M

THE MADDEN–JULIAN OSCILLATION (MJO)

The MJO is a low-latitude intraseasonal oscillation, meaning that it passes through an identified cycle in a period of 60–90 days, quite unlike the annual, biennial or decadal cycles of other oscillations. It is so named for Roland Madden and Paul Julian of NCAR, who discovered the wave in the early 1970s. Its identification and possible forecasting is of considerable importance in long-range predictability of tropical and subtropical weather as well as short-term climate variability.

The MJO is a feature of the tropical atmosphere–ocean system that plays a significant role in precipitation variability. It is characterized by anomalous rainfall conditions that can be either enhanced or suppressed. The beginning of the cycle, the anomalous rainfall event, usually appears first over the Indian Ocean and Pacific Ocean. It remains identifiable as it moves over the very warm water of the western and central parts of the Pacific Ocean. On meeting the cooler waters of the eastern Pacific Ocean it usually becomes less defined, only to reappear over the tropical Atlantic Ocean and Indian Ocean. As noted this cycle can last 1–2 months.

The higher or lower than normal precipitation of the event are associated with both surface and upper-air conditions as they relate to ascending and descending air. A knowledge of the cycle characteristics would enable a clearer understanding of tropical rainfall variability, the role of the cycle in relation to Pacific Ocean and Atlantic Ocean tropical cyclones and the impact of the oscillation upon middle-latitude precipitation. In their study of the MJO's role in hurricanes in the Caribbean Sea and Gulf of Mexico, Maloney and Hartmann (2000) found a distinct relationship. They noted that hurricanes are four times more likely to occur when, in the ascending phase, the rising air and surface westerly winds are conducive to formation of the storm.

The Climate Prediction Center (CPC) of NOAA (http://www.cpc.ncep.noaa.gov/) shows how the tropical oscillation impacts rainfall in the Pacific northwest of North America. The evolution of this event is often referred to as the "pineapple express", so named because a significant amount of the deep tropical moisture traverses the Hawaiian Islands on its way toward western North America.

The MJO is thought to play a significant role in the formation and frequency of hurricanes. If an easterly wave, the initial formation feature of Atlantic hurricanes, meets the cloudy sky phase of the MJO the conditions for hurricane formation improve. The MJO is, of course, moving in the opposite direction to the easterly wave. Unfortunately, although hurricane researchers of the National Hurricane Center have been informally using the MJO, the linkage is currently insufficient for accurate modeling and forecasting.

John E. Oliver

Bibliography

Ångstrom, A., 1935. Teleconnections of climate changes in the present time. *Geografiska Analer*, **17**: 242–258.

Bjerknes, J., 1966. A possible response of the atmospheric Hadley circulation, to equatorial anomalies of ocean temperature. *Tellus*, **8**: 820–829.

Madden, R.A., and Julian, P.R., 1971. Detection of a 40–50 day oscillation in the zonal wind of the tropical Pacific. *Journal of Atmospheric Sciences*, **28**: 702–708.

Maloney, E.D., and Hartmann, D.L., 2000. Modulation of hurricane activity in the Gulf of Mexico by the Madden–Julian Oscillation. *Science*, **284**: 2002–2004.

Slingo, J.M., Rowell, D.P., Sperber, K.R., and Nortley, F., 1999. On the predictability of the interannual behavior of the Madden–Julian Oscillation and its relationship to El Niño. *Meteorology and Geophysics Abstracts*, **51**: 20264.

Cross-references

Cycles and Periodicities
El Niño
North Atlantic Oscillation
Oscillations
Pacific North American Oscillation
Quasi-Biennial Oscillation
Southern Oscillation
Teleconnections

MARITIME CLIMATE

A maritime climate occurs in regions whose climatic characteristics are conditioned by their position close to a sea or an ocean. Such regions, also known as oceanic climates or marine climates, are considered the converse of continental climates.

Land and sea differences

About 72% of the Earth is covered by water. The continents and oceans basins are unevenly distributed over the Earth's surface, with the northern hemisphere containing some 65% of the total continental surface. The vast area covered by water and its relative distribution play a very important role in the climates that occur not only at sea, but also, to varying degrees, on land.

Properties of water and land

The climate at any location can be determined by evaluating the amount of energy the surface receives and the way in which it is budgeted. This energy budget for the oceans differs from that of land and, as a result, produces a different climatic regime. The basic physical parameters that give rise to the differences are given in Table M1.

