dairy industry were particularly hard hit, with losses of cattle and milk production; and the long-term consequences of milk shortages affected ice-cream makers, cheese suppliers and all relying on dairy products. The livelihoods of maple syrup producers were placed in jeopardy because of extensive ice damage to trees.

Telecommunications and electrical services were tested beyond design limits by ice accumulations, radio and television stations were put off the air, airports were closed and rail services suspended. Damage to hiking and snowmobile trails had impacts on the leisure industry, and it was estimated that most Christmas trees intended for the 1998 season were damaged beyond salvage.

Fog

Fog is a hazard to land, sea, and air transportation. Many multiple-vehicle accidents occur when visibility is severely restricted on highways. In 1974 fog was estimated to have cost over £12 million on roads in the United Kingdom (Perry, 1981). The worst aviation accident on record occurred in March of 1977 in the Canary Islands when two wide-bodied jets collided on a runway in heavy fog, killing 583 people. Possibly the worst fog disaster occurred in May of 1914 when the *Empress of Ireland* sank after a collision in the St Lawrence estuary with the loss of 1078 lives (Whittow, 1980). The collision between the *Stockholm* and *Andrea Doria*, and subsequent sinking of the latter, in sea-fog off Massachusetts in July 1956, is another example of a dramatic impact of what can appear as an innocuous phenomenon.

The aviation industry loses millions of dollars each year when fog causes aircraft diversions and delays, inconveniencing thousands of passengers and incurring increased operating costs. Costly adjustments, such as automatic aids for aircraft, are possible and many fog dispersal techniques have been tried. Supercooled fog can be modified but accounts for only a small proportion of fogs. Warm fog can also be modified using thermal energy, but high costs of installation and operation of dispersal systems and the problem of pollution have discouraged their use (Kunkel, 1980).

Visibility is one of the most difficult of meteorological phenomena to forecast. Several objective fog forecasting techniques have been tried, but there is still no entirely reliable method. Publicity of the hazards and driver education remain the surest ways of reducing the dangers of fog on highways.

Air pollution

Many studies have pointed to an association between air pollution and ill-health, particularly for diseases such as bronchitis and lung cancer, and less so for cardiovascular ailments and nonrespiratory tract cancers (e.g. Lave and Seskin, 1970).

The first noticeable effect of photochemical pollution is a stinging of the eyes due to peroxyacetylnitrate gas. Ozone, also formed in photochemical reactions, damages lung tissue, increases death rates as a result of swellings in lung passages, and reduces athletic performance. Sulfur dioxide can lead to infections of the lower respiratory tract, especially among the elderly, the very young, and those already weakened by illness. Carbon monoxide may cause drowsiness in low concentrations, extending to severe headaches, nausea, and collapse in high concentrations. Large cities such as Beijing, Bombay, Cairo, Jakarta and Mexico City are now the most susceptible to hazards arising from air pollution.

Air pollution impact can be reduced, chiefly by not locating pollution sources in badly ventilated areas known to experience frequent low-level temperature inversions. A common approach to air pollution from motor vehicles is emission control, either through the fuel or through exhaust abstraction systems, although the efficiency of these is debatable and their public acceptance often poor. Design of buildings and layout of city streets can help promote greater turbulence and better diffusion of pollutants. Good traffic flow promotes better urban air quality, but traffic engineering and air pollution control aims are frequently at variance. Implementation of fuel conservation policies, use of alternative fuels, and pollutant-emission regulations are steps toward minimization of total pollution emissions and thus of the pollution hazard. On the other hand, it may be better and cheaper to achieve clean air from the outset through sensible planning rather than through remedial steps later. Transport of air pollution by atmospheric circulation means that the problem is one for legislators at the national or continental level.

Acid deposition from the atmosphere is a major worldwide environmental concern, particularly in parts of northern and western Europe, North America and China (Mason, 1990). It originates from the release of sulfur and nitrogen oxides by industry and transport. Oxidation and hydrolysis of the oxides produce sulfuric and nitric acids or related sulfates and nitrates, which are transported in and eventually removed from the atmosphere in rain, snow, dew, frost, fog, gases, or particles. There is evidence that acid rain may be responsible for long-term adverse effects on the environment. These effects may include acidification of rivers, lakes, and groundwater, with damage to all components of the aquatic ecosystems; acidification and demineralization of soils; changes in agricultural and forest productivity; corrosion damage to buildings, monuments and water supply systems; degradation of water supplies; and reductions in biodiversity, with acid-tolerant species flourishing in severely stressed ecosystems (Bridgman, 1997).

Biomass burning for forest clearing, fossil fuel burning and "slash-and-burn" agricultural practices in the tropics of south-east Asia, the Amazon Basin and central Africa create a wide range of pollutants, including smoke components and dust. Such emissions can be transported long distances downwind. The Asian brown haze, largely composed of particulates and sulfate aerosols from south, southeast and east Asia, significantly reduces solar radiation and has possible impacts including a reduction in rainfall, reduction of light available for photosynthesis and adverse health effects (Anon, 2002). Cooling of the land surface; a stronger and more frequent thermal inversion trapping more pollutants; and an overall reduction in evaporation and precipitation, with implications for water quality and availability, are possible indirect effects of the brown haze.

Climatic changes

The importance of climatic changes to the world's future and of human dependence on a stable climate to maintain present patterns of agriculture has been highlighted by many authors (e.g. Schneider, 1976; Bryson and Murray, 1977; Roberts and Lansford, 1979; Lamb, 1982).

Climatic inputs provide the essential resource base for agriculture; even with present technology the world is not immune to the effects of climatic variation. In 1972, when the global climate was particularly unfavorable for food production, millions starved. The increasing demands of a growing population mean that there may be more frequent imbalances between food supply and need, especially in regions of increasing pressure for cultivation of climatically marginal land. The results could be depletion of grain reserves, malnutrition, starvation, and political unrest. Parry et al. (2001) have examined the possible impacts of climate change in some key areas of risk: hunger, water shortage, exposure to malaria transmission, and coastal flooding. They concluded that millions more people will become at risk and that it will be necessary to find a blend of mitigation, to buy time, and adaptation, which in turn can raise thresholds of tolerance.

There has been much speculation about what might happen to frequencies, intensities and distribution of climatic hazards in the future, although not all changes in risk would necessarily be attributable to climate change; societal shifts such as population growth, demographic moves to more at-risk locations, and the growth of wealth have all made for greater vulnerability (Changnon et al., 2000). Some projections suggest an increase in the risk from droughts, floods, heat waves, tropical cyclones and storms (e.g. Changnon and Changnon, 1992; Bruce, 1994; Yarnal et al., 1999). There is evidence that diseases such as malaria and dengue fever, carried by mosquitoes, are undergoing resurgence and redistribution, possibly in response to global warming (Epstein et al., 1998). Mosquito-borne diseases are being reported at higher elevations and higher latitudes, consistent with a spread of warmer, wetter conditions. The distribution of agricultural pests could also shift, with implications for food security and pointing to the need for the public and policy makers to be aware of the possible biological consequences of climate change (Harrington and Woiwood, 1995).

There is still considerable uncertainty about the nature of climatic trends, much depending upon the rate at which greenhouse gas concentrations increase and on the particular models and scenarios used to predict future climate. It has been suggested, for example, that there is no clear evidence of long-term trends in global tropical cyclone activity, with indications that the broad geographic regions of cyclogenesis and of regions affected by tropical cyclones will not change significantly (Henderson-Sellers et al., 1998). Increasing trends in weather-related losses in recent decades have led to a popular view, nurtured by greater media coverage, that hazard frequency and intensity are increasing, but societal changes may be the primary cause (Kunkel et al., 1999). The lack of high-quality long-term data makes it difficult to determine changes in extremes, and observations of the impact of global warming are based on very short records for analysis (Easterling et al., 2000).

Climatologists can assist in many ways to plan for severe weather, now and in the future (Hunt, 1990): for example, with advice at design and planning stages of projects to reduce possibilities of weather-related disasters; with advice on land-use planning and design of structures to minimize risks from climatic hazards; and with improving forecast and warning capabilities and awareness and coping strategies. Whatever changes may occur in the future, and whatever adjustments they may necessitate, there is a need for greater recognition by the public, policy makers and planners that change is possible and preparation is necessary (e.g. Glantz, 1979; Wilson, 1981; Robinson and Hill, 1987; Changnon et al., 1995).

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Cross-references

Acid Rain
Air Pollution Climatology
Applied Climatology
Bioclimatology
Desertification
Drought
Lightning
Thunderstorms
Tornadoes
Tropical Cyclones

CLIMATE MODELING AND RESEARCH CENTERS

The following provides an overview of the main world centers where climate research and climate modeling are carried out.

Australia

Bureau of Meteorology Research Centre, Melbourne (BMRC) – Department of the Environment and Heritage, Commonwealth of Australia

The BMRC has six research groups focusing on different aspects of climatic and meteorological modeling, including model development, model evaluation, data assimilation, weather and climate forecasting and marine and ocean forecasting. Attention is given to the Australasian region in the development of numerical models, long-range forecasting, tropical meteorology and meteorological observation systems.

The research objectives of the center are generally designed to support the operations of the Bureau of Meteorology as well as to produce research that would be of interest to the scientific community. The center maintains close ties to the World Meteorological Organization (WMO), the Commonwealth Scientific and Industrial Research Organization (CSIRO) and is involved with the Cooperative Research Centre program of the Commonwealth Government.

Commonwealth Scientific and Industrial Research Organization (CSIRO), Melbourne – Atmospheric Research

The CSIRO is the major science research institution in Australia. It has several divisions that produce research in all aspects of the environment, of which the Atmospheric Research Division (ARD) is but one. The ARD operates three laboratories, located in Canberra, Victoria and Aspendale, and is grouped into three major programs: Earth systems modeling program; pollution program; and measurements, processes and remote sensing programs. Most of the research conducted within this organization is applied and covers topics that are specific to Australia's interests. The CSIRO collaborates with the BMRC, as well as participating in such programs as the International Geosphere–Biosphere program and the World Climate Research Program.

Canada

Canadian Centre for Climate Modelling and Analysis (CCCma), Victoria

The CCCma is a division of the Climate Research Branch of the Meteorological Services of Canada and is located on the campus of the University of Victoria. This center specializes in modeling coupled sea-ice and atmospheric modeling, climate variability, climate predictability and the carbon cycle, among other areas. They developed the Atmospheric General Circulation Model (AGCM2), which models equilibrium climate change simulations using variations of sulfur aerosols. This model has been used in paleoclimate studies, passer tracer studies, and predictability studies. The model has also been used to produce operational seasonal forecasts for the Canadian Meteorological Centre and by other collaborators. A newer model (AGCM3) has been developed and became operational as of January 2003.

The CCCma is involved with many international projects such as the Program for Climate Model Diagnosis and Intercomparison Project (PCMDI), hosted by the Lawrence Livermore Laboratory in the United States. CCCma has contributed to the Intergovernmental Panel on Climate Change (IPCC) reports of 1990, 1995 and 2001 and presently participates in the Climate Variability Predictability Program (CLIVAR) and the Scholarly Publishing and Academic Resources Coalition (SPARC).

France

Institut Pierre-Simon Laplace des sciences de l'environment (IPSL) (Campus du Jussieu)

The IPSL is composed of several laboratories that model different aspects of the environment and have working relationships with other institutions within France, throughout Europe and the rest of the world. Under the auspices and collaborative efforts of all of the laboratories in their institute, they have developed their own ocean-climate models (e.g. the IPSL model, the LMD and LMDZ, OPA-ICE). All of the IPSL laboratories are associated with the National Center for Research in the Sciences (CRNS) of France. Below is a survey of three laboratories related directly to climate modeling. All of them are affiliated with the IPSL.

1. Laboratoire de Meteorologie Dynamicque du CNRS, Paris. The Laboratoire de Meteorologie Dynamicque du CNRS (LMD) is part of the IPSL. It encompasses three sites: l'École Polytechnique in Palaiseau, l'École Normale Supérieure, in Paris and l'Université Pierre et Marie Curie, also in Paris. The LMD has four areas of research: climate modeling, remote sensing, experimental data measurements, and theoretical activities. The research is divided into several different teams located at the three university sites. They have developed the LMD and the LMDZ models of the general circulation of the atmosphere and presently collaborate with numerous laboratories, research centers and universities to develop linkages to study a wide variety of climate issues.
2. Laboratoire d'Optique Atmospherique (LOA) Lille. This laboratory, housed at the University of Lille, is involved with modeling the radiative properties of the atmosphere, especially as relating to its optical properties and its role in the global energy budget. The LOA focuses on the interaction of solar and terrestrial radiation with the atmosphere in the context of the global climate and climate change. It initially began as a research project on solar ultraviolet irradiance and radiative transfer theory, then shortly thereafter concentrated on air–sea interactions and on the remote sensing of the oceans. In the 1970s the laboratory focused its research on the radiation budgets of Venus and Mars. In the 1980s it applied its modeling efforts to the radiative forcing of the Earth–climate system, particularly to the development of radiative models involving the oceans, cloud cover and other aerosols, and the greenhouse gases.
3. Laboratoire d' Océanographie Dynamique et de Climatologie (LODYC). Housed in the University of Pierre et Marie Curie, the LODYC works to foster understanding of the Earth–climate system and the oceans' contribution to it. They have developed the OPA-ICE model that simulates oceanic circulation that can be coupled with other models (like the LMDZ) through the research at IPSL. They are also interested in modeling marine ecosystems in order to understand the role of the oceans in the exchange of carbon between the ocean and the atmosphere.

European Center for Research and Advanced Training in Scientific Computation (CERFACS), Toulouse, France

CERFACS has a set of research teams devoted to various aspects of scientific modeling, including climate modeling and global change, and modeling in fluid dynamics. They developed the OASIS model that is used as the coupler between general circulation models of the atmosphere and ocean circulation models in the research done by the laboratories of IPSL.

Germany

Alfred Wegener Institute for Polar and Marine Research (AWI – Bremerhaven, Germany)

The AWI is an institute devoted to modeling the Arctic and Antarctic environments and for using and developing general circulation models for polar and marine research. They also devote some attention to modeling for research in middle-latitude regions. The AWI has four main units, one of which is devoted to the climate system. In the climate unit, they have sections focusing on research in large scale and regional circulations, and on the physical and chemical processes of the atmosphere.

Deutsches Klimareschenzentrum (DKRZ) – German Climate Research Center, Hamburg

The DKRZ is a limited corporation funded in part by the Federal Ministry for Research and Technology, and is a service center for those participating in climate research in Germany. It was founded in 1987 at the University of Hamburg and is responsible for maintaining the infrastructure necessary for basic and applied research in climatology. It also cooperates with the European Climate Computer Network to allow for climate research to be pursued in both the larger and smaller nations of Europe.

Max Planck Institute for Meteorology (MPI-Met), Hamburg

Located in the university district of Hamburg, the MPI-Met contributes research into climate model development, investigates new methodologies of observation and measurement of the Earth–climate system and examines human–climate interactions that affect policy-making decisions. The institute is composed of three scientific divisions: biogeochemical systems, climate processes and physical systems, and one research group which itself has two scientific divisions: a regional climate group and a socioeconomic modeling group. The scientific divisions' activities focus on model formulation and application and are at present developing procedures to couple the models produced by the individual scientific departments and research groups with the goal of establishing a comprehensive model of the Earth–climate system.

United Kingdom

UK Universities Global Atmospheric Modelling Programme (UGAMP)

UGAMP is a network of university research centers that are primarily focusing on the development of large-scale numerical global climate models. The Natural Environmental Research Council (NERC), who fund the program, initiated UGAMP in 1987. There are nine affiliated university sites including London, Cambridge, Reading, Southampton, Rutherford (Appleton Lab), Oxford, East Anglia, Leicester and Edinburgh. Two universities in particular house the major climate and modeling centers while the other universities contribute to their work. These two centers are: the Center for Global and Atmospheric Modelling (CGAM) located at the University of Reading and the Atmospheric Chemistry Modelling Support Unit (ACMSU) located at the University of Cambridge. The CGAM is the main group that supports the UGAMP network of research sites. It emphasizes model development for large-scale atmospheric circulations and has partnerships with modeling centers and activities in the rest of Europe, the United States, and Japan, as well as with their colleagues in the United Kingdom. They are responsible for maintenance and diagnostics of the General Circulation Models formulated by the UGAMP community. ACMSU essentially works to provide the chemical data and codes that the rest of the UGAMP community will need in model development.

Hadley Center for Climate Prediction and Research (HC) – London

The Hadley Center for Climate Prediction and Research is the climate research arm of the UK Meteorology Office. It uses a variety of models including GCM, coupled ocean–atmosphere models, regional climate models and atmospheric chemistry models. The research is divided into teams focusing on the following themes: model development, model paramaterization, atmospheric chemistry and ecosystems, global and regional climate prediction, quantifying model uncertainties, rapid climate change, environmental stress, extreme events, climate monitoring, database development and interannual and interdecadal forecasting.

United States

Geophysical Fluid Dynamics Laboratory (GFDL), Princeton, New Jersey

The GFDL is a modeling center that is funded through the US Department of Commerce and the National Oceanic and Atmospheric Administration (NOAA). It has six research groups in it: Climate Dynamics, Hurricane Dynamics, Atmospheric Processes, Mesoscale Dynamics, Climate Diagnostics and Oceanic Circulation. There is also a technical support division that maintains the computing environment for the laboratory. They also maintain the Atmospheric and Oceanic Sciences Program (AOSP), a collaborative effort with Princeton University and funded by NOAA. This center is interested in developing atmospheric models at small and regional scales, as well as large-scale global circulation models. They have produced several models, notably the GFDL R15 and GFDL R30, which are coupled general circulation models of different spatial resolutions and the Modular Ocean Model, the most widely used global oceanic circulation model.

Lawrence Livermore Laboratory (LLL), Livermore, California

The LLL, operated under the US Department of Energy, includes the Program for Climate Model Diagnostic and Intercomparison Project (PCMDI) in which various atmospheric (Atmospheric Model Intercomparison Projects, AMIP), coupled (Coupled Model Intercomparison Projects (CMIP) and Paleoclimate models (PMIP)) are analyzed and compared in order to improve model development and reliability. These projects facilitate a community-based protocol that allows comparison and diagnostics of these models to be made. Models formulated by researchers in research and modeling centers throughout the world participate.

MIT Center for Global Change Science Climate Modeling Initiative (CMI), Cambridge

The CMI, developed as an outgrowth of the existing MIT Center for Global Change Science to concentrate efforts in climate modeling in order to contribute to understanding the larger context of global change science. The main attention initially was to improve climate prediction by ascertaining the limits of predictability. Funding has come from both private and public sources.

One of the objectives of the CMI is to use a hierarchy of models to simulate different portions of the Earth–climate system so that certain specific research goals can be met. They are also interested in addressing the chaotic behavior of the climate system in order to reduce the errors of prediction and to minimize errors in setting initial conditions for the models. Currently the CMI is developing a new coupled-atmosphere–ocean general circulation model that will incorporate and enhance an existing model developed by the Max Planck Institute (MPI-Met) with their collaboration.

NASA Goddard Institute for Space Studies (GISS), New York

GISS is a research institute designed to research a number of disciplines in Earth and space sciences. The institute works closely with Columbia University's Earth Institute and Lamont–Doherty Earth Observatory. Many GISS scientists are also closely involved with Columbia's Center for Climate Systems Research. The primary objective of GISS in the atmospheric sciences is to produce research leading to the understanding of global climate change. The research tends to revolve around the following themes: climate modeling development, climate forcings and impacts, Earth observations, paleoclimatology, radiation studies, atmospheric chemistry and planetary atmospheres.

Most of Goddard's research in climate modeling involves the development and use of Global Circulation Models and coupled GCM and ocean models. Some research is also conducted using energy balance and radiative transfer models. The research primarily focuses on using models geared toward understanding climate sensitivities to a variety of potential perturbations, such as greenhouse gas emissions, aerosols and solar variability. The research is also applied to understanding the possible impacts to society that climate changes associated with these possibilities may produce.

National Center for Atmospheric Research (NCAR): Climate and Global Dynamics Division (CGD), Boulder, Colorado

NCAR is a center devoted specifically to all areas of the atmospheric sciences. It is directed under the auspices of the University Corporation for Atmospheric Research (UCAR) and funded through the National Science Foundation. The major climate modeling division of NCAR is the Climate and Global Dynamics Division (CGD). This division is responsible for developing global circulation models of varying resolutions and scope. The Climate System Model (CSM), which consists of coupling models of the atmosphere, oceans, the land surface and sea-ice without the benefit of artificially providing flux adjustments, is the primary model that is being currently used. The CGD division of NCAR works closely with other divisions in NCAR, with the university research community and with other national and international laboratories to develop simulations using the CSM for use in other projects.

Other modeling and research institutions

There are many modeling centers and research institutions that are smaller in size, narrower in scope, or are part of larger organizations who do not have, as their primary function, climate or atmospheric research. The following is a list of laboratories and institutes and their locations that have climate modeling and research teams or divisions. Their contributions to model development and diagnosis cannot be underestimated.

Ames Research Center (NASA), Moffett Field, California
Argonne National Laboratory (Energy), Chicago, Illinois
Brookhaven National Laboratory (Energy), Upton, New York
The Jet Propulsion Laboratory (NASA), Pasadena, California
The Pacific Northwest National Laboratory, Richland, Washington
The Scripps Institute for Oceanography, San Diego, California
Woods Hole Oceanographic Institute, Woods Hole, Massachusetts

Robert Mark Simpson

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National Center for Atmospheric Research, Tim Kileen, Director, 2003. http://www.ncar.ucar.edu/ncar/about.html
National Aeronautics and Space Administration, Goddard Institute for Space Studies, James Hansen, Director, 2003. http://www.giss.nasa. gov/about/ and http://www.giss.nasa.gov/research/modeling/
Program for Climate Model Diagnostics and Intercomparison, Lawrence Livermore Laboratory, 2003. http://www-pcmdi.llnl.gov/
Universities Global Atmospheric Modelling Programme, National Environmental Resarch Center, 2003. http://ugamp.nerc.ac.uk/
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Cross-reference

Climate Data Centers

CLIMATE VARIATION: HISTORICAL

The basis of history is documentation, and given that database we can attempt to draw some conclusions; one might even try to use this historical record, in combination with the instrumental material of the last century or two, to anticipate the future. Experimental science should, ideally, provide the basis for prediction. In closed systems, like the chemists' test tube, prediction can be made with some degree of certainty. But in the open systems of the planet Earth–its atmosphere, ocean, magnetic fields, moon, sun and other planets–all these bodies and systems are interacting.

The trouble with the meteorological records is that they are too short to capture the "big picture." Instrumental data provide a numerical, quantitative basis for working out the dynamic equations of atmospheric motion. They also offer a quantitative way of comparing day-to-day variables, the contrasts between winter and summer: a definite improvement on our intuitive, sensory, and memory-based impressions of hot and cold, moist and dry. In combination with computer-based storage facilities and sophisticated methods of statistical analysis, some hitherto unknown, or unproven, features are now beginning to emerge from the meteorological materials, such as the 14-month Chandler wobble pulse and the 18.6-year lunar nodal cycle. But what of the longer periodicities and peculiarities of the climatic machine? The agreed base for a standardized climatic average for any observing station, by convention of the World Meteorological Organization, is only 30 years. The climatic variation of the geologic record should indicate clearly that the day–month–year basis of instrumentally based climatology only scratches the surface. But the geological picture has a tendency to drift (sublimely) away from the human scale of things. This is where the historical record helps to fill in an essential gap. It takes us from the instrumental field of "hard numbers" to ancient diaries, tax records, narratives, and eventually to the geological data sources, where the quantitative values must be calculated, albeit by highly sophisticated analytic techniques, not least the measurement of the time factors.

Definition

Historical climatology is seen in a variable context by different workers. (A sampling may be seen in several collections: Rotberg and Rabb, 1981; Wigley et al., 1981; Mörner and Karlén, 1984; Roberts, 1989, 1998; Burroughs, 1992; Shindell et al., 2001). Let us define historical climatology as the study of climate through the time-range of civilized *Homo sapiens*, during the period in which humans have developed the arts of writing and the construction of permanent dwellings and other structures relating to their maintenance and culture. Harding (1982) has provided a useful volume on climate in a context of Holocene archeology.

That time-range varies from region to region. In Britain it is about 3500 years (Stonehenge), but in Egypt it is more like 6000 years. The world's oldest continually inhabited town is said to be Jericho, in the Jordan Valley, which was established 10 000 years ago, or 8000 BC. This date, roughly is also the conventionally adopted date for the Pleistocene–Holocene boundary, agreed by a commission of INQUA–the International Union for Quaternary Research (Olausson, 1982; Fairbridge, 1983). Thus, the 10 000 years of the Holocene Epoch become the logical and ultimate time frame of historical climatology.

Pfister (1980) recognized two categories of data that are appropriately integrated to construct a historical record of climate for any region and interval of time within the designated range: (a) *documentary proxy data* and (b) *field data*.

(a) The use of documentary proxies has been made a special, life-long study by Hubert Lamb (1972, 1977), who integrated them closely with the instrumentally based meteorological standards. The documents provide varied types of information and are of varied quality. They range from diaries, which usually have the virtue of homogeneity (a single observer, living in a given place, and dated with reasonable accuracy), to records of particular events (often found in monastic documents, such things as eclipses, comets, auroras, floods, droughts, killing frosts), and eventually to folklore or sagas (weak in chronology but often helpful if used in conjunction with field data). The diaries may contain actual weather observations (temperatures, wind, precipitation), or they may generate proxy indicators (such as crop planting, harvest times, with yields and costs). Tax records are very illuminating, when interpreted by a skilled specialist (e.g. Pettersson, 1912, 1914, 1930).

(b) Field data are mostly geological, botanical, or glaciological, where the chronology is established on an incremental basis, usually year by year or perhaps century by century. The time-series may go backward from the instrumental era or it may be "hanging" chronology, intended to be integrated "some day." The source material is generally stratigraphically layered (following the *Law of Superposition*), as in the seasonally varved bedding of a lake, or ice layers in a glacier, or the speleological accumulations of a limestone cave. It may be concentrically banded as in the inorganic growth rings of stalactites and stalagmites or in tree rings (*dendrochronology*). Event indicators include such things as volcanic ash layers (tephra, hence *tephrochronology*, frost or burn rings in dendrochronology, and coarse sediments in thick varves, indicating major flood years) or thickness increase in the dark layers of varves (thus higher organic productivity). Seismic events (earthquakes) may be recorded by rock falls in cave deposits (archeological data often record such rude interruptions to the human occupancy), or by "neotectonic" faults or slumps in stratigraphic sequences.

World-class standard time-series

In this entry are listed selected examples of world-class historical standards within the Holocene Epoch that deserve maximum attention from climatologists. If twentieth-century climatologists ignore these mutually interlocking records, they do so at their own risk. History does not repeat itself, and *time's arrow* flies only in one way, but those who ignore history are condemned to repeat its errors. The mutual interlocking of chronologies has, up till recently, been a matter of hypothesis. But now the sheer volume of data is enough to prove that certain events and trends are of global significance, and cannot be dismissed with a shrug or sneering aspersions as to accuracy of dating or whatever. The error bars are often large, but the ever-narrowing ranges of those error bars now permit conclusions based on the convergence-of-evidence line of reasoning (Windelius and Carlborg, 1995).

As to accuracy of chronology, one may point to the vast accumulation of tree-ring data that is now being extended to the whole world (Stockton et al., 1985). Rings are counted year by year, and some sequences reach back to 13 000 BP and more. Errors can be made but, with multiple samples and multiple workers, the margin of error is gradually being refined to the ±1-year level or better. As will be noted below, de Geer's counting of annual varves, which started a century ago, was strongly attacked during his lifetime, but now a Swedish team of devoted workers has repeated his more than 10 000-year record. Small refinements and corrections have been made, to be sure, but the basic chronology is correct (Cato, 1985). More recently, ice cores have furnished a third system of year-by-year counting (Dansgaard et al., 1971). Again errors can occur, ablation can skim off a layer or two, but there are over 100 ice cores over 100 m deep from both hemispheres, which cannot all be wrong.

All three proxy calendars are recorded in *sidereal years*, documenting the seasonal regularity of the Earth's revolution about the sun. The invaluable radiocarbon-dating method, for all its built-in potential for errors, is now calibrated to the sidereal record, and so can be used for bringing "floating" chronologies into the firm records (Klein et al., 1982; Stuiver, 1982). With time measured in sidereal years we can now use astronomical parameters to retrodict planetary–solar–lunar events (Fairbridge, 1984a).

Ten world-class categories of examples of standard time-series are now discussed: ice cores, dendrochronology, lake varves, palynologic sequences, fluvial sediments, beach ridges, marine varves, the volcanic signal, the paleomagnetic signal, and the documentary archives.

Standard time-series

1. Arctic and Antarctic ice cores

One of the most promising fields of research in the scope of historical proxies is the drilling and study of glacial ice cores. The drilling reached 2164 m at Byrd Station (Antarctica) and 2035 m at Dye 3 (Greenland) and dates events back to more than 100 000 (Barry, 1985). The first analyses generated an invaluable ^{18}O–^{16}O isotope curve that is believed to reflect the temperature (and salinity) of the evaporated sea-surface waters (Dansgaard et al., 1971; Lorius et al., 1979). As remarked by Hecht (1985, p. 20), the reproduction of the "Dansgaard profile" in Flint's textbook (1971), "became a standard climate series against which other climatic records could be compared."

An examination of cores both in the Arctic and Antarctic showed that particulate contamination was very much higher during the glacial phases than during interglacials. At first it led to the idea of volcanic forcing of the major glaciations (Bray, 1976). A closer look at the particles, however, showed that they were dominated by desert dust, reflecting the worldwide aridity and greater meridional upper air flow of the glacial phases.

Volcanic activity, however, is recorded in the ice cores (Hammer et al., 1980). The invention of a rapid pH-measuring device opened the way to identifying specific annual layers of high acidity, reflecting volcanogenic SO_2 and CO_2 injections into the upper atmosphere. Precision counting of the ice layers, year by year for the last few millennia, has been rewarded by the discovery that within 12 months of almost all the major historic eruptions there is an acidic *spike* in the cores. Its intensity is presumed to be more or less proportional to the gas production in the eruption. Some eruptions generate only a little gas. Numbers of such spikes already have been found that lack known volcanic events, which only goes to show how little is really known of the Holocene sequences in many parts of the globe (See the AD 535 eruption in the Sunda Strait: Keys, 1999).

Acidity and particulate pollution are not the only variables that can be measured in the ice cores, in addition to the basic ^{18}O signal. Chloride, nitrate, and sulfate have been analyzed by Herron and Herron (1983) from Dye 3. The isotope beryllium-10 (^{10}Be), which is comparable to radiocarbon insofar as it is generated in the upper atmosphere by cosmic ray collisions, it may well serve to measure, inversely, the solar-activity modulation of the geomagnetic field.

Within the ^{18}O ice-core records, there is a prominent cyclicity, on the order of 90, 180, and 360 years. Unfortunately the sampling accuracy is not yet sufficiently refined to give a precise year-by-year variance. Nevertheless, these periodicities are dramatically close to those recorded in the beachridge studies (see subsection 6 below), as well as in sunspots and in known planetary cycles (Fairbridge, 1984a).

Precise attention to the microparticle deposition over the last millennium at the South Pole has shown that accumulation was maximal from AD 1450 to 1860 during the Little Ice Age (Mosley-Thompson and Thompson, 1982). It also provided a basis for calculating annual increments of snow. The Little Ice Age also was marked by the minimal rate of snow accumulation (1657–1686), whereas the warm phases of 1057–1086 (Little Climatic Optimum) and 1867–1896 were marked by maximal snow accumulation. On Greenland the pronounced warming of the Little Climatic Optimum (or Medieval Warm Period) made colonization feasible for the Viking voyagers, and the demonstrated cooling of the fifteenth century brought those colonies to their tragic end (Dansgaard et al., 1975). Even their Icelandic settlements nearly collapsed (see Ogilvie in Wigley et al., 1981).

A newly developed infrared laser spectroscopic technique has made it possible to analyze tiny gas bubbles in the ice and their CO_2 values (Neftel et al., 1982). This has led to the very important discovery (in the light of the present fossil-fuels controversy), that during the last glacial the CO_2 level in the atmosphere was 210 ppm, as compared to the modern 340 ppm. Several specialists regard CO_2 fluctuation as a biological consequence of climate change, not a cause (e.g. Newell and Hsiung, 1984); because of increased upwelling and phytoplanktonic productivity, larger quantities of CO_2 are withdrawn from the atmosphere (and hydrosphere) during cold epochs.

2. The California dendrochronology

Counting tree rings has been carried from a fine art to a sophisticated science in the US Southwest, and is now being extended to the world (Stockton et al., 1985). The crowning achievement has been the radiocarbon analyses of the dendrochronologically dated Bristlecone Pine series from the Sierra Nevada by Hans Suess and a large team of associates. Over eight millennia of records now document the fluctuations of the ^{14}C flux rate through the entire period. This provides an inverse quantification of solar emissions (see Sunspots), as discussed by Damon et al. (1978) and Eddy (1977). The actual ring width (and density), as an indicator of favorable growing climate, correlates precisely with the radiocarbon value in selected areas (Sonett and Suess, 1984). Refined studies have been accorded the rings of the last 2000 years (Stuiver, 1982), and later about 10 000 years (Stuiver and Braziunas, 1993) which permits correlation with the long-term climate proxies of northwest Europe. For a long time tree-ring counts in northwest Europe were in hanging chronologies, that is to say, not interconnected, but the whole sequence has now been unified (Hughes et al., 1982).

The main feature about tree-ring analyses for climatic interpretation (*dendroclimatology*) is that although one particular tree provides only a record of its own locality a regional picture can be constructed when a considerable number of samples are analyzed and integrated. Excessively large sample areas, however, are likely to smooth out the interesting signals, so that large numbers of individual tree-growth areas should be treated separately (Currie, 1984).

Tree-ring analysts are now extending their work to all continents. All latitudes, however, cannot be treated, because certain climatic belts do not furnish distinctive annual rings, as in the year-round wet regimes of the equatorial regions. In some settings the tree grows its thickest rings in the wettest years, but in others it grows them in the warmest years. In some low-latitude mountain areas the vertical zonation (and thus the meaning of the signal) will change dramatically several times in the course of the Holocene Epoch. Nevertheless, the dendroclimatology of the Holocene offers the most universally useful and potentially productive source of long-term climatic data and deserves the most energetic investigations.

3. The Swedish varve chronology (and other varves)

Nearly a century ago the Baron Gerard de Geer (1858–1943) began measuring the thickness of the annual sediment layers in Swedish clay pits (used for the brick industry). The clays were laid down in late glacial and postglacial lakes in distinctive couplet layers (varves, from the Swedish name), a dark band representing winter freezeover and a light band the summer meltwater input. The thickness of the light material is a measure of the melt season in the watershed area, and of the dark layer the organic productivity, and thus a good climatic proxy. Some layers contain volcanic ash impurities from eruptions in Iceland or elsewhere and can thus be used for teleconnections.

De Geer completed an integrated count of thousands of lake deposits, and thus provided a chronology for the entire Holocene, a time scale that for a long time represented the world's only absolute (year-by-year) chronology. In northwestern Europe all other stratigraphic systems, from pollen analysis to archeology, to the stratigraphy of the Baltic, were dated by de Geer's method. His methodology was strongly criticized during his lifetime, but subsequently the Swedish Geological Survey has resurveyed the entire sequence, essentially confirming the old standard, but introducing small corrections here and there (Cato, 1985). It was de Geer's chronology that made it possible to date the sea-level fluctuations represented by the raised beach (strandline) deposits of southern Sweden. Those same sea-level changes were then identified in the South Pacific and dated by de Geer's chronology; subsequently, shells from those beaches were radiocarbon-dated and (after isotopic correction) found to be identical with Swedish standard (Fairbridge, 1961, 1981).

Varved lake clays also are found in Finland, Russia (Lake Saki), Iran (Lake Van), Patagonia, Canada, and the United States. Teleconnections between such far-removed areas are fraught with difficulties, but provided that an approximate correlation can be demonstrated, perhaps within a 100-year error range, then a "vernier tuning" can be applied, as demonstrated by Schove (1978, 1983). The method is based on the twentieth-century observation that the well-known 26-month QBO or quasi-biennial climate cycle sometimes breaks down, especially during sequences of weak solar cycles, to become a quasi-triennial cycle. The length and timing of such breakdowns are globally synchronous so that when they are isolated in hanging chronologies (tree rings, varves, ice cores, or whatever), the century error can be reduced to 1 or 2 years. With the contemporary interest shown by meteorologists and oceanographers in the QBO and ENSO (El Niño–Southern Oscillation), this is an attractive way of extending their statistical data base to 10 000 or more years by using the historical proxies.

Varved marine layers (see subsection 7 below) are much less well known than the fresh-water accumulations, because they are mostly in deep, isolated basins, requiring very costly core-sampling expeditions.

4. Palynology of peat and lake sequences

The science of palynology, usually treated as a branch of botany, treats with the collection, sampling, identification, correlation, and chronology of pollen and spores. Stratigraphic treatment of cored sections provides a continuous paleoecologic record for a specific region that can be compared with actual data (natural contemporary floras and their growth patterns). This comparison permits standardization of results in terms of climatic parameters (Birks and Birks, 1980).

Pioneer studies of this sort were carried out in the nineteenth-century by two Scandinavian botanists, A. Blytt (a Norwegian) and R. Sernander (a Swede), who subdivided the 10 000 years of Holocene time into a series of biozones: *Preboreal, Boreal, Atlantic, Subboreal,* and *Subatlantic*. For convenience they are often identified by the Danish system using Roman numerals: respectively, IV through IX (zones I–III are late glacial Pleistocene).

Mangerud et al. (1974) proposed that, because the local Scandinavian terms of Blytt and Sernander were so useful in designating paleoclimatically determined divisions of Holocene time elsewhere, they should be employed as global stratigraphic time labels, as *chronozones*, and they have been formally defined in terms of radiocarbon years.

Probably the most significant climatic event in the Holocene (Nilsson, 1983), terminating the warmer *Hypsithermal* or

Climatic Optimum phase (Atlantic chronozone, with its mixed-climax forests), was first identified from the palynological data as the Ulmus (Elm) Decline about 5000 ^{14}C-years BP (5800 sidereal years BP. All across northwest Europe this dramatic decrease in the elm forests has been attributed by some botanists either to Neolithic farming or the disease (like the one affecting elms in the twentieth-century). However, the same decline has now been traced by Soviet palynologists all across Asia as well (in Velichko, 1984) and evidence of a simultaneous climatic deterioration and neoglacial advances (Denton and Karlén, 1973) in Arctic Canada, Greenland, and southern hemisphere lands combine to show that it was a global phenomenon. In the temperate belt of eastern North America, according to Davis (1983), the time boundary 5000 ^{14}C-years BP. (5730 Cal. Yr) is marked by a widespread hemlock *(Tsuga)* decline, with a concurrent rise of hickory *(Carya)*. In the western United States, in the central Rockies, from 5000 ^{14}C-years BP on, the tree line became distinctly lower, and is further evidence of the cooling climate. These examples are simply citations documenting a global event of climatic nature. A vast literature has already built up, documenting this one and many others during the Holocene.

A remarkable site is a peat swamp at Ozegahara moor, Japan, where a 4–5 m core has been obtained covering a continuous time-series for 7600 sidereal years (Sakaguchi, 1982). The closely sampled and sharp fluctuations offer a unique record that correlates well with the better-known history of Europe (Roberts, 1998).

Of importance for climatology has been the rising level of confidence in recent decades, that shows an appreciation of the reality of these fluctuations documented by the palynological stratigraphy. Instead of regarding them as local events, experimental error, or whatever, the chronological precision is becoming steadily more refined, and from the repeatability of the pollen diagrams it is becoming quite clear that we are looking at finely tuned paleoclimate records. Care must be taken, especially in the more densely populated areas, to avoid anthropogenic influences, but such problems hardly arise in such remote and sparsely inhabited areas as Finnish Lapland, eastern Siberia, northern Canada, and Patagonia. Yet in all those regions the strongly fluctuating palynologic records proclaim the varied climatic history of the last 10 000 or so years.

5. Nile floods

A unique data source is provided by the documentary record of the Nile floods, which have two peaks each year: the high summer peak reflects the northern hemisphere summer monsoonal rains in Ethiopia; the low, winter peak reflects the equatorial rains in East Africa and the overflow of Lake Victoria. Inasmuch as Egypt's agricultural wealth has always been gauged by the height of the annual Nile floods, and thus for tax reasons, the records have been carefully kept since the foundation of Islam (AD 622), with some gaps caused by social upheavals and invasion. Nevertheless, the record is enormously helpful, providing not only a tropical climatic standard, but also some global signals (Fairbridge, 1984b; Currie, 1995). Historical events such as persistent droughts have been correlated with it all across North Africa from Senegal to Ethiopia. One remarkable example was experienced during a succession of very low Nile floods in Egypt from 827 to 848, but in 828–832 the Nile is reported to have frozen over (presumably in the delta); this correlation of dry cycles with low temperatures is frequently noted in the subtropics. In some epochs the dry cycles match sunspot minimums, but in others there is a phase shift to match the sunspot maximums.

The lowest Nile flood of the twentieth century (the year 1913) corresponded to the highest lunar declination in a millennium; similar low extremes were reported all across Africa. Particularly interesting is the fact that in some centuries there is a strong correlation with the lunar nodal cycle (18.6 years), but in others the solar (11 years) signals appear. This question has been discussed by Currie (1984, 1995). In the Senegal River flood–drought sequence Faure and Gac (1981) found a 30-year periodicity that could represent a beat frequency of these two, or perhaps the 31-year perigee–syzygy cycle of the moon.

Prior to the Islamic era the Nile records are more scattered, but pharaonic data are quite informative (Bell, 1975), especially when taken in conjunction with stratigraphic material. The latter can be dated by an integration of several methods: radiocarbondating of freshwater shells (mollusca) and charcoal (from fireplaces) and by association with dated archeological cultures or with established structures such as temples.

Evidently the pharaohs, for all their interest in both taxes and immortality, had little idea of how much variance the annual flood could offer. Some of the temple walls show little white bands (of calcium carbonate) marking flood levels up to 3 m *above* the flood plain on which they were built. At Semna (above the Second Cataract) the writer measured the present maximum flood height at 8 m *below* a high watermark inscribed about 1800 BC during the reign of Amenemhat III. Such fluctuations correlate with the ups and downs of world sea level (Fairbridge, 1962), itself a proxy for sea-surface temperature (SST) and for glacier advance and retreat (glacial eustasy).

The Nile deposits (with their radiocarbon and historical data series) provided the first quantitative demonstration of the falsity of a long-accepted climatic premise: that chronologically the glacial epochs equaled pluvials (wet or snowy cycles), and interglacials equaled interpluvials (i.e. dry periods). The delusion goes back in the past century to biblical scholars who observed the wadi gravels of the Dead Sea region and elsewhere, associating them with the great rains of the Noachian flood. High lake levels in East Africa and high terraces on the Nile were easily drawn into the same net. On logical, climatological grounds, objections were raised in 1913 and 1928 by Albrecht Penck (citations and discussion in Fairbridge, 1962). A cold glacial-phase ocean did not lead to more evaporation, but less. Besides, it was eustatically smaller, and large areas were covered by sea ice, a 13% reduction in effective sea surface. Net evaporation may have dropped 40%. East African lakes were in fact low during glacial phases. The middle Nile valley became a desiccated inland delta because of the reduced discharge, and the high terraces seen today are relics of hyperarid phases when discharge failed to reach the Mediterranean; indeed, the middle valley filled up to 40 m with silt. Glacials were *not* pluvials; there was ice-age aridity in Africa (Fairbridge, 1962, 1976b).

A similar inland delta has been discovered on the Okawango of northern Botswana and another on the Niger, in Mali, upstream from Timbuktou. In the Sudan and Egypt the aridity ended with the overflow of Lake Victoria into the Nile around 13 000 BP, and this can be matched by similar pluvial activity all across North Africa to India and northern Australia. The monsoon, which had been brought to an end during the glacial phase by the year-round high-albedo and high-pressure system established over Central Asia by the growth of ice caps, began to reassert itself around 13 000 BP. Heavy seasonal rains, a real pluvial, set in as part of the usual monsoonal pattern. Oxygen-18 isotopic studies and planktonic indicators in the surroundings of the Arabian sea confirm the story from the oceanographic side (Prell, 1984).

During the Holocene, from time to time, throughout the entire tropical region, there also were strong fluctuations of the climatic regime (Fairbridge, 1976b), just as is indicated by the Nile record. They are particularly interesting from the human point of view, because this same belt also happens to include the Cradle of Civilization. Climatic forcing of human migrations seems as real here as in central Asia (Huntington and Visher, 1922), although sociologically oriented historians seem to have a marked distaste for climatic determinism. There is no doubt that the nomadic Neolithic people were forced out of the north African savannas by the advance of the climatic (*not* human-induced) desertification of the Sahara. It was surely not a question of overpopulation or politics around 6000 BP! At this same time began the intensive settlements along the Nile that were associated with a meteoric rise in building, engineering, and irrigation skills. Equally well, during certain catastrophic climatic deteriorations, such as the desiccation of the Indus Valley around 3000 BP, one sees the total obliteration of a major culture, in this case Mohenjodaro (Bryson and Murray, 1977). Care needs to be taken, of course, to avoid psychological and viewpoint errors (Bryson and Padoch, 1980), but the large number of investigators is a fairly healthy insurance against jumping to conclusions.

6. Coastal beach ridges of Hudson Bay and elsewhere

Around the coastlines of the world certain areas are favored by gentle offshore gradients and large supplies of gravel, sand, or silt so that a prograding shore tends to develop. In regions of crustal subsidence such as the Mississippi or Rhine deltas, those shoreline deposits are progressively drowned and buried. In rather stable regions such as the shores of Western Australia, West Africa, and Brazil, the prograding systems develop sheaves of beach ridges across flat coastal plains. In uplift regions such as the glacioisostatic rebound coasts of Hudson Bay, the Canadian Arctic, Spitsbergen, Scotland, and Scandinavia those beach ridges are progressively uplifted, so that in protected places they form dramatic staircases of shore terraces. This third category is the best for the study of historical time-series, because the crustal uplift is a slow, steady, non-fluctuating motion and relatively easy to measure. The residual value in the elevation, which with the width ratio of the terraces, is a measure of past sea levels and degree of storminess.

The staircase beach ridges of Richmond Gulf on eastern Hudson Bay (Quebec) have been measured in the field in numerous traverses (thesis work of Hillaire-Marcel), and associated formations have been dated by radiocarbon with many cross-checks, so that chronology is considered well established (Fairbridge and Hillaire-Marcel, 1977; also discussion in Fairbridge, 1983). The area is near the former center of the maximum ice loading of the Laurentide (New Quebec) ice sheet and today is the site of maximum deglacial isostatic uplift. The oldest Marine Limit of 8300 BP is uplifted to around 315 m elevation. The Earth's crust here is still rising at about 8 mm per year. The rising eustatic sea level of the last post-glacial (Flandrian) phase caused sea water to intrude the Hudson Bay beneath the continental ice, leading to buoyancy and a very rapid calving, so that within a few hundred years of 8300 ^{14}C years BP the entire bay was emptied of icebergs and normal marine deposits began to accumulate. Concordant dates for this event are obtained all around the bay. A shallow-water marine inland sea of very large size (520 000 km^2) was created, at first several hundred meters deep but today with an average depth of only about 100 m. It is thus hydrographically rather comparable to the Baltic, and subject to strong fluctuations of water level due to atmospheric pressure, wind, and seiche effects; with the addition of tides (of the order of 1 m) the annual water level fluctuation may thus be as much as 2.5 m.

The staircase of raised beach ridges is for the most part uniform, with 185 steps distributed from the present beach up to 315 m. There is a gradually increasing individual height toward the top, a phenomenon due to the initially rapid isostatic rebound, now gradually decreasing. The mean age difference between beaches is 45 ± 1–5 years. On average, they are 1 m high and 5 m wide, but exceptional ones are 3–4 times higher. The individual ridges represent relatively short periods (3–10 years) when the degree of storminess was high, those episodes being separated by longer intervals when only fine sands were moved by the waves and relative sea level was rather stable. The Hudson Bay, at the present time, freezes over for 7–9 months of the year, so it is only summer storminess that is involved. Meteorological and tide-gauge records provide a guide to this alternation between stormy and calm periods during the present century, but the duration of the ice-free to frozen-in states is documented rather well, back to the early 1700s, thanks to content analyses of the old fur-trading log books of the Hudson Bay Company (Catchpole and Ball, 1981).

On the south western Hudson Bay, a more gentle slope permitted the preservation of intermediate beach ridges at 22 and 11-yr intervals (Grant, 1993). Prior to about 3000 BP the higher isostatic emergence rate amplified the contrasts (Figure C18).

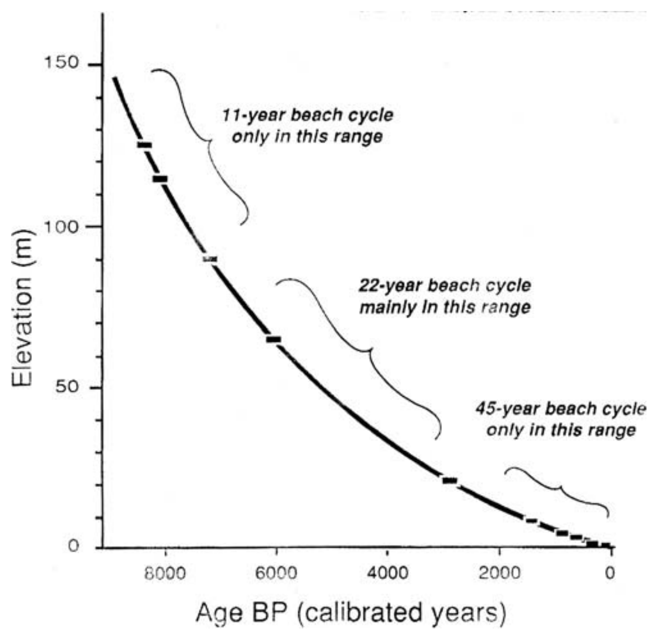

Figure C18 Holocene beach-emergence cycles in southwestern Hudson Bay, showing gradual decrease of glacio-isostatic recovery over the last 8000 years (courtesy: Douglas R. Grant, Geological Survey of Canada). Note: Progressive increase of storminess cycles from 11-year prominence, in the early stage, before 7000 yr BP (calibrated), through 22-yr before 3000 BP, down to 45 yr in last 3000 years. In the eastern coast, the 45 yr and longer cycles occur throughout the sequence (Fairbridge & Hillaire-Marcel, 1977).

Maximum beach building corresponds to longest periods of summer ice-free conditions and to the incidence of frontal storm systems across the bay. When the jet stream lies well to the south, the Arctic high-pressure cells move sluggishly across northern Canada from northwest to southeast, often remaining stationary over northern Quebec, causing cold easterly to northerly winds on the bay. In contrast, when the jet trajectory runs farther north, the frontal systems cross the bay, bringing in warm, wet, southerly and southwesterly air; a considerable storm fetch and wave set-up is then generated along the eastern margin of the bay.

From studies of pollen indicators and dune sand outbreaks in the same region, Filion (1983, 1984) discovered a correlation between the cold-dry climatic cycles, forest fires, dune building, and anomalous negative sea-level fluctuations that have a cyclicity of about 360 years, as noted earlier in the beach ridges by Fairbridge and Hillaire-Marcel (1977). Both amplitude and period are very similar to the Rocky Mountain tree-ring record of LaMarche (1974).

For convenient identification the ridges are numbered from no. 1 to no. 185, with 1 tentatively dated from tide-gauge and meteorological data as AD 1962. Beach 2 would be 1917, and so on. To calculate sidereal BP dates we create an imaginary beach 0 at 2007 which is 57 years AP (after present, the arbitrary astronomic and geochemical datum being 1950). Then for any beach number (n) we calculate: $(n \times 45) - 57 =$ date (in sidereal years BP)

The most astonishing feature of these beach ridges is their regularity. No stochastic variance would seem possible for such a uniform cycle of 45 years. When plotted as a simple graph, with the isostatic component removed, it is evident that there is also a further modulation into sets of somewhat higher and somewhat lower ridges. The length of this secondary long-term oscillation seems to have at least two components, of the order of a half and a third of a millennium. What manner of cycles are these?

In view of the dramatic demonstrations by a number of workers (Currie, 1984 and others) that both solar and lunar cycles are clearly recorded in long-term historical series, we need to compare them with both the standard oscillations (i.e. the lunar nodal precession at 18.6 years and the solar activity cycle of 11–22 years, together with their harmonics) and various higher-order astronomic alignment periodicities. In the first paper (Fairbridge and Hillaire-Marcel, 1977), a double-Hale solar cycle of 45 years was suggested, and it is indeed observed that 11-year solar cycles are commonly arranged in groups of four (with increasing sunspot numbers). A 45-year periodicity is seen in eastern North American summer temperatures, in sea levels, and in the secular drift of the geomagnetic field. A careful analysis of Canadian Arctic winter temperature and pressure data (meteorology plus proxies) by Guiot (1985) show a phase coherence with Currie's solar and lunar results, as well as a volcanic dust signal. As mentioned earlier, the 18.6-year lunar tidal effect not only appears in sea-surface temperatures and sea tides but also triggers some earthquakes and volcanicity. Currie's work shows that its phase fluctuates by 180° regionally and over long time-series, and this was confirmed by Guiot for the northern Arctic region (including the area of the north magnetic pole). When the solar signal is strong the lunar one is often weak, and vice-versa. Maximum winter temperatures often correspond to the even-numbered (south positive) solar cycles that are most often weaker than the next cycle. Over the last centuries the solar (22-year) signal has been strongest over the Hudson Bay. However, after having established the most probable twentieth-century dates (tide- and weather-based evidence) for the Richmond Gulf beaches 1 (1962) and 2 (1917), it is noticed that 1962 matches the lunar perigee–syzygy cycle (31-year) peak and 1917 matches the 18.6-year lunar–solar tidal maximum. The relationships of the lunar cycles to sea level and storm-flooding events are described by Wood (1978, 2001).

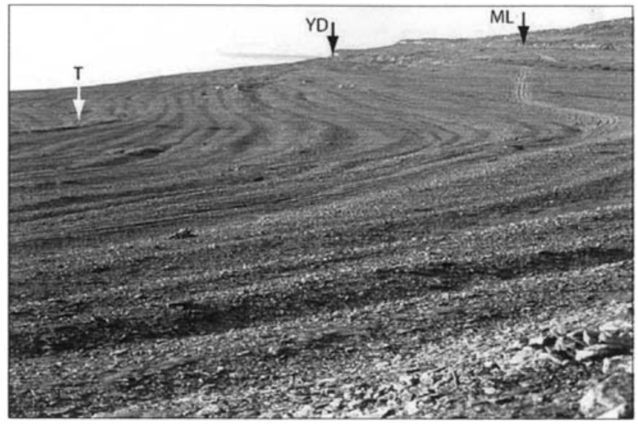

Figure C19 Emerged beach ridges facing the Barents Sea in northern Russia (Møller et al., 2002), the dates ranging from late Pleistocene to early Holocene, and the storminess frequency is less than on Hudson Bay. The "Marine Limit" (ML) here is probably Alleröd, and the Younger Dryes (YD) is dated 9190 BC (Cal. 11,140 BP).

A systematic exploration of Arctic coasts was undertaken by the writer and associates (Fletcher et al., 1993), with comparable results except that in northern Norway and Russia (Møller et al., 2002) ice-borne boulders marked cold cycles with advance of Arctic sea ice (Figure C19).

Referring now to the long-term climatic oscillations on the Hudson Bay shown by Fairbridge and Hillaire-Marcel (1977), it is curious that the solar–planetary influences (178-year and 356-year cycles) seem to be most prominent since about 3000 BP, but prior to that the 556-year (50 × sunspot cycle), and 558-year lunar cycle (28 × nodel cycle and precession of lunar perigee) seem to be associated with larger fluctuations. From evidence in the Baltic history, Pettersson (1912, 1914, 1930) reached the conclusion that the long-term lunar cycles played a leading role in the regional oceanographically forced climatic behavior but recognized that they also were gravitationally tied-in to the sun. A long-term correlation with the Baltic herring fishery introduced an interesting economic overtone.

In nonglacial areas, dated series of prograding beach ridges have been described in Alaska, Mexico, and South Australia. Several hundred-odd beach ridge series north and south of Perth, Western Australia, again disclosed a 45-year periodicity with a larger fluctuation of *ca*. 360 years (Woods and Searle, 1983). This is remarkably interesting from the climatological viewpoint, because they are evidently responding to southern hemisphere weather systems as well as northern triggers.

7. Marine varves

Normally all deep-sea deposits are homogenized to about a 3000-year smoothing by bioturbation (burrowing worms and so on), but in a few isolated basins anoxic conditions isolate the bottom waters and annual layers are found, e.g. the Adriatic Sea, Red Sea, Cariaco Trench, and Gulf of California. Well-studied records were those from the Santa Barbara Basin (Pisias, 1978). By analyzing the climatogenic significance of plankton carried by the California Current, he established sharp fluctuations in sea-surface temperatures over the last 8000 years. The major cycles are of the order of 500–1000 years, and disclose, for example, a variance of the mean February (midwinter) temperature of as much as 15°C. This astonishing range is apparently due to sort of northern hemisphere El Niño effect, whereby the normally cold south-setting California Current is replaced in some periods by warm tropical water. Pollen analyses disclose that the southern California landscape has been correspondingly vegetated at cold times by a dry, chaparral cactus and coastal sage scrub flora–the Sonora Desert type, which today exists 500 km farther south in Baja California. At warm times this flora was replaced by a mixed deciduous and pine forest, reflecting warm–wet conditions with southerly winds and summer rains.

Most significantly, the weakening of the California Current appears to match northward shifts in the Kuro Shio in the western Pacific, indicated in Japan by northerly appearance of reef-building corals in the littoral terraces, assumed to be partly eustatic (Taira, 1980). Chronologically, they seem to approximate those dated by Fairbridge (1961) and others from Australia and elsewhere in tropical seas. Climatologically, it appears that during the major warm cycles, the subtropical highs expand and migrate somewhat poleward, accompanied by a reduction in the equator–pole thermal gradient. Trade winds weaken and the equatorial Pacific westerly air flow increases. Sea-surface temperature rises and we have a super-El Niño effect. The major warm cycles were most prominent during the mid-Holocene (Fairbridge, 1993). In the coastal shell middens of Peru a remarkable shift from warm to cold-water mollusca is observed in the late Holocene.

8. Volcanic signal

As developed by Lamb (1972, 1977) and much favored by meteorologists and journalists in the present century, the concept of violent, dust- and gas-producing volcanic eruptions is important climatologically because they often distribute a dust veil into the stratosphere that effectively screens incoming solar radiation, thus cooling the Earth's surface. In the four largest of such eruptions of the recent historical era, Asama (Japan) in 1783, Tambora in 1815, Krakatau in 1883, and Agung in 1963, the effect may have been a planetary cooling that lowered the mean global temperature by as much as 0.5°C for up 2 years (Rampino and Self, 1982). Tambora dramatically made 1816 the ill-famed "Year without a Summer" (Stommel and Stommel, 1983). No longterm climatic effects have been demonstrated, but sociological effects were, in places, profound, ranging from forced migrations from New England, to famine in central Europe and Bengal. Even more dramatic results followed a Sunda Strait eruption in AD 535 (Keys, 1999), although its effects were mainly felt in AD 536.

Volcanic eruptions have been traced back and catalogued as far as possible for the whole 10 000-year interval of the Holocene Epoch (Simkin et al., 1981), but their climatic potential has been variously appraised. They have even been elevated to the role of controlling the primary cooling in the 100 000-year glacial–interglacial cycles (but see criticism by Huntington and Visher, 1922; Schwarzbach, 1950, 1974). Evidence for that hypothesis seems to be entirely in the wishful-thinking category. Volcanism, as a source of CO_2, might even lead to warm cycles, as suggested long ago by Fritz Frech (see discussion in Schwarzbach, 1974, p. 291). Seen in a global context, volcanic eruptions are too brief and discontinuous to play a major role in long-term climatic cycles–either heating or cooling, although their effect on regional populations can be apocalyptic (Winchestor, 2003).

Although they clearly modulate climatic fluctuations on a less-than-decadal basis, the major volcanic eruptions over the last few centuries, for which we possess good chronological data, all seem to have taken place during but not at the beginning of cooling cycles. Lag times are of the order of 10–20 years. This anomalous reversal of possible cause and effect was commented on long ago by Huntington and Visher (1922), who pointed out that a month-by-month global study by Arctowski showed that the great eruption of Katmai, in June 1912, coincided with a general cooling cycle. The volcanic school submit that only equatorial volcanoes should count because of the lower tropopause in low latitudes, but neither Agung, nor Krakatau, nor Tambora–all near the equator–had lasting effects.

Tambora, in 1815, was one of the largest eruptions and it was certainly associated with cooling, but the whole of the preceding decade was one of global cooling. Pettersson (1914) noted that the three coldest winters ever recorded in Stockholm were in 1805, 1809, and 1814. In 1809 a Russian army attacked Sweden by crossing over the Baltic on the ice from Finland, the general reporting to the czar an air temperature measured at −36°C. Napoleon's retreat from Moscow was in December of 1810. For several winters even the southern Baltic was frozen over as far as the Kattegat. The East River (a salt-water channel) and lower harbor at New York City also froze over. The years leading up to 1815 were marked by one of the worst droughts in East Africa, subtropical droughts generally being associated with worldwide cooling phases. The year 1810 also marks the lowest point in mean sunspot numbers in the last two centuries.

An alternative hypothesis has therefore been offered: "Can rapid climatic change cause volcanic eruptions?" (Rampino et al., 1979). Climatic stress, notably atmospheric angular momentum for short pulses, and ice build-up and decay for longer cycles, is proposed as a trigger for crustal stress. Sudden accelerations in the atmosphere's angular momentum (Rosen et al., 1984) are accompanied by changes in spin rate as well as in the rate of secular drift and intensity of the geomagnetic field. It seems hardly a random coincidence that the two largest changes in spin rate and geomagnetic field intensity in the twentieth century match the declination maximums of the lunar nodal cycle (1913, 1969) when that cycle correlates precisely with either sunspot maximums or minimums (Courtillot et al., 1982).

Long ago, Köppen 1914 proposed a correlation between air temperatures, sunspots, and volcanicity. Köppen had noticed as early as 1873 that a temperature–sunspot correlation could be seen in low-latitude records but not elsewhere. The cumulative factors of lunar periodicities combine to give 1913 the maximum crustal stress build-up in more than one millennium. The cool cycle at this time and the volcanic eruptions (of which Katmai in June of 1912 was only one) do not seem to be random occurrences. A statistical survey of large earthquakes in California (Kilston and Knopoff, 1983) shows a clear link to lunar periodicities, including the 18.6-year cycle, probably through tidal loading on the continental shelf. A similar periodicity is observed in oceanic tidal height, and in temperature (SST), reflecting enhanced water exchange. Russian geologists have also observed the same 18.6-year period in the volcanic eruptions on Kamchatka (Shirokov, 1983).

During the Maunder Minimum (the sunspot dearth of 1645–1715) the incidence of global volcanic activity was twice as high as during the interval 1715–1800, and in the two decades 1680–1700 it was six times higher than for 1740–1760 (Fairbridge, 1980, p. 390). The coldest decades indicated by the Greenland ice-core oxygen-18 proxies were 1660–1680, preceding the volcanicity peak (see also Little Ice Age).

None of these complex relationships demonstrates conclusively a cause-and-effect situation. If the lunar declination cycle is locked into the sun's barycentric orbit and to solar radiation, as urged by Pettersson (1914), the 18.6-year period in historical climate proxies may also involve a solar pulse. The volcanic signal in long-term historical proxies thus deserves more far-reaching investigations, not so much as the cause of important climatic fluctuations but as a consequence, and also for their taphrochronological value, where they have a potential of providing a "golden spike" or a "silver thread" that runs through multiple proxy records, not least being the human observations of the actual eruption. Archeological documentations of the burial of villages or towns should be integrated into the record.

9. Paleomagnetic signal

The Earth's magnetic field is just beginning to be recognized as a major link between solar particulate radiations and their role in modulating terrestrial climates. This important geophysical parameter has been carefully studied and documented over a longer period than any other, initially because of its use in navigation. In 1692 Edmund Halley recognized that there was a secular westward drift of the magnetic field of some 0.2° per year, involving not only declination but also changes in inclination and total intensity. Because of the eccentric arrangement of the dipole and quadripolar fields, the actual readings of these variables at the Earth's surface are quite complicated and constantly changing; for navigational use they needed to be continuously monitored and mapped. Most significantly, the north and south magnetic poles (nonantipodal) are found to migrate some 5–10 km per year.

Prior to the documented historical period, i.e. the last five centuries or so, the record has been extended back by a technique known as *archeomagnetism*, whereby the former field characteristics may be measured in bricks or fireplace clays of archeologically established ages that have been heated and had cooled through their Curie point, thus preserving a fossil relic of the magnetic field of that time. The key observation was made in 1906 by a French geophysicist, Bernard Brunhes, studying the cooling of modern bricks (see Imbrie and Imbrie, 1979). Pioneer work on historical bricks was done much later by the Czech Geophysical Institute in Prague (Bucha, 1970). Recent studies with refined techniques confirm the general pattern, making the values stratigraphically useful.

The most striking result of this work was the discovery of a slow secular rise and fall of the total intensity field during the Holocene. At about 6000 BP (the Climatic Optimum of the Holocene Epoch), the dipole field moment strength was about 10% weaker than its present value. This variable is particularly interesting in connection with the flux rate of radiocarbon as determined from dendrochronological series (Suess, 1981), magnetic intensity being inversely proportional to radiocarbon value. That is to say, when the magnetic field intensity is weak, the influx of the weakly radioactive ^{14}C isotope is high, and vice-versa.

The Earth's spin rate is influenced by both exogenetic and endogenetic variables. Most emphatic of these variables is the mass transfer from glacial to interglacial mode and vice-versa; ice build-up near the poles shifts the moment of inertia and increases the spin rate, whereas deglaciation slows it down. Angular momentum is conserved by differential motions of the Earth spheres and by changes in the lunar orbit, thus affecting ocean tides. Changes in spin rate affect the angular momentum of all fluid systems, from the outer core, and asthenosphere, to the hydrosphere and atmosphere. The last is the most fluid and reacts most rapidly. The atmosphere is also most subject to extraterrestrial energy input and thus is most likely to disclose evidence of the two disturbing factors, external and internal (see also subsection 8 above).

The Earth's principal magnetic field, the dipole field, appears to be generated largely by the dynamo-like turbulence in a 100-km layer of low strength in the semiliquid outer core. Inasmuch as the spin rate of the mantle and crust are repeatedly subject to sudden changes, a differential torque is generated. The nearest familiar analog in human experience is the automatic transmission and gearshift of the family car.

The geomagnetic field appears, in fact, to be modulated by three principal phenomena, generating cycles or fluctuations of variable length and intensity. These are still subject to much research but appear to be as follows: orbital variations, magnetic pole migration and westerly drift, and solar wind modulation of the Earth's magnetic field. All three appear to play distinctive roles in the Earth's climate either by affecting the geomagnetic field intensity or by the geographic location of the magnetic pole. The relevant climatic link is through their control over the amplitude and location of incoming solar particulate radiation, which modifies the atmospheric chemistry and dynamics. The three control mechanisms are briefly summarized here.

Orbital variations.

Most important in this context is the variation of the obliquity of the ecliptic (*ca.* 41 000 years; Williams, 1993), together with some input from the eccentricity cycle (*ca.* 96 000 years) and the precession cycle (*ca.* 23 000 years and 19 000 years). These orbital variations, as already mentioned, correlate very precisely with geological proxies (Imbrie and Imbrie, 1979), but it has been questioned whether insolation alone has the ability to produce the large glacial–interglacial fluctuations. The question is usually skirted by pointing to the powerful feedback mechanisms that come into play, but the geomagnetic factor should also be included in the model.

Studies of the paleomagnetic intensity of Pleistocene deep-sea cores suggest higher fields in the cold, high spin-rate intervals, but the glacial-cycle sediments themselves carry more terrigenous (and magnetic) particles, which weakens that argument. Nevertheless, cold episodes during the Holocene also correlate to higher intensity signals, without contamination, so that the interrelationship is probably nonetheless true. What triggers the reversals in the geological past is still a controversial problem, but it seems that rapid changes in the spin rate and tilt must accompany the intervals of rapid crustal break-up and redistribution. Long periods of no magnetic reversal seem to coincide with times of mild climates (e.g. the late Cretaceous). Conversely, the times of disturbed crustal distribution and cycles of ice-age glacier loading must affect both tilt and spin.

During the course of the Holocene Epoch there have been appreciable variations in the secular orbital parameters. In particular, the tilt angle is now nearly 2° less than at the beginning of the Holocene (Berger, 1978). This secular change may be the principal component in Bucha's archeomagnetic variation. Mörner (Mörner and Karlén, 1984, p. 489) has presented the concept of a paleogeoid, linked to a phase transition at the core-mantle boundary, that is paralleled during the last 10 000 years by the curves of archeomagnetism and ^{14}C production.

Magnetic pole migration and westerly drift.
Within a period of about 1800 years the eccentric dipole field migrates westward, as does the weaker quadripole field, but more slowly, giving it a relatively eastward component, having a period of about 2800 years. The magnetic pole migrates at about 5–10 km per year. Bucha (1984) illustrates the position of the geomagnetic north pole over the last 18 000 years based on values obtained in Czechoslovakia, Japan, and the United States, its trajectory appearing to follow a quasi-ovoid form over the period of historical measurements, with a four-leafed clover form in its longer records.

To obtain more detailed records of total intensity, declination, and inclination, a vigorous campaign of lake coring has been undertaken in Europe, North America, and Japan (e.g. Creer, 1981; King et al., 1983). Records are pushed back to beyond 15 000 BP. They show regular fluctuations of all three variables but, because of the field asymmetry, the pattern varies slightly from one region to the next and with a general lag corresponding to west drift. Eight inclination cycles occurred during the last 10 000 years.

It was proposed by Fairbridge and Hillaire-Marcel (1977) from evidence of alternating calm and storminess cycles in the Hudson Bay that the more westerly sites of the magnetic pole generated a higher continentality situation (cold and calm) in the Canadian Arctic, in contrast to the more easterly pole situations (more oceanic sites around Spitsbergen), which coincided with warmer temperatures, less summer ice, and increased stormwave action (Fletcher et al., 1993).

Bucha (1984) believes that during brief but highly energetic solar flares the site of the magnetic pole and its associated auroral oval favors the development of low-pressure systems and storminess, which in turn helps to maintain milder temperatures, and that this condition applied to the North Pacific–Aleutian area during the late Pleistocene (18 000–12 000 BP) This would explain the lack of ice sheets in eastern Siberia and Alaska at the time. In contrast, according to Bucha, the remoteness of the North Atlantic region from the north magnetic polar area during the last glaciation maximum permitted a stable high-pressure situation favoring meridional air flow and the build-up of major continental ice sheets in eastern North America and northwestern Europe. If these hypotheses are correct, then not only would historical Holocene climatic fluctuations but also the glacial–interglacial relationships have a geomagnetic component in the 1800-year range and its subharmonics.

When the geomagnetic field is in an extended weak cycle, as during the mid-Holocene, and is modulated by unusually weak, short-period solar cycles (Damon et al., 1978), there will be an enhanced effect of galactic cosmic radiation, marked by prominent spikes in the ^{14}C flux record (Suess, 1981; Stuiver and Braziunas, 1993). The cosmic rays lead to the nucleation of high cirrus clouds (Roberts and Olson, 1973). This stratospheric cloud screen would furnish a high-albedo, cooling screen for the midlatitudes and Arctic.

During magnetic excursions, field intensity drops away dramatically and the result can be catastrophic cooling events as shown by evidence from the Pleistocene Lake Biwa, Japan (Kawai et al., 1975). In contrast, when the long-term magnetic field is strong, as seems to be the case during glacial cycles when the spin rate is high, the solar signal will be largely suppressed.

Solar wind modulation of the Earth's magnetic field.
This exogenetic factor provides a variable modulation of the internal field that can be identified in two independent ways. On a day-to-day basis, eruptions of giant solar flares (see Sunspots), observed astronomically, can be traced through a series of magnetosphere and ionosphere interactions, ultimately to their role in atmospheric chemistry.

The major flare events trigger upper atmosphere chemical reactions, particularly in the magnetic pole area, that in the troposphere at the 500-mb level lead to a dramatic warming (20–30°C) within 24 hours, accompanied by a drop in air pressure and the initiation of giant frontal weather systems that bring in a flow of warm humid air from as far away as the North Pacific. Such events trigger displacements of the jet stream and related atmospheric systems throughout the high- and mid-latitude belt of the northern hemisphere (Bucha, 1984).

One such solar–terrestrial event will not, of course, generate a climatic change, but the rate of incidence and magnitude of major solar flares is a variable that correlates roughly with the 11-year solar cycle. Landscheidt (1984) relates this variable to the crescendos of the torque stress developed by the spin and orbital positions of the planets. It is, therefore, an astronomically predictable quantity.

The most complete long-term record reflecting geomagnetic–climatic relations, now extending over almost the whole 10 750-year length of the Holocene, is provided by the radiocarbon flux-rate measurements made from dendrochronological series (Suess, 1981; Damon et al., 1978; Stuiver and Braziunas, 1993). These represent the ^{14}C isotope product of cosmic particle impacts on nitrogen in the stratosphere. The actual flux rate depends on two principal factors: (1) variation of the geomagnetic field strength in an inverse relationship–that is, when the field is weak, the flux rate is high; and (2) variation in the strength of the solar wind, which is modulated by planetary motion dynamics on the sun's photosphere (see Sunspots). A third but less frequent phenomenon is the modulation of galactic cosmic radiation by supernova events. Damon et al. (1978) commented on a very important aspect of these interrelationships already noted above in the subsection "Magnetic pole migration and westerly drift." When the internal geomagnetic field is weak, as in the mid-Holocene about 6000 BP, the amplitude of ^{14}C flux-rate departures is up to five times higher than their level during the high-field intensity intervals. These large fluctuations are noted at about 500–1000-year intervals (Figure C20). A similar oscillation may be seen in many of the standard climatic time-series noted in this entry: sea-level fluctuations, palynological diagrams, ice-core data, hydrological indicators, and so on.

From all these indications it is evident that the solar (i.e. exogenetic) component is most important in terrestrial climate series in the 500–1000-year time frame. However, on the 10 000–100 000-year scale the endogenetic components (controlled by spin-rate and tilt factors) are paramount.

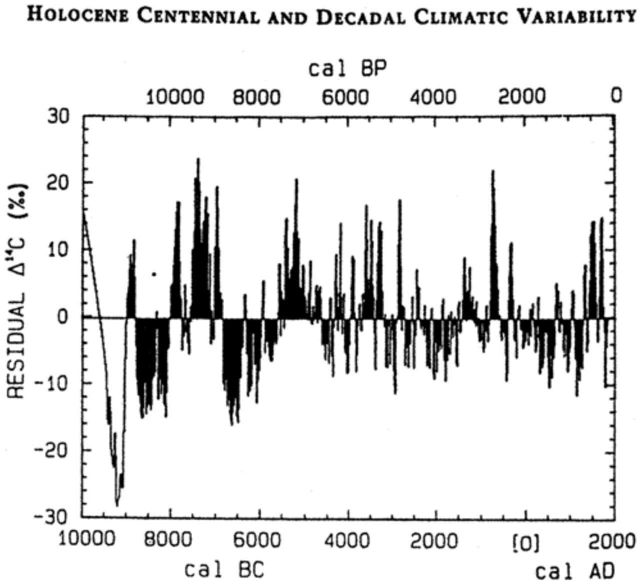

Figure C20 Solar variability, indicated by C-14 flux (inverse to solar activity) based on tree rings, and partly on corals (Stuiver and Braziunas, 1993). Maxima correspond to temperature minima in the Greenland ice cores. The post-Younger Dryes (cold spell, not shown) was followed by the dramatic warming of the Pretoreal and its several oscillations. Glaciers tend to respond rather sluggishly, so that eustatic sea level shows a retardation (Fairbridge, 1961).

10. Climate–aurora–sunspots documentation of ancient China

A gratifying feature of the new cultural orientation in modern China is the liberal approach to science and an appreciation that the written archives of that country contain the largest storehouse of proxy climatic data outside of Europe (Chu, 1973; Wang, 1979). Studies of the aurora and sunspot records were undertaken by Schove (1955), assisted by expatriate Chinese scholars, and in this way the first long-term (2600-year) reconstruction of solar history became available. Subsequent improvements and reprints of the key literature have been published in Schove (1983). The aurora have been separately treated by Siscoe (1980). Spectrum analyses by Jelbring (1995) broadly confirm the Schove model.

A preliminary study of the long-term variables in China's climate has been offered by Ren and Li (1980), who came to the conclusion that its cyclicity was in phase with the All-Planet Synod, an approximately 178-year alignment of planets in quadrature. At such times, as demonstrated by Newton, the sun is displaced as much as 1.5×10^6 km from the barycenter (center of mass) of the solar system, its spin rate accelerates (up to 5% from the usual 27 days) and its photospheric convection is retarded, resulting in a decrease in solar emissions (see Sunspots). On Earth these sunspot-dearth periods are associated with cooling spells (see Little Ice Age).

A study of diary and yearbook records for the period AD 1470–1979 by Wang and Zhao (1981), using content analysis, has disclosed six spatial patterns for droughts and floods, using 25 stations in eastern China, arranged in east–west belts in more or less meridional distribution. The first eigenvector relates to long-term variations of the Southern Oscillation Index, which is linked to the intensity of the western Pacific subtropical high. The second is linked to the same high, specifying whether its influence extends north or south of the Yangtze River. The third relates to the high-frequency variations of the East Asian Circulation Index.

The linkage to the Southern Oscillation makes these figures of global significance, especially for the Indo-Pacific subtropics and regions affected by the El Niõ Southern Oscillation, which in the recent example (1982–1983) means practically the whole world. When the Southern Oscillation and Walker Circulation indices are both high, in the mid-Pacific there is low sea-surface temperature (SST), the Pacific Hadley cell weakens and the summer monsoon is strong in southeast Asia (Yoshino and Xie, 1983), and the Yangtze discharge level is high.

In an investigation of winter thunder from 250 BC to AD 1900 by Wang (1980), it was possible to show a correlation with cold airmasses associated with unusual winter front development. A close match was found with temperature records collected by Chu (1973). It is significant that the patterns also correlate with similar historical records from Japan based on the dates of the blossoming of cherry trees (an important ceremonial event) and the freezings of Lake Suwa (Arakawa, 1957).

Application of modern computer technology and statistical theory to the Chinese data for the Beijing (Peking) are was begun by Hameed et al. (1983), who found in the proxy drought–flood records since AD 1470 signals of both the 18.6-year lunar nodal periods, probably a tidally induced modulation, and the 11-year solar emission cycle. These conclusions were dismissed by Clegg and Wigley (1984), but in a study by Currie and Fairbridge (1986) the results of Hameed et al. were supported, and, in addition, the curious bistable phase reversals found elsewhere in the world (Currie, 1984) appeared once more. Currie employs the maximum entropy method (MEM) to develop power spectra, and then constructs waveforms for the principal cycles as indicated. This permits a visual comparison of phase and amplitude, which can then be conveniently compared with historical events.

Testing and modeling

In order to derive a useful scientific product from the vast store of historical data, documentary and field, evidence that are either available at the present time or merely potential and unexploited, it is desirable to develop programs of testing and modeling. This rich area of research endeavor has hardly been explored up to the present moment, except in the most sporadic or generalized way. Intuitive matching has been the usual, nonrigorous methodology, and the lack of adequate statistical procedures has been brought out by Pittock (1978, 1979, 1983), Currie (1984), and others. It is petulant and childish however to make fun of the pioneer investigations, many of which rested on the most limited or crudest of databases. Yet, almost miraculously, many of their preliminary conclusions have turned out to have more than a grain of truth, as may be seen from a careful reading of Lamb (1972, 1977), for example.

In the preceding sections dealing with standard time-series, 10 examples of historical data types covering the last 10 millennia are presented. They have been derived conceptually from a short contemporary base line of instrumentally observed processes, furnished by meteorology, oceanography, geology, botany, astronomy, and astrophysics. Building up on this minuscule chronologic base line we are able to construct an array of time-series of varying quality but with sufficient common threads that the researcher gains confidence.

The last 10 millennia are today not a sea of unknowns. During the approximately 100 years that have elapsed since Blytt and Sernander constructed the first climatologically oriented chronology (covering the last 15 000 years), based on Scandinavian pollen analysis, we have now come to possess three independent time scales that have the potential of providing year-by-year chronologies and that are mutually verifiable. Probably their error bars are still in the range of ± 10 years, but the three methodologies have the potential of refinement and correction; they also can be extended to other (favorable) regions around the globe. An immense field of potential research is thus open and still largely unexploited.

The first and most ambitious attempt to develop a global model of the glacial–interglacial climatic fluctuation generated the CLIMAP (1976), which, for all its faults, was a remarkable first attempt. It exploited a largely oceanographic–paleontologic database to derive a broad picture of an instant in time during a glacial-phase maximum. Further endeavors have been made for modeling distinctive intermediate (Holocene) stages. Using the three year-by-year chronologies, supplemented by input from the other seven signals, an exciting vista is now open for establishing what *did* happen. The geologists have a pet phrase: What did happen, can happen! From that base comes the potential of explanation and prediction.

Summary of historical climatic epochs

In the space of one encyclopedia entry it is impossible to review the entire 10 000 years of climate history on a world basis. Here we concentrate on the best-known Scandinavian and northwest European data series with only brief references to events elsewhere. No book exists that summarizes the record, but several useful references deal with specific fields of research. Three excellent multiauthored volumes cover the Holocene history of coastal changes and deposits around the Baltic (Gudelis and Königsson, 1979), the North Sea (Oele et al., 1979) and the Irish Sea (Kidson and Tooley, 1977). The general vegetational and climatic history of Europe has been summarized by Frenzel (1973). A single-volume, multiauthored work has been dedicated to the Holocene of the United States (Wright, 1983) and the Holocene is included in a companion volume on the Quaternary of the USSR (Velichko, 1984). Detailed investigations are covered in the journal *The Holocene* (vol. 1, no. 1, 1991).

There is now an abundance of professional papers but, alas, the much-needed syntheses are for the most part lacking. The explanation for this lack of what might appear to be a key area of scientific and humanistic literature is associated in part with the inherent costs and problems of dating, and in part with the multidisciplinary training that would be needed to study the whole field. The specialists are botanists, or malacologists, or archeologists or whatever; unfortunately very few scientists are trained today outside of their specialized fields. In as much as the Holocene Epoch is a segment of geologic time, there is primarily a problem in geological chronology. Its climate history is a derivative product that must be extracted from a proxy record that basically is stratigraphical and preserved in geologic strata, or ice, or tree rings.

Reasons of priority dictate that, all things being equal, the first classification worked out by any team of scientists for an interval of geologic time is the one used as a basis for future workers to build on. In the case of the Holocene it may appear at first sight to be anomalous that the Blytt–Sernander botanical classification is commonly adopted (see earlier discussion under subsection 4: Palynology). This unusual action becomes understandable when it is remembered that the geologic basis for chronologic ordering is, first, the superposition rule, second, the fossil content of strata, and third, any absolute (geochemical) method of dating that may be appropriately applied to the formations in question. In this case it was the integration of pollen studies by Blytt and Sernander, in stratigraphic sequences of lake or swamp deposits, and provided with the varve counting by de Geer (see subsection 3), that furnished the first standard geologically based record of the last 10 000 years of Earth history.

This serendipitous combination of distinctive botanical fossils, sequential strata, and a precise year-by-year chronology placed Scandinavia in an extraordinarily advantageous position for presenting a world-standard historical record (Mangerud et al., 1974). Thanks to written documents and early scientific instrumental records, it would be additionally very helpful if this region could be selected for various stratotypes, i.e. chosen areas that contain well-preserved and well-analyzed records. It is unquestionably true that several other parts of the world *could* provide detailed, continuous stratigraphic records of the last 10 000 years of the Earth's history, but up till now no area has been proposed that has had the same rich, multidisciplinary background of study. Continuously cored lake or swamp deposits are now available from Britain, France, Nova Scotia, Minnesota, Japan, India, Colombia, and elsewhere. In time it is hoped much of the world will be brought into this chronological framework.