Water has a much greater capacity for absorbing and storing heat energy than does other Earth materials. The basis of this difference is the high specific heat of water. Specific heat is defined as the amount of heat required to heat a unit mass by a unit of temperature. In the example given in Table M1, the temperature of water would be raised only one-fifth that of the same amount of a given soil upon applying the same unit of temperature.

The reflectivity of a surface, its albedo, varies over both land and water. Generally, the lighter, smoother, less transparent the surface, the higher the albedo (a perfect reflector has an albedo of 100%). Land surface albedos vary because of the great range of surface types; a desert surface may have an albedo of 25% while an asphalt parking lot may be only 5%. The major difference in albedo over water is the angle of inclination of the sun. If the sun is high in the sky very little reflection occurs, but as the angle increases so the albedo increases. At a sun inclination angle of 90°, water albedo is less than 2%; at 10° it becomes 35%.

Thermal conductivity is the amount of heat that can pass through a given thickness of material in a given time (using standardized units). It is roughly proportional to the density of the material, causing the surfaces of dense material to heat slowly. If transparency is considered along with conductivity, then a marked difference is seen between land and water. Transparency is a measure of how much radiant energy (such as sunlight) can penetrate a surface. Since opaque a materials, granite rocks for example, have no transparency, radiant energy is transformed to heat energy at the surface. The transparency of water permits energy to penetrate to various depths with most infrared energy absorbed close to the surface; with increasing depth other portions of the electromagnetic radiation energy are absorbed until, at about 10 m, the unabsorbed blue portion of the spectrum is reflected back. This property gives the sea its blue color.

Table M1 Selected properties of air, soil and water

Material	Specific heat	Albedo (%)	Thermal transparency	Conductivity
Air	0.17	6	Very low	High
Soil	0.2	5–20	Variable	Low
Water	1.0	5–10	Medium	Medium

The heat budget

To these physical parameters must be added the dynamic elements that make up the heat budget of a surface. These include:

1. The amount of energy used to evaporate water is much greater over the ocean than the continent. This means that a large part of available energy goes to change water to water vapor. This latent heat is later released when water vapor changes back to water.
2. The large portion of available energy over the ocean that is used for evaporation, means less is available for sensible heat, the heat energy that is distributed through the atmosphere through turbulent transfer. This causes air over the land to be warmer than that over the ocean.
3. As noted, oceans absorb radiation to a substantial depth; over land the penetration of energy is limited to often less than a few feet. The oceans are mobile and the absorbed energy may be advected to other areas through currents and drifts.

The combined result of the physical and dynamic differences cause temperature variations in oceans to be much less than those of a land surface. The ocean is an equable environment that absorbs, transports and eventually releases stored heat to the atmosphere. Oceans thus warm up more slowly than landmasses but, because they conserve the heat, they cool down more slowly. Temperatures in the oceans are conservative, those over landmasses, which warm and cool quickly, are extreme. This maritime effect will be most clearly seen in extratropical regions where seasonal extremes are expected. In tropical latitudes the seasonality is more a function of precipitation than of temperature.

The influence of oceans

Surface ocean currents

Surface ocean currents are wind-driven and, like the wind, are subject to considerable variation. The actual pattern of world ocean currents can be related to the global circulation of winds and the shape of landmasses. Just as the subtropical high-pressure zones are a dominant feature of the atmospheric circulation so are the oceanic gyres (oceanic circulation cells) associated with them. On viewing a generalized map of ocean currents of the world the dominance of the subtropical gyres is seen.

The direction in which currents flow is important. Those flowing poleward, the warm currents, transport heat energy from low to high latitudes, while the cold currents, flowing from high to low latitudes, bring cool water to warmer areas. The relative impact of this movement upon temperature, particularly coastal temperatures, is illustrated by temperature distribution over the North Atlantic Ocean. The 10°F (−12°C) isotherm is located at about 50°N on the east coast of North America but extends to almost 70°N in western Europe. This is largely the result of the influence of the warm waters by the North Atlantic Drift.

Upwelling

The top waters of the ocean, perhaps down to 100 m, are mixed by motion of the wind and waves. Below this mixing layer there is a general decline in temperature until the coldness of the bottom waters is reached. The strong temperature contrast between top and bottom waters inhibits vertical mixing unless cold water is forced up to the surface. This upwelling occurs in a number of locations, often where surface water is swept away from the coast by divergence or offshore winds.