The smallest units of geological time, derived in the case of the Holocene from the original Scandinavian pollen-based stratigraphy, are categorized as *chronozones*. Five of these units, unequal in duration, are used to classify the epoch. In different parts of the world these boundaries are not always conveniently applicable to local records, and local terminologies need to be established. The purpose of a set of chronologic labels is simply that it is helpful to have a reference standard that is universally known to scientists. The local records should never be forced into a straitjacket of any assumed universality. Many natural boundaries are diachronous; this is particularly true when we are dealing with a eustatically rising sea level, where the beginning of a transgression may be hundreds of years earlier than its culmination. Ocean water warms or cools slowly; glacier melting involves latent heat and other delaying factors; again there will be lags or retardation. The purpose of agreed time planes is simply to coordinate our data within an internationally understood framework of time intervals, often called *epochs* or *ages* in an informal sense. (It may be noted, according to the internationally agreed stratigraphic code, that each term has a specified meaning in the chronostratigraphical heirarchy; and *epoch* is thus longer than an *age*, and shorter than a *period*, e.g. Holocene Epoch, Quaternary Period.)

The Blytt–Sernander scheme of pollen stratigraphy was applied to every chronozone back to the first formations to emerge from the melting ice fronts of northwest Europe; that is, from about 14 000 BP, ranging up to the present. The earlier zones (14 000–10 000 BP) are classified as late Pleistocene (Weichselian or Wisconsinan glaciation). In this entry we only consider the Holocene, which includes five principal units. Several are now subdivided, giving 10 units in all.

The latest Pleistocene corresponds to the first part of the very rapid deglacial process, when the midlatitude glaciers were melting at an astonishing rate, peaking in two warm phases (Bölling: 12 500–12 000 BP; and Alleröd, 11 800–11 000 BP) that were separated by three cooler fluctuations, known as the *Tundra* ages (Oldest, Older, and Younger Dryas). Heavy snowfalls consequent upon the oceanic warming (the Simpson principle), were probably responsible in part for glacial readvances in the mountains during those colder intervals. In the last one, the Loch Lomond Readvance in Scotland, glaciers actually reached down to sea level. Even in the tropical oceans signs of these cool fluctuations are seen in the oxygen-18 records.

Thus the stage was set for the beginning of the Holocene, the boundary being established, by international agreement, as the time when warm marine fossils replaced cold ones in the uplifted deposits of southwest Sweden (Olausson 1982; Fairbridge, 1983) the time was roughly 10 000 BP (by radiocarbon) ±300 years. (This time equivalent in sidereal years is believed to be approximately 8000 BC, but precise calibration is still controversial; Cato, 1985.) In other parts of the world another boundary might have seemed better. In tropical regions the dramatic postglacial warming was generally about 13 500 BP. In northern Canada or Alaska about 8000 BP or even 6000 BP might seem more appropriate. The local boundaries are always diachronous, so that only the somewhat arbitrary chronostratigraphic boundary should be shown as a horizontal plane on our diagrams, a "time line" that has been agreed to by a commission of the International Union for Quaternary Research (Olausson, 1982).

Holocene chronozones

We will now briefly review the 10 chronozones of the Holocene. The ages will be given first in sidereal (calendar) years, expressed as BC or AD, then in BP equivalent (before 1950), and finally in radiocarbon years, indicated as b.p. (lower case). Calibration tables for conversion of one to the other were provided by Klein et al. (1982), and subsequent authors, simplified in Roberts (1998). When precise astronomic equivalents are required, one should remember there is no year zero in the AD–BC system, so that the astronomical year − 100 is 101 BC and conversion to BP made by adding 1949 to the BC date, thus 2050 BP. Inasmuch as radiocarbon dates normally carry at least a ±100-year range of error, it is usual to round off calendar equivalents of such dates to the nearest 10 or 20 years. The chronozones are listed with abbreviated initials and with the Danish pollen zones (Roman numerals). Botanical aspects are treated in Birks and Birks (1980), for the North American setting by Davis (1983), for the Scandinerian setting by Berglund (1986).

Preboreal (PB, IV) 9350–7500 BC (11 300–8900 BP, 10 400–9650 b.p., calibration doubtful)

Glaciers still covered most of North America north of the St Lawrence and the Great Lakes, most of Scandinavia north of Stockholm and all of the central Alps. Sea level, at minus 40–45 m, had reached about halfway across the continental shelves in mid- to low-latitudes, introducing maritime air and leading to a rapidly rising humidity. The Baltic area was invaded by the Yoldia Sea. Marine formations of this age are isostatically uplifted today in the Oslo Fjord to 221 m and to lower elevations in southwest Sweden. In Finland the Yoldia strandline reaches 220 m. Tundra vegetation was quickly replaced by birch and pine forests across western Europe. In North America, boreal forests (mainly spruce) followed closely on the retreating ice but in the northern plains was replaced by deciduous forest by 9500 b.p. The Champlain Sea, which had invaded the St Lawrence lowlands somewhat earlier, brought maritime influences to the lower Great Lakes; whale bones and other marine fossils are found from Quebec to Michigan. (Note: radiocarbon dates given for the early Holocene are not precise.)

Early Boreal (BO-1, V) 7500–6900 BC (9450–8750 BP, 9650–8550 b.p.)

Glaciers were in universal retreat. The *climatic optimum* (Europe) or *hypsithermal* (North America), a diachronous rise of air temperatures to higher-than-present levels, reached southern Sweden, Britain, and much of North America except for the still-ice-covered areas of Keewatin and Quebec. Vegetation in northern Europe now became dominated by pine forests, but in favored areas in Sweden the hazel was flourishing, a thermal indicator requiring 2°C to 2.5°C warmer summers than today. Isostatic uplifts isolated the Baltic (the Ancylus Lake) and emptied the St Lawrence Lowland largely of the Champlain Sea, leaving the Goldthwait Sea in its lower valley. In the plains of the United States, prairie grasslands began to replace deciduous forests.

Late Boreal (BO-2, VI) 6800–6200 BC (8750–8150 BP, 8550–7700 b.p.)

Glaciers in Scandinavia now shrank to the principal mountain axis but the Ancylus Lake persisted until 8300 BP. In North America the Keewatin Ice Sheet and Laurantide (New Quebec) Ice Sheet became separate, with Hudson Bay open to the sea. Ice remnants still remained on the high ground of the Maritime Provinces and Maine. A strong zonal (westerly) air flow began to affect most of Europe and North America and northern Mexico, with Gulf air limited to the southeast United States. Alaska and most of Canada were influenced by a strong Arctic airmass. In northern Europe, pines and spruce were progressively replaced by mixed oak forests.

Early Atlantic (AT-1, VII-a), 6200–5500 BC (8150–7450 BP, 7700–6600 b.p.)

Glaciers became limited to high mountains in Scandinavia and to residual patches in northern Quebec and Keewatin. Sea level rose to within 20 m of its present level, and a warm-water marine transgression invaded the Baltic (the Littorina Sea). Progressive isostatic uplift has left former Littorina strandlines (high eustatic oscillations) in parallel belts around the coasts. Mixed oak forests spread across most of Europe and North America. Strong westerly air flows maintained highly maritime climates. Even in northern India winter rains were almost as important as the summer monsoon. The AT-1–AT-2 boundary is marked in Greenland ^{18}O ice cores by a distinctive cooling indication, and likewise in the California tree-ring ^{14}C series.

Late or Main Atlantic (AT-2, VII-b), 5500–4200 BC (7450–6150 BP, 6500–5300 b.p.)

The last ice patch disappeared from northern Quebec. Only mountain glaciers and polar latitude glaciers persisted for the rest of the Holocene. Around 6500–6000 BP eustatic sea level reached its present height and has only fluctuated somewhat

(+1 m to +3 m) since then. The late Atlantic is marked by the 3- m *Early Peron* shore terrace in the tropic and the *Nizza* terrace in the Mediterranean. Successive uplifted Littorina strandlines rim the Baltic. Climatic optimum warmth now reached its farthest northerly influence. In northern Canada the boreal (spruce) forest–tundra boundary almost reached the Arctic shore. In eastern North America and Europe mixed oak–elm forests reached climax states. From the Swedish flora, in 1910, Andersson determined that the mean temperature was 2.5°C above present. This value is about 20% of the total glacial–interglacial range for this latitude. In the tropics the East African rains were much heavier than today and the Nile floods were on average three times above today's. The Early Kingdom pharaonic culture was established in Egypt and the Mohenjodaro civilization developed along the Indus valley.

Early Subboreal (SB-1, VIII-a), 4200–2300 BC (6150–4250 BP, 5300–4000 b.p.)

In most latitudes this chronozone marks the beginning of the fall in global temperatures from the peak of the climatic optimum or hypsithermal. Its beginning is marked all across Eurasia, from Britain to Siberia by the *Ulmus* (elm) decline, with a dramatic change in the climax mixed forests. The progressive cooling is well shown by the gradual reduction in tree-ring widths; in the Bristlecone Pine series (White Mountains, California), they reached an initial low point about 2800 BC (matched by a high level of ^{14}C flux, evidence of low solar activity). This corresponds to the *Bahama Emergence* (−3 m) on tropical coasts. In the upper Midwest (Elk-Lake core, Minnesota), the same interval is marked by a 300-year period of reduced varve thickness and drop in diatom (*Melosira*) frequency; it was followed by an abrupt rise in varve thickness to a peak around 2000 BC (late Subboreal). Mountain glaciers showed repeated advances, corresponding to sherp ^{14}C flux peaks.

Late Subboreal (SB-2, VIII-b), 2300–900 BC (4250–2850 BP, 3800–2900 b.p.)

What is sometimes called a *mid-Subboreal* mild phase generated warmer climates for about five centuries in Scandinavia and elsewhere. Then from 1400 to 1000 BC the Bristlecone Pine and most pollen series show a major climatic deterioration, marked by a high ^{14}C flux; it was paralleled by the *Crane Key Emergence* in the tropical seas, followed shortly after by the *Pelham Bay Emergence* about 800 BC, both being certainly glacioeustatic in view of important neoglacial readvances in the interval 1350–450 BC (Denton and Karlén, 1973). In northern Canada the forest–tundra border retreated. Culturally this epoch was marked in Europe by the Bronze Age.

Early Subatlantic (SA-1, IX-a), 900 BC–AD 175 (2850–1775 BP, 2775–1700 b.p.)

In northern Europe this chronozone was characterized by cool to mild and moist conditions, accompanied by the general spread of beech forests. The Mediterranean saw the rise of the Etruscans, and of the Roman Empire. Sea level dropped from a high around +2 m in the classical Greek period, to a low around −1 m around the first century AD, followed by a progressive rise (late and post-Roman transgressions). Culturally and economically this sequence had a significant impact: Roman port facilities all around the Mediterranean are today at least partly submerged and commonly built over by Medieval structures. The coastal salt-pan economy was ruined by the rising sea level and salt mines were developed inland (Spain, Austria, Germany, Dead Sea), coastal farmers, particularly in France and the North Sea area, were forced inland or started to build dikes.

Mid-Subatlantic (SA-2, IX-aa), AD 175–750 (1775–1200 BP, 1700–1150 b.p.)

Corresponding to the late Roman/Byzantian cultural phase, the coasts of northern Europe were extensively inundated of the Mediterranean region was plagued by droughts and famines, which in the Middle East favored the rise of Islam.

Late Subatlantic (SA-3, IX-b), AD 750–present (1300–1 BP, 1300–1 b.p.)

In broad terms the youngest chronozone is divided into three: the *Medieval Warming* or *Little Climatic Optimum* (q.r.) from AD 950 until about 1250, the *Little Ice Age* (with two maximums, divided by a slight fifteenth-century warming (about 1390–1540), and finally the late nineteenth–twentieth-century warming that brings us up till today. The ^{14}C flux according to the California dendrochronological series remained fairly stable until about AD 1250, after which a precipitous decline is registered with minimums about AD 1350 and 1700, after which, with a minor low around 1810–1820, it rose to modern levels; after 1950 it becomes grossly contaminated by nuclear bomb pollution, followed by anthropogenic burning of fossil fuels and deforestation.

Sea level rose during the Medieval Warming to about 0.5 m above present, as may be seen from raised coral platforms and shell banks in the tropics (Rottnest Submergence in Australia), and from anomalously high Viking beaches in Sweden. It fell to at least −0.5 m during the Little Ice Age, after which its fluctuating upward trace is recorded by the Amsterdam tide gauge installed in 1682. With decadal fluctuations of around 5 cm, it rose to a peak around 1770 and then sank again to a low about 1820, since when it has been mostly rising (after correction for subsidence; Fairbridge, 1961, p. 104). The Little Ice Age regression has been called the *Paria Emergence*, from South American data.

As mentioned in the entry on the Little Ice Age, the most striking development of that interval was the enormous expansion of sea ice in the Greenland Sea, which around 1600 reached a peak of 26 weeks of ice cover on the north coast of Iceland each winter. The effect on the albedo was appreciable and long-term average temperatures in western Europe fell by 1–2°C. Around 1600 the north magnetic pole came to lie northeast of Iceland, and the low solar activity phases like the Wolf, Spörer and Maunder minimums led to excessive fluxes of cosmic radiation (shown by high ^{14}C and ^{10}Be values). According to the Roberts and Olson (1973) model, the geomagnetic screen is weakened during those solar activity minimums and the amplified cosmic ray flux created frequent high-altitude cirrus cloud cover, thus generating a primary stratospheric albedo.

Summary

Why study the historical geological record of the last 10 000 years? To the climatologist, trained on the straightforward hard data of meteorological instrument observations, who would perhaps ask that question, it might be useful to summarize some key points:

1. The current Quaternary Ice Age (of the last 2 million years or so) alternates rather abruptly between two modes: glacial and interglacial. We are living in one of the latter, the Holocene Epoch of geology, its lower boundary being 10 750 ± 300 BP.
2. During these 10 millennia, sea level has achieved its present level glacioeustatically, that is, mainly by the melting of midlatitude continental ice sheets. The last traces of the latter were gone by 6000 BP.
3. Following the Milankovitch theory of orbital control of terrestrial insolation, the last peak of effective radiation was about 9000 BP, around which time the sea-surface and low-latitude land temperatures (in areas not dominated by the residual ice sheets) a *climatic optimum or hypsithermal* climatic regime developed. Gradual deterioration (global Cooling and desiccation) began after about 5000 BP.
4. The last 5000 years have been marked by remarkably non-linear climate, glacial, and sea-level record. Fluctuations are mostly in saw-toothed patterns of various scales, progressive cooling with abrupt deviations to a generally warm mode; the upper thermal bound is episodically interrupted by cold spikes which vary in length from 1 year to a century or more. The amplitude of the fluctuations has oscillated or even increased, not decreased toward the present (Figure C21).

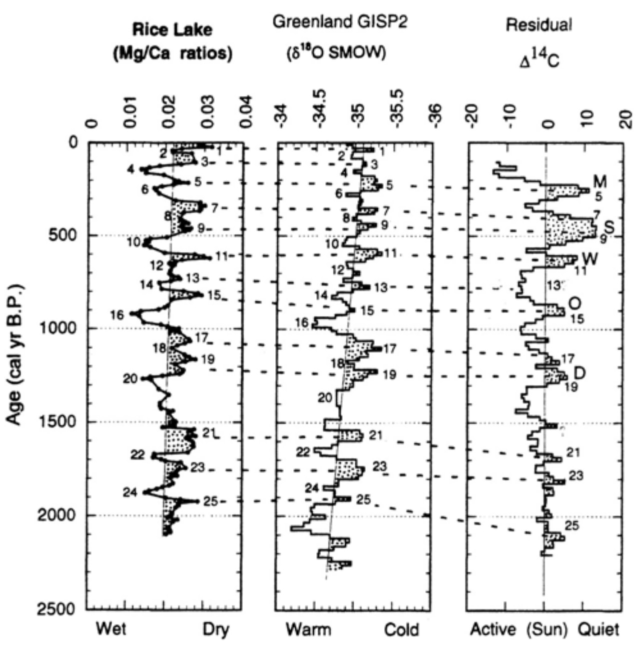

Figure C21 Three examples of cyclic climatic signals spanning the last 2000 years. Left: a lake core in western Canada (Rice Lake). Middle: an ice from Greenland (GISP-2). Right: bidecadal residual C-14 from tree rings (Stuiver & Reimer, 1993). With slight modifications for simplicity (R.W.F.). Abbreviations: **M** (Maunder Minimum: AD 1645–1715); **S** (Spoerer minimum: AD 1420–1530); **W** (Wolf minimum: AD 1280–1340); **O** (Oort minimum: AD 1010–1050); **D** (Dark Age minimum: AD 660–740). Each represents low levels of solar activity (sunspots), corresponding high O–18 levels in ice cores, i.e. cold episodes, and cool-dry climates in western Canada.

5. The nature of the cooling episodes is highly varied and subject to considerable research effort. The longer ones clearly relate to solar radiation (cf. ^{14}C data). More or less random spikes of 1–2 years correlate with specific incidents of volcanism, particular explosions leading to a gross loading of the *dust veil index* (DVI).
6. More or less regular quasi-biennial or quasi-triennial climate cycles relate to equatorial statospheric winds and the Walker circulation; the El Niño and Southern Oscillation are well-recognized expressions of these systems, which are controversial in origin, variously explained as stochastic (endogenetic thermodynamics), or as produced by solar–lunar forcing (exogenetic: radiational and tidal).
7. Time-series of millennial and multimillennial length furnished by ice cores, varves, and tree rings provide a sound statistical basis for exploring the nature of longer climate cycles. The solar (11- and 22-year) and lunar nodal (18.6-year) periods are now well established, but are associated with *noise* generated by complex feedback processes, ranging from magnetic and electrical fields to oceanic circulation and lagging sea-surface temperatures.
8. Long solar–lunar periodicities of the order of 45, 79, 93, 180, 360, 558, 1112, 1850 and more years have been studied only in a very preliminary way. Clearly, there is a strong astronomical pulse recognizable in all the long time-series but far more investigation is still needed.
9. The role of CO_2 in climate change, both long-term and authropogenic, is complex and remains to be resolved.
10. Finally, a word about the future. We cannot yet confidently make predictions, even for a few months. For the curious and the dedicated, however, the historical record offers a vast reservoir of information that is simply waiting to be tapped. The stratigraphic and other natural repositories have only just begun to be explored.

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Cross-references

Climate Change and Ancient Civilizations
Cycles and Periodicities
El Niño
Climate Change and Global Warming
Ice Ages
Little Ice Age
Southern Oscillation
Sunspots
Tree-Ring Analysis

CLIMATE VARIATION: INSTRUMENTAL DATA

Climate is often regarded as the average state of weather. Although this definition is not incorrect, it can lead (incorrectly) to an interpretation of climate as a constant, unchanging phenomenon. On the contrary, the climate system is characterized by constant change. Climatic variations result in the modification of the average, variability, and frequency of extreme values of temperature, precipitation, pressure and other meteorological variables. Such changes, which may happen in all temporal and spatial scales beyond those of individual weather events, are defined by the Intergovernmental Panel on Climate Change (IPCC) as *climatic variability*. Statistically significant changes in the mean state of climate or its expected variability, which persist for an extended period (decades or longer), are referred to as *climatic change* (IPCC, 2001).

Causes of climatic variability and change include: natural internal processes such as dynamic and thermodynamic interactions between the atmosphere and oceans; volcanic eruptions; external forcing, like e.g. changes of intrinsic solar radiation, or persistent anthropogenic changes in the composition of the atmosphere, or changes in land use. Such factors may affect climate directly by altering the local atmospheric heat budget and/or indirectly by modifying the planetary wind and ocean current circulation by which local climate to a large extent is regulated. The effects of climatic variation are numerous and often affect socioeconomic and natural systems.

Geographically, climatic changes are so extensive that they deserve to be treated as an integrated planetary phenomenon. Indeed, a hemispheric or global viewpoint is required in order to evaluate cause-and-effect relationships. On a worldwide basis, variations of the mean state of surface climate can be described with good confidence from instrumental observations since the mid-nineteenth century, although analyses of surface variability and extremes, and the climate of the upper atmosphere, are more limited.

Climatic analyses require homogeneous data, and this is particularly important for analyses of climate change and variability. A homogeneous climatic time series is defined as one where variations are caused only by variations in climate. Unfortunately, most long-term climatological time series have been affected by a number of non-climatic factors that make these data unrepresentative of the actual climate variation occurring over time. These factors include changes in: instruments, observing practices, station locations, formulae used to calculate derived quantities, and station environment, among others. Some changes cause sharp discontinuities while other changes, particularly a change in the environment around the station, can cause gradual biases in the data. All of these inhomogeneities can bias a time series and lead to misinterpretations. Climate researchers must carefully remove all these non-climatic biases or inhomogeneities or at least determine the possible error they may cause before drawing conclusions from data (Aguilar et al., 2004).

The available instrumental record broadly covers the period of anthropogenically induced increases in CO_2 and other greenhouse gas concentrations and the contemporary global warming. Global warming has led to a more extreme climate, an increase in global precipitation and atmospheric water vapor content, an increase in ocean heat content and a sea level rise, a reduction in mountain glacier extent and snow coverage, a shortening of the seasons of lake and river ice and a systematic decrease in spring and summer ice in the Arctic regions, among other effects.

Observational data contribute to the knowledge of paleoclimatic variations, helping to calibrate the relationship between proxy data (tree rings, ice cores, boreholes, sediments, corals, historical documents) and meteorological variables. Finally, instrumental data are needed too for the climate models which can predict the future evolution of climate and assess the goodness-of-fit of their results. The next section will focus on the analysis of temperature and precipitation evolution.

Climatic variations in the instrumental record

Global temperatures can be reconstructed through instrumental data back to the mid-nineteenth, century, as shown in Figure C22 (IPCC, 2001). The average global land surface temperature has increased by 0.6°C since 1860 (Folland et al., 2001). Warming is found to be significant for the aforementioned period annually and in all seasons. Independently compiled, homogenized and analyzed datasets (Peterson and Vose, 1997; Hansen et al., 1999; Jones et al., 2001) agree and show two remarkable warming phases, 1910–1945 and 1976 to present, as the most distinctive features of the observational record. Sea surface temperatures have experienced similar pulses, although variations and trends have been slightly less pronounced. As for land data, different compilations of oceanic data agree on similar results (Quayle et al., 1999; Jones et al., 2001; Reynolds et al., 2002).

The late nineteenth century was the coldest period in the instrumental record. This is true not only for the global estimates extending back to 1860, but also when some longer temperature records for central England, Fennoscandia and central Europe, going back to the eighteenth century, are considered (Jones, 2001). The early twentieth century showed positive anomalies, mainly concentrated in the North Atlantic and nearby regions, with a contrary cooling tendency prevailing in parts of North America, southern Eurasia, and the southern hemisphere.

The ongoing warming is qualitatively consistent with some climate model projections driven by anthropogenic forcing (Mitchell et al., 2001), and has a nearly global extent, although some oceanic areas and Antarctica (excluding the Antarctic Peninsula) are still cooling. Global temperatures – including land and ocean – have been rising at about 0.165°C per decade between 1977 and 2001. Trends are larger over land than over oceans and, as a consequence, the northern hemisphere warming (0.223°C per decade) is much larger than that observed for the southern hemisphere (0.106°C per decade). The highest regional year-round warming rates are found over Europe and the Arctic, and – on a seasonal basis – winter and spring over North America and Eurasia show the largest increase (Jones and Moberg, 2003), probably in relation to the enhanced westerly flow caused by the maintained positive phase of the North Atlantic/Arctic Oscillation.

The described warming trend places the 1990s decade as the warmest in the observational record, and probably the warmest (at least in the northern hemisphere) of the last 1800 years as derived from multi-proxy data reconstructions (Mann and Jones, 2003). This happened even with the short-lived cooling influence of the 1991 eruption of the Pinatubo volcano (Parker et al., 1996). The year 1998, affected by the strong 1997/1998 El Niño, is, up to this writing (late 2003),

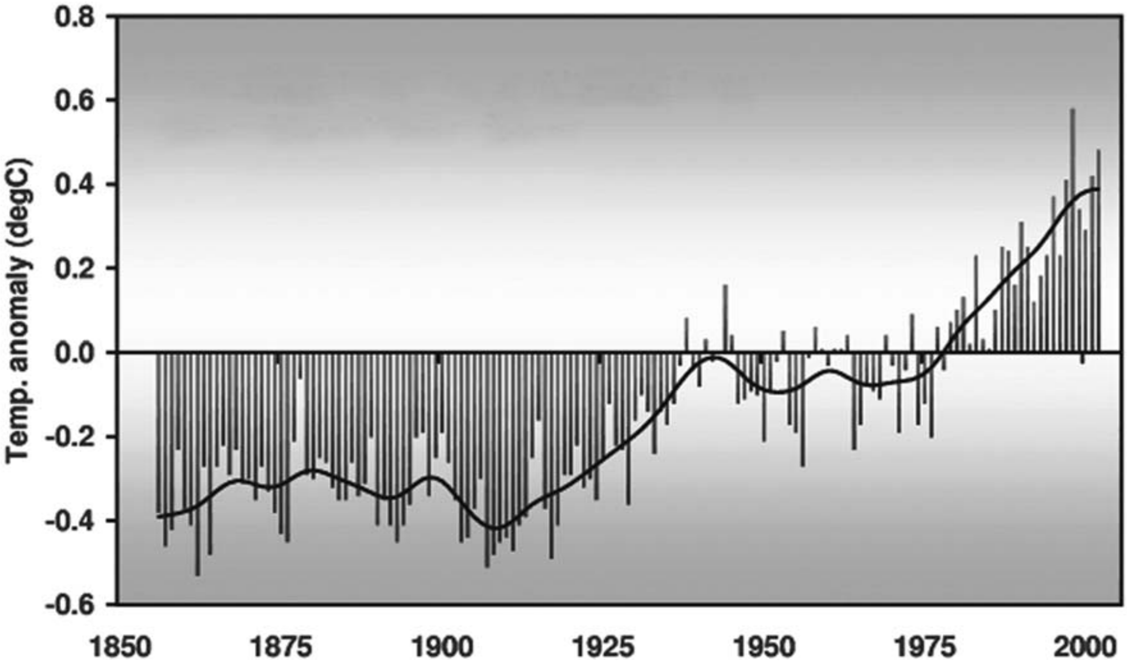

Figure C22 Global air temperature (HadCRUT2v), 2002 anomaly +0.48°C (second-warmest on record). Reproduced by kind permission of the Climatic Research Unit, University of East Anglia, Norwich, U.K., to whom any queries should be directed (cru@uea.ac.uk). Copyright: Climatic Research Unit.

the warmest year on record, with an anomaly of roughly + 0.5°C with respect to the 1961–1990 normal period. The El Niño – Southern Oscillation (ENSO) phenomenon has shown an increased occurrence, persistence, and intensity of its warm phase (El Niño) in recent decades, which enhances the global positive anomalies of the climate. The intensities of the 1982/1983 and the 1997/1998 El Niños were unprecedented in the observational record. The most recent event led to much warmer global temperatures, in spite of the offsetting effect of the 1982 El Chichon volcanic eruption, in part because of a warmer starting point in 1998. The initial decade of the twenty first century, continues with the warming trend started in the mid-1970s. The year 2002 was the second warmest in the instrumental record and the warmest for latitudes poleward of 30°N (Waple and Lawrimore 2003). If the current trend is maintained a future El Niño event will easily lead to a new record high global temperature.

During the second part of the twentieth century the DTR (diurnal temperature range or difference between daytime and nighttime temperature) has been reduced on a global basis by 0.1°C per decade. This happened, in agreement with an observed increase in global cloudiness, because three-quarters of the observed warming over land areas has been attributed to nighttime (usually daily minimum) temperatures and the remaining to daily maxima (Easterling et al., 1997). Nevertheless, several regions do not follow the planetary pattern – among them, eastern North America, middle Canada, portions of Europe, parts of Southern Africa, India, Nepal, Japan, New Zealand, the western tropical Pacific Islands, Australia, Antarctica and Spain display no change or slight increase of the day–night temperature differences (Karl et al., 1994; Horton, 1995; Brunet et al., 2001).

In the 50 years of available radiosonde data, upper air tropospheric temperatures (roughly the lowest 10–15 km of the atmosphere) have experienced warming, in agreement with the surface observations, and also in agreement (for thermodynamic reasons) with the observed stratospheric cooling. The extra-tropical northern hemispheric troposphere concentrated its warming since the mid-late 1970s, while the austral mid to high-latitude troposphere started warming earlier in the record (Lanzante et al., 2003).

Increases in global temperatures are very likely to be reflected in changes in precipitation and atmospheric moisture via induced changes in the atmospheric circulation; a more active hydrological cycle; increases in the water vapor holding capacity throughout the atmosphere (Folland et al., 2001); and changes in the leading modes of interannual variability, e.g. ENSO, the Arctic Oscillation, and the North Atlantic Oscillation. Different rain-gauge datasets (Peterson and Vose, 1997; Rudolf et al., 1999; New et al., 2001) and satellite measurements, available since the late 1970s (Xie and Arkin, 1996; Huffman et al., 1997; Doherty et al., 1999; Huffman et al., 2000) can be combined to produce global and regional estimates of changes in precipitation (New et al., 2001). The later authors reported a global precipitation increase of 0.89 mm per decade, mainly the result of an increase of 40 mm from 1901 to the mid-1950s, when global precipitation peaked. This trend was not globally uniform. Between the late nineteenth century and the mid-twentieth, precipitation in the tropics and along the east coasts of the continents decreased by roughly 10%. Some investigators believe this decrease occurred quite abruptly around the turn of the century. Pressure data point to a slight, gradual weakening of the general circulation during the same period, in the northern hemisphere at least, and to a contraction of the zonal westerly wind belt in temperate latitudes toward higher latitudes (analogous to its annual contraction from winter to summer).

The magnitude of the century-long trend (9 mm) appears to be modest when compared to the interdecadal and interannual variability of precipitation. Spatially, larger increases came from the mid and high northern hemisphere latitudes (40–80°N). The year 1998 was the wettest year on record for latitudes exceeding 55°N (Folland et al., 2001). Precipitation in the Northern Subtropics did not show significant trends and was rather characterized by strong subdecadal variability. Finally, areas between 20°N and 40°N experienced a decrease of 6.3 mm per decade, most of it after the mid-1950s. The southern hemisphere tropics and midlatitude precipitation remained steady in the last 100 years, when the austral subtropics showed an increase of 3.6 mm per decade (New et al., 2001).

Besides the analysis of the mean state of climate and its variability, it is crucial to understand the evolution of the extreme values. Extremes tend to have a notorious impact on natural and socioeconomic systems. Relatively modest changes in the mean state of climate can lead to large changes in the frequency of extremes (Katz, 1999). Despite a good degree of spatial differences, and the lack of data for significant parts of the southern hemisphere, it is safe to say that a significant proportion of the global land area was increasingly affected by significant changes in climatic extremes during the second half of the twentieth century. During this period, days with cold temperatures have diminished and, consequently, the number of frost days has been reduced; the frequency of extremely cold days has been reduced, and the number of days with extremely warm temperatures seems to be increasing (a result of increased variances), as well as the number of days with heavy rainfall and the total amount of precipitation coming from wet spells (Frich et al., 2002).

Causes of climatic variation

The immediate meteorological causes of observed local climatic variation involve changes of the general atmospheric circulation pattern. In this respect many regional anomalies in the world-average variation can be understood. However, it is unlikely that the net world-average variations of temperature and precipitation are traceable to circulation changes alone; rather, they are caused by more fundamental environmental changes that can affect both climate and circulation. Ultimate causes of climatic variation have not yet been identified with certainty, but they are thought to be several. The most likely among these fall into the first four of the following five main categories.

1. *Air–sea interaction*. Since the atmosphere is closely coupled with the oceans both dynamically and thermally, the relatively much slower rates of mixing and circulation of the oceans, together with their enormous heat-storage capacity, open up many possibilities for the oceans to stimulate long-term variations of climate. Air–sea interactions are, in fact, a probable source of much of the observed variability of climate on all scales of time from years to centuries to millennia. In general, the longer the time scale of variation involved, the deeper are the oceanic depths involved in the interaction. On interannual time scales a principal mode of air–sea interaction is the ENSO phenomenon, centered in the tropical Pacific Ocean, but whose effects on climate are of worldwide extent (Rasmusson and Wallace, 1983). The strong El Niño event of 1982–1983 was accompanied by extremely wet weather in parts of the Americas and, concurrently, by severe drought in parts of Australasia.
2. *Explosive volcanic activity*. In modern times, explosive volcanic eruptions like that of Krakatoa in 1883, violent enough to inject large amounts of fine ash and sulfurous gases into the midstratosphere (altitudes of 20–30 km), have occurred with an average frequency of several per century. Some such eruptions have produced veils of ash and sulfuric acid aerosols that have spread worldwide in the stratosphere with lifetimes of 2 or more years. Because such volcanic veils scatter and absorb significant amounts of incoming solar radiation while having little effect on outgoing terrestrial radiation, they are capable of cooling worldwide climate by a fraction of a degree Celsius over a period of years (Newell and Deepak, 1982). The warming of world climate during the 1920s and 1930s can be attributed in part to the absence of such volcanic eruptions at that time, and the cooling that culminated in the 1960s can be partly attributed to a renewal of eruptive activity that included the major eruption of Agung in 1963. The more recent major eruption of El Chichon in Mexico in the spring of 1982 was expected to result in a climatic cooling in the years following. This cooling, however, is not readily apparent in the temperature record (see Figure C22), a fact that may be explained by the contrary warming effect of the great El Niño event of 1982–1983 in which enormous quantities of heat were transferred to the atmosphere from the warmer-than-normal tropical Pacific Ocean.
3. *Changes of atmospheric composition*. Changes of the gaseous composition of the atmosphere may alter the terrestrial heat balance if the gases are selective absorbers of radiation such as water vapor, carbon dioxide (CO_2), and ozone (O_3). Carbon dioxide has accumulated in the atmosphere in the past two centuries, partly owing to vast deforestation for agriculture and fuel wood and sharply increasing use of fossil fuels. The net accumulation of CO_2 in the atmosphere has by now increased from preindustrial levels of about 280 parts per million by volume to levels of about 360 ppmv (as of 2003). Other trace gases of industrial origin, such as nitrous oxide and chlorofluorocarbons as well as methane, are also increasing. These other gases behave in the manner of CO_2, to cause a "greenhouse" warming of climate. Their combined effect on climate is thought to have been a warming since 1880 of about 0.6°C (0.9°F); further warming at an accelerating pace will likely continue into the future as long as these gases continue to accumulate as expected (IPCC, 2001). Changes of "greenhouse" gas concentrations to date are a likely cause of much of the net warming of climate observed in the past century (see Figure C22).
4. *Solar radiation changes*. Solar radiation may vary in several respects, including the total solar constant, its ultraviolet component, and solar wind intensity. Long-term changes of total solar constant, if real, would clearly affect climate but evidence of such changes other than those of a fraction of 1% in the past few years of direct satellite measurement is as yet unverified (Foukal, 2002). Possibly larger changes may have occurred over the 80–90-year Gleissberg cycle of solar activity or in the course of longer historical solar disturbances such as the Maunder Minimum, a time of unusual solar inactivity between about 1650 and 1710. Changes of ultraviolet emissions are known to parallel the 11-year solar cycle, and may influence surface climate indirectly via their effect on upper atmospheric conditions such as ozone amount. Changes of solar wind intensity may likewise be capable of altering the behavior of atmospheric circulation through subtle

effects on upper-atmospheric temperatures and winds and the "ducting" of Rossby-wave energy in the troposphere (Geophysics Study Committee, 1982). To date, however, such presumed solar–climate relationships lack adequate observational and theoretical confirmation.

5. *Impacts by large asteroids*. Though extremely rare (with time scales of hundreds of thousands to hundreds of millions of years), impacts by comets or asteroids have the potential to cause massive climatic changes via mechanisms similar to volcanic eruptions. Such impacts are thought to have resulted in a few mass extinctions in the geological record. No such event has ever been recorded in the instrumental record.

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CLIMATE VULNERABILITY

To be vulnerable is to be "susceptible of receiving wounds or physical injury" (*Oxford English Dictionary*). The phrase "climate vulnerability" raises an immediate question: what is it that might be injured? Climate might be affected by the actions of or changes in natural and social systems. Or natural and social systems experience impacts from changes in climate. Or the phrase may mean both of these interpretations.

Vulnerability of climate

Climate – the assemblage of atmospheric properties and behaviors, in the mean and with their inherent and characteristic variances – responds to the state, and changes of state, of the hydrosphere, cryosphere, and biosphere. Conversely, and reflexively, those natural systems are conditioned by the state of global climate. As a simple example, the poles are frozen due to the fundamentals of global climate; the extent of Arctic pack ice responds to episodes of hemispheric warming or cooling; hemispheric climate responds to the state of the surface of the Arctic basin. Most models of climate consider coupled natural systems, describing sensitivities and feedbacks, but not "vulnerabilities" in the colloquial sense of the term. In the case of social systems, where there are direct and indirect effects from alterations in atmospheric composition and of biosphere, it does make sense to think of climate being vulnerable. Much research in global climate change at the turn of the millennium focuses on these processes, patterns, and implications (see, for example, the IPCC Working Group I assessment; Houghton et al., 2001).

Strongest among the evidence that human activity can effect change in the climate system is the relationship of global temperature and "greenhouse gases". Concentrations of carbon dioxide, methane, nitrous oxides, and other gases have increased exponentially during the industrial era, i.e. since the beginning of the eighteenth century. Theory about atmospheric chemistry and radiative forcing of climate suggests that those changes should produce atmospheric warming, and models suggest a good fit of observed global surface temperatures with the records of gas concentrations. Models of likely socioeconomic futures, greenhouse gas emissions and resultant concentrations in the atmosphere, suggest that, even with marked reductions in greenhouse gas production, the atmosphere and thus the climate system will remain perturbed for a century or more (cf. the summary by Albritton and Meira Filho, 2001, and the more detailed reviews within Houghton et al., 2001).

The notion that human alteration of surface features at a regional scale might change the climate is not new (Brooks, 1970, originally published 1926). In late nineteenth-century America, for example, a popular idea was that "rain follows the plow"; conversion of the vast central grasslands to agriculture would increase precipitation of the area, to the benefit of agriculture (Lawson, 1974). In the latter half of the twentieth century, speculations about the effects of other major landscape alterations were made, often without aid of global circulation models: redirection of northward-flowing rivers in Asia; Amazon deforestation; expansion of desertification in Africa. Models are less successful at – or perhaps have paid less attention to – reproducing the effects of land surface changes on climate, perhaps because of the complexity of the climate system and because most attention has focused on atmospheric chemistry and the impacts of climate change on natural and human systems.

Vulnerability to climate

The dictionary definition cited at the head of this article implies a straightforward model of impacts: the characteristics of an entity (or system) are changed in a deleterious sense due to the action of outside forces. Application of the concept in natural sciences such as ecology led to notions of degree of vulnerability. Entities might, for example, not exhibit any evidence of impact until external actions (uniquely or cumulatively) surpassed some threshold beyond which the entity could not continue as it had been. If the entity returned to its original condition after an impact had occurred, it was said to be resilient. In some usage, resilience is the ability to withstand change, to persist. Thus, vulnerability and resilience are related concepts, although they are not – as might be the usage in colloquial speech – antonyms. More on this point follows in the paragraphs below.

The concept of vulnerability has broadened considerably in the work on climate change and adaptations that emerged toward the end of the twentieth century. In the Third Assessment Report of the IPCC, Working Group II specifically addressed climate impacts, adaptations and vulnerability (McCarthy et al., 2001). The 19 sectoral and regional chapters of this volume provide the most comprehensive review of the topic. The overview chapter (Schneider and Sarukhan, 2001), the final chapter that synthesizes vulnerability to climate change and reasons for concern (Smith et al., 2001), and the extensive bibliographies therein, are of particular relevance to this item. The other chapters and their bibliographies should not be ignored. Much of what follows in this article is drawn from this volume.

Definitions

The IPCC's working definition of climate vulnerability is "the degree to which a system is susceptible to, or unable to cope with, adverse effects of climate change, including climate variability and extremes" (McCarthy et al., 2001, p. 995). That is, vulnerability is "the extent to which a natural or social system is susceptible to sustaining damage from climate change" (Schneider and Sarukhan, 2001, p. 89). These definitions are notably more epansive than the classical concept stated at the outset of this article, incorporating fundamental ideas from earlier definitions in the hazards and development (poverty) literatures. Whereas the simple definition contained the idea of exposure to a hazard and resultant impacts, the IPCC's definition integrates hazard, exposure, consequences, and adaptive capacity. In this sense it is akin to the concept of "risk" in the risk-assessment literature.

What makes an entity or system vulnerable to climatic impact? Vulnerability is a function of both the climate change or variability to which the system is exposed, and the nature of the system in terms of its sensitivity to climatic control and its adaptive capacity. Exposure is a function of the climatic event or change – its character or direction, its magnitude, its rate of onset, and its duration.

Sensitivity is "the degree to which a system will respond to a given change in climate, including beneficial and harmful effects" (Schneider and Sarukhan, 2001, p. 89). Effects may be direct, such as crop yield responses to changes in temperature or precipitation; or they may be indirect, via intermediary actions, such as coastal flooding by sea-level rise due to climate change. Thus impacts might be sequential, nested, or hierarchical, and vulnerability may be immediate or delayed, and muted to lesser or greater degree (cf. Kates, 1986).

Adaptive capacity is "the ability of a system to adjust to climate change (including climate variability and extremes) to moderate potential damages, to take advantage of opportunities, or to cope with the consequences" (McCarthy et al., 2001, p. 982). Adaptations in social systems may be made by adjustments in practices (institutions), structures (infrastructure), or processes (mitigation). Many natural systems, or elements thereof, may be limited in adaptive capacity because of human alteration of the globe. For example, ecosystems will not adjust en masse, but as the suite of species responses; some species may be unable to adjust fast enough, or to find suitable habitat because of fragmentation and destruction of ecosystems that has already occurred.

"[A] highly vulnerable system would be a system that is very sensitive to modest changes in climate, where the sensitivity includes the potential for substantial harmful effects, and for which the ability to adapt is severely constrained" (Schneider and Sarukhan, 2001, p. 89). Many researchers believe "[s]ocial systems generally are more resilient than natural systems because of the potential for deliberate adaptation" (Schneider and Sarukhan, 2001, p. 91). Natural systems cannot access coping mechanisms or strategies beyond the autochthonous, whereas social systems can draw upon or invent purposeful adaptations. Synergies among purposeful adaptations may also yield additional coping capacity.

A system with low vulnerability is often termed resilient. Resilience results from adaptive measures, whether in reaction to actual impact or in anticipation of potential impact. The Resilience Alliance (2002) goes so far as to state that "[t]he antonym of resilience is often denoted vulnerability". Adger (2003) calls resilience "the ability to persist and the ability to adapt". A system's resilience is determined by: the magnitude of perturbations it can absorb without changing its overall function; the system's capacity for self-organization; and the system's capacity for learning and adapting (Adger, 2003, p. 2). It is not correct, however, to think of persistence merely as stasis, such that the system returns to the pre-existing state that prevailed prior to impact. Such was the classic systems view of the steady state. Rather, the current concept of resilience recognizes change as an essential characteristic, as the system adapts to new conditions, incorporating learning and adjustment.

Implications for assessment

The first generation of climate change assessments looked like an impacts model with (simple) feedback (see Kates, 1986, for impact and interaction models). Adaptation comes at the end of a causal chain between climate and impact, and is not an integral part of the system from the start. Such assessments, as in the Second Assessment Report of the IPCC (Watson et al., 1995), addressed the following questions: How will climate change? What first-order impacts will there be on natural systems? What will be the effects of those impacts on humans? How might human society adapt? How might the adaptations alter the impacts? (Leary, 2003).

Climate change assessments of the second generation, as the Third Assessment Report typifies, resemble adaptation models with synergies between natural and social systems (cf. Kates, 1986). Vulnerability and adaptation occupy a central place in these assessments. The latest generation of assessments builds upon the earlier in four major ways. First, vulnerability and adaptation to climate change are assessed: who is vulnerable, the source and nature of that vulnerability, and capacities for coping with current variability and extremes (and implications for the future). Second, stakeholders are integrated with scientists in the process: who needs to know what, and how can they be involved in production of credible and relevant information? Third, risk-assessment approaches are utilized. Fourth, scenarios of future environmental and socioeconomic conditions are integrated with climate change scenarios (Leary, 2003).

The emphasis on vulnerability permeates the Third Assessment Report to a much greater degree than previous reports. This focus can be seen by the attention paid to: adaptation to climate change; links between climate change, sustainable development, and equity; vulnerability to changes in climate variability; discontinuous responses to climate change; emphasis on ranges of outcomes using subjective probabilities; and other matters.

Vulnerability of agriculture

Agriculture is arguably the social system in which vulnerability might be easiest to define and earliest to be noted. Food production is of course important to existence and well-being, as well as livelihood and social structure. Agriculture exists at the interface of environmental and social systems, and forms a nexus with water resources not easily subdivided into components.

Agricultural production is, at base, affected by water and nutrient availability, and temperature. Human factors – genetic manipulation whether by traditional breeding or biotechnological engineering; artificial inputs; and so on – modify but do not replace these natural controls. Global temperature increases may allow cultivation in subpolar areas now inhospitable because of short growing seasons, if soil conditions in those areas are amenable. In other areas, however – tropics, drylands – heat or drought stress might increase, limiting productivity. Many of the least-developed countries are in the tropics, and have limited adaptive capacities for socioeconomic and political reasons. At moderate global warming (*c.* 1°C), subpolar and midlatitudes may benefit; above *c.* 2.5°C that productivity is modeled to diminish, and tropical decreases are more severe. Thus, it is possible to imagine an increased disparity in food production between developed and developing nations under scenarios of climate change for the twenty-first century (Schneider and Sarukhan, 2003; Smith et al., 2003).

Water resources are controlled to a great extent by precipitation and (temperature-related) evaporation. Climate scenarios suggest precipitation increases in high-latitude and some equatorial regions, which would decrease water stress, but decreases in many midlatitude, subtropical, and semiarid areas, which would increase water stress. These changes imply respective positive and negative impacts for rain-fed and irrigation agricultural productivity in these regions. With growing population, urbanization, and improving living standards, there will be increased demand for urban and municipal uses of water, thus increasing competition for scarce water at the same time that agricultural demand increases. There is marked disagreement in the literature about prospects for meeting food demands. Some see growing populations as necessitating great expansion of agricultural lands into marginal areas (and climate change may also require shifts in agricultural areas or make expansion infeasible). Others say intensification of activity, with technological improvements, will meet demand, satisfy growth in nutritional standards, and save land for ecosystem functions (but socioeconomic factors mitigating against agricultural lifestyles, growing urbanization, and competition for water may confound these prospects) (Smith et al., 2003).

Differential vulnerability

Some physical systems, by virtue of their size, location, or sensitivity to climate factors, are highly vulnerable to anticipated direction and magnitude of climate change. Among those so identified are: tropical glaciers, such as in the Andes Mountains of South America; and small lakes of interior drainage, such as in central Asia. Similarly, unique and threatened biological systems include: montane systems, where high-elevation species have nowhere higher to go; prairie wetlands and remnant native grasslands of North America, because of extensive alteration and fragmentation; coldwater fish habitat of northern latitudes; ecosystems overlying permafrost in circum-Arctic locations; ice-edge ecosystems of polar latitudes; neotropical cloud forests; coral reefs and mangroves ecosystems of tropical latitudes; and ecotones between existing ecosystems (Schneider and Sarukhan, 2001).

On a global scale some human systems are thought to be particularly vulnerable by virtue of poverty, isolation, size, or dependence on ecosystems of limited extent or diversity. Among those so identified are: small island states, especially atoll nations in the Pacific and Indian Oceans; and indigenous communities with low levels of technological development. Within developed nations the hazards literature indicates differential vulnerability between and within populations. Socially or economically disadvantaged groups – the young, the elderly, females, ethnic minorities, the poor, among other social strata – are found to be particularly at risk; that is, vulnerable. Further, vulnerability may be situational; that is, it may vary depending on the kind, character, or combination of hazards (Bohle et al., 1994; Buckle et al., 2000).

In general, "[s]ystems that are exposed to multiple pressures (synergistic effects) usually are more vulnerable to climate changes than systems that are not" (Schneider and Sarukhan, 2001, p. 91). Such pressures include current and projected demands on resources, unsustainable management, pollution, and fragmentation of natural areas.

Has climate made a difference?

There is a large literature of long standing about impacts of and adaptations to extreme weather events. Extreme weather, however, does not constitute climate, although scenarios of future climate suggest that present-day extremes may be more frequent (Albritton and Meira Filho, 2001). Assessments of vulnerability to recent climate variability are hampered by the limited range of climatic experience in the recent record. While responses to events such as those anticipated in the future – drought, say – may provide some indication of impacts and adaptations, we just have not had the magnitude and duration of changes that scenarios suggest might be in our future. However, the IPCC reviews suggested that many ecosystem changes that had been reported during the twentieth century were either associated with observed climatic changes, or in the directions expected as responses to directions of climate change (Schneider and Sarukhan, 2001; Smith et al., 2001).

Because climate change, system vulnerability, and adaptation involve changes in natural and social systems over time, as well as the synergistic and cumulative effects of multiple events, there is a need to aggregate over time and to accumulate over occurrences. There is a smaller body of literature that looks to the historical record for evidence of impacts of climate on society (Wigley et al., 1981; Rotberg and Rabb, 1981; Jones et al., 2001). Some papers in these sources concern themselves with the search for appropriate methodologies. How are we to know that social changes over historical time are the results of impacts by and adjustments to climate, apart from social, political, and economic developments? One methodological example is Parry's (1978) multi-stage "postdictive" approach: climate sensitivity of agricultural production was modeled by relating crop yields to various climatic factors; a history of climate was produced from the paleobotanical and other records, independent of the record of human activity; impacts on agriculture were postdicted by applying the sensitivity analysis to the climate record; the reconstructed record of impacts was interpreted in terms of implications for human settlement at the margins of sustainable agriculture; and the historical record of settlement was then compared to the postdicted impacts for verification.

The historical and archeological evidence suggests that societies, especially those at the margins of sustainability, have exhibited vulnerabilities to climate variability, yet have also learned to adapt in some ways (among the adaptations selected by pre-modern societies has been abandonment of inhospitable areas; cf. Burton et al., 1983, for the categorization of choices available to society as adaptations to hazards). More research using historical materials, with refined methods based on the above postdiction model, should combine with the more quantitatively detailed modeling of interactions from recent records, to produce a fuller understanding of vulnerabilities.

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Cross-references

Agroclimatology
Archeoclimatology
Climate Change Impacts: Potential Environmental and Societal Consequences
Desertification
Greenhouse Effect and Greenhouse Gases
Models, Climatic

CLIMATE ZONES

A climate zone is a world area or region distinguished from a neighbor by a major physical climatic characteristic that is of global scale. Climate zones are bounded by limits that parallel lines of latitude to form "belts" that mostly extend around the globe; the word "zone" is actually derived from the Greek word meaning belt, and it is from classical Greek times that the concept of zones is derived.

History of the zonal concept

Beginning in the sixth century BC, Greek scholars identified zones of the Earth based upon astronomical knowledge and position of the overhead sun as related to changes in length of daylight. These illumination zones were called *klimata* (a word that is the origin of our word climate) and differentiated (1) a torrid zone (where the noonday sun was never far from overhead), (2) a temperate zone, and (3) a frigid polar zone. It was in the fourth century BC that Aristotle identified the parallels that gave the identified zones actual boundaries. He named the Tropics of Cancer (23½°N), and Capricorn (23½°S), and the Arctic (66½°N) and Antarctic (66½°S) Circles as lines separating the zones. Aristotle, however, also adhered to the viewpoint that the Torrid Zone was too hot and the Polar Zone too frigid for human habitation.

Claudius Ptolemy (AD 100–170) made use of earlier Greek ideas and, as part of his extensive geographic writings, divided the Earth into seven zones based upon length of daylight.

Ptolemy's zones of the Earth remained in effect for many centuries. Although their utility was questioned by such Arabic scholars as Ibn Hauqal in the ninth century, and modified by Idrisi in the twelfth century, they still formed the basis of the highly popular work of the Dutch Renaissance geographer Bernhardus Varenius (1622–1650). He presented a system of geography in which a table of zones relating day length at the summer solstice to latitude was titled *climata,* climates to the English-speaking world.

Temperature zones

Edmond Halley was perhaps the first to suggest that heat rather than day-length hours provided a more meaningful way to identify climatic zones; in his 1686 work he came to the conclusion that winds were caused by solar heating. Halley's suggestion to identify zones by temperature could not be undertaken until actual temperature data for world areas became available. This did not occur until 1817 when Alexander von Humboldt used observed data to draw the first isothermal map. By 1848 H.W. Dove was able to prepare maps of average temperatures of individual months. A new zonality was born; that based upon heat rather than day length.

The first modern map of climatic zones was devised by A. Supan in 1879. The division was based upon mean annual temperature and the temperature of the warmest month (Table C13a). Shortly after, in 1884, Wladimir Köppen (whose climatic classification remains widely used today) gave a detailed listing of temperature zones based upon the number of months above or below selected thresholds. As Table C13b shows, this introduces the concept of duration of a given climatic factor, quite different from the use of mean annual values.

Numerous other authors suggested alternate temperature zones of the Earth, but all show a similarity in that hot,

Table C13 Temperature zones: selected examples

(a) Supan's 1879 thermal zones

	Annual mean temperature		Mean temperature of warmest month	
	°C	°F	°C	°F
Hot	Over 20	Over 68	Over 10	Over 50
Temperate	Under 20	Under 68	Over 10	Over 50
Cold	Under 20	Under 68	Under 10	Under 50

(b) Köppen's 1884 thermal zones

	Duration in months of critical temperatures		
	Above 20°C (68°F)	10–20°C (50–68°F)	Below 10°C (below 50°F)
Tropical	12	—	—
Subtropical	4–11	1–8	—
Temperate	Less than 4	4–12	Less than 4
Cold	—	1–4	8–11
Polar	—	—	12

(c) Herbertson's 1905 thermal zones and economic belts

Economic belt	Temperature limits (°C)
No Crops	<10
Very few crops	0–10
Temperate Climate Crops	10–20
Tropical Crops	<20

(d) Miller's 1951 thermal zones

	Number of months 43°F or warmer
Warm temperate	12
Cool temperate	6–12
Cold	3–5
Arctic	1–2
Polar	0

temperate and cold zones are identified. In some instances, as in the case of Herbertson's 1905 thermal regions, the identified zones were used to delineate other variables, including economic zones as shown in Table C13c. As part of the development of climatic zones, many new concepts were introduced. For example, in 1951 A.A. Miller used the idea of month–degree and accumulated temperature. This identifies the length of the growing season for broadleaf trees and extends the relationships between zonal climates and vegetation growth. The idea of relating climatic zones to vegetation zones was not new; Köppen, for example, had relied heavily upon the work of plant geographers in devising his climatic zones and classification.

While the identification of thermal zonation initially relied upon the new availability of data, it soon became apparent that the zone designated "temperate" contained some of the most extreme conditions on Earth, and was in fact highly "intemperate" in terms of temperatures. This led to a number of variations in the way in which zonal maps were prepared. For example, Figure C23, taken from a 1901 publication by Dryer, is based upon the following rationale: "The tropics and polar circles do not divide the face of the earth into zones of temperature but of insolation. The true temperature zones are bounded by isotherms. By drawing isotherms of 70° and 30° for January and July upon one map, we obtain a set of zones which are not shifting but fixed, and reveal in striking manner the temperature conditions of the globe." Despite attempts to provide a more meaningful delineation of the temperate zone, it is now more appropriate to call the area between the tropics and the Arctic and Antarctic circles the mid- or middle-latitude climatic zone when dealing with anything other than the Greek concept of temperate.

Circulation zones

The availability of actual climatic data, that permitted the construction of thermometric zones in the late nineteenth century, also permitted identification of other zonal concepts, especially those based upon atmospheric circulation. In 1735 Hadley had postulated a simple atmospheric circulation and later the general motion he identified became known as a Hadley Cell. The Hadley Cell was incorporated into models of global circulation by, for example, Maury (1855), Ferrell (1856 and 1889), and Thompson (1857). While in such representations the use of the term "zone" is limited, the term "belt" is used and, essentially, each of the other circulation characteristics is a zone; thus, the Horse Latitude calms, the Trade Wind zone, and the Westerly Wind zone are zonal concepts based upon wind patterns.

The simplicity of the zonal concept of circulation, while attractive, had great limitations. During the twentieth century, meteorologists and climatologists have striven for more realism in model construction and the broad zones identified by the three-cell zonal concept proved unacceptable. New ideas, in-depth studies and, in some cases, actual hostility to the zonal concept demoted its representation to highly general introductory books about the atmosphere. In meteorology today, the term "zonal" invariably implies a zonal circulation type in

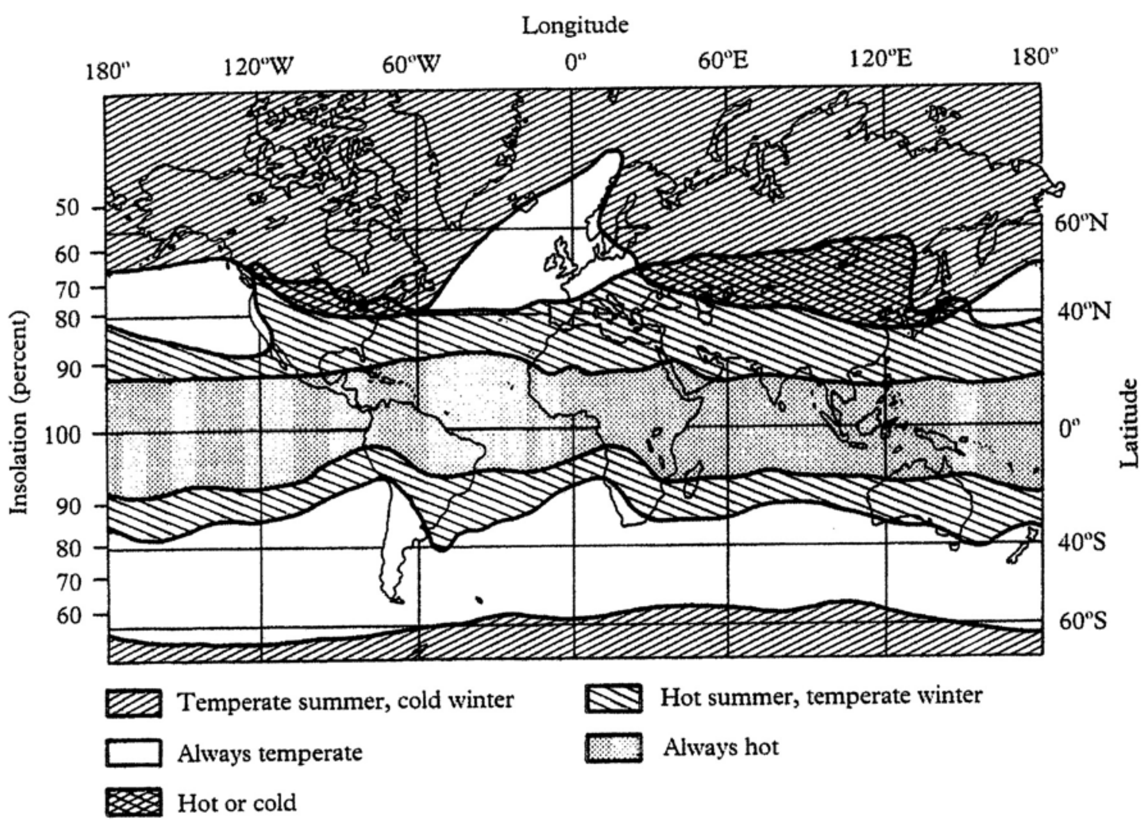

Figure C23 Redrawn from a 1901 map by Dryer, the map represents an earlier attempt to provide meaningful limits to temperature zones using winter and summer isotherms.

Table C14 de Candolle's (1874) plant zones

Symbol	Physiologic plant group	Temperature limit (°F)
*	Megistotherms	Over 86
A	Megatherms	68–86
B[a]	Xerophiles	
C	Mesotherms	59–68
D	Microtherms	32–59
E	Hekistotherms	Below 32

* Found only in high temperatures of geologic past or in hot springs.
[a] Plants adapted to dry conditions.

which flows, especially those in the upper air westerlies, were essentially in an east–west pattern. Nonzonal flow, or meridional, indicates a more significant north–south component.

Related zonal systems

The nineteenth-century development of climatology was strongly influenced by botanists. Köppen's boundaries for climatic zones were initially based upon vegetation zones identified by de Candolle, who had published his work in 1874. As Table C14 indicates, his physiologic plant groups are based upon the same temperature limits as those used by Köppen in his classification of climates.

The concept of vegetation zonation was, at least in part, emphasized in the monumental work of the German botanist Schimper, published in 1898. Schimper noted that certain plant orders and families are limited by high or low temperature limits, as found in the tropics and polar realms. As a result, the plant cover of the Earth is arranged in more or less parallel zones of different systematic characteristics. He further noted, however, that temperature zones cannot be equated to floristic and ecological characteristics, and these are identified regionally rather than zonally.

A similar zonal relationship exists in relation to the identification of soils. The Russian soil scientist Dokuchaev is credited, in a 1898 publication, with identifying soil zones of the northern hemisphere. These are (a) the Boreal or Tundra zone, (b) the Taiga or Forest Zone, (c) the Black-Earth Zone, (d) the Dry Subtropical Zone and (e) the Red-earth or Laterite Zone. This organization of hemispheric soil types resulted in a much-used soil classification based upon climatic processes. It is only in recent years, with the production of the soil taxonomy system, that it has been replaced.

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Cross-references

Atmospheric Circulation, Global
Climatic Classification
Climatology, History of
Doldrums
Horse Latitudes
Temperature Distribution
Zonal Index

CLIMATOLOGY

Climatology is the science of climate, which in turn is the composite of all weather events. It should be noted at the outset that climate fluctuates on all time scales: monthly, yearly, decadally, centennially, and millennially. Thus, climate is a statistical collective. It has often been described in terms of mean values of particular climatic elements, but it encompasses a wide range of values, including occasional extremes. Although atmospheric conditions and variations follow well-known physical laws, the day-to-day variations and those for longer intervals can, for many purposes, be treated like quasirandom variables. They can be represented by statistical distributions applicable to stochastic universes.

Climatology has multiple aims. In its historical development first efforts were directed toward a geographical inventory of climates, which led to attempts at climate classification. As a physical science, developments led to theories of the causation of global as well as regional and local climates. Because of the influence of climate on human activities, much effort has been devoted to applying the knowledge gained about climate for human welfare, including attempts to develop schemes for prediction of future climatic conditions. A major motivation for this has been to assess the potential for anthropogenic alterations of climate.

Areas of study

The wide range of the physical controls and the impacts of climate have led to the development of specialized realms of study in climatology. The major realms are considered here.

Physical climatology

The incorporation of physical principles in explaining climate is examined under the title of physical climatology, a subject area that deals with mass and energy exchanges at or close to the surface of the Earth. The focus is on energy and water balances and budgets and upon boundary layer studies.

Energy budget studies concern the availability of solar energy at a location, the transformations of that energy and its transfer through the atmosphere. Because the amount of solar radiation striking the Earth varies over both time and space, and because the Earth's surface is made up of materials of varying properties, then the physical climatology of any particular surface is unique. The patterns and exchanges are complicated by the special physical attributes of water and the exchanges of energy when water changes from one form to another.

These energy and water exchanges are dealt with in terms of boundary layer studies. Differing surfaces of the Earth result in highly variable boundary characteristics and hence different types of energy/mass exchanges and eventually different climates. Studies concern natural surfaces as well as those modified through human activities. The most modified type of surface occurs in cities; as a result, urban climatology has become an important component of the study of physical climatology.

Dynamic and synoptic climatology

The term dynamic climatology was first used by Tor Bergeron in 1929 when he outlined how the concept of airmasses and fronts might be adapted to develop a comprehensive dynamic climatology. Since then, the approach used by dynamic climatologists has evolved into studies based upon atmospheric motion characteristics and the thermodynamic processes that produce them. Of major importance today is modeling the dynamics of selected systems.

Most modeling is used to study the reaction of climate to forced changes in various physical parameters; these are the sensitivity studies. The most popular of these at the present time are models related to the increase of greenhouse gases and the attendant change of global temperatures. At the same time important studies in dynamic climatology concern such topics as sea surface temperatures and the role of mountains in determining dynamics of the atmosphere.

Synoptic climatology may be defined as the study of climate from the viewpoint of atmospheric circulation with emphasis upon the connections between circulation patterns and climatic differences. The term was first used in the 1940s when military services became concerned with the climatology of weather types and surface/air transport. While dynamic climatology is often global in scope, synoptic largely concerns hemispheric and local climatologies. In relation to the former, hemispheric teleconnections, such as ENSO events, have been investigated; at the local scale many types of studies, such as the relationships between circulation features and severe weather, are of interest.

Regional climatology

As its name implies, this concerns the climate of regions. The size of the region can vary but is often synonymous with identified geographic regions; the Great Plains, the Paris Basin, Southern China, providing apt examples. The manner in which the climatology is presented can range from highly descriptive, such as in the classic climatography of the early twentieth century, or can be based purely upon derived climatic indices, such as in component analysis.

To compare the climate of regions and to derive analog climates, regional climatologies often use a climatic classification. Earliest classifications were often based upon a single variable, such as temperature or daylight hours. By the nineteenth century increased data availability permitted a more quantitative approach and the use of more than one variable. Vegetation distribution strongly influenced the development of climatic regions in classification systems, with that devised by Wladimir Köppen proving the most durable. The Köppen approach provides a unique shorthand system based upon numeric rules for classifying climates, and the system remains in wide use today. Many other climatic classification systems have been devised, most for special-purpose uses such as human comfort or crop distribution. Such systems, based upon observed data, are termed empiric classifications. Regionalization by cause is dealt with in genetic classification systems, often based upon airmass climatology. To demonstrate the nature of classification, Table C15 shows a 13-category scheme with some prototypes shown as examples.

Applied climatology

Applications of climatology have risen rapidly in recent years. Aside from agriculture, the construction industry makes major use of climatic data. Temperature, wind, and humidity statistics permit rational design of buildings and dwellings, as well as their heating and cooling plants. Siting of factories and power plants requires analysis of wind and vertical atmospheric stability parameters to judge the pollution potential of effluents. In the case of nuclear power plants, climatological analysis is mandated. For major structures such as bridges, broadcast towers, and electric transmission towers, data on extreme wind speeds, low-level icing, and exposure to lightning are needed, although they must often be inferred from locations other than the building site. The transference of climatic information from an observing site to a "silent" area has become an important part of applied climatology. A combination of the use of interpolated data from a climatological observing network and site-specific interpretations from a series of synoptic weather maps is an applicable technique.

Climatological information also enters into the housing problems of farm animals. Climate affects the design of clothing and its seasonal distribution and marketing. The effect of climatic conditions has had a major impact on transportation and packaging and storage of goods, especially foodstuffs.

A major role for combined climatic–synoptic analysis is in the planning of dam construction, both for water-resource conservation and for flood control. Floods still remain a major hazard and their probability of occurrence has entered into calculations for setting insurance premium rates. Estimates of extreme values of precipitation leading to floods, as well as for extreme values of other damaging weather events, has resulted in the use of a variety of statistical techniques. These are attempts to simulate the frequency distribution of past rare events and high values. There are a number of such extreme value distributions. Such extreme value probability analyses can be used for predictions without date. They can inform a planner or construction engineer as to what extreme value of a particular climatic element can be expected at least once in 50 or 100 years. Such information, while not indicating when this value will occur, will at least permit the necessary safety factors to be included in the design. Aside from wind speeds, maximum precipitation amounts, and extreme temperatures, this type of analysis is also particularly useful for estimates of high snow loads, especially on places of public assembly.

The causation of climate

In regard to the causation of climate, the principal factor is the astronomical position of the Earth with respect to the sun, which is the source of all energy. The distance between the Earth and the sun governs the total energy received from the sun. There are seasonal differences in that distance and

Table C15 Basic climatic elements at selected stations

Climatic category	Location	Temperature (°C) W	C	Y	Mean precipitation (cm) J	F	M	A	M	J	J	A	S	O	N	D	Y
Glaciated	Ice cap 77°56′W Greenland	−7	−35	−23													<10
	South Pole, Antarctica	−28	−59	−49													<15
	Barrow, Alaska	4.3	−27.7	−12.2	0.41	0.38	0.31	0.25	0.33	0.71	2.11	2.03	1.40	1.32	0.69	0.51	10.4
Tundra	Ostrov Dikson, USSR	4.8	−27.5	−12.3	2.0	1.3	1.7	0.9	1.1	2.3	3.2	4.6	4.2	2.1	1.4	1.8	26.6
	Nome, Alaska	9.7	−15.3	−3.3	2.6	2.4	2.2	2.0	1.8	2.4	5.8	9.7	6.8	4.3	2.9	2.5	45.4
Short summer with frost	Salekhard, USSR	13.8	−24.4	−6.7	2.4	2.0	2.4	3.2	3.9	5.1	5.7	5.7	5.4	4.6	3.1	2.9	46.4
	Bismarck, North Dakota	22.3	−12.8	5.4	1.1	1.1	2.0	3.1	5.0	8.6	5.6	4.4	3.0	2.2	1.5	0.9	38.5
Cold winter, adequate precipitation	Moscow, USSR	17.8	−10.3	3.6	3.1	2.8	3.3	3.5	5.2	6.7	7.4	7.4	5.8	5.1	3.6	3.6	57.5
	Cincinnati, Ohio	24.9	0.9	12.9	9.3	7.1	9.9	9.2	9.7	10.6	9.1	8.3	6.9	5.7	7.5	7.0	100.3
Distinct winter, moist	Lyon, France	20.7	2.1	11.4	5.2	4.6	5.3	5.6	6.9	8.5	5.6	8.9	9.3	7.7	8.0	5.7	81.3
	San Francisco, California	16.7	10.4	13.8	11.6	9.3	7.4	3.7	1.6	0.4	0.1	0.1	0.6	2.3	5.1	10.8	52.9
Mild, winter rain	Rome, Italy	24.7	6.9	15.6	7.6	8.8	7.7	7.2	6.3	4.8	1.4	2.1	7.0	12.8	11.6	10.6	88.1
	Elko, Nevada	20.9	−5.2	7.4	2.9	2.3	2.1	2.1	2.4	1.8	1.0	0.8	0.9	1.9	2.3	2.6	23.1
Cool, arid	Semipalatinsk, USSR	22.1	−16.2	3.1	1.4	1.5	1.7	1.9	2.2	3.0	3.2	2.3	2.1	2.2	2.7	2.2	26.4
	Singapore, Malaya	24.7	22.9	23.9	25.2	17.5	20.0	19.6	17.4	17.1	16.7	19.1	17.9	20.8	25.1	26.6	243.0
Hot, rainy	Mlänje, Zambia	24.5	17.5	22.0	31.3	31.9	30.8	12.5	5.9	6.4	3.8	3.3	3.0	5.8	18.3	26.8	179.6
	Barra do Corda, Brazil	27.4	24.2	25.8	19.0	20.8	21.4	14.4	6.0	1.6	0.7	0.7	2.3	4.1	7.0	11.7	109.7
Hot, seasonal rain	Kinshasa, Zaire	27.0	17.0	26.5	32.1	13.9	18.1	20.9	13.4	0.5	0.0	0.4	3.3	13.7	23.6	17.1	137.8
	Salta, Argentina	21.5	10.6	16.8	17.6	14.9	9.4	2.5	0.6	0.3	0.2	0.4	0.5	2.5	6.1	12.1	67.1
Warm, seasonal rain	Windhoek, Namibia	23.5	13.0	19.0	7.7	7.3	8.1	3.8	0.6	0.1	0.1	0.0	0.1	1.2	3.3	4.7	37.0
	Guaymas, Mexico	31.2	17.9	25.0	1.2	0.7	0.3	0.1	0.0	0.3	3.9	6.0	5.1	2.2	0.7	1.7	22.1
Warm, semiarid	Karachi, Pakistan	31.4	17.7	25.9	0.7	1.3	0.5	0.2	0.1	0.9	10.1	4.7	2.3	0.8	0.3	0.6	22.1
	Yuma, Arizona	35.3	13.0	23.2	1.0	0.7	0.5	0.3	0.1	0.0	0.5	1.1	0.6	0.7	0.6	0.9	6.8
Warm desert	Cairo, Egypt	29.0	14.0	22.0	0.4	0.3	0.5	0.2	0.2	0.1	0.0	0.1	0.1	0.2	0.5	0.1	2.7
	Mount Washington, New Hampshire	9.3	−15.0	4.6	13.0	16.6	14.2	13.9	14.8	16.5	17.2	18.3	16.2	15.5	19.5	17.9	193.5
Mountain type	Sonnblick, Austria	1.6	−13.2	−6.0	11.5	10.8	11.2	15.3	13.6	14.2	15.4	13.4	10.4	11.8	10.8	11.1	149.5

Notes: W = mean temperature of warmest month; C = mean temperature of coldest month; Y = mean temperature of the year.
* Precipitation not measurable because of drifting snow.

long-term changes in the eccentricity of the Earth's orbit. Of equal importance is the inclination of the Earth's axis with respect to the ecliptic. This too is subject to cyclical changes; it is presently 23.5°. This inclination causes the seasons, with the poles during the respective summer seasons of the hemispheres inclined toward the sun and away from the sun in winter. Thus the polar regions change from complete daylight to complete darkness during the course of the yearly revolution of the Earth around the sun. The axis tilt therefore governs the amount of solar radiation as the seasons change. This fact was well known in ancient times and is reflected in the Greek word *klima*, which means slope.

The cyclical elements in the astronomical position of the Earth – namely, the *precession* of the spring point (~21 000 years), the tilt of the Earth's axis (~41 000 years), and the change of the *eccentricity* of the Earth's path (~97 000 years) – lead to different radiation conditions at various latitudes. Determination of these elements has led to formulation of a model of climatic changes by Milutin Milankovitch. The model permits calculation of radiation energy received as a time series prior to the present. For the past half million years, during which continents and oceans have been in their present position, the radiation curve coincides with the glaciations and interglacials derived from geologic and isotopic evidence. Whether or not the model is adequate to explain the initiation of glaciation remains controversial. Evidence for changes in solar energy output exists, but quantitative measurements have been available only in recent years from satellites; prior to that, sunspot observations were used as corollaries to climatic fluctuations since the eighteenth century, but statistical tests indicate that less than 10% of the variance can be accounted for, both for temperature and precipitation.

If the sun ceased radiating energy to the Earth, it has been estimated that atmospheric motions would cease in about 2 weeks. Satellite determinations of the energy received from the sun at the boundary of the atmosphere gives values between ~1360 and ~1378 $\mathrm{W\,m^{-2}}$. Partitioned over a spherical surface this fixes the solar energy amount entering the atmosphere at ~340–344 $\mathrm{W\,m^{-2}}$.

As solar radiation enters the atmosphere and penetrates to the surface, a complex system of interaction comes into play. Atmospheric gases, clouds, and conditions of the surface partition the solar radiative flux. At the surface reflectivity plays a key role; it varies from the high reflectivity of areas covered with snow and ice to intermediate reflectivity (albedo) of desert and vegetated land areas, and low albedos of water surfaces. The high albedo of the frozen surfaces contributes to their maintenance. The low albedo of water leads to large heat absorption in the oceans, which become the principal secondary energy source of the atmosphere. These effects have led to the designation of climate as an atmospheric–oceanic–cryospheric system. In the energy exchanges the interception of outgoing long-wave radiation from the Earth by atmospheric gases, principally water vapor and carbon dioxide, keeps the Earth at temperatures favorable for biota. The depletion of the solar radiation in various spectral regions is shown in Figure C24, which also shows the wavelengths in which radiation escapes from the Earth to space. The atmospheric gases are nearly opaque to the longwave infrared radiation from the Earth, except for a few "windows" in the spectral absorption of water vapor and carbon dioxide. It is these few escaping infrared fluxes that make satellite scanning of atmospheric and surface temperatures possible.

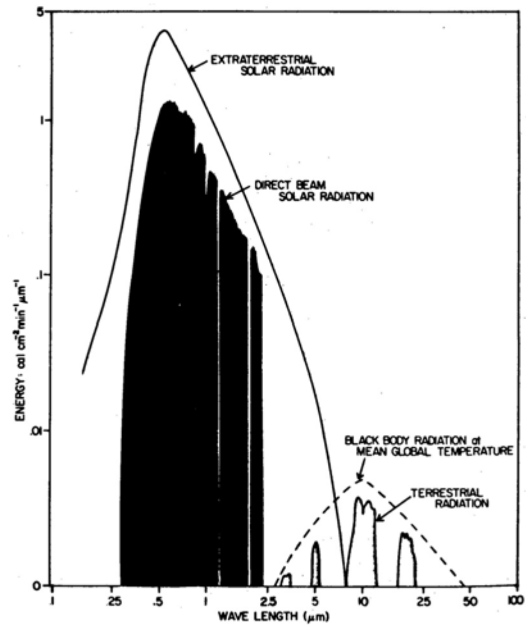

Figure C24 Spectral distribution of solar beam (left); shaded area showing flux reaching the Earth's surface, depletion caused by absorbing gases; terrestrial radiation to space (right); shaded areas showing the only fluxes escaping through the so-called atmospheric "windows" in the water vapor and carbon dioxide absorption bands.

A relative partitioning of the various radiative fluxes of Earth and atmosphere (in percent of the incoming solar radiation) are shown in Figure C25. This diagram also shows clearly that more radiative energy transactions take place in the atmosphere than at the surface. Radiation is the primary element in the climate-forming processes. For this reason energy balance models have gained a dominant place in modern climatology. They sum up all energy fluxes of an area or locality and incorporate incoming solar radiation, albedo, longwave outgoing radiation, as well as the terrestrial processes of evaporation, which adds latent heat to the atmosphere, and sensible heat transfers. On a hemispheric scale, some of these components are shown in Figure C26 for hemispheric summer and winter as zonal averages. High-energy incomes are indicated for summer, but no solar energy is received in polar regions during winter. The net energy income is positive for the whole summer hemisphere but negative for most extratropical regions in winter. The latent heat values stay at the high level throughout, and hence play a role as a major heat source in the atmosphere.

Global circulation

Satellite observations have given better insight into the net radiation balance of various parts of the globe. Not only the areas of high latitudes show net heat loss, but large areas at high altitude in Asia as well. Some of the North African and Arabian desert regions show radiation balance deficits. The tropics and major ocean areas are the areas of positive

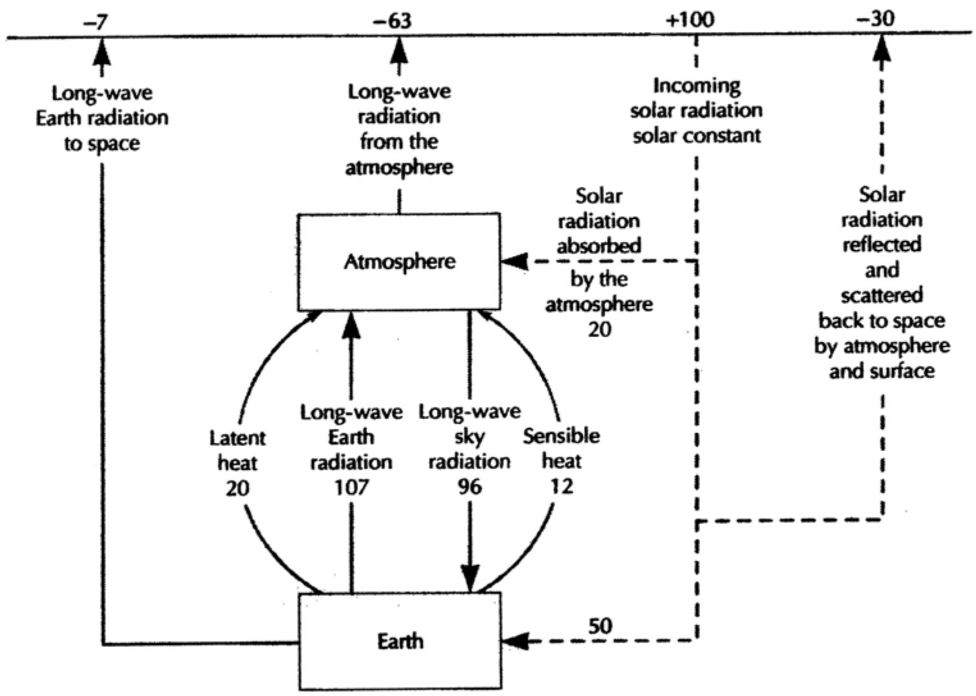

Figure C25 Radiative fluxes in and through the Earth's atmosphere, in percent of the solar radiation received at the boundary of the atmosphere.

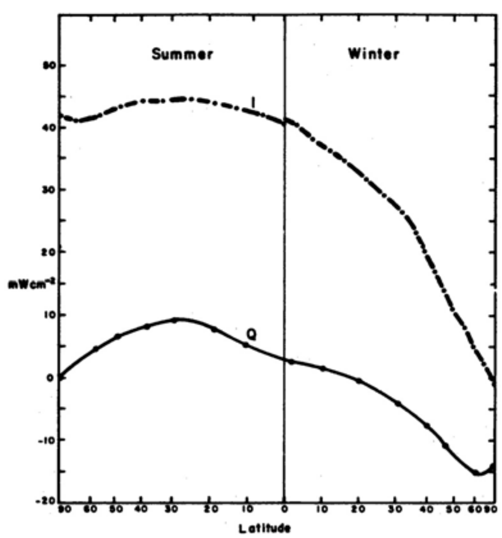

Figure C26 Zonal distribution of incoming solar radiation flux I and net radiation (gain or loss) Q, for hemispheric summer and winter.

radiation balance. These differences are the driving force of the global circulation in atmosphere and oceans. Changing air pressures transport heat from the areas of surplus to those of deficit and try to smooth out the differences. Radiation imbalance is the principal ingredient of climate. About half of the heat transfer from low to high latitudes takes place in the ocean. Oceanic currents are very important in climatogenesis of many areas of the globe. In the atmosphere the heat contrasts lead to a multicellular circulation, which follows the principles of fluid dynamics as modified by the rotation of the Earth. This leads to the existence of three meridional circulation cells in each hemisphere, causing different dynamic processes. Between the equator and the tropics, or subtropics, is a direct circulation of air from warmer to cooler areas, the so-called Hadley cell, with low pressure near the equator and high pressure in the subtropics. The resulting wind system produces the trade winds over the oceans, northeasterly in the northern hemisphere, southeasterly in the southern hemisphere. Near the equator in an area of weak winds – the doldrums – the trade winds create a convergence zone with cloud bands and convective precipitation. This intertropical convergence zone migrates somewhat, following the seasonal changes.

In the polar regions another direct circulation cell exists. The cold air forming there breaks out in irregular bursts toward the warmer areas. This is delineated by a notable discontinuity, the polar front, toward the lower latitudes. Its thermodynamic interaction with the warmer air forming in the lower latitudes, usually generously endowed with latent heat from evaporated water vapor, causes migrating pressure systems in an indirect cell, the Ferrel cell. These bring various midlatitude areas under rapidly changing weather regimes with alternations of warm and cold airmasses and wet and dry spells. A schematic representation of the atmospheric circulation is shown in Figure C27, which shows another globally important circulation that is zonal in character. This is imposed over the other circulations in the tropical area and is rather weak and slow, but it has important temporal effects on both atmospheric and oceanic currents. This so-called Walker circulation affects the intertropical convergence zone and the monsoonal circulations.