The upwelling process, while locally important in influencing ocean temperatures, is really part of a centuries-long circulation of the oceans that involves both surface and deep currents. Upwelling represents the return of cold waters to the surface circulation while downwelling is the return of surface water to the depths.

The influence of upwelling, together with cold ocean currents, will affect the rainfall regime of some coastal regions. By inhibiting convection, cold offshore waters result in desert climates along some coastlines. Although the dominant air is maritime, precipitation is minimal. The Atacama and the desert of Baja California provide examples.

Maritime airmasses

Air derives its thermal and moisture properties from the surface over which it originates. Air originating over the oceans, maritime tropical (mT) or maritime polar (mP) air, is generally characterized by (a) relatively small changes in the annual range of temperature in the source region and (b) a high moisture content derived from its origin over the oceans. These characteristics are, of course, reflective of the ocean properties already outlined. The relative frequency of maritime air received by a location will play a significant role in determining its climate.

Precipitation

Maritime precipitation in the westerly belts is marked by the cyclonic activity characteristic of the polar fronts. These are most intense in the winter months, but the fall season is likely to be wettest. The amount of evaporation into the maritime airmasses is related to both water and air temperature over the ocean. Accordingly, since the warmest conditions of the winter season occur in the fall, the latter is marked by the heaviest precipitation. In midwinter and spring a cold continental airmass may become more important.

There is marked diurnal variation in maritime precipitation, having its maximum at night or in the early morning, in contrast to the continental type, which tends to occur in the afternoon or early evening (Haurwitz and Austin, 1944), although admittedly many variants are known. There may, for example, be a maritime type in winter and a convectional, continental type in summer (especially in the lower-latitude stations). In the tropics there is greater radiation over the ocean at night, although the sea surface remains warm; thus the nighttime temperature gradient is steeper than that of the daytime, and convectional rain is common at night.

Humidity

Over the oceans humidity will reach a maximum in the afternoon and a minimum in the early morning, reflecting the diurnal air temperature cycle. The vapor pressure is controlled in part by the air temperature and in part by the turbulent transport that tends to disperse the humid layer upward, but the latter is less important over sea than over land, where a double cycle tends to develop over the 24-hour period. Relative humidity is lowest in the high-pressure belts of each hemisphere. In the completely maritime stations (all-year) relative humidity is likely to be over 80% at all seasons

Climates near the oceans

The discussion of the factors that influence climates over the oceans can now be related to the climates that occur in regions close to the oceans. It has been emphasized that the oceans are conservative in terms of temperature changes; such is also the characteristic of maritime climates.

The equable maritime climate is found in small, isolated oceanic islands that are most representative of the influence of maritime airmasses and oceanic climate controls. Table M2 shows examples.

There is an appreciable climatic difference between east and west coast climates in middle latitudes. In continental areas the influence of the oceans on climate is experienced in the west coast locations. This is a result of the general circulation of the atmosphere which, outside of the tropics, is dominated by a westerly air flow. Thus the most frequent airmasses that influence the west coast of middle latitudes are from maritime source regions. Those on the east coasts experience air that moves from a continental source.

Ocean currents, which themselves reflect atmospheric circulation, influence both temperature and rainfall in some west coast locations. The cool air of the offshore water lowers annual temperatures while the resulting stable air provides little rainfall.

Continentality and oceanicity

To measure the effect of a continental landmass on climate (i.e. a minimum impact of oceans), climatologists use the concept of continentality. As early as 1888 it was suggested that, by measuring the average annual range of climate and adjusting for latitude, continentality could be quantified. Later workers used this idea as illustrated by the formula used by Gorzynsky:

$$K = [1.7(A/\sin\varphi)] - 14$$

where K is continentality, A is annual thermal amplitude and φ is latitude.

Many modifications of this formula have been suggested. Using the formula where continentality is given by

$$[1.7A/\sin(\varphi + 10)] - 14$$

Conrad (1946), showed that Thorshaven in the Faeroe Islands would have a continentality index of 0.