Although the general atmospheric and oceanic circulations affect the broad-scale distributions of climate over the globe, the fine fabric is created by the differentiation of the surface. Mountains and their orientation with respect to the general circulation are extremely influential. Windward and leeward sides have radically different climates. Proximity to water surfaces, both oceanic and inland, are other factors causing climatic differences. The presence or absence of vegetation and the land use by humans also cause notable climatic differences. Anthropogenic effluents are the cause of changes in atmospheric composition. Their local influences have been explored for decades, but there are potential global repercussions if the concentration of heat-absorbing gases becomes sufficiently large to cause measurable alterations of climate. An uncertain, sporadic factor in the climatic conditions of individual years, and perhaps longer intervals, are volcanic eruptions. These can cause changes in the radiation balance of the globe. Even a limited analysis of climatogenic factors shows so many feedback mechanisms that one can begin to visualize the diversified mosaic of climate in space and time. In Figure C28 some of the various elements of climatogenesis and some of the mutual interactions are reflected. The many agents having effects on climate make their exact representation in climatic models difficult, and hence these remain currently inadequate to represent the rich fabric of terrestrial climates.

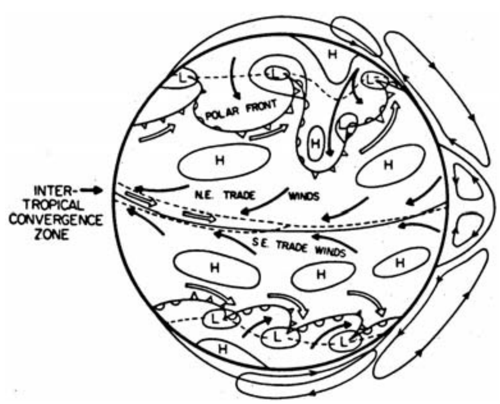

Figure C27 Schematic representation of the global atmospheric circulation showing principal frontal systems, high (H) and low (L) pressure areas and wind systems.

Global patterns

Systematic instrumental observations of the surface climate of the Earth have been carried on in most land areas for over a century, and in some localities for two centuries. In central England over 300 years of temperature observations are available. Time series of the various climatic elements such as pressure, temperature, precipitation, sunshine, cloudiness and wind show notable fluctuations from year to year. Even longer intervals can show marked differences, yet the mean value of

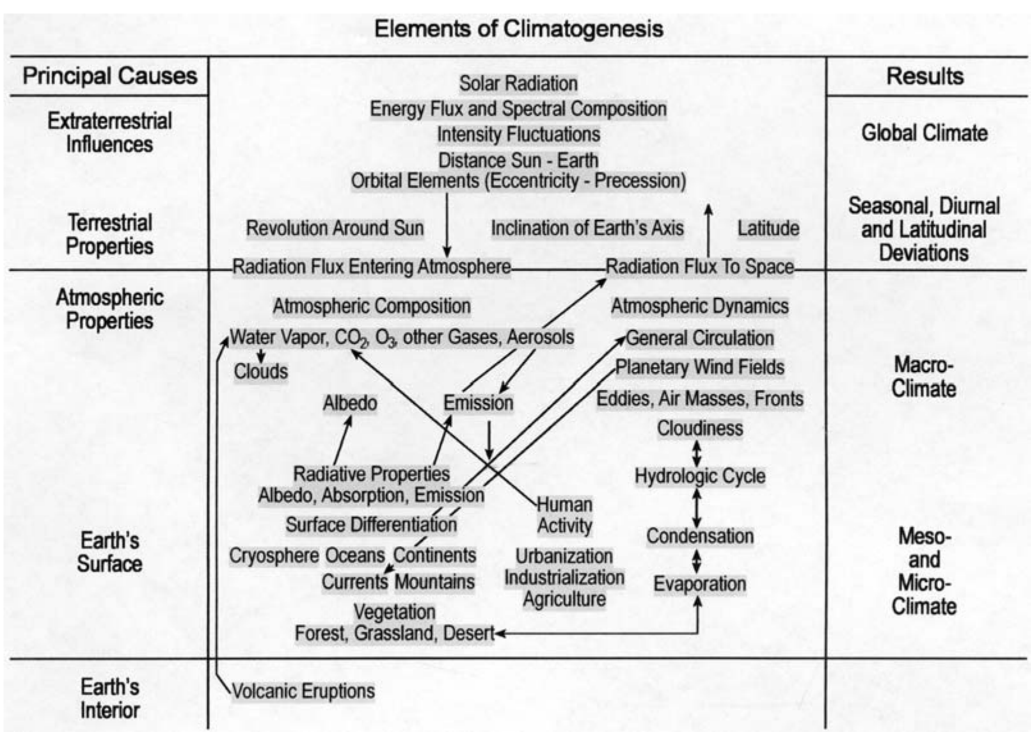

Figure C28 Principal climatogenetic factors on Earth with limited indications of feedback mechanisms.

the various elements has traditionally been used to represent climate. Although this has shortcomings, it permits broad-scale comparisons of geographical areas. The mean pressure distribution, as shown for January and July in Figures C29 and C30, shows the great differences in flow patterns between the two hemispheres caused by the greatly different distribution of land and sea. In January the two great centers of action are the Aleutian and the Icelandic low-pressure systems. These reflect the steady stream of cyclonic disturbances along the polar front bringing alternating outbreaks of cold air and intrusions of warmer subtropical air, generally accompanied by precipitation, to the middle northern latitudes. The sunny subtropical high-pressure cells have migrated southward, as has the intertropical convergence zone, which is now south of the equator. The strong Asiatic anticyclone not only coincides with intense Siberian cold, but its outflow controls the dry winter monsoon in south and east Asia. In the southern hemisphere there are some vestiges of heat lows over the continents and some inflow of moist air from the oceans, but in the great southern west wind belt the winds keep roaring at the 40th parallel.

In July the southern hemisphere has not changed much. The subtropical high-pressure cells have strengthened. Sunshine, and often drought, prevails over the continents. The Intertropical Convergence Zone has migrated north of the equator. The oceans show higher pressure than the continents and oceanic air influx to the land brings monsoonal rains. The classical monsoonal region is east Asia, especially the Indian subcontinent. Even though there are large year-to-year fluctuations, the broad general patterns have persisted.

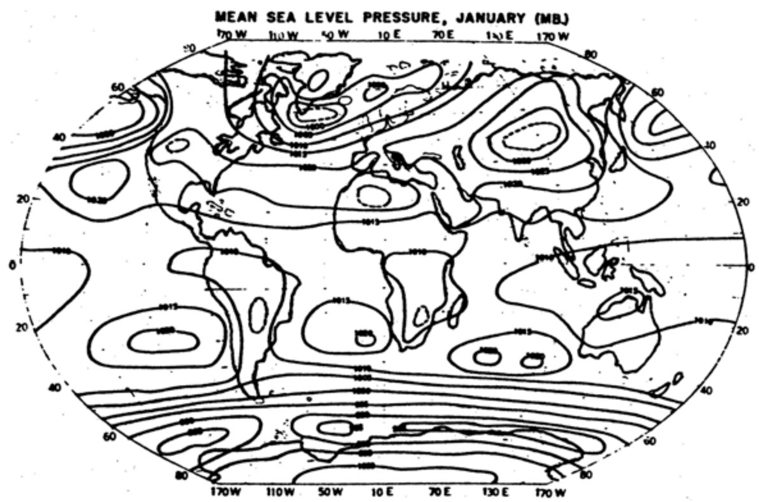

Figure C29 Mean sea-level pressure in January, in millibars (hPa).

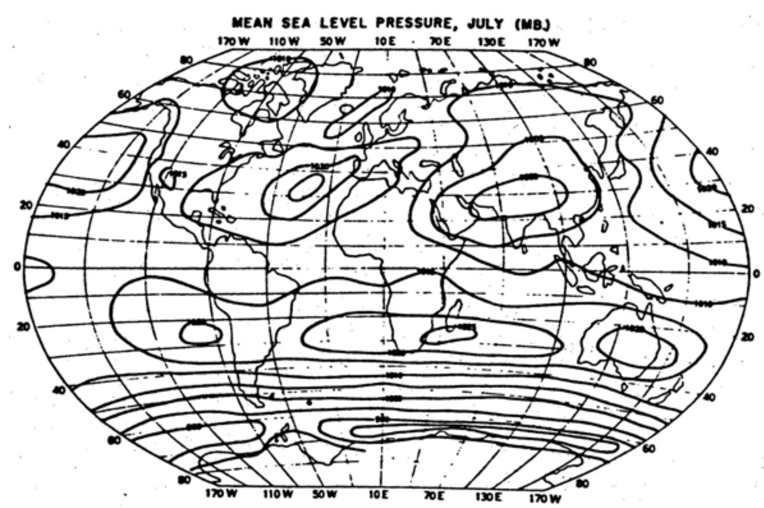

Figure C30 Mean sea-level pressure in July, in millibars (hPa).

These circulation patterns are reflected in the meridional distribution of cloudiness and precipitation (Figure C31). The cloudiness shows three major belts: the subpolar latitudes in both hemispheres and the equatorial zone. The precipitation amounts are somewhat less strikingly distributed latitudinally, but the zone just north of the equator shows a notable maximum. Equally pronounced are the subtropical dry zones where the prevalent high-pressure belts produced by the subsiding branches of the Hadley circulation create vast desert areas, especially in the northern hemisphere. But within the prevailing structure of global circulation, numerous other controls shape and modify the climate. Among these are proximity to the ocean or other large water bodies, and position with respect to mountain ranges. Oceanic currents exercise profound influences. Thus the North Atlantic Gulf stream and its continuation, the North Atlantic Drift, create a notable asymmetry of latitudinal temperature distribution between the east coast of North America and west coast of Europe. Whereas along the American Atlantic shore there is a wide annual range of temperature and cold winters prevail, the west coast of Europe far north into Scandinavia has a much smaller range and mild winters.

Other coastal climates are equally influenced by oceanic currents. Some of them are tied to the broad flow patterns of the general circulation, others are produced by more local air flow. The interaction of wind and currents can produce coastal upwelling of cold water from greater depths at sea. Such coastal cold waters not only have ecological benefits for fisheries, but also modify the regional climate. Cold currents cause coastal dry zones or even deserts and often cause high fog frequency. In perspective, the arid and semiarid areas of the globe cover 48×10^6 km^2 or 31% of the land areas, excluding the Antarctic continent. If one adds the 12% glaciated areas, which usually have very little precipitation, to the 31% of arid areas, the great contrasts of climate on a planet covered 70% by ocean becomes clear.

Precipitation is the climatic element that shows the greatest variability in time and space. Some extremes, together with those of other elements, are shown in Table C16. As the time

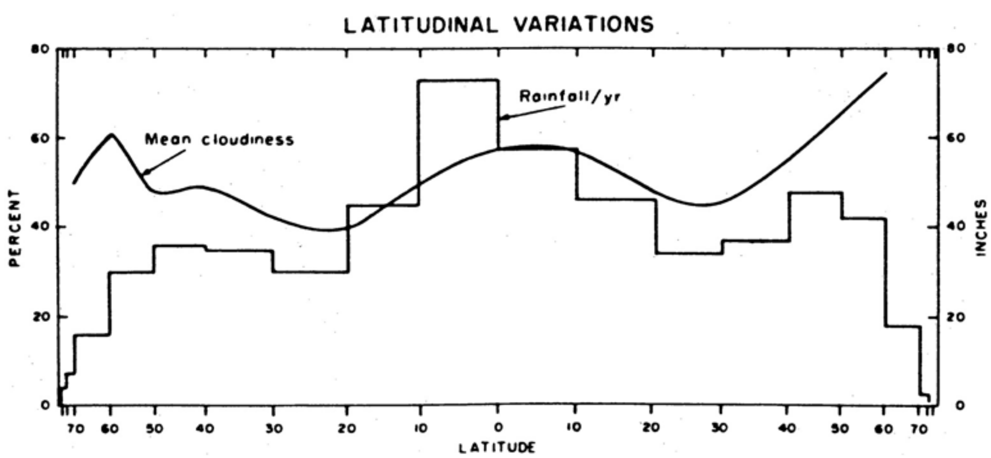

Figure C31 Zonal distribution of mean rainfall (in right scale) and mean cloudiness (percent, left scale) on Earth.

Table C16 Some extremes of climatic elements at the surface of the Earth (observed through 2000)

Element	Measurement	Globe location	Measurement	US location
Temperature (°C)				
Highest	58	Azizia, Libya	57	Death Valley, California
Lowest	−89	Vostok, Antarctica	−63	Snag, Yukon, Canada
Pressure (mb)				
Highest	1084	Agata, USSR	1068	Barrow, Alaska
Lowest	870	Typhoon Tip, Ocean	892	Matecumbe Key, Florida
Precipitation (cm)				
1 minute	3.12	Unionville, Maryland	3.12	Unionville, Maryland
24 hours	187.0	Cilaos, La Reunion, Indian Ocean	109.2	Alvin, Texas
1 month	923.0	Cherrapunji, Assam, India	271.8	Kukui, Maui, Hawaii
1 year	2647	Cherrapunji Assam, India	271.8	Kukui, Maui, Hawaii
Greatest annual mean	1168.4	Mt. Waialeale, Kauai, Hawaii	1168.4	Mt. Waialeale, Kauai, Hawaii
Windspeed (km/h)	372	Mt Washington, New Hampshire	372	Mt. Washington, New Hampshire
Longest rainless period	14 years	Arica, Chile	993 days	Bagdad, California

interval of observations and their spatial coverage increase, new extremes can be expected. It might be noted here that areas subjected to tropical cyclones are particularly apt to experience rainfall and wind extremes. Most of these storms occur in the northern hemisphere, with a peak frequency in fall. Much of their energy is derived from the latent heat of water vapor condensation, which derives from evaporation at the ocean surface, using the large energy income from the warm season. The tropical storms balance the water vapor excess in the northern hemisphere and adjust the vapor pressure to the seasonal temperatures.

Of all the features of global circulation, the seasonal winds called "monsoons" have the greatest climatic significance. About half the population of the globe is affected by monsoonal currents. These are most marked in the northern hemisphere and are most conspicuous in India and Southeast Asia. There are, however, monsoonal currents in West Africa, Australia, Western Europe and southern North America. The basic cause is the seasonal contrast of temperatures between land and adjacent oceans due to changes in radiation balance between the different types of surfaces. This creates high pressure over land and lower pressure over the sea in winter. The reverse is true in summer. Thus the winter monsoon creates outflow of air from the continent and the summer monsoon, inflow into the continent. This flow pattern results in dry winters and rainy summers. In India and Southeast Asia the rain tendency in summer is reinforced by the migration of the Intertropical Convergence Zone to almost 20°N latitude. The early spring heating of the Tibetan Plateau also plays a major role in initiating the summer monsoon. The seasonal shift in wind directions is another principal character of the monsoons, additional to the precipitation patterns (Figure C32).

Airmass frequencies are useful adjuncts to describe climatic conditions other than by numerical values of climatic elements. Prevalence of a given airmass indicates a stable, uniform climate whereas frequent airmass changes indicate a turbulent climate. Higher frequency of maritime airmasses over continental airmasses indicates a marine influence and probability of a rainy climate. If the continental types dominate, a dry climate is likely to prevail. Similarly, if the polar airmasses dominate, a

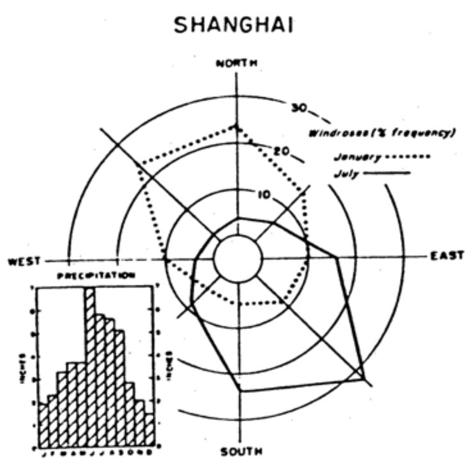

Figure C32 Wind roses for midwinter and midsummer showing the monsoonal wind shift in east Asia at Shanghai and the associated rainfall distribution (lower left).

cool or cold climate is indicated. High frequency of tropical airmasses will be associated with a warm climate. Seasonal changes in the frequency of various airmasses readily indicate the annual course of climate.

Scale and time in climatology

It need be appreciated that climatology can be considered at almost any spatial scale and over any time period that is long enough to establish a climatic record.

Scale

The smallest area scale is considered as microclimatology, which is characterized, for example, by the climate that might occur in an individual field or around a single building. A microclimate may extend horizontally from less than 1 m to 100 m, vertically from the surface to 100 m.

A local climate comprises a number of microclimates that make up a distinctive surface such as a forest or a city. The size of a local climate may extend horizontally from 100 m to 10 000 m and vertically to 100 m. A mesoclimate may contain a number of individual landscape types but often having a similar physiographic component. Thus the neighboring states within the Great Plains may be part of the same mesoclimate, with a vertical extent ranging to 6000 m. Macroclimatology concerns the climates of continents and, eventually, global climates.

Time

Climatology considers many time scales. In terms of the reconstruction of past climates, studies vary depending upon the type of information available. Given that reliable and extensive climatic data have been available for little more than 100 years, climatologists rely upon proxy data for interpreting past climates. Table C17 provides a listing of the common types of proxy data used. The level of sophistication of interpretation of proxy data is a function of time. The more recent the event, the more information that is available. Thus, the representation of climates of, for example, the Paleozoic is much more general than that of say, the Pleistocene. When people began to observe the conditions around them, and eventually describe them in writing, proxy data had a new dimension added. The type of information used to reconstruct conditions of, for example, the Little Ice Age, illustrates how such data are used. Contributing to reconstruction of such climates are descriptions of unusual events, at least in terms of today's climate; the cited frequency of river freezings, reported crop failures, writings describing "years without summers", contemporary paintings and etchings showing wintry conditions, are some examples.

Timed predictions of climatic developments have been attempted for decades, so far with only limited success. Nearly all of the schemes are based on persistence, lag correlations, and teleconnections of climatic elements. The emphasis has been on temperature and precipitation projections for a month or season. Some skill is shown for temperatures but very little for precipitation. Relations to solar activity have proved elusive and no reliable prediction models for energy emission from the sun have so far been found. Periodicities have also been ardently searched for. There are some indications that a small amount of the variance in both temperature and precipitation that might be explained by solar fluctuations in long climatic

Table C17 Palcoclimatic data sources

Data Source	Variable measured	Potential geographical coverage	Period open to study (years BP)	Climate Inference
Ocean sediments (cores, accumulation rate of < 2 cm/ 1000 years)	Isotopic composition of planktonic fossils; benthic fossils; mineralogic composition			Sea-surface temperature, global ice volume; bottom temperature and bottom-water flux; bottom-water chemistry
Ancient soils	Soil type	Lower and midlatitudes	1 000 000	Temperature, precipitation, drainage
Marine shorelines	Coastal features, reef growth	Stable coasts, oceanic islands	400 000	Sea level, ice volume
Ocean sediments (common deep-sea cores, 2–5 cm/1000 years)	Ash and sand accumulation	Global ocean (outside red clay areas)	200 000	Wind direction
Ocean sediments (common deep-sea cores, 2–5 cm/1000 years)	Fossil plankton composition	Global ocean (outside red clay areas)	200 000	Sea-surface temperature, surface salinity, sea-ice extent
Ocean sediments (common deep-sea cores, 2–5 cm/1000 years)	Isotopic composition of planktonic fossils; benthic fossils; mineralogic composition	Global ocean (above $CaCO_3$ compensation level)	200 000	Surface temperature, global ice volume; bottom temperature and bottom-water flux; bottom-water chemistry
Layered ice cores	Oxygen-isotope concentration (long cores)	Antarctica; Greenland	100 000+	Temperature
Closed-basin lakes	Lake level	Lower and midlatitudes	50 000	Evaporation, runoff, precipitation, temperature
Mountain glaciers	Terminal positions	45°S to 70°N	50 000	Extent of mountain glaciers
Ice sheets	Terminal positions	Midlatitudes, high latitudes	25 000 to 1 000 000	Area of ice sheets
Bog or lake sediments	Pollen-type and concentration; mineralogic composition	50°S to 70°N	10 000+ to 200 000	Temperature, precipitation, soil moisture
Ocean sediments (rare cores, > 10 cm/ 1000 years)	Isotopic composition of planktonic fossils; benthic fossils; mineralogic composition	Along continental margins	10 000+	Surface temperature, global ice volume; bottom temperature and bottom-water flux; bottom-water chemistry
Layered ice cores	Oxygen-isotope concentration, thickness (short cores)	Antarctica; Greenland	10 000+	Temperature, accumulation
Layered lake sediments	Pollen type and concentration (annually layered core)	Midlatitude continents	10 000+	Temperature, precipitation, soil moisture
Tree rings	Ring width anomaly, density, isotopic composition	Midlatitude and high-latitude continents	1000 to 8000	Temperature, runoff, precipitation, soil moisture
Written records	Phenology, weather logs, sailing logs, etc.	Global	1000+	Varied
Archeological records	Varied	Global	10 000+	Varied

After various sources.

time series, is generally less than 10% of the total. In the Western United States a coherence of the extent of drought areas, as found by tree ring analysis with the double sunspot rhythm (Hale cycle), has been found; it has only limited prediction value. Similarly, the persistent quasibiennial oscillation found in many time series of climatic elements seems to wax and wane in time and space to prevent practical predictive use. The Southern Oscillation, a large-scale pressure see-saw in the South Pacific Ocean, is repetitious on a 5–7-year time scale. It influences the intensity of the Indian summer monsoon and the occasional warm water current El Niño off the Peruvian coast.

Of considerable interest to climate researchers are the identification of climatic analogs, periods of climate in the past that can be considered analogous to those of the present or of the possible future. Thus warm climates of the past, such as the Eemian interglacial (125 000–130 000 years BP) or the Holocene "climatic optimum" (5000–6000 years BP) have been used in attempts to assess future greenhouse climates. For the limited period for which observed climatic data are available, climatologists use numerous statistical methods to search for trends, periodicities or iterations. The specter of global warming, as promulgated in the 1980s, has resulted in extensive examination of observed data with attempts to standardize and validate them.

The forecasting of climate is a difficult task. At present there are two main methods of attempting to determine what lies ahead. The first uses climatic analogs which, as earlier noted, are time periods in the past that resemble conditions as they may soon be if the Earth's climate were to warm. Second, climate models are generated. Climatologists use various types of models but of particular importance at present are general circulation models (GCM). Modern GCM are a product of computers' calculating capability for GCM, attempt to solve equations of motion, thermodynamics and conservation for moisture and mass in a defined global area through various atmospheric levels. A number of models have been run, and while problems still exist, the output of such models is becoming more acceptable.

Climatic data

All of the understanding within the science of climatology must rest upon the availability of data. For climates preceding the instrumental period actual measurements are not available, and proxy data are used. Since about 1850, especially for land surfaces in the northern hemisphere, monitoring and data collection have gradually improved. Today, routine measurements of the atmosphere are accumulated for many world locations and are stored in archives. For most climatic analysis the statistics derived for a 30-year period are adequate. Unless otherwise noted, the cited average data of most locations represents the most recent 30-year block, ending in a decennial year. Thus 1951–1980 observations are replaced by those of 1961–1990 and so on.

For examination of climatic trends and periodicities, periods longer than 30 years are needed. Often, errors enter into older data from such things as change of observing location or modified instruments. Such is true in many urban locations where the development of an urban heat island causes the observed data to differ from surrounding areas. Climatologists check historic data carefully before assuming it is error-free. Despite the gains made in collection, quality control and archiving of climatic data, there should be improvements especially over the oceans and in underrepresented global areas. Climate is a basic global resource, and for managing and comprehending this resource, climatologists must have good data. Certainly, without such a base, the important task of predicting future climates becomes even more difficult.

H.E. Landsberg and J.E. Oliver

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Cross-references

Applied Climatology
Bioclimatology
Cycles and Periodicities
Dynamic Climatology
Climatology, History of
Hydroclimatology
Microclimatology
Scales of Climate
Synoptic Climatology
Water Budget Analysis

CLIMATOLOGY, HISTORY OF

Since the mid-1980s the content and research topics of climatology have broadened considerably. Less-developed topics, such as teleconnections and periodicities, have attained new significance. Additionally, during this period the ready availability of data and information via the worldwide web and the enhancement of data processing and modeling have also produced new research directions. This more recent history is dealt with in many pages of this volume. As a result the historical perspective provided here provides information up to the time prior to the explosive development of paleoclimatology and the paradoxes of global warming. It deals with the history of climatology to the mid-1980s.

Introduction

In one sense climatology has no special history. Outdoor people – farmers, foresters, sailors – have always been aware of their environment, its physical and chemical dimensions, and its biological effects. As Brooks (1948, p. 155) said: "Anyone who has lived here [Charlottesville, Virginia] a decade or two can give you a pretty clear picture of the climate", but records and instrumental readings are needed when comparisons or explanations are wanted. Measurements of state, e.g. temperature of air or soil, and of fluxes, such as solar radiation, are required for these purposes: (1) to picture the variability of a climate and to assess risks of extreme conditions for personal, agricultural, or economic planning; (2) to compare one place with another or to describe a remote, perhaps never-visited place, and answer the perennial question, what is it like there? (3) to understand how climate is formed, how it relates to plants, soil, landforms, and water bodies and the rest of the environment; and (4) to define the roles that solar energy or marine influences play, and what is the outcome of the eternal conflict between local, indigenous processes, and inflows of energy or water from outside. These kinds of understanding make it possible to apply science to models of crop growth, moisture stress, and stream-flow, and for climatology to contribute to its sister sciences of hydrology and meteorology, ecology, forestry and agronomy, the geophysical and geological sciences, as well as to the management of regional resources and national economic health.

Early observations

The Greeks, who lived under cloudless summer skies, were aware of how the height of the daytime sun differed in the desert to the south of them and in the cold lands to the north. Their zoning of climate (= angle of inclination of the sun to the surface of the Earth) is reasonable, emphasizing the role of the Earth–air interface (Tweedie, 1967). Later expressed as day length, this zonation of the globe persisted into the period of Arabian science (Blüthgen and Weischet, 1980). Meanwhile, the Chinese were living where a cloudy, wet summer is the productive season of the year, and they thought of their environment in terms of abundance or failure of the rains. In about the same period as the Greeks, they apparently were recording variations in their environment in terms of the rainfall that determined harvests and tax revenues. They recorded floods from the second century, dry periods from the seventh century, and severe winters from the eleventh century (Needham and Ling, 1959).

The Mediterranean center emphasized the sun and latitude, which seemed to be associated with the range from the warmth of the deserts of northern Africa to the cold of Scythia and to the miserable winters beyond the Alps where the legionnaires huddled around the warm springs of Baden. The effect of the sea was recognized, especially at places exposed to the steady winds used in navigation. During the same time period that the Greeks built the Tower of the Winds, the Chinese began to record wind direction, which was associated with air streams delivering water; they measured rainfall even earlier than the thirteenth-century records in Korea (Needham and Ling, 1959), measured snowfall, and apparently organized networks that reported these readings to the central government. The Greeks gave the winds names and qualities (Wright, 1925), Cicero spoke of the marine climate of England (Khrgian, 1959), and the general effects of the sea and of mountain barriers against polar winds were well known (Wright, 1925). During the late

Middle Ages weather diaries and phenological observations provided climatological information, much of it as yet unused.

When fifteenth-century navigators moved out of the Mediterranean into the lower latitudes of the eastern Atlantic, they encountered the steady trade winds, which differed from the seasonal winds of the Mediterranean or the monsoons of the South Asian seas (Landsberg, 1985). When continuing southward they encountered another belt of trade winds, the low-latitude elements of the global circulation. Halley, who had been studying the relation between land elevation and barometric pressure, turned in 1686 to the winds associated with pressure readings and recognized a thermal driving force for the trades (Frisinger, 1977), and in 1735 G. Hadley developed the first general circulation model (Lorenz, 1983).

As settlers moved westward from Europe and eastward from Russia into new lands, they encountered more severe climates. The cold winters and overheated summers of North America forced changes in the pattern of farming and subsistence (US Department of Agriculture, 1941), and seventeenth-century reports on Siberia discussed whether or not grain could be grown (Khrgian, 1959). Concern about the effects of the climatic environment on plant and human health promoted the use of instruments, especially the barometer and thermometer, which were just being developed in Western Europe. These were first employed by scientists in different fields who shared an interest in their environment; some of the earliest observations were made at astronomical observatories (Blüthgen and Weischet, 1980), and it may be recalled that many eighteenth-century discoveries in chemistry arose out of interest in the atmosphere. These early measurements, made carefully and systematically, gave climatology numerical data to enjoy centuries before geology, geomorphology, ecology, or other field sciences had such materials. In fact, the word *climatology* came to be used for the ordering of any variable displaying an annual regime. We have a climatology of Atlantic shelf water, of wind stress on the oceans, of clouds, and of upper-atmosphere conditions. Time series of early measurements have been recently constructed by Landsberg (1985).

The nineteenth century

At the beginning of the nineteenth century A. von Humboldt, during his expeditions in the New World, noted differences in climates, mapped them by isotherms, and recognized the effect of altitude on air temperature in the Andes (Leighly, 1949a; Khrgian, 1959). This work brought the end of the mathematical formalisms that had so oversimplified climatic distributions, although Fourier series continued to be published as summaries of diurnal and annual regimes at individual places (Leighly, 1949a).

As records at particular places accumulated, it became possible to describe the climate of a city, as L. Howard did for London in 1818 (Landsberg, 1981), E. Renou for Paris in 1855, and M.F. Spasskii for Moscow in 1847; some pointed out urban–rural differences, others the variability from year to year. Single-station observations of the seventeenth and eighteenth centuries expanded into networks (Rigby, 1965), the system in the Rhineland beginning in 1780 was "decisive for the history of climatology" (Landsberg, 1985, p. 58). Not much later (1818), the Surgeon General of the United States began receiving and publishing regular climatic reports from military posts (Landsberg, 1964). This network and those in several states, which later merged into the Smithsonian network, served as models for a Russian system that started in 1839 (Landsberg, 1964). The value of these networks can be seen in early handbooks on climate (Hellman, 1922), which dealt with plants (1822, 1827), forestry and agriculture (1840, 1853), that were generally associated with air temperature, as by H.W. Dove (Khrgian, 1959). The climate of mountains, e.g., Humboldt in 1831 and 1843, introduced the question of sea-level reduction of shelter temperature readings, which continued, parallel to the question of the decrease in temperature of the free air with height, through much of the century. The measurement networks provided data for national surveys of climate like those of L. Blodget for the United States (Leighly, 1949a) and K.S. Veselovskii for Russia (Khrgian, 1959), which represented a high degree of literary skill as well as concern for practical applications. Both were published in 1857.

Extension to the worldwide scale had already taken place for temperature with Humboldt's isotherm maps of 1817, L.F. Kämtz's (1831–1836) discussions of temperature and wind patterns and Dove's polar and tropical air streams (Khrgian, 1959). Humboldt's research in the 1840s on the mountains and climates of central Asia led him to a concept of a solar climate (Flohn, 1971) that was a comprehensive expression of physical climatology. World maps of the winds, pressure, and rainfall followed (Leighly, 1949a); temperature maps became less schematic. Many world maps were products of the Smithsonian Institution in the years before 1870, when it still operated both climatological and synoptic services (Leighly, 1954).

Observations from all over the world, on sea as on land, flowed into the scientific community. J.F. Maury had enlisted the maritime nations in standardized observations at sea in the 1850s and sketched a general circulation model. Later, W. Köppen, at the Deutsche Seewarte in Hamburg, had at hand worldwide flows of data of interest to seamen and merchants. J. von Hann, an editor in Vienna for decades, developed a worldwide network of correspondents whose reports appeared in his journals. A.I. Voeikov, who had traveled widely and worked in many scientific institutions such as the Smithsonian, had similar global contacts and experiences, especially of the influences of climate on agriculture and settlement in the Americas and east Asia. Voeikov, Köppen, and Hann studied the general circulation and, with the scientists of the Smithsonian and other organizations, created the classic period of climatology that occurred during the middle third of the century from about 1855 to 1885 (Leighly, 1949a).

While Maury (1806–1873) preceded the others, it is notable that three of these founders of climatology were born within a few years of one another. Hann lived from 1839 to 1921, Voeikov from 1842 to 1916, and Köppen from 1846 to 1940 (within our own times). Similarly, V.V. Dokuchaev lived from 1846 to 1903 and studied global soils in their climatic context.

By midcentury the most easily measured climatic elements on land, shelter temperature, and rainfall were observed widely; cloudiness, wind, and humidity were observed at fewer places; radiation, notwithstanding its recognized role in the energetics of climate, still remained beyond reach. As observations accumulated, the global circulation of the atmosphere expanded and deepened into a global view of climate. Cartographic analyses, one of the major strands of climatology in the nineteenth century (Leighly, 1949a), drew attention to problems of causation, but much effort still went into futile attempts to reproduce the zonation of shelter temperature and to cast global energy budgets. These were an improvement over the earlier trigonometrical formulas for global temperature but failed because of lack of data on the thermodynamic properties of the underlying surface, which either went unrecognized or were incorrectly assumed, as was albedo for many years (Miller, 1969). Physical laws of climate stated by Spasskii included absorption of radiation, heat conduction, evaporation, and so on (Khrgian, 1959) but could not be investigated because of the lack of accurate measurements of solar radiation and the late discovery of longwave radiation.

Local climates, though not explained, could be described from phenological observations in both their year-to-year variability and their spatial patterns, as well as by tables, maps, and verbal discussion of such variables of state as shelter temperature. These descriptions make up a large part of the classic books that essentially defined the field of climatology (Hann, 1883; Voeikov, 1884). These two gave rise to books in English and Italian (Hellmann, 1922); Voeikov's papers from his travels in Japan influenced the development of climatology in that country (Fukui, 1977). Explanatory power increased as time went on; for example, Voeikov's ideas on the effect of snow cover on climate were introduced in the second edition of Hann's handbook (1897), and near the end of the century a few rough data on radiation began to appear.

New directions in energetics

Maury (1861) outlined the components of ocean surface energetics, and Spasskii and Voeikov those of the land interface. The gaps mentioned in radiation data remained and augmented the lack of knowledge of properties of the interface, which influences not only the atmosphere as a whole, but particularly its lowest meters in which biological activity and physical interactions are concentrated. As a result, the effects of climate on agriculture, for instance, in any but gross features remained inadequate, and basic understanding of climatic genesis was absent. Voeikov's book and papers pointed out the importance of this interface for climatic energetics, the bookkeeping of solar energy receipts, and included such indicators of a regional climate as the temperatures of lakes and rivers, presence and effects of a snow cover on air and soil climate, and mutual interactions between vegetation cover and the lower air. Following Russian observations of soil freezing and temperature, Homén's heat-budget measurements (1897) in three sites in Finland centered on heat fluxes in the substrates. A number of investigators subsequently concerned themselves with heat storage in soil and water bodies. Radiation observations at the surface of the Earth – although considered as early as L.W. Meech's mathematical work in 1857 (Miller, 1969) – were just appearing. Nineteenth-century radiation work had been chiefly a matter of instrument development (Khrgian, 1959) on both sides of the Atlantic, much of it near the 60th parallel north (St Petersburg–Leningrad, Helsinki, Stockholm, and later Tartu). Radiation studies in Vienna entered climatology after the work of J. Stefan and L. Boltzmann, and a long period of productive work in that city and the Alps began. By the 1920s the capability to measure radiation fluxes converged with studies on the survival and growth of plants, especially in central Europe (Munich, the Alps) because plant growth could not be explained by shelter temperature. Field studies in dissected terrain by foresters, ecologists and meteorologists resulted in data that, in other hands, could sometimes be given a physical interpretation. In doing this R. Geiger (1927) invented the discipline of microclimatology, which in subsequent decades proved invaluable to climatology at larger spatial scales (Miller, 1969). Another biological connection was made by A.E. Douglass's work on tree rings (Fritts, 1976), in collaboration with archeologists trying to extend the climatic record back in time. Brooks (1926) associated climate changes with drops in energy gained from solar radiation resulting from volcanism or high polar albedo.

Also at this time, the feeling that recent changes in climate might be measured in terms of changes in glaciers led Scandinavian scientists to examine mass balances of snowfields, where summer melting rates express both solar and advected energy. Energy budgets were developed by A. Ångström in the 1920s and subsequently by H.U. Sverdrup (1942), H.W. Ahlmann, and C. C. Wallén (1948–1949).

Sverdrup also did notable work on turbulent processes at the interface, which had been studied by Schmidt (1925) and others, completing the roster of energy fluxes in climate. Leighly (1938, 1942) applied energetics reasoning to assess the influence of advection on the regional climates of California and the Great Lakes area. By 1940 Albrecht had calculated budgets for several parts of the Earth, including one from detailed observations brought back from an expedition into Central Asia. Unfortunately, cities remained refractory subjects for energetic analysis, but their mosaics of climates were described in detail in terms of state variables – principally air temperature (Kratzer, 1937), radiation and microscale interface temperature (courtyards, streets, walls) by Viennese climatologists.

Energetics, which is central in the calculations at each grid point in large-scale climate models and provides a tool to explain the genesis of climate, forms in addition a link between the climate as environment and the environed organisms. Energy budgets were cast for plant leaves in the 1950s and for animals in the 1960s. For the first time such physically significant characteristics of the Earth's vegetation cover as the geometry, mass, radiative properties, and ventilation of crop and forest canopies were measured and introduced into energy budgets (Gates, 1962; Rauner, 1972; Ross, 1975). Such budgets were able to assess the effect of forest on the atmosphere, and also to explain and predict the way living organisms utilize time and space. Embodying the same radiative and turbulent fluxes that figure in the interface budget, they defined the true impact of environment on an organism's thermoregulation, productivity, water and nutrient needs, behavior during the diurnal and annual cycles, and even its life strategy and evolution. Climatology took on a new immediacy and precision.

The water flux

The fluxes of water in climate, difficult to approach on a global basis because of lack of data over the oceans, never gave rise to the studies of merely state variables that had slowed progress in climatic energetics. Perhaps European scientists, like Dove, considered water a secondary factor, and this attitude persisted through much of the nineteenth century. It had been known since the late seventeenth century that rainfall supplied enough water to support rivers, not only, as E. Halley thought (Middleton, 1965), to mountains in the center of land masses, but even in lowlands like the basin of the Seine (Biswas, 1970). However, without knowledge of the partitioning and conversions of water at the interface, late nineteenth-century efforts to evaluate the true role of moisture in climate and to bring it into climatic typing and classification were in vain, even after it had become clear by 1900 that rainfall by itself was an inadequate variable. Trials of many combinations of rainfall and shelter temperature were made to make various schemes of classifying climates fit vegetation better (Leighly, 1949a). The most notable of these were the successive stages Köppen went through from 1874 up until the 1930s (Leighly, 1949a; Khrgian, 1959). None of these classifications, however, turned out to have lasting physical relevance; partly, at least, because of the error in the assumption that the kind of observations established in the first half of the nineteenth century provided sufficient data for solving the problems of climate (Leighly, 1954).

Agricultural climatology was, from the beginning, an important part of the subject and was well summarized in 1941 by *Climate and Man*, a yearbook of the US Department of Agriculture. This compendium discussed agricultural settlement of major climatic regions along with chapters on corn, cotton and other crops affected by the conventional elements of shelter temperature and rainfall, especially as measured by the network of cooperative stations, then at its peak. Because administrative shifts later separated this work from interface and synoptic climatology, a hiatus followed. *Climate and Man* marked, along with the Köppen-Geiger *Handbuch der Klimatologie* of the 1930s and especially Conrad's volume (1936) on the dependence of climatic elements on terrestrial influences, a summing-up of the early twentieth century.

The lag in the hydrologic dimension of climate behind energetics and microclimatology ended in the 1930s when flood-mitigation programs and droughts in the United States revealed the deficiencies of direct rainfall-to-runoff transfers. R.E. Horton and others emphasized soil moisture as a dimension of drought and a condition of infiltration during rainstorms. Evaluating soil moisture, in turn, called attention to evapotranspiration. This long-sought convergence of water and energy burst into prominence in 1948, when important papers were published by C.W. Thornthwaite, H.L. Penman, and M.I. Budyko (see references in Miller, 1977). These three papers, which recognized the role of energetics, made it possible for the first time to provide an adequate representation of moisture in climate (Tweedie, 1967). Their energetics aspect stimulated research on all the exchanges between the interface and the atmosphere (Thornthwaite, 1961), especially radiation. Lettau's partitioning (1952) of sensible heat between substrate and atmosphere supplemented this work on latent heat and rounded out the theory for interface effects on climate, despite the fact that observational data remained scanty in most kinds of terrain. These methods have been adapted to faster computation methods in such applications as drought indices, probabilities of soil-moisture stress, streamflow modeling, flood frequencies, irrigation scheduling, and crop-growth models. Water-budget methods developed in Thornthwaite's Laboratory of Climatology were applied to reforested uplands, coastal marshes and estuaries, stream flow, and biological phenomena such as decomposition of litter (Muller, 1972; Mather, 1978). Ocean climatology used the methods of interface energetics to outline large-scale patterns of evaporation and hence salinity as a factor in ocean dynamics. Sverdrup (1942) discussed radiation, energy budgets, and sensible-heat exchanges with the air, all related to evaporation.

Circulation

Large-scale climatology benefited from advances in meteorology following 1920, when frontal and air-stream analyses (Namias, 1983), combined with upper-air observations, sparked research in a science stagnant for half a century (Bergeron, 1959) that had been a drag on climatology whenever the two were associated in governmental weather-forecasting services. Some studies of airmass formation, e.g. the classic by Wexler (1936), added to knowledge of climate genesis by virtue of developing interface energy budgets; Flohn (1971) applied energy budgets of the Arctic vs. Antarctic to explain the strength and steadiness of the circulation of the southern hemisphere and budgets of Tibet to evaluate the effects of the injection of heat into the middle troposphere. Many investigators, especially in England and Germany, found air streams useful entities for typing climates; calendars of synoptic air flow pursued the beginnings that Köppen in 1874 (Leighly, 1949a) and others had made toward stratifying climatic events by direction and vorticity of air flow. These became a comprehensive synoptic climatology in World War II (Jacobs, 1947; Hare, 1955; Court, 1957) and served as frameworks in regional climatologies like that of the Paris Basin (Pédelaborde, 1957), or baseline climatic data for New Orleans (Muller, 1977). On a larger scale, circulation analyses provided indexes related to meltwater generation in anticyclonic conditions and to glacial accumulation (Leighly, 1949b). Circulation types over the northern hemisphere (Dzerdzeevskii, 1968) and vapor transport in monsoon climates (Yoshino, 1971) were evaluated. They provided a basis for climatic classifications (Flohn, 1971), revisions in general circulation models (Lorenz, 1983), world maps (Hendl, 1963), and study of climatic genesis and fluctuations (Flohn, 1971; Lamb, 1972, 1977). Fluctuations of large magnitude presumably reflected circulation changes, which could be associated with profiles of pollen and other sediments and statistical properties of tree-ring measurements (Fritts, 1976).

Evaluation of atmospheric vapor transport (Benton and Estoque, 1954) provided an independent check on assessments of evapotranspiration at the underlying interface and not only had hydrologic value but also showed the ties of a regional climate to other parts of the world. Such studies assessed the strength of advection from other regions *vis-á-vis* local factors, as for example, the Andean antiplano, a nearly autochthonous climatic region (Hendl, 1963) that receives little advection (Gutman and Schwerdtfeger, 1965). Postcirculations are sketched on the basis of plate tectonic data on locations of the continents as far back as the Cambrian (National Research Council, Geophysics Study Committee, 1982; Budyko, 1982).

Expanding applications of climatology

Applications of climatology increased greatly during World War II, particularly in the US Air Force. The worldwide distribution and diverse nature of these applications, including the design of equipment and clothing to meet climatic conditions anywhere in the world, had important effects on the whole field. Among other things, they showed the inadequacy of the old hope that a single classification could meet all needs of science (Landsberg, 1958). Networks expanded, archives grew and were put into computable form and machine methods for sorting and summarizing data became standard. This effort was aided by a series of initiatives by the World Meteorological Organization's Committee on Climatology, first chaired by C.W. Thornthwaite. Large-scale statistical programs that had started in the late 1930s with marine data and research on drought, developed machine methods that made possible frequency analyses on an unheard-of scale, as well as correlations, principal-component analysis and other ways to identify major features of a regional climate. Statistical techniques were transferred from agricultural and biological research and developed within climatology (Court, 1951). As storage and computing techniques grew more powerful, however, the gritty problem of assuring the quality of field measurements, especially those not immediately and directly utilized in forecasting the weather, persisted, and in spite of efforts of the World Meteorological Organization, records have lost homogeneity.

Concern in the 1970s about variations in climate that threatened world food production revealed the fact that serious questions of uniformity of series and representativeness of stations had grown up, questions that had been thoroughly examined a century before but forgotten in many countries. The struggles to maintain even a skeleton network of benchmark stations to detect changes in temperature of the kind Mitchell (1963) investigated revealed the difficulty of these historic problems of reliability. The question of change was also examined from the 56-year record of circulation types (Dzerdzeevskii, 1962) and upper-air records. The study of climatic changes at longer time scales, i.e. in the preinstrumental periods, which had been stimulated by nineteenth-century research on glaciation, was again aided by geological research beginning in the 1960s on oceanic sediments (CLIMAP Project Members, 1976) and plate tectonics (Frakes, 1979; National Research Council, Geophysics Study Committee, 1982; Budyko, 1982). For example, the movement of the Australian plate into the latitude of the subtropical anticyclones caused a decrease in rainfall, changes in the structure and fire regime of the vegetation cover and its microclimatic habitats. These geological and geochemical studies indicate that carbon dioxide and water vapor have been components of the atmosphere since the earliest times and stabilize the planet's energetics against "greenhouse" overheating or a cooling to a condition of complete glaciation or a "white earth" (Budyko, 1982, p. 131). The Earth has a long climatic history, only glimpsed in part at the present time.

A large program in energetics at the Voeikov Main Geophysical Observatory in Leningrad (Budyko, 1956, 1971) began with physical research on energy and water fluxes at the interface, tested the results in expeditions to desert, forest and mountain locations, and applied them to calculate national and global patterns of distribution (Budyko, 1963) and to analyze changes (Budyko, 1982). This expansion to global scale paralleled the climatologically based soil science of V. V. Dokuchaev (Khrgian, 1959). The results of the energetics program provided better data for research because of the foundation in adequate measurements of physical exchanges at the interface, a consideration lacked by some earlier research on global climatology (Miller, 1969). This program was supplemented by energetics research in several regional hydrometeorological institutes, e.g. that at Tashkent (Aizenshtat, 1960), where soil climate had been measured for half a century and radiation for several decades; at Tbilisi, another old observatory, at Kiev, and at Tartu, for radiation (Ross, 1975); and in a program developed by Dzerdzeevskii (1962) and Rauner (1972) to examine the influence of forested regions on the atmosphere. Many of the global patterns became inputs to large-scale computer models of the circulation that examined possible changes caused by alterations in atmospheric composition, interface characteristics such as sea ice and continental snow cover or albedo, remotely sensed over extensive areas or in the intake of solar energy. Examining the annual march of climate affords a useful approach to its physical causes, a method used earlier (Leighly, 1938) and well adapted to quantitative computation techniques exploring solar forcing of climatic fluctuations.

Meteorological observations that expanded during and after World War II into the low latitudes and in the 1950s into the polar regions, filled in gaps in the patterns of world climates and added variables desired since Hann and Voeikov, but not earlier measured in routine programs. The results are summed up in the many volumes of Landsberg's *World Survey of Climatology* (1969–1984) and made possible worldwide water budgeting in several countries. Intensified observations of radiation and other heat fluxes in the USSR and Western Europe served for both applied purposes and definition of large-scale patterns, which were also shown by improved upper-air observations, the measurement of sea ice, snow cover, and interface temperature at sea by remote sensing, with its potential for synchronous and macroscale surveillance. Short field programs over the low-latitude oceans in the 1970s might have climatological significance for the light they cast on the phenomenon that is the major exception to the astronomical regimes that otherwise dominate climatology: the Southern Oscillation and El Niño, with their wide teleconnections.

The climatology of the energy budget influenced ecology and agriculture; one-third of Chang's text (1968) is devoted to radiation and photosynthesis and more than a third to evapotranspiration; shelter temperature receives only part of a chapter. Agricultural climatology advanced in practicality in most countries, including China, where much use of radiation data is made (Huang, 1981), and in many states of the United States, which developed their own observational networks to support these methods. A phase of this work was the growth of "near real-time" climatology, which called on the relevant observations of antecedent atmospheric and soil conditions to make direct recommendations on the management of rainfed cropland, fire hazards in wildlands, irrigation of crops, resources of polar lands of Canada, and to predict crop harvests. This time dimension of application was expressed most clearly in mesoscale and regional climatology, though the basic physical and biological relations were generally microscale.

Because most applications of climatology, as distinguished from attempts to improve weather forecasting, dealt with management of water resources or crop production, which operate at regional scales, information at this scale was most useful. For example, new data on mountain climates in practical energy and water terms came from programs in snow hydrology in three mountain ranges of the western United States, to provide data for reservoir design and scheduling, and from investigations at the timberline in Austria and Switzerland to determine how to regenerate forests in avalanche source areas. These programs included multidisciplinary work on the soil, topographic, and biological conditions associated with the energetics of the water and biomass yields of mountain climates.

Lettau's climatonomy (1969) centered on the energy budget, using as forcing functions rainfall and solar radiation, which are independent at the regional scale, to develop the regimes of the state variables of soil moisture and interface temperature and of the flux of latent heat. These quantities are basic in agricultural climatology as it developed in several midwestern states of the United States, making use of observational programs in radiation, soil temperature, and soil moisture, as well as a dense network of recording rain gauges. Concern in several states about energy supplies in the 1970s encouraged research in wind and solar energy climatology at regional and mesoscales. Basic factors in small-scale climatology were classified by Yoshino (1975) as topography, local air streams and atmospheric disturbances, and interface characteristics of open land, forests, and cities. Urban regions received attention beyond the old heat- island idea as their pollution thickened, but while descriptions of urban climates included more physical considerations (Landsberg, 1981) it was generally in spatially averaged terms, although patterns of rain and pollutants were examined in the Metromex program in 1971–1976 (Changnon, 1981).

Addition of chemistry to the usual energetics and hydrology of climatological analysis caught observation programs and climatologists unaware, and much research on diffusion of pollutants neglected important climatological considerations. Some pollutants, such as ozone, were found to be transported on a regional scale, and some, like acid rain, following C.G. Rossby's revival of interest in precipitation chemistry on even larger scales; modeling these transports and studying their effects entered the area of global climatology. Concern about possible effects on climate of increased atmospheric concentrations of carbon dioxide and other radiatively active substances was expressed in the 1950s, but served less to bring about better observations of, say longwave radiation, than to justify computer-assisted models of global circulation. While these models portrayed zonal averages better than regional climates, where water resources and food production are most likely to be affected by fluctuations, they are expected to help study the climates of the past and foreshadow those of the future.

Summary

Seven strands were identified by Leighly (1949a) throughout the development of climatology since 1800: empirical formulation, classification, physical explanation, synoptics, cartographic analysis, description, and variability. Empirical formulations have nearly disappeared as an important activity of climatologists, as has classification of climates. They were compensated for by a strengthening of physical science, especially energetics and hydrology, which were compatible with the newer concerns of chemical and ecological climatology and with more sophisticated applications. Interface exchanges and tropospheric synoptics shape climate, and both increased with advances in meteorology after about 1920. Climatic, water, and agricultural atlases shifted to portray the elements specific to the needs of particular applications of climatology, and as new physical relations became known the relevant observations (e.g. solar energy) were made, duly mapped, and given verbal description. Many descriptions were written to explain applications in water management, food production, and construction, both in design and planning, and in operation. Climatic fluctuations became better defined with the lengthening of the instrumental record, more accurate deductions from information in the preinstrumental period, and from geology; this knowledge has gained importance with concern over the frequencies of severe winters and multiyear droughts, as well as the possibility of permanent shifts resulting from human activity.

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Cross-references

Applied Climatology
Archeoclimatology
Bioclimatology
Climate Classification
Climatology
Microclimatology

CLOUD CLIMATOLOGY

A cloud is an assemblage of liquid water particles and/or ice particles in the atmosphere. Almost all clouds form when moist air is cooled until it becomes supersaturated, i.e. the relative humidity is above 100%. If the temperature is above freezing, supersaturated water vapor then condenses onto cloud condensation nuclei to produce liquid particles. If the temperature is below freezing, the supersaturated water vapor deposits on ice nuclei to form ice particles. If there are not enough ice nuclei present, supercooled liquid particles will form even though the temperature is below 0°C. When the temperature is below −40°C, ice particles will form without ice nuclei and supercooled liquid particles will spontaneously freeze. Under certain conditions cloud particles will combine and/or grow large enough to fall as precipitation.

Clouds play a key role in the climate system. They are intimately connected to precipitation and have a large influence on the transfer of energy within the atmosphere. Latent heat is released when water vapor is converted to liquid water or ice and when liquid water is converted to ice. This added heat increases the temperature, and hence buoyancy, of the air in which a cloud forms. Correspondingly, the temperature of the air decreases due to conversion into latent heat when cloud or precipitation particles melt or evaporate. Latent heating and cooling associated with clouds and precipitation drive atmospheric motions, especially in the tropics. Clouds also have a strong influence on the transfer of solar and terrestrial radiation within the atmosphere. Because they are generally brighter than the surface of the Earth, clouds reflect more solar radiation back to space. This decreases the amount of energy absorbed by the Earth, thus causing it to be cooler than would be the case were it cloudless. On the other hand, clouds are generally colder than the surface and consequently emit less thermal infrared radiation than does the surface. This decreases the amount of energy emitted to space, thus causing the Earth to be warmer than would be the case were it cloudless. The greatest reduction in emitted thermal radiation occurs for clouds high in the atmosphere, because they are coldest. Globally, the reduction in absorbed solar radiation by the Earth caused by clouds outweighs the reduction in emitted thermal radiation by the Earth caused by clouds. This result is a net cloud cooling effect.

Cloud types

The most basic categorization of clouds is by the altitude range in which they occur. Low-level clouds have bases less than 2 km above the surface. Mid-level clouds occur between approximately 2 km and 6 km above the surface. High-level clouds typically occur more than 6 km above the surface. Each altitude range is further divided into genera, or types of clouds, according to their morphology. Low-level genera are stratus, cumulus, and stratocumulus. A stratus cloud exists as a uniform grayish layer with little apparent structure. Cumulus clouds develop vertically as mounds or towers separated from one another. Stratocumulus cloudiness occurs as a layer with distinct clumps. Mid-level genera are altostratus and altocumulus. Both altostratus and altocumulus clouds occur in layers, but differ in that altocumulus has distinct cloud elements and altostratus has a more uniform appearance. High-level genera are cirrus, cirrostratus, and cirrocumulus. Cirrus clouds are detached from one another and appear white and often feathery. A cirrostratus cloud occurs in a whitish and transparent layer. Cirrocumulus cloudiness exists as a whitish layer composed of very small but distinct cloud elements. Two cloud types that produce rain or snow are cumulonimbus and nimbostratus. A cumulonimbus cloud, typical of thunderstorms, resembles a cumulus cloud and may grow very high, but is nevertheless classified as a low-level cloud type. Cumulonimbus clouds sometimes have tops shaped like anvils and occur with thunder and lightning. Nimbostratus cloudiness exists as a gray and frequently dark layer and is distinguished from stratus cloudiness by the occurrence of rain or snow. Although nimbostratus clouds are thick, and can occupy a large range of altitude, they are often classified as a midlevel cloud. Table C18 lists these 10 basic cloud types. Fog, essentially a cloud that touches the ground, can be identified as an additional cloud type. Because specific meteorological processes produce different types of clouds, cloud morphology qualitatively describes the local atmospheric environment. The global distributions of different types of clouds are closely related to the locations of various climate zones of the Earth.

Table C18 Cloud types

Altitude range	Genera (cloud type)	Abbreviation
Low	Stratus	St
	Cumulus	Cu
	Stratocumulus	Sc
	Cumulonimbus	Cb
Middle	Altostratus	As
	Altocumulus	Ac
	Nimbostratus	Ns
High	Cirrus	Ci
	Cirrostratus	Cs
	Cirrocumulus	Cc

Cloud observations

The primary sources of cloud data are visual observations made by people at the surface of the Earth and satellite observations made from orbit. Human observers report cloud fraction, the fraction of sky-dome covered by clouds, and what cloud types are present. They have a clear view of low-level clouds but often cannot see higher clouds because lower clouds obscure them. Satellites have an unobscured view of high clouds but not lower clouds. Since human observers view the hemispheric sky-dome and include cloud sides as part of cloud cover, the cloud fraction they report is generally not identical to the cloud fraction reported by satellite. The difference is usually greatest for cumuliform clouds (cumulus and cumulonimbus) and least for stratiform clouds (stratus, stratocumulus, and fog). People have made regular visual observations of clouds for many decades in connection with weather forecasting, and it is possible to use these data to study past changes in cloud fraction and cloud type. Unfortunately human observers are not everywhere at all times on the Earth, and there are large regions of the globe, especially over the ocean, where no surface observations of clouds are available. However, a small number of satellites can observe clouds over every part of the globe several times a day. The International Satellite Cloud Climatology Project (ISCCP) (Rossow and Schiffer, 1999) provides cloud fraction and other information from July 1983 onwards. Unlike human visual observations, satellite observations cannot unambiguously identify morphological cloud type. Satellites instead report the reflectivity of clouds and the amount of emitted thermal radiation. This information is useful for determining how clouds affect the transfer of visible and infrared radiation within the atmosphere.

Satellites directly measure radiances, the amounts of visible, infrared, or microwave radiation upwelling in the direction of the satellite from various locations on the Earth. Meteorological satellites typically cannot resolve, i.e. distinguish, features less than several kilometers in size. The minimum area resolved by a satellite is called a pixel. Clouds smaller than the size of a pixel cannot be distinguished; the pixel is assumed either to be completely clear or completely cloud-filled. A common method for identifying clouds by satellite makes use of the fact that clouds are generally brighter and colder than the underlying surface. Consequently more visible radiation will be reflected and less thermal infrared radiation will be emitted by a cloudy pixel than by a cloud-free pixel. The threshold method used by ISCCP identifies pixels as cloudy if their visible radiance is greater by some threshold, and/or if their thermal infrared radiance is less by some threshold, than the expected visible and thermal infrared radiances of a cloud-free pixel at that location. Increasing the threshold will cause some true cloudy pixels to be identified as cloud-free, and decreasing the threshold will cause some true cloud-free pixels to be identified as cloudy. Clouds are also difficult to identify over snow and ice using the threshold method since the surface can be brighter and colder than the clouds. Satellite cloud fraction within a particular area is defined as the number of cloudy pixels divided by the number of all pixels. Cloud reflectivity is obtained from visible radiances using additional information about surface and other atmospheric properties at the location of the cloudy pixel. Because the amount of thermal infrared radiation emitted by a cloud is quantitatively related to its temperature, cloud top temperature is obtained from thermal infrared radiances, also using additional information about surface and atmospheric properties. Matching cloud top temperature to a profile of atmospheric temperature as a function of altitude provides cloud top height. Unlike the case for clouds observed from the surface, clouds detected by satellite are classified into altitude ranges according to the height of cloud top. ISCCP arbitrarily defines middle-level clouds as those with tops between approximately 3.2 km and 6.5 km above sea level and high-level clouds as those with tops more than 6.5 km above sea level.

Geographical distributions of cloud types

Figure C33 shows annual mean cloud fraction of all clouds over the entire globe, as reported by ISCCP. There is more cloudiness at middle latitudes than at low latitudes, except near the equator. The geographical variations in cloud fraction are largely related to the large-scale circulation of the atmosphere. Oceanic low-level trade winds flowing from the northern and southern hemispheres come together near the equator in the Intertropical Convergence Zone (ITCZ). Conservation of mass requires upward motion to balance the converging trade winds, and the upward motion causes air to cool and produce substantial cloudiness and precipitation. Latent heat release provides buoyancy that strengthens the upward motion. The climatological ITCZ can be identified by the relatively large fraction of middle- and high-level cloud near the equator, as seen in Figure C34. The ITCZ is broad over the Indian Ocean and western Pacific Ocean and is narrow and slightly north of the equator over the Atlantic Ocean and eastern Pacific Ocean. Over continents, the area of equatorial precipitation and middle- and high-level cloudiness is very widespread. Enhanced cloudiness marks tropical rainforests in South America and Africa. Air moving upwards in the ITCZ diverges and flows out toward the subtropical latitudes of the northern and southern hemispheres and then downward to complete the circuit of the trade winds, also called the Hadley circulation. The downward-moving air warms, thus decreasing its relative humidity. Figure C34 shows that very little middle- and high-level cloudiness exists over most regions between 10°

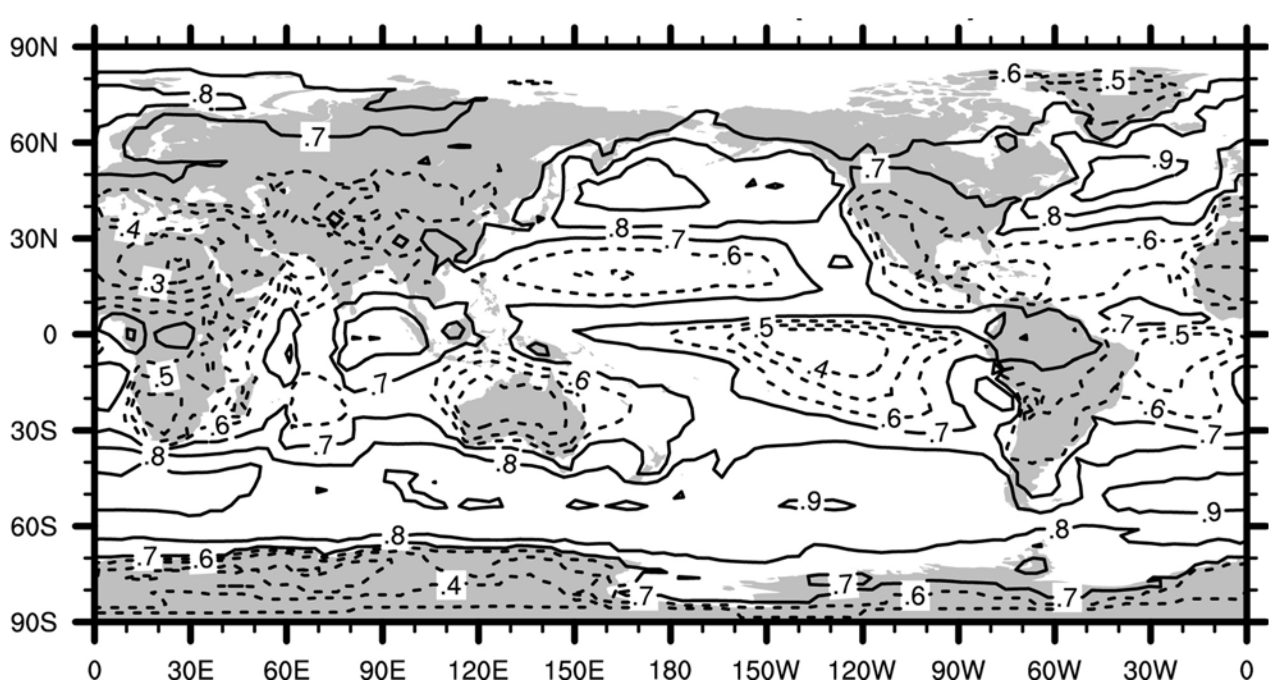

Figure C33 Annual mean cloud fraction of all (total) cloudiness obtained from the International Satellite Cloud Climatology Project and averaged from the years 1984–1999. Contour interval is 0.1, and values equal to 0.6 and lower are dashed.

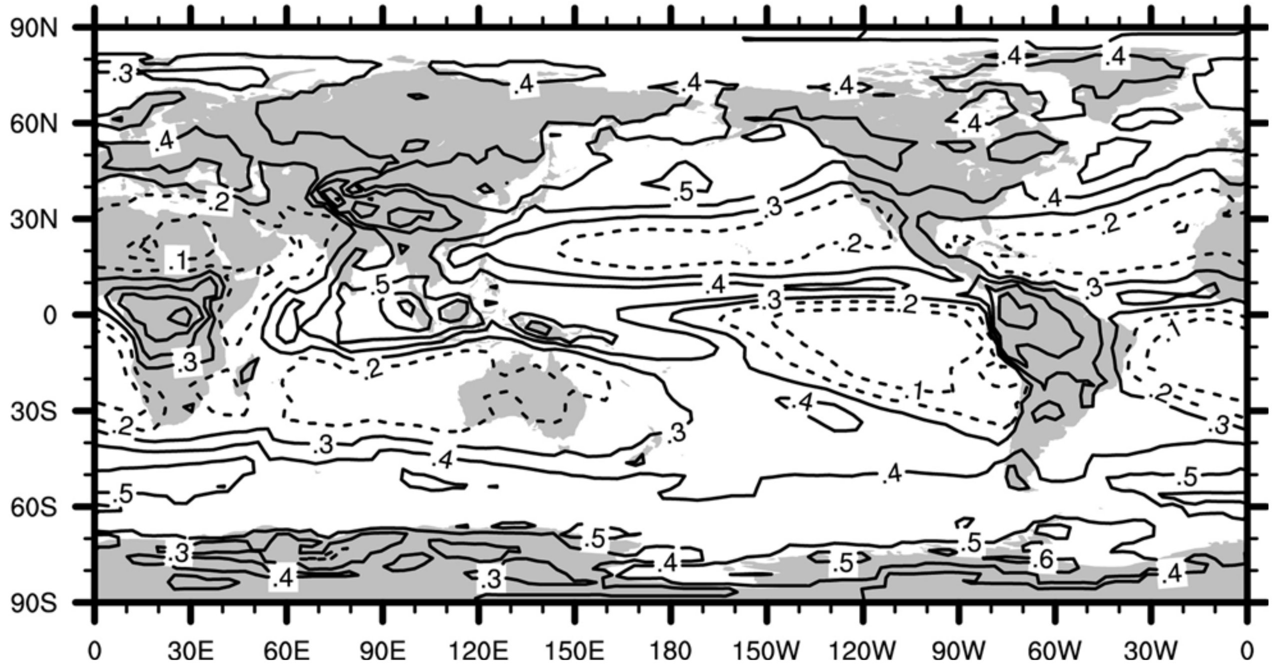

Figure C34 Annual mean cloud fraction of middle- and high-level cloudiness obtained from the International Satellite Cloud Climatology Project and averaged from the years 1984–1999. Contour interval is 0.1, and values equal to 0.2 and lower are dashed.

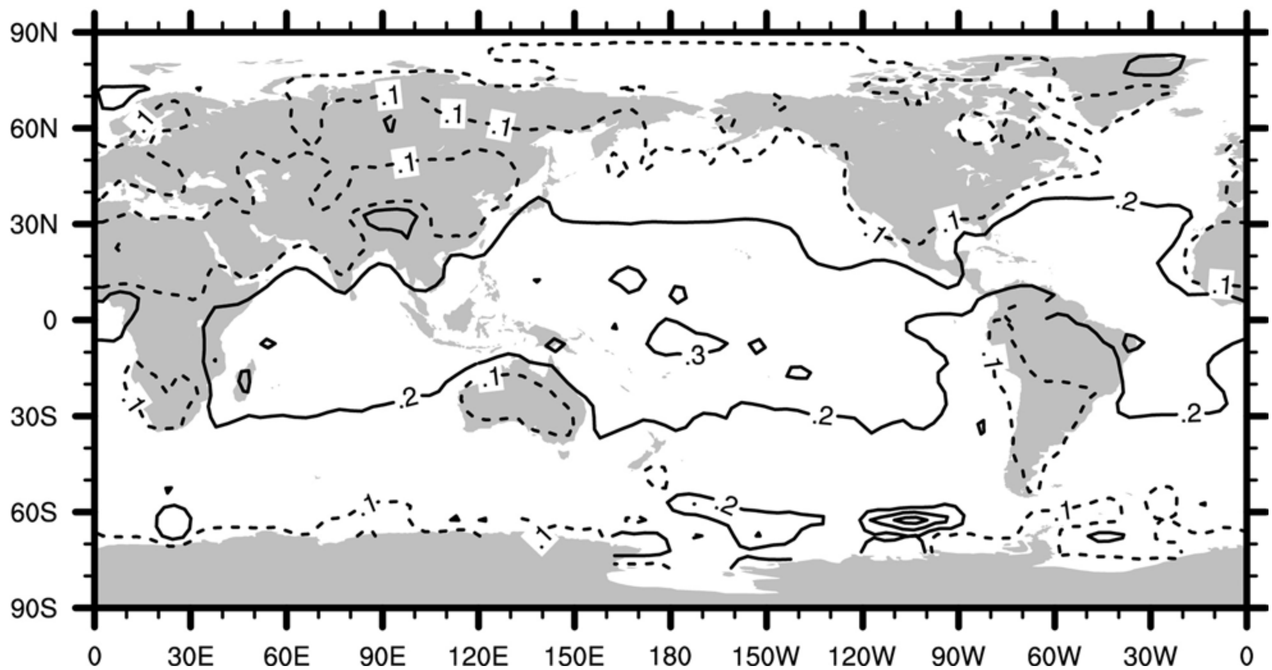

Figure C35 Annual mean cloud fraction of low-level cumulus and cumulonimbus cloudiness obtained from surface observations in the Extended Edited Cloud Report Archive and averaged from the years 1971–1996. Contour interval is 0.1, and the value equal to 0.1 is dashed.

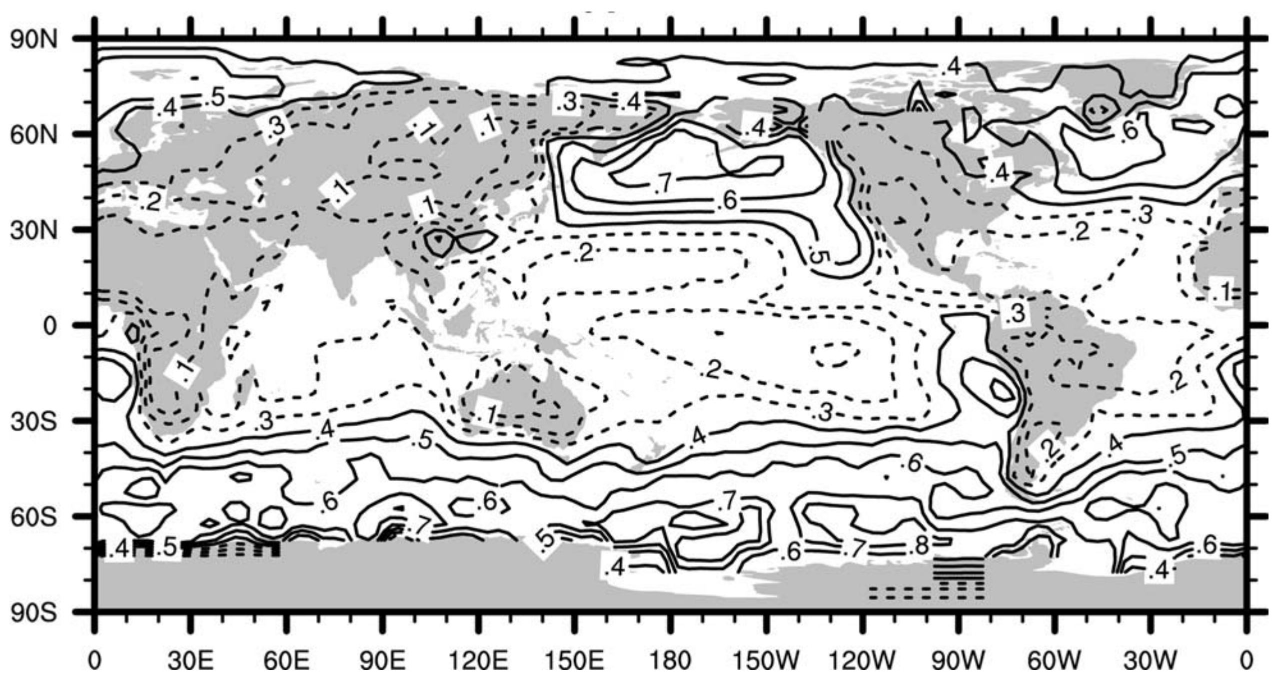

Figure C36 Annual mean cloud fraction of low-level stratus and stratocumulus cloudiness and fog obtained from surface observations in the Extended Edited Cloud Report Archive and averaged from the years 1971–1996. Contour interval is 0.1, and values equal to 0.3 and lower are dashed.

and 35° latitude. Middle- and high-level cloud fraction is greater poleward of 35°, especially over oceans. This is the latitude zone of the storm tracks, which are stronger over ocean than over land. Upward motion associated with extratropical cyclones passing through the storm track produces large amounts of cloudiness at middle and high levels.

Figures C35 and C36 show global distributions of low-level cumulus-type cloud fraction and low-level stratus-type cloud fraction, respectively. The source of cloud data for these plots was the Extended Edited Cloud Report Archive (Hahn and Warren, 1999), a collection of visual observations made by humans on ships or at land meteorological stations. Due to greater moisture availability, more low-level clouds occur over oceans than over continents. Cumulus-type clouds primarily occur equatorward and stratiform clouds primarily occur poleward of 30° latitude. A notable exception to this rule is the presence of large amounts of persistent stratocumulus clouds over eastern subtropical oceans, particularly off the coasts of California, Namibia, and Peru. In these regions the downward motion of the Hadley circulation is stronger, and the sea surface temperature is colder, than elsewhere in the subtropics. The marine boundary layer, the layer of the atmosphere directly affected by the ocean, is kept shallow by the strength of the descending branch of the Hadley circulation and kept cool and moist by the relatively cold underlying water. These conditions are conducive for the production of horizontally extensive layer clouds. Outside of stratocumulus regions, the tropical and subtropical ocean is warm and generates plumes of buoyant moist air that intermittently rise up from the surface. The plumes cool as they rise and form cumulus or cumulonimbus clouds when saturation is reached. Stratiform clouds occasionally occur over the tropical ocean in association with large cumulonimbus cloud systems. The cold oceans at middle and high latitudes are favorable for the production of stratiform clouds. Widespread uplift associated with storms also produces many stratiform clouds. Cumulus clouds sometimes occur during winter over middle-latitude oceans when cold continental air flows offshore over relatively warmer water and generates many buoyant plumes.

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Cross-references

Adiabatic Phenomena
Coastal Climate
Condensation
Lapse Rate
Latent Heat
Thunderstorms

COASTAL CLIMATE

For some distance on both sides of the 450 000-km-long world shoreline, meteorological processes and resulting climates differ from those farther seaward and landward. The coast is in essence a transitional zone between marine on one side and terrestrial on the other, a transitional zone or band that varies in total width as well as in the relative width of its two parts, land and water. The distinctiveness of this band varies with the amount and rate of meteorologic and climatic change across it. Although most of its characteristics are a blend of marine and terrestrial phenomena, there are a few characteristics, such as land and sea breezes and coastal fogs, that can be truly labeled coastal.

The climatic characteristics of the coastal zone most often have been considered a landward invasion of adjacent sea air and have been referred to as oceanic, marine, maritime, littoral, and coastal. The first three designations have been loosely used, and the last two appear only rarely in the literature. Maritime is the term that comes closest to our usage here of coastal (see Fairbridge, 1967). With the rapid evolution of the concept of the coastal zone, and especially because of the recent tendency to include the seaward segment of the coastal zone as one of its integral parts, the terms coastal meteorology and coastal climatology are now more appropriate than any of the others.

Climatic controls

Three of the most important climatic controls on Earth are variation of insolation with latitude, distribution of land and water, and surface configuration. Coastlines cross virtually every parallel on Earth except those immediately surrounding the two poles. Although coasts do not occur south of 78°S latitude because of the presence of Antarctica and its ice shelves, nor are they present north of 82°N latitude because of the presence of the Arctic Ocean, they nonetheless have a wide range of latitudinally influenced characteristics.

Whereas the juxtaposition of any two environments (e.g. forest/grassland, mountain/plain, and city/country) give rise to climatic modification, the land/water combination is the most distinctive on a worldwide basis. There are marked variations in the heat and moisture budgets of these two surfaces because of contrasts in albedo, evaporation rate, transparency, surface mobility, heat capacity, and sensible heat flux. Moreover, there are important modifications in circulation patterns because of variations in surface irregularities across the interface, including variations in the actual relief as well as in the mobility and spatial characteristics of roughness forms. Possibly the two most important differences are in wetness and roughness – the sea is almost always wetter and smoother than the land.

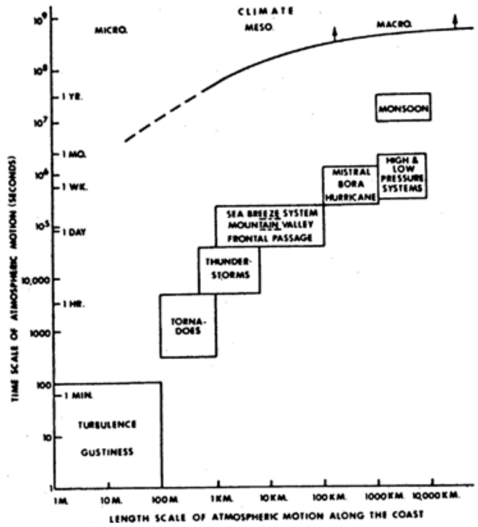

Figure C37 Time-length scales of atmospheric motions.

These controls are especially important in affecting atmospheric motion in the coastal zone. Pertinent motions have a great range in scale both spatially and temporally (Figure C37), and although all of these phenomena have some coastal expression, only a few have actually been considered from that standpoint. Studies to date most often have dealt with the impact of these levels of atmospheric motion (especially turbulence, sea breezes, and hurricanes) on the coastal environment rather than with the motion itself as a coastal phenomenon. Such an emphasis reflects the fact that coastal meteorology and climatology are integral parts of the total systems approach to the study of coastal environments.

Boundary-layer meteorology (microscale and mesoscale meteorology)

The boundary layer is the region of the atmosphere that is directly affected by friction caused by interaction with the Earth's surface. Transport in this layer in nearshore and estuarine environments is important from several points of view. For example, the wind stress or momentum flux is one of the most essential driving forces in shallow-water circulation. Heat and convection are the origins of some localized coastal weather systems. Sensible heat and water vapor fluxes are necessary elements in radiation and heat budget considerations, including the computation of salt flux for a given estuarine system. Experiments in these environments ranging from the tropics to the Arctic have produced a large number of conflicting drag and bulk transfer coefficients for both deep and shallow waters. There are several reasons for the discrepancies, but the most important is that early methods did not take into account the simultaneous contribution of wind, dominant waves, and atmospheric stability. This difficulty was removed by the development of a wind/wave interaction method of determining wind stress from commonly available wind and wave parameters (Hsu, 1976).

Aerosol transport

Atmospheric particles, particularly sea salts, have become an increasingly important subject for investigation in recent years.

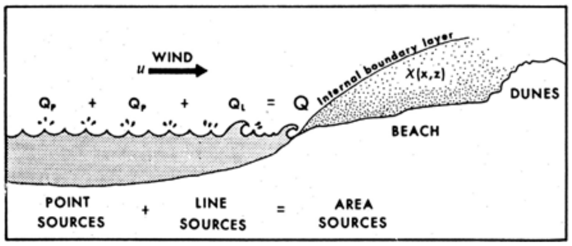

Figure C38 Schematic representation of point and line sources for sea salt and the effect of the internal boundary layer (from Hsu, 1977).

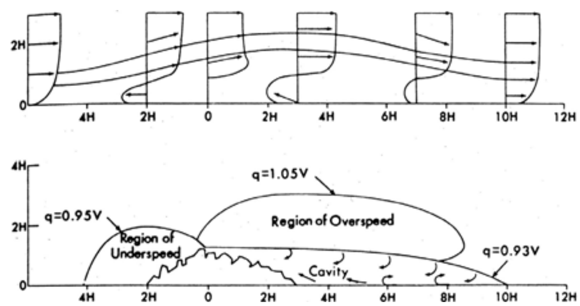

Figure C39 A basic flow model and a similarly shaped ice ridge. q is the local resultant mean velocity and V is the reference velocity in the uniform stream (from Hsu, 1977).

Sea-salt aerosols are a significant source of condensation nuclei. The quantity and quality of sea-salt particles deposited on land may be important in determining the physical and chemical characteristics of coastal soils and plants. The generation of aerosols depends on many meteorological and oceanographic factors. Among those in the coastal region are wind speed, direction, duration, fetch (which govern sea state and whitecap distribution), and subaqueous bathymetry (which controls the breaking wave condition in the surf zone). Aerosol contribution from these point sources (whitecaps) and line sources (surf zone) can be considered as originating from an area source (Figure C38). Experiments measuring the vertical distribution of sea salt in the atmospheric boundary layer over beaches have shown that on a coast influenced by synoptic onshore winds, such as a beach–dune complex, mixing depths are less thick than they are under sea breeze conditions (Hsu, 1977).

Sand dune and ice ridge air flow

When air passes over an obstacle it must adjust to a new set of boundary conditions. Major characteristics associated with this adjustment are the displacement zone, the wake, and the cavity. Measurements of the wind field in the region of coastal sand dunes in the tropics and ice ridges in the Arctic were combined with data from laboratory experiments in the preparation of a basic flow model. This model (Figure C39) is aiding the study of the modification of the structure of airflow over selected coastal topographies and is furthering the understanding of low-altitude atmospheric dispersion characteristics in the coastal zone.

Sea breeze

The sea breeze, a coastal local wind that blows from sea to land, occurs when the temperature of the sea surface is lower than that of the adjacent land (Defant, 1951; Hsu, 1970). It usually blows on relatively calm, sunny, summer days, and alternates with an oppositely directed, usually weaker, nighttime land breeze. As a sea breeze regime progresses, coriolis deflection causes the development of a wind component parallel to the coast. Coastal air circulations of this kind have been known and recorded since the time of Aristotle, about 350 BC, and have been studied more or less scientifically since the seventeenth century. Of all coastal meteorological phenomena, these have received by far the greatest amount of attention and are continuing to do so. For example, during the First Conference on Coastal Meteorology in 1976 about 40% of the papers dealt in some way with the sea breeze.

Synoptic meteorology and its coastal impact (macrometeorology)

Associated with the readjustment of airflow at the sea/land interface are other parameters such as temperature and humidity. As considered above at the microscale and mesoscale, these are quite distinctive. Although their effect is less obvious at the larger scale, they nonetheless have a significant impact on synoptic and planetary circulation systems.

Data support for the conduct of studies at this scale is scarce. Few meteorological data are routinely obtained at coastline stations; the major exception is at the all too few US Coast Guard stations where observations of temperature, precipitation, and wind are made. The most useful data, however, come from coastal airports, even though these are usually located several kilometers inland.

Such data have been used in the analysis of the weather input and environmental response in the coastal zone in Louisiana (Muller, 1977). This approach classified Louisiana coastal weather into eight inclusive synoptic types: Pacific High (PH), Continental High (CH), Frontal Overrunning (FOR), Coastal Return (CR), Gulf Return (GR), Frontal Gulf Return (FGR), Gulf Tropical Disturbance (GTD), and Gulf High (GH). Because each type can be identified on weather maps, climatic calendars were devised that are useful in the analysis of mean properties by months and seasons, climatic variation from year to year, and geographical comparisons of one coastal region with another (Muller and Wax, 1977).

Synoptic weather type changes have been found to be related to water level and salinity changes in the coastal marshes and estuaries of Louisiana (Borengasser et al., 1978). When the effects of astronomical tides were "filtered" from water-level data, it was found that water levels tended to drop during the 24-hour periods associated with a progression from the Frontal Gulf Return type to the Pacific High, Continental High, or Frontal Overrunning types. In contrast, water levels were shown to rise during other types of progression. Such analyses are only beginning but appear to hold promise for other nontropical coastal areas as well.