Table M2 Climatic data for some maritime stations

Island	Warmest month	Coolest month	Annual range	Precipitation (mm)
Fiji	26.6	22.9	2.7	3026
Midway	25.9	18.4	7.5	1176
Waitangi	14.9	7.8	7.1	851
Dutch Harbor	11.7	−0.3	12.0	1443

While maritime climates may be expressed by low values of continental indices, oceanicity indices have also been proposed. The first, in 1905, is given by

$$O = 100[(T_o - T_a)/A]$$

where O is oceanicity, T_o and T_a are the mean monthly temperatures for October and April, and A is mean annual range of temperature.

Although the scientific validity of the empiric measures has been questioned (Driscoll and Fong, 1992), they do provide a general comparison of the oceanic and continental influences.

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Bibliography

Conrad V., 1946. Usual formulas for Continentality and their limits of validity. *American Geophys. Union Transactions*, **27**: 663–664.

Currey, D.R., 1974. Continentality of extra-tropical climates. *Association of American Geographers Annals*, **64**: 268–280.

Driscill, D.M., and Fong, J., 1992. Continentality: A basic climatic parameter examined. *International Journal of Climatology*, **12**: 185–192.

Hartmann, D.L., 1995. *Global Physical Geography*, San Diego, CA: Academic Press.

Haurwitz, B., and Austin, J.M., 1944. *Climatology*. New York: McGraw-Hill.

Kerner, F., 1905. Thermisodromen, versucheiner Kartographischen Dartstellung des jährlichen Ganges der Lufttemperatur. K. K. *Geographische Gesellschaft, Wien*, **6**(3).

Köppen, W., 1936. Das Geographische System der Klimate. *Handbuch der Klimatologie*, **1**: 1–44.

Landsberg, H., 1958. *Physical Climatology*, 2nd edn, Dubois, PA: Gray Publishing Co.

Lockwood, J.G., 1974. *World Climatology.* New York: St Martin's Press.

Schroedor, M.J., 1967. Maritime air invasion of the Pacific Coast: A problem analysis. *American Meteorological Society Bulletin*, **48**: 802–808.

Thompson, R.D., 1998. *Atmospheric Processes and Systems*. London: Routledge.

Thornthwaite, C.W., 1948. An approach toward a rational classification of climate. *Geographical Review*, **38**: 55–94.

Trewartha, G.T., 1961. *The Earth's Problem Climates*. Madison, WI: University of Wisconsin Press.

Trewartha, G.T., and Horn, L.H., 1980. *An Introduction to Climate*. New York: McGraw-Hill.

Cross-references

Climatology
Continental Climate and Continentality
Evapotranspiration
Lakes, Effects on Climate
Land and Sea Breezes
Wind and Wind Systems

MATHER, JOHN R. (1923–2003)

John Russell Mather was born in Boston on 9 October 1923. After serving as a weather forecaster in World War II, he earned his BA (1945) at Williams College and a BS (1947) and MS (1948) in Meteorology from MIT.

The earliest days of his long distinguished career were spent at the C. Warren Thornthwaite Laboratory of Climatology as one of the first group of graduate students to become intimately involved in the research of the laboratory. The laboratory, which began operation in 1946, was originally located at Seabrook Farms in New Jersey and, in the spring of 1954, was relocated to Centerton, New Jersey. Mather reveled in the research being conducted there, writing that he enjoyed the job because he could go into the woods and not sit at a desk all day. His investigations while at the laboratory included topics such as micrometeorology, evapotranspiration, soil tractionability, soil moisture, climatic classification, and the movement of radioactive strontium in the soil.

Mather is probably best known for his work with the water budget, defined as the daily or monthly accounting of moisture inflows, outflows and storages at a particular place or over a geographical area for any period of time. He felt that the water budget was the basic tool of applied climatology, providing quantitative information on such things as moisture surplus, deficit, retention and runoff.

Mather earned his PhD from Johns Hopkins University in 1951 and remained at the laboratory for 16 years. When Thornthwaite died in 1963, Mather assumed the presidency of the laboratory, and continued his research there for 9 more years.

It has been said that Thornthwaite had the ideas but Mather put them into practice. In many instances this was true; but Mather took those ideas and refined them, making them his own. He expanded Thornthwaite's 1940 version of the water budget, and in 1955 he co-authored with Thornthwaite, *The Water Balance*, which made the ideas more useful for a wide range of soil and vegetative conditions. Thornthwaite had planned to write a book entitled *Physical and Applied Climatology*. The book was never written; it remained for Mather to do it in 1974. His book, *Climate Fundamentals and Applications*, was the fulfillment of Thornthwaite's plan. In it Mather hoped to show how humans could adjust their activities to the atmospheric environment. Chapters focused on hydrology, water resources, agriculture, clothing, human comfort, human health, air pollution, architecture, and industry.