In recent years there has been increasing focus on coastal storms, classification (Dolan and Davis, 1992), synoptic patterns (Davis et al., 1993), and some of the environmental and economic impacts (Davis and Dolan, 1993). Very important environmental issues still not well resolved focus on questions about increasing or decreasing frequencies and magnitudes of

Table C19 Climatic characteristics related to coastal landscapes

		Climatic conditions[a]				
		Temperature (°C)		Precipitation		
Climatic type and percent of world coastline	Typical locality	Mean maximum warmest month (tm)	Mean minimum coldest month (tm)	Mean annual depth, cm (P)	Mean annual no. days $\geqslant 0.004$ cm (FP)	Winder concentration of precipitation[b] (R)
Rainy tropical (20%)	Inner tropics	30–32	19–23	198–310	134–185	14–49
Subhumid tropical (10%)	Border tropics	30–33	15–21	104–140	61–114	17–38
Warm semiarid (2%)	Tamaulipas, Venezuela	32–34	13–19	53–71	42–60	9–40
Warm arid (5%)	Horn of Africa, Sonora	33–37	11–20	13–25	10–32	36–94
Hyperarid (4%)	Cool-water coasts of subtropics	24–34	9–14	<5	1–4	42–100
Rainy subtropical (6%)	East coasts, lat. 20–35°	29–32	6–9	114–147	93–142	29–49
Summer-dry subtropical (7%)	Mediterranean	27–31	6–9	43–69	54–103	74–87
Rainy marine (1%)	W. coasts, Tasmania, New Zealand	17–20	4–7	109–206	166–187	51–69
Wet-winter temperate (2%)	Oregon, Washington	17–22	0–6	99–170	120–198	67–78
Rainy temperate (9%)	NE United States, W. Europe	20–27	−7–2	66–112	127–188	41–54
Cool semiarid (1%)	Bahia Blanca	19–31	−3–9	30–53	45–87	37–52
Cool arid (2%)	Patagonia	21–26	1–7	10–15	24–41	54–88
Subpolar (6%)	Gulfs of Alaska, Bothnia	15–23	−23–−13	46–104	106–184	32–50
Polar (25%)	Arctic Sea border	9–13	−34–−13	18–66	91–131	30–49

Source: From Bailey (1976).

[a] Each pair of numerals in the body of the table refers to data from climatic stations at the 25th and 75th percentiles of the frequency distribution appropriate to the climatic type and element. As only long-period means have been entered into the frequency distributions, the data in the table above show the spread in average conditions of climate in the most representative parts of the several climatic regions. Because approximately equal spacing was employed in the station network, it is also true that the data illustrate, for a given climatic type, conditions in about 50% of the aggregate length of coastlline affected by that climatic type.

[b] The winter concentration of precipitation is defined as the percentage of the mean annual total that falls in the winter half-year. October through March in the northern hemisphere, April through September in the southern hemisphere. The computation was not carried out for those places where the difference between the mean monthly temperatures of the warmest and coldest month was less than 3°C.

severe and extreme events in terms of global change. For tropical storms and hurricanes there is clear evidence that regional frequencies of events vary significantly along the Atlantic coast of the United States and appear to be associated with ENSO and North Atlantic Oscillation indices (Muller and Stone, 2001).

Coastal climates

Whereas the atmospheric processes thus far considered are measured in intervals of seconds to weeks, there remain longer time periods to be considered – periods representative of climate (Figure C39). Climate, as a synthesis of atmospheric events, is unique for each location on Earth. Yet vast areas have climates that are sufficiently similar to justify consideration together. In most classifications continental types are extended directly to the shore, so that their coastal ramifications are indistinguishable (Wilson, 1967).

The major exception to date is the work of Bailey (1960, 1976), who developed a classification of coastal regions through the utilization of data from coastal stations. Defining his climatic types so that they closely approximated the distribution of coastal vegetation, Bailey (1976) arrived at 14 types (Table C19). The relative length of coastline represented by each type varies from 1% and 2% for five types to 20% and 25% for rainy tropical and polar climates, respectively. Only about one-fourth of the coastline of the world has a temperate climate.

The longitudinal gradients of the 14 coastal climates in Bailey's classification are generally weak. In contrast, the climatic gradients across the coastal zone are frequently steep. That there is a distinction between oceanic, coastal, and continental conditions has been demonstrated by Bailey (1976) through comparison of the data in Table C19 with that from non-coastal areas. When the mean annual range of temperature is related to mean annual temperature, the intermediate nature of the coastal zone is evident (Figure C40). There are decided interhemispheric contrasts in the curves, contrasts that are especially evident in the seasonal swings (A of Figure C40) of temperature.

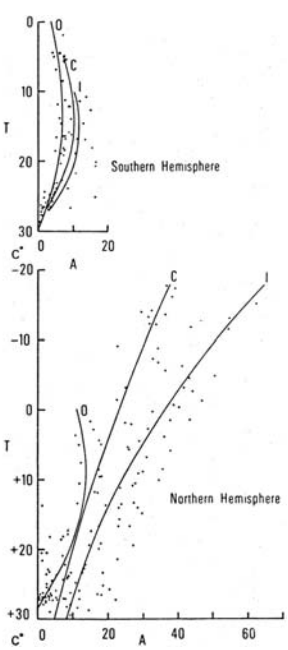

Figure C40 The relation between mean annual range of temperature (A) and mean annual temperature (T) in each hemisphere for oceanic (O), coastal (C), and interior (I) places (from Bailey, 1976).

Bailey's classification of coastal climates, like most climatic classifications, is based on temperature and precipitation. Dolan et al. (1972), however, used airmass climatology as their basis for separating the coastal climates of the Americas. The airmass characteristics they considered significant are: airmass seasonality, airmass source region, nature of the surface over which the airmass moves, and the confluence of airstreams at the coast. Further conditions strengthening this approach are: each airmass possesses a set of secondary characteristics, including the traditionally utilized meteorological elements; the airmass is modified upon crossing the coast; and the boundary between airmasses (i.e. a front) is distinct and separates natural climatic complexes. Although the two approaches are different, the mapped distributions of coastal climates are nonetheless similar in both.

The future

Coastal meteorology and coastal climatology are in their infancy as research disciplines, but the future for both looks bright. Many research organizations are beginning to place greater emphasis on the meteorology and climatology of coasts, albeit an emphasis that is still related mostly to the demand for data by other coastal sciences. In addition to the topics briefly discussed here, research is under way on storm surges, upwelling, tidal stirring, coastal fog, coastal dunes, and coastal frontogenesis, among others.

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Cross-references

Coral Reefs and Climate
Lakes, Effects on Climate
Land and Sea Breezes
Local Climatology
Maritime Climate
Ocean Circulation
Sea Level Rise

COMMERCE AND CLIMATE

The saying that "everyone talks about the weather but no one can seem to do anything about it" may seem like an appropriate statement for the vast majority of people even today. In many sectors of the nation's economy the effect of weather and climate variations are well known yet the "weather" variable is the least understood in terms of all factors involved in weather-sensitive decisions (Maunder, 1970). Enhanced use of climate information has come about in recent years through improved understanding of the weather and climate, enhanced technology, and an interdependent global community that is ever more reliant on decisions impacted by weather and climate issues.

The "value" associated with the use of climate information in aspects of "commerce" (herein related to retail sales, the commodities market, and insurance sector) has been difficult to determine over time due to the lack of available data and investigations by either climatologists or businesses. Since the 1970s atmospheric scientists in three sectors – government agencies, universities, and commercial meteorology firms – have become increasingly engaged in activities with business decision makers (Smith, 2002). What triggered this increased interest in knowing more about the benefits and costs associated with weather and climate? First, as climate data and information became more timely and accessible, users could easily incorporate them into timely weather-related decisions. Second, since the 1950s weather forecasts at all time scales, hours to seasons, have improved and become more accurate. Third, weather-related losses in the United States and around the world have increased dramatically, forcing government officials and insurance risk managers to re-examine climate risks and adjust accordingly. Recent studies (Dutton, 2002; Murnane et al., 2002) have indicated that the value of weather and climate information has increased and will continue to do so in this highly urbanized, industrial, and interdependent world. These new challenges have created new areas of growth for atmospheric scientists.

For the United States, Dutton (2002) estimated that nearly one-third (approximately $3.0 trillion in 2000) of all private industry activities were weather-sensitive. Commercial activities (retail trade) accounted for $894 billion and he estimated that nearly 100% of those activities were weather-sensitive. Changnon (1999) noted that United States retail sales increased nearly $5.6 billion as a result of warmer than average conditions across the northern tier of states during the El Niño winter of 1997–1998. There is uncertainty in the economic value of weather in commercial activities but it is clearly significant.

Increased understanding of oceanic/atmospheric interactions and their relationship to regional climate variability around the world is considered one of the major climate advances since the 1970s. As a result, monthly and seasonal climate outlooks, which provide users with probabilistic forecasts of temperature and precipitation anomalies, are now being integrated into financial strategies used to hedge weather risks in energy, retail, and other industries (Dutton, 2002).

Retail sector

When examining the influence of weather/climate on the retail sector two time scales have been considered. Many studies have examined whether short-term day-to-day sales of a specific item are influenced (and to what degree) by weather conditions (whether it rains or not). For example, Zeisel (1950) examined the relationship of summer hot periods to beer sales, while Linden (1959) analyzed the influence of precipitation events on umbrella sales. Longer-term seasonal sales have generally been related to seasonal climate (either temperature or precipitation) anomalies such as those associated with El Niño (Changnon, 1999).

In any retail assessment it is difficult to document non-weather factors such as whether a shopper has the resources (money) to make purchases. To minimize this uncertainty in order to draw a clearer picture of how retail sales associated with one item are related to specific weather events or climate anomalies, studies have examined the retail–weather relationship over a number of years. This approach provides a range of outcomes and allows one to examine sales during warm versus cold periods and wet versus dry periods (Linden, 1962).

When examining the weather influences on all retail sales it is also important to consider that regional differences in weather across a country as large as the United States may in fact balance out regional profits in one area and losses in another. For example, during the 1997–1998 El Niño winter retail sales for the United States were $4 billion above average; however, the gains were not experienced nationwide. Warmer and drier winter conditions along the northern tier of the United States helped to enhance retail sales as people were more likely to venture out of their homes to shop, whereas retail sales from southern California eastward along the Gulf Coast and into Florida sagged due to cooler, cloudier, and wetter winter conditions (Changnon, 1999). Similar regional increases in retail sales were experienced in the eastern two-thirds of the United States during the record warm winter of 2001–2002 (Changnon and Changnon, 2002). On the other end of the scale, the record cold and snowy winters of the late 1970s in the Midwest and Northeast United States had a largely negative impact on retail sales (exceptions being winter clothing) in these regions (Changnon and Changnon, 1978). In summary, when examining national retail sales, one must keep in mind that there will often be regional "winners" and "losers" as some aspects of the retail sector are enhanced by pleasant weather while are others hurt by a specific weather event or climate anomaly.

Commodities sector

This sector has witnessed some of the most dramatic changes in how it values weather and climate information. The commodities market involves trading of futures commodities such as oil, grains, meat, etc. (purchase now for something expected at a later date). Agricultural commodities have always been directly impacted by two conditions: present and forecasted weather events (Dutton, 2002). Mitigative strategies, such as contracts between different parts of the decision chain (resource, producer, processor, distributor, consumer) and a third party, were developed as a type of insurance plan for traders (Figure C41). Any commodities trader knows that the "weather factor" is ever changing, and that because commodities are grown and traded worldwide, one must examine weather issues and impacts around the globe (Davis, 2002). In the United States the weather-sensitive sector dominating commodities trading in the mid-1980s was agriculture. However, due to the deregulation of oil, natural gas, and electricity industries in the early 1990s, the need for weather and climate information in the United States energy sector has since increased and is now equal to agricultural needs.

Many decision makers in weather-sensitive industries realize that in this interdependent world they must better manage the weather and climate "risk" to reduce profit volatility and deal with increased competition. These efforts are becoming more comprehensive and require quantitative answers (Dutton, 2002). As part of the commodities sector, the weather risk market has generated considerable interest in new approaches to management of the financial aspects of weather and climate risk. The recent development of weather "derivatives", as a financial instrument to deal with seasonal climate anomalies, requires the use of very accurate climate information (Dischel, 1999, 2002; Dutton, 2002).

How does a weather derivative (contract) work? For instance, you own a Chicago company that is negatively impacted by very cold winter temperatures and desire protection. Once the risk associated with a particular seasonal climate anomaly (e.g. average winter temperatures in Chicago at 6°F below average) is determined, a contract can be established between you and another party. That is, one party concerned about this weather risk purchases an insurance-type policy that will pay (cover losses) if the average winter temperature is $\geq$6°F colder than average. According to the Weather Risk Management Association (WRMA, 2002) the total national value of seasonal weather derivatives executed between parties over-the-counter was about \$2 billion per year in 1998–2000 and \$4 billion in 2001. This climate-based tool has dramatically altered how those involved with the commodities markets manage weather and climate risks and, as a result, it has opened new and extremely important opportunities for atmospheric scientists (Dutton, 2002).

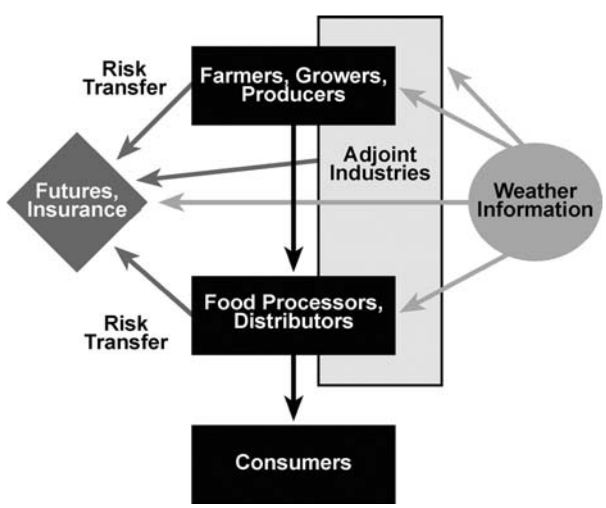

Figure C41 Flows of agricultural products, weather information, and transfers of financial risk (Dutton, 2002).

Insurance sector

Since the late 1950s insurance companies have asked atmospheric scientists to assist them with the difficult task of determining weather-related risks to both crops and property. Initially, crop-hail insurance companies, which had experienced severe and extremely variable losses during the 1950s, wanted to gain a greater knowledge of the hail risk they faced at the local and regional scale. Using a statewide index of annual hail frequency and the annual "loss cost" value calculated for each state by the insurance sector, the hail risk calculations were completed (Changnon and Stout, 1967). These initial analyses, which were updated in the late 1990s (Changnon and Changnon, 2000), showed that, except in the western Great Plains and northern Texas, the hail risk had generally decreased or not changed in most agricultural regions of the United States (Figure C42). The research conducted by these meteorologists has assisted insurance decision makers in establishing the degree of risk and hence the premiums to be levied. Importantly, atmospheric scientists can help explain the causes of weather-related losses.

When examining property and casualty-insured losses related to weather catastrophes since the 1950s, one is impressed with the steady increase in total losses (Figure C43) and may have initially associated this increase with global warming. However, when losses for each catastrophe were adjusted for changing conditions (changes in property values, cost of repairs, growth in the size of the fixed property market, and relative change over time in the share of the total property market that was insured against weather perils), the trend in losses looked very different (Figure C44). Years with high values occurred when a major tropical storm/hurricane hit the United States (e.g. estimated losses for Hurricane Andrew in 1992 adjusted to 1996 dollars were \$30.5 billion).

Assessment of climate extreme data (hail, tornadoes, thunderstorms, floods, hurricanes, etc.) and insurance data led to important findings. Although annual insurance losses have

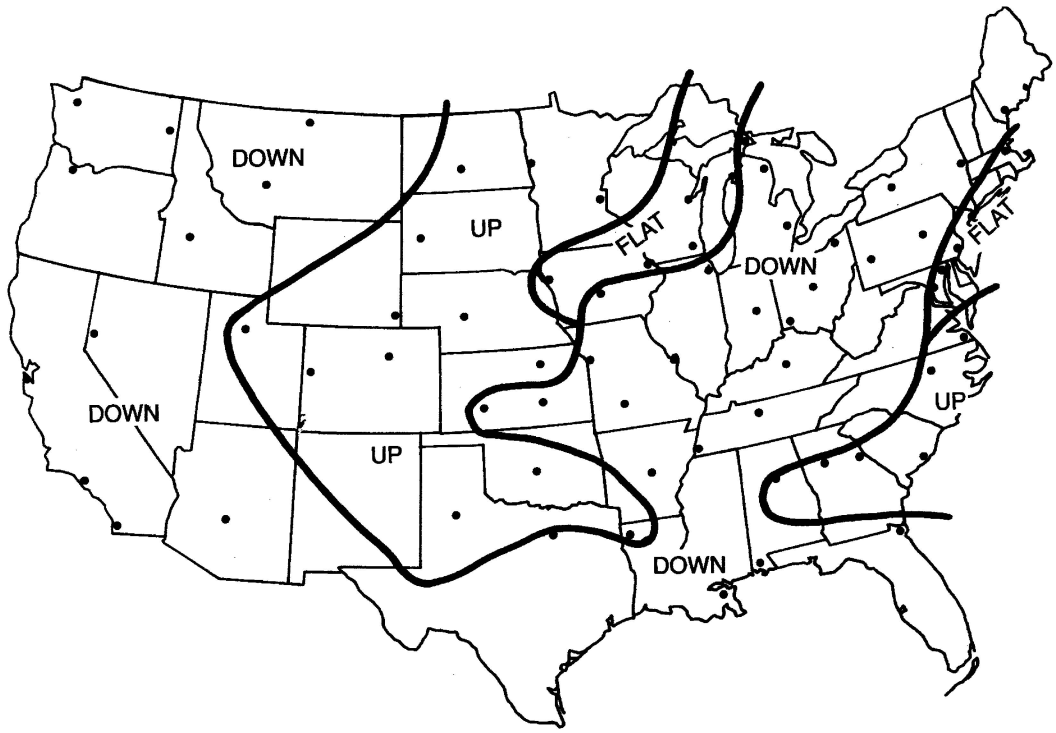

Figure C42 Regions defined based on linear trends of the 100-year hail-day values. Up is for increasing trends significant at the 5% level, flat is essentially no trend up or down, and down is for decreasing trends significant at the 5% level (Changnon and Changnon, 2000).

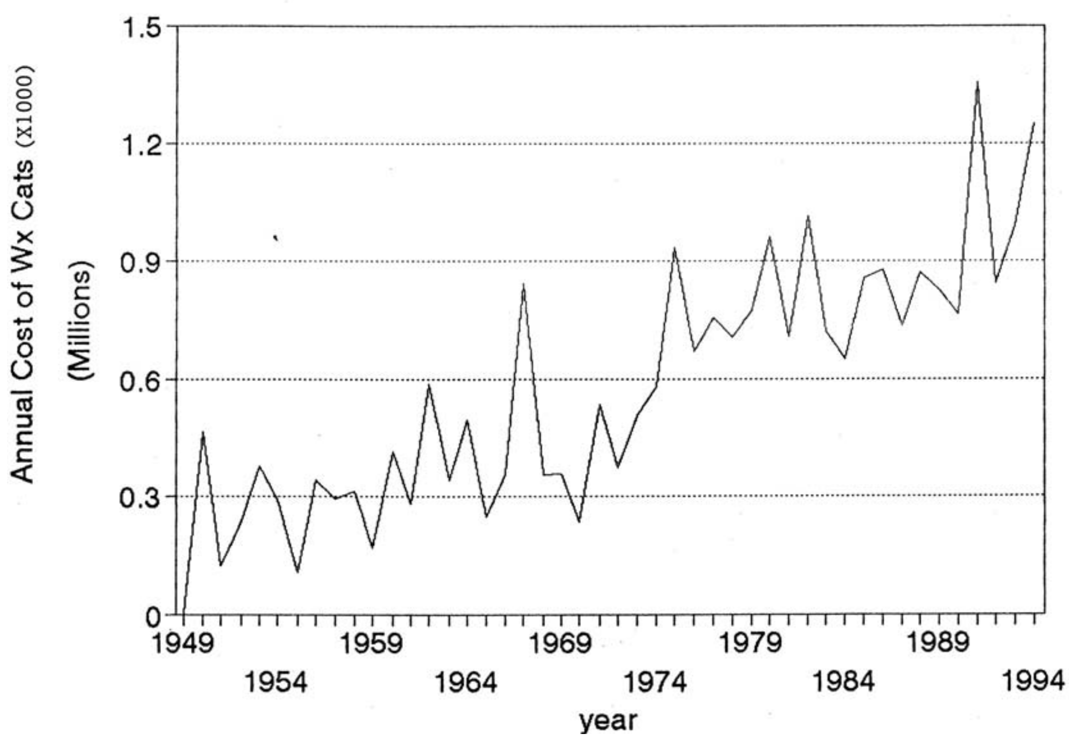

Figure C43 Annual losses produced by $10 million to $100 million catastrophes during 1949–1994 (Changnon and Changnon, 1998).

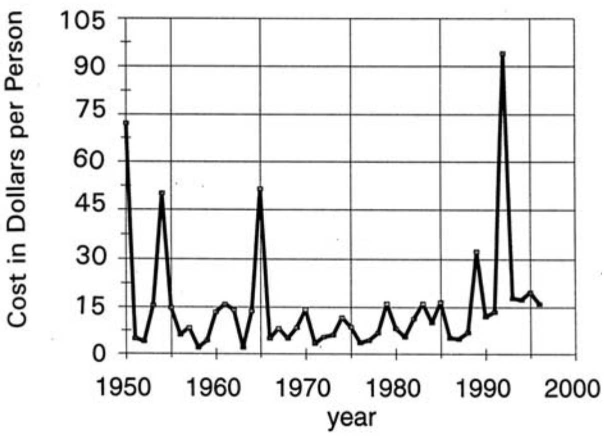

Figure C44 Annual losses caused by catastrophes causing < $10 million in insured losses normalized by dividing annual losses by the annual US population during the 1950–1996 period. Values are dollars per person (Changnon and Changnon, 1999).

increased, and will continue to do so into the future, the changes during the past 50 years appear to be more closely related to societal changes than to changes in weather extremes (Changnon et al., 1999). Most climate extremes in the United States do not exhibit steady, multidecadal increases over time, as found in the insurance losses. Societal (demographic) changes – moving to coastlines, cities, and mountains) – explain some of the loss, along with growth of population and wealth. These societal shifts suggest that the "targets" for damaging weather have changed and have become more vulnerable to huge financial losses (Changnon and Changnon, 1999). Furthermore, more climate "impact" data are needed if climatologists working with insurers are to continue to understand and forecast insured losses.

Summary

There are a number of messages that need to be learned by climatologists if they are to become more effective in relating weather and climate information to commercial activities described herein. Many of the fundamental points listed below have been identified earlier by other atmospheric scientists (Maunder, 1970, 1973). However, given increased knowledge of the atmospheric sciences, enhanced technology, and increased loss amounts related to weather catastrophes, a number of these fundamental points have been modified to reflect current issues. The overall theme of this list is that interactions and research conducted with and for commercial business decision makers (Figure C45) should be "user-centered" (Changnon, 1998). The concept that atmospheric scientists totally understand (without user interaction) and solve the weather-related issues facing the commercial sector must be tossed aside. It is time climatologists and business decision makers got together. Following are five lessons to enhance commerce in the future.

1. Climatologists must work closely with the user in this changing world.
2. Climatologists should understand the value of climate information and develop relevant tools for commercial decision makers.

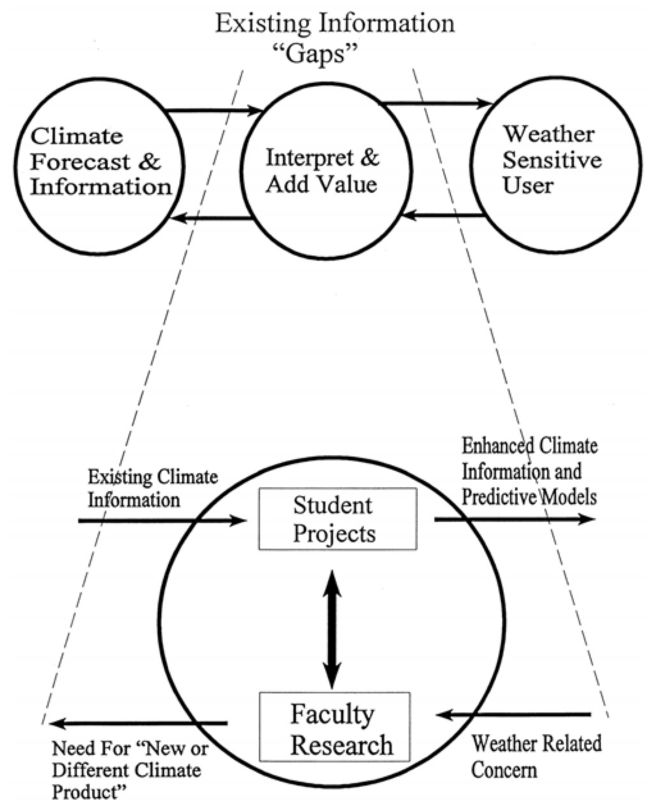

Figure C45 Bridging the information "gap" between climatologists and users (Changnon, 1998).

3. Climatologists need to separate climate variations from those of nonclimate factors involved in decision models.
4. Climatologists need to understand that uncertainty exists in any climate-related decision and this must be explained to users.
5. Climatological products need to change as the commercial world evolves.

Increased resources (public and private) for maintaining important climate databases and information are necessary if long-term benefits are to occur. Due to the dynamic nature of both the atmosphere and commerce, true benefits will be found only through long-term relationships. Basic reference unbiased climate datasets need to be maintained by the government to help users determine and minimize weather risks (Murnane et al., 2002).

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Cross-references

Applied Climatology
Agroclimatology
Climate Affairs
Climate Hazards
Tourism and Climate

CONDENSATION

Condensation is the phase change of water from its vapor to liquid. When moist air is cooled sufficiently, it attains a temperature at which saturation occurs – the dewpoint. Further cooling produces supersaturation, a condition in which relative humidity exceeds 100%. Supersaturation seldom occurs in the atmosphere, therefore condensation is the result of cooling. The condensed water may appear as dew or frost on the ground or as a suspension of water droplets in the air, fog or cloud.

Table C20 Heat transfers associated with phase changes of water

Phase change	Heat transfer	Type of heat
Liquid water to water vapor	540–590 cal absorbed	Latent heat of vaporization
Ice to liquid water	80 cal absorbed	Latent heat of fusion
Ice to water vapor	680 cal absorbed	Latent heat of sublimation
Water vapor to liquid water	540–590 cal released	Latent heat of condensation
Liquid water to ice	80 cal released	Latent heat of fusion
Water vapor to ice	680 cal released	Latent heat of sublimation

Condensation is brought about in three ways: (1) adiabatic expansion, which occurs when air rises to progressively lower pressure levels; (2) contact with a cold surface, for example, if the temperature of the surface is below the dew point, condensation may occur on that surface; and (3) mixing, for example, when two moist airmasses of very different temperatures are mixed, condensation may result.

If the air were perfectly clean, supersaturation of several hundred percent could occur. Air, however, contains impurities that act as nuclei on which condensation can occur. These condensation nuclei can be conveniently considered in three categories:

1. Aitkin nuclei with radii between 5×10^{-3} and 2×10^{-1} μm.
2. Large nuclei with radii between 0.2 and 1 μm.
3. Giant nuclei with radii greater than 1 μm.

The sizes and concentration of the condensation nuclei influence such factors as visibility and the precipitation process.

The condensation process, as a phase change, is responsible for the transfer of heat energy in the atmosphere. When water melts or in vaporized heat is absorbed. When the reverse process – condensation – occurs heat is released. Table C20 provides a summary of the energy exchanges that occur.

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Cross-references

Adiabatic Phenomena
Atmospheric Nuclei and Dust
Evaporation
Evapotranspiration
Latent Heat
Phase Changes

CONTINENTAL CLIMATE AND CONTINENTALITY

A continental climate has characteristics associated with areas within a continental interior as opposed to coastal locations under the moderating influence of water. Among the most significant of these climatic properties are large annual temperature ranges, which can be attributed to the distinctive thermal differences between land and water. Heat capacity is the ratio of the amount of heat energy absorbed by a substance to its corresponding temperature increase. Specific heat refers to the amount of energy necessary to raise the temperature of an exact mass of a substance by a certain quantity. It takes the energy of 4190 joules to increase the temperature of pure water by one degree Celsius. Thus, water has a high specific heat causing it to cool and heat slowly. Conversely, land has a low specific heat (five to six times less than that of water) causing it to cool and heat faster than water. Additionally, land surfaces lack the evaporative cooling effect of water and, unlike water bodies that distribute heat energy both horizontally and vertically, the solar radiation received by land heats only a shallow surface layer. As a result of these characteristics, continental locations tend to have cold winters and warm to hot summers.

Climatic characteristics

Continental climates also have large diurnal temperature ranges. Because these regions tend to be located in the interior of a continent, they are far removed from sources of moisture, reducing the availability of water vapor and subsequent cloud development. Clear skies allow insolation to quickly heat the land surface by day, causing temperatures to rise, while at night the absence of clouds enhances radiational cooling as temperatures rapidly decrease. Reno, Nevada, is an example of a continental location with a large diurnal temperature range. The mean maximum temperature in July is 92˚F; however, the average minimum temperature during the same month is 47˚F, yielding a daily range of 45˚F.

A lack of significant moisture sources also causes continental climates to have relatively low annual precipitation amounts. Maritime airmasses that originate over distant water bodies typically lose much of their moisture before arriving at a continental interior. However, higher temperatures during summer increase potential evapotranspiration, thus contributing water vapor to the regime and improving the possibility of precipitation. Table C21 lists the mean annual temperature, annual temperature range, and annual precipitation of selected continental stations.

Continentality

Continentality refers to the degree to which a place on the planet is affected by the temperature and moisture characteristics of a large, interior landmass. With the exception of latitude and its effect on solar radiation receipt, continentality is the most meaningful climatic control, and it is practical, for certain purposes, to separate every climate into a maritime or continental category (Driscoll and Fong, 1992). Further, Kopec (1965) asserts that it is possible to characterize a location based on its proportion of continentality, that is, its rank between perfect continentality and the total lack of continentality. D'Ooge (1955) notes that continentality quantifies the degree to which a region is influenced by continental characteristics. In the classical sense of the term, Stamp defines continentality simply as a climate with a wide range of diurnal and annual temperatures, as in the geographical center of a continent (Oliver, 1970).

The degree of continentality at a particular location is typically measured through the use of a derived climatic index. For example, the index of continentality at Prairie du Chien, Wisconsin, is 50%, which implies that the effect of continentality on the local climate is 50% greater at Prairie du Chien that on the island statio of Pago Pago (Samoa) where the index of continentality is 0%. The converse of continentality is "oceanicity." A value of 0 represents. the climate on an Earth completely covered with water (Kopec, 1965).

Continentality indices

Several climatic indices that account for the daily or annual range of temperature and which include some allowance for latitude have been derived. One of the earliest attempts to assign a numerical value based on geographic location and temperature came from Forbes (1859), who correlated the dependent variable, temperature, with the independent variable, latitude, in a formula expressed as:

$$T = -17.8° + 44.9° \cos^2(\phi - 6°30')°\mathrm{C}$$

where T represents temperature and ϕ represents latitude.

Table C21 Temperature, precipitation, and continentality of selected stations

Station	January mean temperature (°C)	July mean temperature (°C)	Annual range (°C)	Continentality value (%)	Annual precipitation (cm)
Beijing, China	−04.4	26.1	30.5	54	62.23
Calgary, Canada	−10.6	16.8	27.4	39	42.42
Fargo, North Dakota	−13.9	21.7	35.6	58	47.45
Fort Simpson, Canada	−27.2	17.2	44.4	65	33.27
Irkutsk, Russia	−21.1	15.6	36.7	57	37.59
Moscow, Russia	−10.0	18.9	28.9	40	62.99
Sverdlovsk, Ukraine	−17.8	16.8	34.6	50	42.41
Verkhoyansk, Siberia	−50.0	13.3	63.3	96	13.46

In 1905 Kerner developed a measure of the maritime influence on a location, or oceanicity, which he referred to as the thermoisodromic ratio. Kerner's formula is expressed as:

$$O = 100[(T_o - T_a)/A]$$

where O represents oceanicity, T_o represents the monthly mean temperature for October, T_a represents the monthly mean temperature for April, and A is equal to the average annual temperature range (Oliver and Fairbridge, 1987).

Among the earliest indices specific to continentality is that of Gorczynski (1920), whose formula is expressed as:

$$K = 1.7(A/\sin\phi) - 14$$

where K represents continentality, A is the average annual temperature range, and ϕ is latitude.

Conrad (1946) later amended Gorczynski's equation, suggesting that the sine of the latitude plus a constant of 10 would allow for use of the index at low latitudes. Conrad's index represents a reliable formula, which is widely accepted and expressed as:

$$K = 1.7[A/\sin(\phi + 10)] - 14$$

where K represents continentality, A is the difference between the mean temperature of the warmest and coldest month in degrees Celsius, and ϕ is station latitude in degrees.

Conrad's formula is perhaps the most frequently used of all continentality indices. Based on temperature records published in *Climatological Data* for 1952, Fobes (1954) applied Conrad's index in an assessment of the continentality of New England. Conrad personally assisted in the study, and the researchers concluded that northern stations in Maine possess the highest continental index values in the region with the effect of continentality diminishing as one approaches the coasts. Index values for New England range from 30% along the coasts to 50% in the north.

D'Ooge (1955) examined the western United States using Conrad's index of continentality and temperature data from 1952. He found the maritime influence to be much greater than in New England with the 40% continentality values extending approximately 500 miles inland. The highest values are in eastern Montana and North Dakota (60%). However, the values are still less than the 100% calculated by Conrad (1946) for Verkhoyansk, Siberia, which represents the maximum continentality value on the planet.

Trewartha (1961) applied Conrad's index to an analysis of North America and found continentality to be much weaker on the West Coast than on the East Coast, with continentality values of 5–10% for the Pacific Coast, 25–30% along the Gulf Coast, and 30–40% along the Atlantic Seaboard. The core area of maximum continentality (60%) is located west of Hudson Bay in central Canada. Figure C46 depicts continentality values of North America as derived by Conrad's index.

Mapping continentality

Using data from the United States Weather Bureau for the period from 1931 through 1952, and Canadian data for the period from 1921 through 1950, Kopec (1965) analyzed continentality around the Great Lakes. Conrad's index reveals that the highest values for the region (60%) can be found northeast of the Great Lakes. As expected, the Great Lakes act as a moderating influence on the surrounding region with the lowest continentality values located along the shores. Kopec (1965) explains that the Great Lakes delineate a zone of reduced

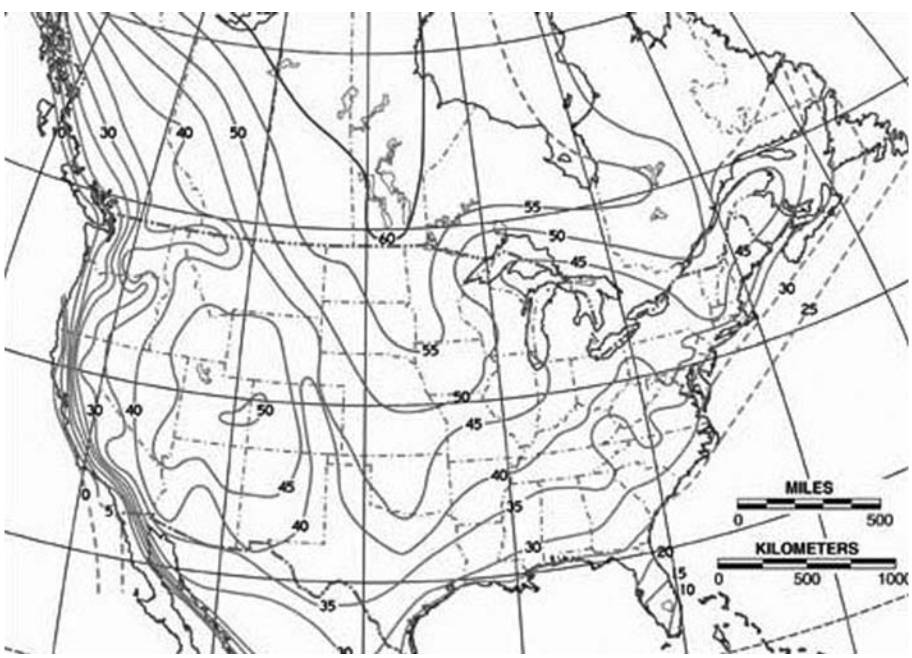

Figure C46 Continentality values of North America (after Trewartha, 1961).

continentality values situated within a region that would otherwise represent the core of continentality for North America. In Siberia a comparable role is played by Lake Baikal. Even small lakes, as in Switzerland, a mild effect is noticeable.

More recently, Conrad's continentality index is among the formulas selected for a major study conducted by the Climate Criteria Examination Team (CET). The CET acts under the auspices of the Atmospheric Radiation Measurement Program (ARM), a multi-laboratory, interagency program created in 1989 with funding from the US Department of Energy (DOE). The CET is tasked with identifying and characterizing the geographic regions of the world based on the climatic attributes of latitude, seasonality, and continentality.

Despite its widespread use, the inclusion of the sine of latitude in continentality indices has been criticized because it assumes that contrasts in seasonal radiation receipts are approximately distributed such that dividing by the sine of the latitude negates the influence. This leaves aridity and continentality as the only remaining controls over annual temperature range. However, Driscoll and Fong (1992) insist the technique is flawed since seasonal radiation does not increase monotonically and have derived their own formula based on residuals from the regression line of annual temperature range on latitude. Their index yields a scale for North America that ranges from -10 along the coasts to $+10$ in the heart of the continent. However, the researchers acknowledge that, for regional studies in which the latitudinal area is limited, the use of sine ϕ is essentially correct, and the results are not necessarily invalid (Driscoll and Fong, 1992).

Others have attempted to quantify continentality without including the sine of the latitude in the formula. Oliver (1970) applied an airmass evaluation to the concept of continentality using the formula:

$$K = L \cos A$$

where K is continentality, L is the length of the long axis (in millimeters), and A is the angular deviation away from the vertical on a climograph. Oliver concluded that a continental climate is one that experiences a large daily and annual range of temperature due to the proportionate preeminence of continental airmasses. Despite the ambiguity often associated with the classification of climatic regimes, continentality remains a viable concept that has inspired a wealth of research.

Richard Snow

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Cross-references

Airmass Climatology
Asia, Climates of Siberia, Central and East Asia
Coastal Climate
Maritime Climate
Temperature Distribution

CORAL REEFS AND CLIMATE

Coral reef ecosystems are found in tropical marine environments, and extend between about 30° north and south of the equator. Optimal reef development occurs in clear, nutrient-poor waters where winter temperatures remain above 18°C. Coral reefs are unique in having a strong geological component (calcium carbonate, or limestone) as well as an ecological one (the coral reef ecosystem). Coral reef structures are built by the coral reef ecosystems themselves. Corals, calcareous algae, and many other organisms secrete limestone skeletons that progressively accumulate into complex three-dimensional structures that can attain tens of meters in thickness over a period of several thousand years. The coral reef ecosystem exists as a living veneer over these structures.

Although corals are animals, their success in reef-building is largely founded on their strong symbiotic relationship with microscopic algae, called zooxanthellae. Almost all reef-building corals require this symbiosis to survive. The zooxanthellae grow directly within the otherwise transparent coral tissue, fueled by photosynthesis and benefiting from nutrients provided from the coral animal. The coral animal in turn benefits from the carbohydrates produced by the zooxanthellae. This internal recycling allows corals to thrive in an otherwise nutrient- and food-poor environment. Corals which have these symbiotic algae grow and secrete calcium carbonate much faster than those that do not.

Early predictions of how climate change would affect coral reefs were optimistic. Because coral reefs thrive in warm waters, an expansion of the tropics during global warming was considered a probable boon for coral reefs, allowing them to migrate poleward with the 18°C isotherm. Over the last few decades the negative effects of climate change on coral reefs have dwarfed this optimism.

The most immediate problem facing coral reefs in a high-CO_2 world is global warming, which has so far proved to be damaging to corals and many other reef dwellers. The phenomenon of temperature-induced coral bleaching is the most serious climatic threat to coral reefs. Bleaching is a natural stress response of corals and other organisms that host zooxanthellae, whereby the host organism expels the algae and becomes colorless. A rule of thumb is that corals and many other organisms bleach once temperatures exceed the normal maximum temperature by 1–2°C. Many corals can and do recover from bleaching if the offending conditions are short-lived and of moderate intensity. Until the 1980s, bleaching events were rare and isolated events. In

1982–1983, mass coral bleaching was observed in the eastern Pacific, which resulted in nearly total coral mortality on Galápagos reefs. Mass bleaching events have become increasingly common and widespread during the past two decades, and nearly all have been associated with elevated temperatures. Secondary climate factors in coral bleaching and mortality include water circulation and light. Increased light penetration, particularly of ultra-violet radiation, intensifies bleaching. Vigorous mixing by winds, tides, and currents tends to keep sea surface temperatures in check, and to reduce light penetration. Bleaching in many areas has often been associated with several climatic patterns that occur during El Niño events: decreased cloud cover and increased insolation; increased sea surface warming; reduced winds; and reduced upwelling. In 1997–1998 an estimated 16% of the world's coral reefs were destroyed by bleaching.

Global warming also leads to other changes in the climate system that can affect reefs. Cyclones (hurricanes, typhoons) can destroy coral reefs, and although there is little evidence of an increase in storm frequency, there is some evidence that maximum winds will intensify. Increased precipitation on land can lead to greater sedimentation on many reefs, particularly those near deforested areas. Sea level rise is a consequence of climate change that can benefit some reefs and harm others. Because of the prolific production of calcium carbonate by coral reef organisms, many of the coral reefs that we see today were able to keep pace with the sea level rise of 100 m or more that occurred between about 18 000 and 3000 years ago. Sea level rise is predicted to be less than 1 m over the course of the twenty-first century. This relatively small rise could affect reefs near land if flooding of the coastal zone releases nutrients and sediments that degrade water quality. However, coral reefs that occur away from major landmasses are not considered threatened by a 1–2 m sea level rise.

An additional consequence of increasing atmospheric CO_2 concentration is ocean acidification. Ocean acidification in the tropics is a predictable process, driven directly by rising atmospheric CO_2 concentration rather than by climate change *per se*. The increase in partial pressure of atmospheric CO_2 drives more CO_2 into the ocean, some of which combines with water to form carbonic acid (H_2CO_3), which lowers the pH. Over this century, predicted decreases in surface ocean pH (about 0.25–0.33 pH units) are not likely to cause direct coral mortality, but constitute a creeping environmental problem that affects skeletal formation in corals and reef-building algae, and the overall calcium carbonate accumulation on a reef. Under increased acidity, corals secrete less calcium carbonate, and reef limestone dissolves more readily. It is likely that these effects will cause corals to secrete their skeletons more slowly, which would leave them less competitive for space; or less densely, which would leave them more fragile.

Predicting the effects of future climate on coral reefs depends on how well we can predict climate change, and on how well we can predict coral reef ecosystem response to this change. Corals may adapt to increases in temperature through a variety of mechanisms, including natural selection, and an interesting phenomenon whereby a bleached coral becomes repopulated with a more heat-tolerant variety of zooxanthellae. The recent increase in coral bleaching indicates that these mechanisms are not yet keeping pace with temperature rise. There is no evidence that coral calcification can adapt to changes in seawater chemistry. However, some coral species calcify more rapidly with temperature increase, and for those that survive the negative effects of future warming, a decrease in calcification due to ocean acidification may be offset by increases in calcification due to warmer temperature. Projections of the state of coral reef ecosystems have ranged from moderate to total degradation; all predict significant disruption of ecosystem functioning. Current conservation efforts aim toward removing the controllable stresses on coral (e.g. overfishing, sedimentation, pollution), while encouraging a reduction in fossil fuel emissions.

J. Kleypas

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Cross-references

Coastal Climate
Climate Change and Global Warming

CORIOLIS EFFECT

Any moving object above the Earth's surface tends to deflect from its course because of Earth's rotation. This deflection is known as the Coriolis force, named after a French engineer and mathematician Gaspard Gustave de Coriolis (1792–1843) in 1835. Given that this deflection is an apparent force, it is perhaps more correctly known as the Coriolis effect.

The Earth rotates eastward and has the same rotational velocity. However, places at different latitude have various linear velocities. A point near the equator goes around at 1000 miles an hour, while one near the pole moves only a few miles an hour. An object launched from the equator to the north will maintain the eastward component of velocity of other objects at the equator. When this object travels away from the equator it will be heading east faster than the ground beneath it. Similarly, an object moving to the equator from the north will be moving more slowly than the ground beneath it, and deflects to the right of its true path (apparently deflects to the west).

In fact there is no actual force involved in these displacements, and the ground is simply moving at a different speed than its original ground speed. Therefore this deflection (the Coriolis force) will only affect the direction and not the speed of the moving body. An object moving a north–south path on Earth will deflect to the right in the northern hemisphere and left in the southern hemisphere.

The magnitude of Coriolis deflection is directly related to the velocity of the object and its latitude, and can be expressed as:

$$F_c = 2(\Omega \sin \phi)v,$$

where: F_c is magnitude of Coriolis force per unit volume,
Ω is Earth's angular velocity,
ϕ is latitude and
v is velocity of the object.

The Coriolis effect is zero at the Equator and maximum at the poles.

The Coriolis effect is significant in the dynamics of atmosphere, in which it affects large-scale atmospheric circulation and development of storms. Global scale winds blow in the paths commanded by the Coriolis effect. Wind deflects to the right in the northern hemisphere and left in the southern hemisphere. Stronger winds will have a greater amount of deflection than weaker winds. Wind blowing near the poles will deflect more than that with the same speed near the equator.

Yuk Yee Yan

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Cross-references

Atmospheric Circulation, Global
Wind, Principles

CRIME AND CLIMATE

Scientific inquiries into the relationship between climate and crime began during the early nineteenth century. Usually climate has referred to the monthly or seasonal mean temperatures and crime has been classified into two types: crimes against the person (or violent crimes such as homicide, rape, assault, and robbery); and crimes against property (such as burglary, larceny–theft, and motor vehicle theft).

During the nineteenth century A.M. Guerry in France and A. Quetelet in Belgium, in two independent studies, observed the coincidence of higher frequencies of crimes against persons in warmer areas and during warmer months. Moreover, crimes against property increased in winter months and in cooler areas. Quetelet conceptualized these observations into the *Thermic Law of Delinquency* (Cohen, 1941; Harries, 1980). Neither Guerry nor Quetelet felt that climatic conditions were the principal causes of delinquency. Both criminologists are credited with starting the geographic school of criminology where the ecological facts of crime are pursued – namely, the relationship between crime and social conditions (Gibbons, 1978).

Other European scholars, during the late nineteenth and early twentieth centuries, corroborated Guerry's and Quetelet's findings and agreed that climate was not the principal cause of crime (Cohen, 1941). Yet studies asserting a direct causal linkage emerged and persisted.

E.G. Dexter, studying 40 000 assault cases in New York City between 1891 and 1897, suggested that temperature was the most significant condition aggravating the emotions and erupting into fighting (Cohen, 1941; Harries, 1980). A stronger and more enduring endorsement of the causal relationship came from Ellsworth Huntington (1945) who claimed physiological and psychological conditions vary with climate and weather, hence the propensity for riots and assaults to increase with temperature. Huntington's vehement advocacy of climatic determinism as the cause of crime overshadowed the works of others including Cohen, whose work in 1941 contradicted climatic determinism.

During 1941, using the Federal Bureau of Investigation's Uniform Crime Reports, Cohen observed the expected highs for violence during the midsummer and property crimes during the midwinter. However, the most important finding was that regional variations of frequencies and types of crimes were more pronounced and significant than the seasonal variations. Cohen believed that the regional variations were the surrogates for significant social forces (1941). Despite findings like these the academic community's reaction to Huntington's environmental determinism was so strong and negative that inquiries into the relationship between climate and social phenomena in general, and climate and crime in particular, became almost taboo (Harries and Stadler, 1983).

Gradually, studies began emerging disproving the direct climate–crime linkage or including climatic variables with a host of other independent variables in causal analyses. An example of the former is a study by Lewis and Alford (1975) testing the *Thermic Law of Delinquency* by examining 3 years of monthly assault data across 56 US cities. Beginning with the winter months, the expectation was that the number of assaults would oscillate seasonally with the march of the sun on a south to north vector. Such an orderly progression was not found and, regardless of region, the transition from lower to higher levels of assaults was rather abrupt (Lewis and Alford, 1975). An example of the latter is a study by Van de Vliert et al. (1999) associating riots and armed attacks in 136 countries between 1948 and 1977 with temperature, population size and density, degree of democratic government, socioeconomic development, and a cultural masculinity dimension. They conclude that the culture mediates the effect between temperature and collective violence.

The fast pace of changes in computing and information systems technology since the mid-1980s has decreased the number of inquiries into the climate–crime relationship. Easy access to weather and crime information of varying temporal and spatial scales has spawned numerous studies on the relationship between crime or police activity with weather elements or events (see LeBeau and Corcoran, 1990; Rotton and Cohn, 2000). Despite the changes in focus the *Thermic Law of Delinquency* remains an integral guide for these and other studies.

The relationship between climate and crime is more suggestive than definitive, but climate is viewed as indirectly providing enhanced opportunities for the commission of different crimes. Finally, regardless of the empirical evidence, the public may feel there is a definitive relationship between the two; especially since many residential real-estate companies on their Internet websites list a range of factors describing the quality of life in a place and on the list climate is followed by crime.

James Lebeau

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Cross-references

Applied Climatology
Determinism, Climatic

CULTURAL CLIMATOLOGY

Climate and life are intimately linked. Without life the composition of the Earth's atmosphere today would be totally different and the climate of the Earth would be more like that of Venus – the Earth's atmosphere would be 98% CO_2 (Morrison and Owen, 1996). This intimate link between life and the atmosphere on Earth has led to claims that the atmosphere is effectively managed by the biosphere, leading to the "Gaia hypothesis" (Lovelock, 1979). No-one is sure how life developed on Earth about 3850 million years ago, but after continuous evolution humans emerged about 5 million years ago, and civilization and society as we know it began about 5000 years ago. Hence it is impossible to study climate in isolation from the biosphere and society, and any changes to the Earth's atmosphere caused directly or indirectly by society should be firmly within the remit of climatology. Since the industrial revolution of the nineteenth century society has threatened to change our climate, ironically to increase CO_2 concentrations and other so-called greenhouse gases, the need for effective climate management has never been greater.

Today the links between climate and society are explored under the headings of applied climatology where the climate is seen as both a resource and a hazard that needs to be effectively managed by society. Climate change cannot be managed under the heading of applied climatology as the issue has global consequences and climatologists do not have the social science skills to effectively, manage climate change and social scientists do not understand the vagaries of climate sufficiently. Hence it is time that climatologists were trained to understand how to manage the climate effectively, otherwise it will be left to politicians and economists to decide the future of our climate. This has never been a more important issue than in the assessment and management of climate change. Hence the need for cultural climatology to emphasize this requirement for climatologists who understand how society works, as well as the climate.

Cultural climatology is a new branch of climatology concerned with the study of physical and dynamical processes in the atmosphere at various space and time scales in conjunction with evaluating and understanding climate–society interactions and feedbacks. Climate is considered to be an integral part of culture and conversely culture is an integral part of climate. This climate–culture dialectic gives meaning and value to the study of climatology that has been missing in the past. Given that the ultimate goal of climatology is to apply the field's knowledge to the solution of both environmental and socioeconomic problems, a cultural turn is taking place as climatologists enter the post-normal phase of their science. The nature of this cultural turn, and what constitutes cultural climatology, is discussed in detail elsewhere, together with a consideration of climate and society relationships as these are at the heart of cultural climatology (Thornes and McGregor, 2003).

Culture is a complex and overused word, and is used to encapsulate everything from the "total way of life" of a people (language, dress, religion, music, values, etc.) through to works of art and advertising. Mitchell (2000, p. 14) gives six possible definitions of culture, but the simplest definition of all is: "Culture is all that is not nature". Although the definition of nature is not as simple as it might seem (Macnaghten and Ury, 1998) we can fruitfully examine the relationship between climate, as part of nature, and culture.

The summons to cultural climatology should not be misconstrued as a call for climatologists to abandon the research mainstays of synoptic, dynamic and physical climatology. On the contrary we expect cultural climatologists to be concerned with issues such as global climate system change, establishing what the principal drivers of the climate system are, assessing how the climate system will respond to natural and human-induced changes, answering how society might respond to the opportunities and threats posed by climate change, and evaluating to what extent the changes expected in the climate system can be predicted. This call to cultural climatology should be seen more as a signal to climatologists that opportunities await us at the interface between science and society, an area which climatologists on the whole have felt great apprehension with in the past.

Cultural climatology is just the ticket for catching the climate and society boat of opportunity. A passage on this boat will bring climatologists closer to understanding the physical and societal mechanisms underlying the complex interactions between components of the climate–human system. All students and purveyors of climatology courses need to look beyond the learning and teaching of straight climate processes by considering the multitude of ways in which climate and society may interact. Such a broadening of horizons into the realms of cultural climatology will not only will provide us with learning and research opportunities, but, will provide society with a better understanding of the meaning of climate.

John Thornes

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Cross-references

Art and Climate
Climatology, History of
Folklore, Myths and Climate
Literature and Climate

CYCLES AND PERIODICITIES

Throughout recorded history people have been aware of periodic meteorological events that have influenced their well-being. Because so much of our lives is governed by seasonal rhythms, it is natural that there has been a search for long-term order in the physical world. Today the study of cycles, which may be defined as a series of recurring events that repeat with some sense of periodicity, is a well-defined area of research.

The search for cycles

The search for cycles in weather records has been a constant endeavor of a lot of meteorologists for many years. Whether the product of natural variability of the climate system or the result of interactions between the Earth's atmosphere and oceans and external agencies, including periodic variations of solar activity and the astronomical motions of the Moon and the planets, there has been a long and largely fruitless search for clear evidence of predictable cycles.

Interest in weather cycles has fluctuated over time, but has recently experienced an upsurge in activity for two principal reasons. First, the record-breaking El Niño of 1982/1983 stimulated huge interest in the possibility of events in the tropical Pacific Ocean having an impact on the weather around the globe (Diaz and Markgraf, 2000). The realization that the El Niño was part of atmosphere–ocean interactions that can set up quasi-cyclic behavior in the climate has led to a burgeoning study of such "oscillations". Second, satellite measurements since 1980 have shown that solar irradiance varies in phase with the 11-year cycle in the number of sunspots on the sun (Figure C47). This has led to renewed interest as to how changes in solar activity might affect the climate (Haigh, 2000).

The quasi-biennial oscillation (QBO)

Prior to the 1980s the one oscillation that had established as being a real feature of the climate was the periodic reversal of the equatorial winds in the stratosphere (Baldwin et al., 2001). On average the cycles in these winds have a period of around 27 months, hence the name quasi-biennial oscillation (QBO). First observed in the 1950s, their period has varied from over 3 years to well under 2 years (Figure C48). The QBO shows well-defined characteristics. The wind regime propagates downward as time progresses. The amplitude is greatest at an

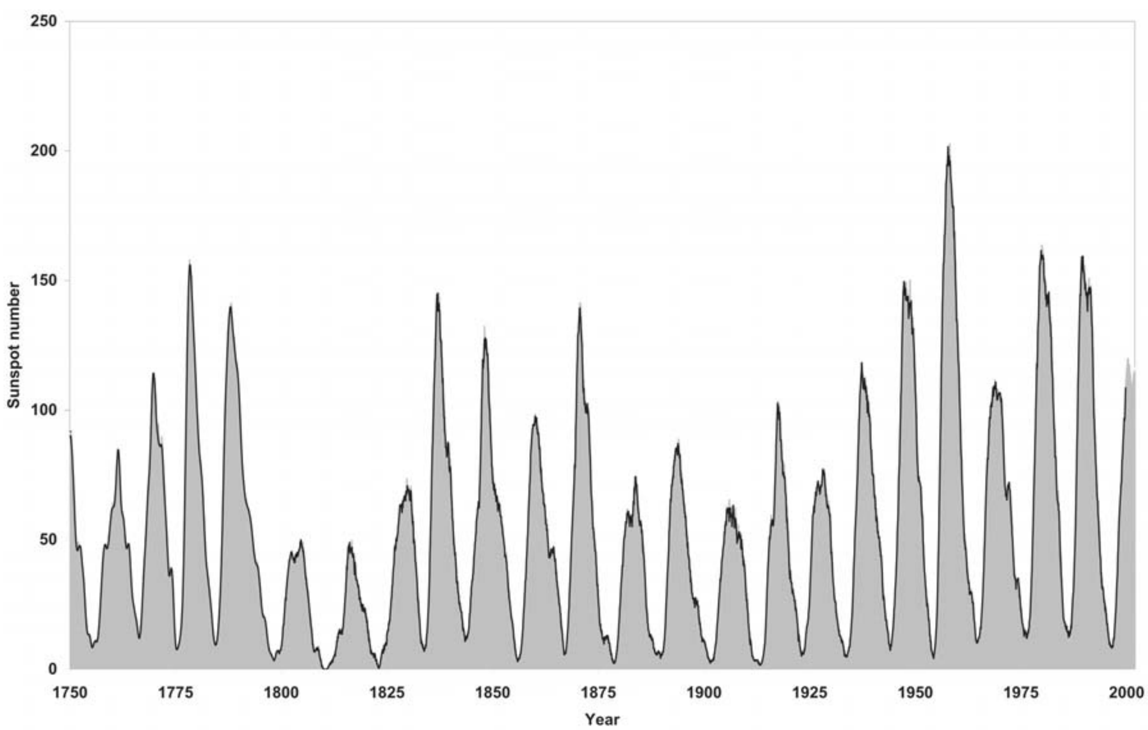

Figure C47 The observed variations in the number of sunspots since 1750, showing the 13-month running mean of monthly numbers. (Data from National Geophysical Data Center, NOAA.)

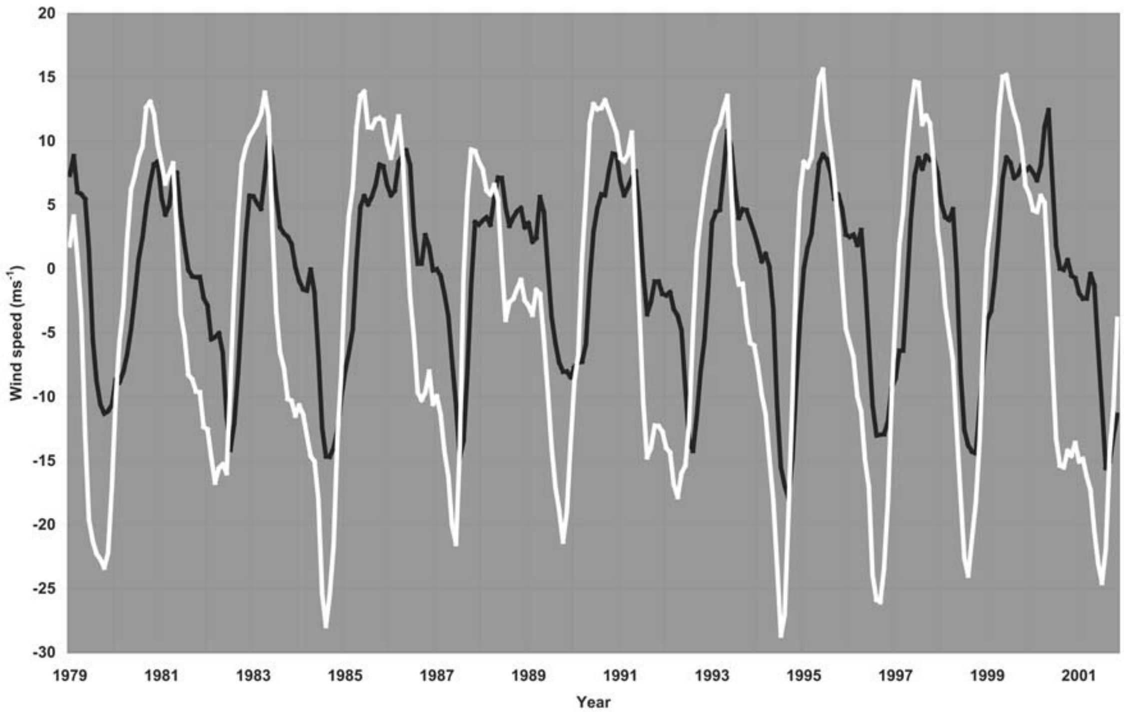

Figure C48 The quasi-biennial oscillation of the winds in the stratosphere over the equator shows a pronounced periodic reversal. The scale of this oscillation is greatest at around 30 hPa (white line) and reduces at lower levels, as shown by the values at 50 hPa (black line) as the periodic feature migrates downward over time, so that the peaks occur first at high levels. (Data from NOAA, available on websites: ftp://ftp.ncep.noaa.gov/pub/cpc/wd52dg/data/indices/qbo.u30.index and ftp://ftp.ncep.noaa.gov/pub/cpc/wd52dg/data/indices/qbo.u50.index.)

altitude of 30 km (a pressure of 30 hPa). The easterly phase of the QBO is stronger than the westerly phase.

The accepted explanation for reversal of the winds in the equatorial stratosphere involves the dissipation of upward-propagating Kelvin and Rossby waves in the stratosphere. These waves originate in the troposphere and lose momentum in the stratosphere by the process of radiative damping. This involves rising air cooling and radiating less than the air that is warmed in the descending part of the wave pattern. Westerly momentum is imparted by decaying Kelvin waves. Rossby waves perform a similar function in respect of the easterly phase. These two sets of waves combine to produce a regular reversal of the upper atmosphere winds.

The tropospheric biennial oscillation (QBO)

The tropospheric biennial oscillation (TBO) is the less well-defined cousin of the QBO. The TBO appears in many meteorological series, but it is nowhere near as distinct as the QBO. Moreover, both within individual series and between different series, its period varies from around 2.1 to 2.6 years in a manner that cannot be directly linked to the QBO. In part this disconnection may be related to the dominant role of the annual cycle in many meteorological series, and especially to the extremes of winter and summer. Because the QBO and TBO move in and out of phase with the annual cycle there is possibility that the weaker oscillation will tend to "phase lock" with the annual cycle and then skip a beat every third year or so. This potentially chaotic behavior has compounded the difficulty of establishing whether the QBO and TBO are manifestations of a common meteorological phenomenon or reflect separate physical processes.

El Niño Southern Oscillation (ENSO)

The El Niño Southern Oscillation (ENSO) is a major contributor to interannual fluctuations in the tropical climate. It is best known for its warm (El Niño) episodes, when much of the eastern tropical Pacific has well above normal sea surface temperatures (SST). These are interspersed with either more normal conditions or cool (La Niña) episodes. This coupling between the atmosphere and the ocean across the equatorial Pacific shows a quasi-cyclic periodicity centered around 3–5 years with the strength of both warm and cool events varying appreciably over time. Detailed analysis of SST for the central equatorial Pacific since 1950 shows that much of the variance is made up of two periodicities: the first being the TBO (period 26 months) and a quasi-quadrennial oscillation (QQO) with a period of 53 months (Figure C49). Here again there is evidence of "phase locking" between ENSO and the annual cycle, as warm (El Niño) episodes tend to peak around the end of the calendar year, thereby making their spacing erratic and difficult to predict.

Other interannual oscillations

The insights provided by modeling ENSO have led to renewed interest in other forms of atmosphere–ocean oscillations around the world. In particular the North Atlantic Oscillation (NAO),

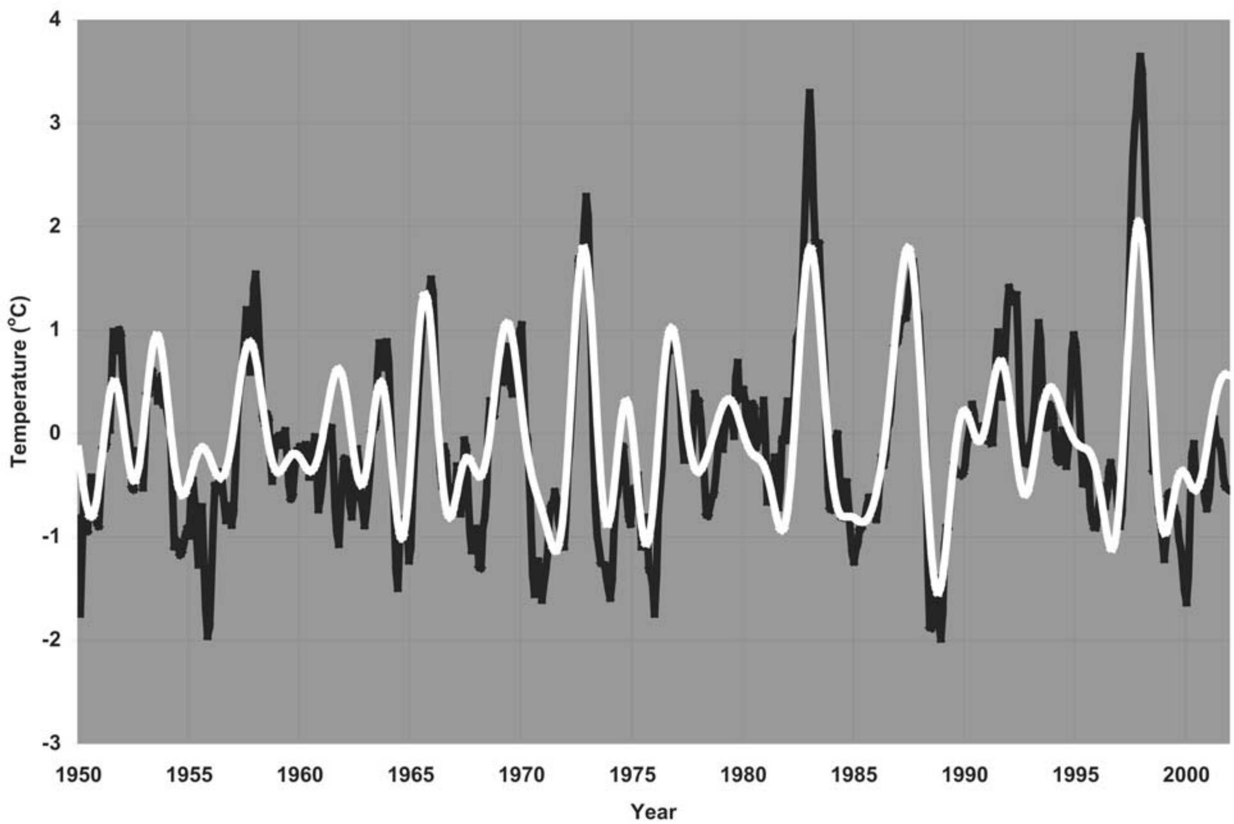

Figure C49 A comparison of the observed bimonthly temperature anomalies in the central equatorial Pacific Ocean (black line) and a reconstruction of this series using a combination of two filtered versions of the data (white line): one centered on 24 months and the other centered on 48 months. This result shows that the data can be largely represented by the combination of a tropical–biennial oscillation (TBO) and a quasi-quadriquennial oscillation (QQO) that is effectively "phase locked" to the warm events peaking at around the end of the calendar year.

the Pacific Decadal Oscillation (PDO) and Antarctic Oscillation have attracted most interest. In certain cases these oscillations were first identified in interannual fluctuations in the pressure difference patterns between mid- and high latitudes. So the NAO is usually measured in terms of the pressure difference between Iceland and the Azores during the months of December to March and is a measure of the strength of the winter circulation of the midlatitudes of the northern hemisphere. Although the NAO shows some evidence of periodicities in the region of 7–9 years and around 20 years, these do not show real statistical significance. The PDO tends to occur with a periodicity of around 20 years.

Decadal oscillation

Many meteorological records show evidence of periodicities around 10–12 years. Often these have been linked with solar activity, (see section on Solar activity), but the case for the link is not always convincing as they are inclined to shift in phase with respect to the 11-year sunspot cycle. So, although seen in both large-scale temperature patterns across the northern hemisphere, notable over North America, and in the oscillations in the Pacific Ocean, including ENSO, these various manifestations do not show any particular coherence.

Bidecadal oscillation

A cycle with around 20 years frequently appears in meteorological records, and in some instances is closely linked with the double-sunspot cycle (known as the Hale cycle). One of the most convincing associations is with the incidence of drought in the western United States. An alternative hypothesis however, is that the 18.6-year lunar tidal cycle may be the cause of this variation. Other impressive examples of bidecadal oscillations are found in Greenland ice-core data, central England temperature records and extratropical Pacific SST, where it is part of the case for the PDO.

Multidecadal oscillations

There is considerable evidence of 65–90-year oscillation in both temperature records and tree-ring data. In particular, for the North Atlantic and its bounding continents, there is a basin-wide pattern of sea surface temperature anomalies. This is probably related to the NAO, but it has now become known as the Atlantic Multidecadal Oscillation (AMO). This periodicity has also shown up in tree-ring data in northern Eurasia, sea level measurements in northern Europe and rainfall observations in the United States. It appears also to be part of a wider phenomenon, as there is evidence of an oscillation in the global climate system of period 65–70 years, although in a record of only 135 years this is the subject of some doubt.

Longer-term cycles

The best-established cycles in the climate are associated with the waxing and waning of the ice ages over the last million years or so. These have periodicities centered on 100 000, 41 000 and 23 000 years (100, 41 and 23 kyr). These cycles appear to be the result of variations in the Earth's orbital parameters, which lead to the amount of solar radiation falling at different latitudes at different times of the year varying in a cyclic manner. These variations result in the amount of snow and ice in polar regions rising and falling with a characteristic 100-kyr sawtooth pattern, which is modulated by the 41- and 23-kyr cycles (Figure C50).

In addition, within the last ice-age cycle there is evidence of a roughly 1500-year cycle that has been linked with variations of the ocean circulation in the North Atlantic. While the most striking evidence of this cycle is during the last ice age, it appears to have played a part in less dramatic climatic changes during the last 10 kyr.

Ocean–atmosphere coupling

The most obvious explanation for the many oscillations observed in the global climate is that they are the product of feedback processes between the atmosphere and the oceans. Because the atmosphere and the oceans respond to various physical stimuli on different timescales, it is possible for them to interact nonlinearly to produce quasiperiodic behaviour. Indeed, it is possible to produce nonlinear models of the atmosphere alone that are capable of producing substantial interannual variability. When combined with the fact that the oceans are more likely to exhibit longer-term fluctuations, it is possible for coupled atmosphere–ocean computer models to produce periodic fluctuations on every timescale from the interannual to the millennial.

Physical explanations of ENSO involve large-scale atmosphere–ocean interactions. During warm (El Niño) episodes the combination of the warming of the ocean off the coast of South America, the associated strong convection and the reduced trade winds all contrive to maintain the abnormal conditions. Similarly, during cool (La Niña) episodes the strong convection in the western Pacific and the strong trade winds should maintain the below-normal SST in the eastern equatorial Pacific. So, in principle, the system should stay in one or other state indefinitely. The reason this does not happen is because of changes in the depth of the thermocline in the ocean due to Kelvin and Rossby waves in the ocean.

Close to the equator downwelling and upwelling Kelvin waves in the depth of the thermocline move rapidly eastwards across the Pacific basin every 2 months or so. Farther from the equator wind fields generate Rossby waves moving westward whose speed depends on the latitude. These speeds are much slower than those of the equatorial Kelvin waves. Near the equator they cross the Pacific basin in about 9 months. Toward the poles the time increases rapidly, to be about 4 years at 12°N

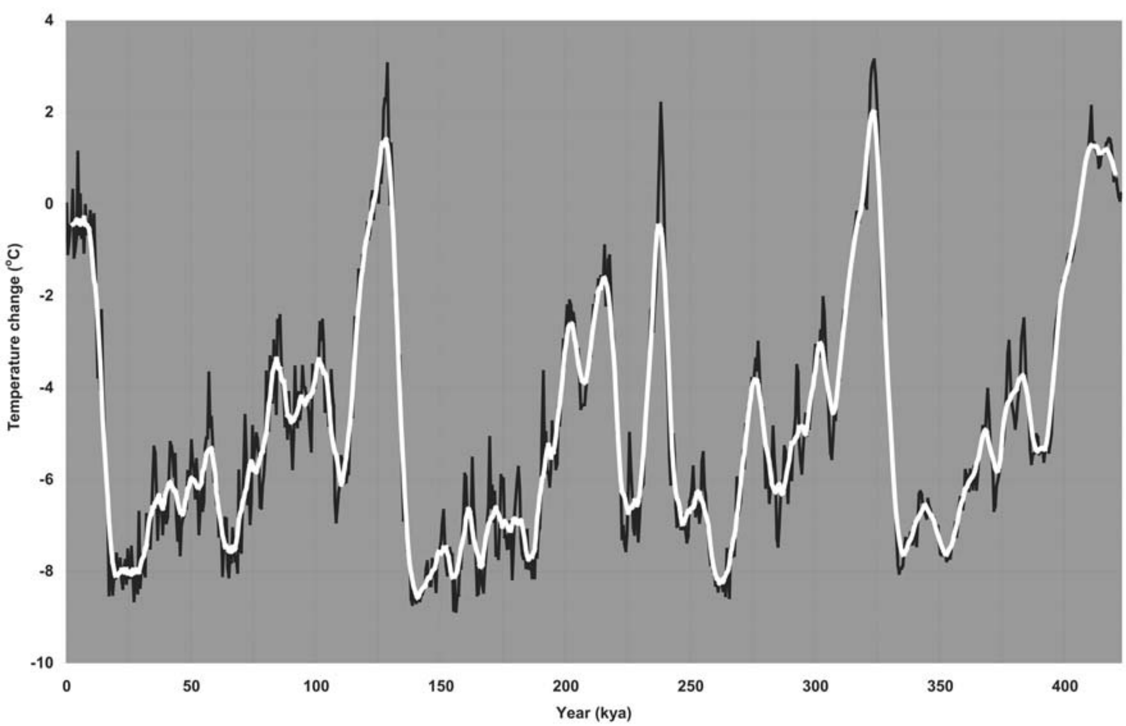

Figure C50 The estimate of the temperature change over Antarctica using measurements of the hydrogen isotope ratio (H/D) in an ice core drilled at the Vostok research station, going back to over 400 000 years ago (400 kya). This shows how the record is dominated by a 100-kyr periodicity over the last four ice ages, while during the glacial periods there is evidence of shorter periodicities of around 20 and 40 kyr. (Data archived at the World Data Center for Paleoclimatology, Boulder, Colorado, USA.)

and S. The combination of these effects leads to the quasiperiodic appearance of warm water in the eastern Pacific (El Niño) every 3–5 years, interspersed with cool (La Niña) episodes.

The unresolved question about atmosphere–ocean coupling is whether the atmosphere or the ocean is the driving force behind interannual periodicities. In principle, the greater thermal inertia of the oceans should exert a dominant influence, and hence make the prediction of behavior months and years ahead a practical proposition. In practice, both the models and experimental forecasting suggest that, with both ENSO and the NAO, the chaotic influence of the atmosphere exerts an overriding influence.

In the longer term changes in the thermohaline circulation of the world's ocean currents may play a part in the 1500-year periodicity seen within the ice age cycle. In particular the pattern of currents in the North Atlantic appears to be especially sensitive to shifts in temperature and salinity of the surface water up around Iceland. Computer models of the oceans show that changes in evaporation, precipitation and freshwater runoff from the surrounding continents can produce sudden switches in circulation patterns. It is proposed that, during the last ice age, such changes, combined with surges of icebergs from the northern ice sheet, led to dramatic swings in temperatures around the northern hemisphere.

Solar activity

Since the seventeenth century, observations on the sun's photosphere have revealed sunspots whose number, size and duration vary over time in a more or less regular manner with a period of about 11 years (Figure C49). The variation in number during each cycle is more than two orders of magnitude greater than for any shorter period. It ranges from virtually no spots during the minimum of solar activity to just over 200 in the most active cycle that peaked in 1957. The instrumental record has now accumulated reliable data since around 1750, and is now in its twenty-third cycle of activity, which peaked at around 120 spots in 2000.

Another measure of solar activity is the number of faculae, which is closely associated with sunspots: they brighten as the number of dark sunspots increases. Their increased brightness is the dominant factor in changing solar output rather than sunspot darkening. A series of satellites have made measurements that provide unequivocal observations of how the sun's output varies with the 11-year cycle in solar activity (Figure C47). These results show that the level of total solar irradiance (TSI) is greatest around the solar maxima and the amplitude of the solar cycle variation is slightly less than 0.1%.

Models of the TSI that combine changes in sunspot number and faculae brightness produce a good fit with satellite observations of the TSI. These models explain not only the observed changes in solar output since 1980, but also the longstanding hypothesis that the cold period known as the Little Ice Age may be related to changes in solar activity. In particular, the colder weather of the seventeenth century appears to have been the result of an almost complete absence of sunspots during this period, known as the "Maunder Minimum".

The small change in the TSI during the last two solar cycles is, however, on its own, an order of magnitude too small to explain observed correlation between solar activity and global temperature trends since the late nineteenth century. Predicted changes in TSI since the seventeenth century, which include the sun's magnetic, solar cycle length, solar cycle decay rate, solar rotation rate and various empirical combinations of all of these, suggest greater changes in the TSI (Haigh, 2000). These models produce widely different estimates of changes since the Maunder Minimum, ranging between 1 and 15 $W\,m^{-2}$ although most estimates lie in the range 2.5 to 5 $W\,m^{-2}$. Even so, this may be sufficient to explain much of the observed correlation between global climate change and solar variability during the last three to four centuries.

The TSI also varies with wavelength. The greatest changes are in the ultraviolet (UV) region. These wavelengths are largely absorbed in the atmosphere and this could enhance the impact on the weather. In particular, the amount of UV affects the amount of ozone formed in the stratosphere. This, in turn, alters the radiative balance of both the stratosphere and lower levels of the atmosphere. Computer models, which include realistic changes in UV and ozone, show a significant effect on simulated climate. Solar cycle variability may therefore play a significant role in regional surface temperatures, even though its influence on the global mean surface temperature is small. So it is possible, by both radiative and dynamical means, that variations in solar UV output can produce disproportionate changes in the atmosphere.

Possible links between the sun's magnetic field, solar activity and the Earth's weather involve a complex set of mechanisms. The magnetic polarity of sunspots alternates between successive cycles. Known as the Hale magnetic cycle, this 22-year cycle could be the key to the amplification process. Furthermore, the sun's magnetic flux at the Earth rose by 130% between 1901 and around 1990. While it exhibits periodic fluctuations apparently linked to 11-year sunspot cycle, these are not precisely in phase with it and are less regular.

The direct effect of fluctuations of the sun's magnetic field on the weather may relate to the quantity of energetic particles emitted from the sun. Second, the overall strength of the sun's magnetic field alters the Earth's magnetic field, and with it the amount of cosmic rays (energetic particles from both the sun and from elsewhere in the universe) that are funneled down into the atmosphere.

The impact of variations in the solar magnetic field depends on the energy of the cosmic rays as they enter the Earth's atmosphere. Cosmic rays of low energy have their origin in the sun and are absorbed high in the atmosphere. Galactic cosmic rays (GCR) are of much higher energy and have an appreciable impact on the troposphere. When the sun is more active GCR are less able to reach the Earth and so their impact on the lower atmosphere is inversely related to solar activity.

Cosmic rays produce various chemical species, which alter the concentrations of radiatively active molecules such as ozone (O_3), nitrogen dioxide (NO_2), nitrous oxide (N_2O) and methane (CH_4). These species are most likely to be seen in the stratosphere where their impact will be similar, but less significant to the changes caused by solar UV variations. In addition, the formation of ions will affect the behavior of aerosols and cirrus clouds that have a direct radiative impact and also alter the amount of water vapor throughout the atmosphere. These changes could lead to shifts in the radiative balance of both the stratosphere and troposphere and so produce long-term fluctuations in the temperature.

There is also a question of whether GCR alter cloud cover at lower levels. An analysis of satellite observations of global total cloud cover, between 1984 and 1991, suggested a correlation with cosmic ray flux. Increased GCR flux appears to cause total cloud amounts to rise, and this cools the climate. The interannual variations in cloudiness are, however, difficult to distinguish from parallel changes caused by warm and cold ENSO events.

There is, however, a possibility that charged particles are more efficient than uncharged ones in acting as cloud condensation nuclei (CCN), even though only a small proportion of aerosol particles are capable of acting as ice nuclei, depending on chemical composition or shape. Recent work has provided the first observational evidence of cosmic ray-induced aerosol formation in the upper troposphere. In theory, the higher level of charge carried by such aerosols can enhance the formation of ice nuclei. By changing precipitation rates or radiative balance, the changes in the clouds then affect atmospheric dynamics and temperature.

Tidal forces

The other obvious periodic influences on the weather tidal forces, which can have a direct effect on the movement of the atmosphere, the oceans and even the Earth's crust, are the Earth's tides. The nature of these links varies in complexity. Tidal effects in the atmosphere are relatively predictable and measurable, but tiny compared with normal atmospheric fluctuations. In the oceans the broad effects can be calculated, but estimating changes in the major currents is much more difficult. In particular, recent satellite altimetry studies have come up with some interesting results that may shed new light on the climatic implications of the dissipation of tidal energy. Prior to this work it was widely assumed that this energy, representing the effect the moon receding from the Earth at a rate of some four centimeters a year, was dissipated in the shallow waters of the continental shelves around the world. It is now reckoned that about half this energy is fed into deeper water where it exerts a significant effect on the strength of the major ocean currents and hence the transport of energy from the tropics to polar regions. So the 18.61-year lunar cycle affects the strength of the principal ocean currents.

As for movements of the Earth's crust, the problems are compounded by possible links with solar activity. The direct influence of the tides could influence the release of tectonic energy in the form of volcanism. Since there is considerable evidence that major volcanic eruptions have triggered periods of climatic cooling, this would enable small extraterrestrial effects to be amplified to produce more significant climatic fluctuations. In addition, there is evidence that intense bursts of solar activity interact with the Earth's magnetic field to produce measurable changes in the length of the day. Such sudden tiny changes in the rate of rotation of the Earth could also trigger volcanic activity. It should be noted that, while there is no evidence that their occurrence was in any way related to this effect, the three climatically important volcanoes in the second half of the twentieth century (Agung in 1963, El Chichon in 1982 and Pinatubo in 1991) were spaced in such a way as to have a confusing impact on the interpretation of any solar or tidal effects.

The Earth's orbital parameters

The explanation of the periodicities in the ice ages relies on the Earth's orbital parameters producing latitudinal and seasonal variations in incident solar radiation. In particular, the variations in the amount of sunlight falling in summer at high latitudes in the northern hemisphere appear to affect the amount of build-up of ice sheets over the continental landmasses. These changes are controlled by the precession of the equinoxes (the 19- and 23-kyr periodicities) and the variations of the tilt of the Earth's axis (the 41-kyr periodicity) and the 100-kyr periodicity due to the eccentricity of the Earth's orbit. There is, however, a problem with this proposal. While the first two periodicities are probably sufficient to trigger significant climatic changes, the 100-kyr eccentricity periodicity is the weakest of the orbital effects. Nevertheless, the 100-kyr ice age cycle is the strongest feature of the climatic record in the last 800 000 years.

A solution to this problem is to use a differential model in which the rate of change of the climate is a function of both the orbital forcing and the current state of the climate (Imbrie et al., 1992, 1993). Not only is this a more realistic representation of climatic behavior, but it also contains a nonlinear relationship between the input and the output that has important physical consequences. For instance, if the model is sensitive to changes in ice volume that are reflected in the time constants for the growth and decay of the ice sheets and the lag between the changes in solar radiation falling in summer at high latitudes of the northern hemisphere, it can be shown to achieve a reasonable representation of the observed long-term behavior. Various models have tuned the parameters to achieve the best fit between the calculated ice-volume changes and the oxygen-isotope record to give a reasonable match with past changes. The dependence on empirically derived time constants, which are broadly linked to the physical behaviour of the ice sheets, is a measure of the nonlinear response of the global climate to changes in the orbital parameters.

A more sophisticated approach is to draw on the fact that there is evidence that the North Atlantic appears to have exhibited three different modes of circulation during the last ice age (Paillard, 1998). As a result of this behavior it appears that the global climate is capable of switching between three distinct regimes (e.g. interglacial, mild glacial and full glacial). Switches between these regimes could well be controlled by a combination of changes in insolation, resulting from the Earth's orbital parameters, combined with changes in ice sheet volume. By defining the conditions for the transition between the three regimes a model has been produced that is capable of reproducing with remarkable accuracy the changes in ice sheet volume over the last million years.

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Cross-references

El Niño
Maunder, Edward Walter and Maunder Minimum
North Atlantic Oscillation
Ocean–Atmosphere Interaction
Oscillations
Sunspots