Mather began teaching part-time at the University of Delaware in 1961. When Thornthwaite died that same year, Mather assumed the presidency of the laboratory and remained as president until 1972 when he left to devote his full efforts to the department at the University of Delaware. Thornthwaite's plans for his Laboratory of Climatology were outlined in a 1943 publication. They included, among other things, the training of climatologists on the graduate level. Students, because of the collaborative nature of the research, would need a broad background in all of the sciences. Thornthwaite's ideas became the foundation on which Mather, as chair, and the faculty built the Geography Department at the University of Delaware.

Mather felt that the best career decision he made was to move to the University of Delaware. He saw creating the new department and hiring new faculty as the most challenging job he faced in his life. As Mather wrote in the department history, three decades after Thornthwaite had written about an institution for climatological research, one had been created at the University of Delaware with the rigor in mathematics and the sciences he envisioned.

Permeating Mather's teaching and research was his desire to make climatology come alive, be dynamic and relevant. He hoped his students would be able to apply the principles of climatology to solving problems concerning humans and their environment. The descriptive, encyclopedic approach to climatology, he felt, was wrong; it was "dry as dust."

Mather's list of publications while at the University of Delaware is extensive, touching a wide variety of topics. He studied large- and small-scale applications of the water budget on climate, soils, vegetation, sea level, stream flow, irrigation, agricultural yield, forest products; the effect of land-use changes on crop yield; the effect of urbanization on the water table; the impact of society on the water budget of a particular area; and natural hazards such as coastal storms. His book, *The Genius of C. Warren Thornthwaite, Climatologist–Geographer*, co-authored with Marie Sanderson, is a major contribution to the history of climatology. In it he and Sanderson detail the life of Thornthwaite and his contributions to the discipline.

Mather's own service to the discipline is extensive: American Geographical Society Council member from 1981 to 2000 and secretary of the Council from 1982 to 2000; President of the Association of American Geographers in 1991–1992; Director of the Center for Climatic Research at the University of Delaware from 1972 to 1991; and State Climatologist for Delaware from 1978 to 1991, to name just a few examples.

Mather received many awards: the University of Delaware's Excellence in Teaching Award in 1989; the Commander's Award for Public Service, US Department of the Army, in 1990; the Francis Alison Award for distinguished scholarship at the University of Delaware in 1991; the Association of American Geographers Career Achievement Honors Award in 1998; the Association of American Geographers Climate Specialty Group's Lifetime Achievement Award in 1999; and the American Geographical Society's Charles P. Daly Award in 1999 for "valuable distinguished geographical labors."

Mather retired from the University of Delaware in the fall of 2000 as *professor emeritus*. He died on 3 January 2003, at the age of 79.

Mather was a "people" person who enjoyed interacting with colleagues and students. His legacy is huge; his ideas living on in his students and publications. He called Thornthwaite the "Father of Applied Climatology" and the Laboratory of Climatology the "mecca" for climatologists from around the world. Because of his position as a valued partner of Thornthwaite, Mather spread the gospel of the water budget to succeeding generations. He was eulogized as "an able leader, whose contributions to geography and especially physical geography have rarely been equaled."

Sandra F. Pritchard Mather

Selected publications

Mather, J.R., and Thornthwaite, C.W., 1955. The water budget and its use in Irigation. *1955 Yearbook of Agriculture*, US Department of Agriculture.

Mather, J.R., Thornthwaite, C.W., 1955. *The Water Balance*. *Publications in Climatology*. Laboratory of Climatology, Vol. VIII, No. 1.

Mather, J.R., 1960. The laboratory of climatology – an institute for climatic research. *Weather*, **XV**(5).

Mather, J.R., 1972. Thornthwaite's legacy to climatology. *Publications in Climatology*, Laboratory of Climatology, Vol. XXV, No. 2.

Mather, J.R., 1974. *Climatology: fundamentals and applications*. New York: McGraw-Hill.

Mather, J.R., 1980. Hydrologic cycle, Evapotranspiration, Thornthwaite climatic classification. Three articles prepared for and published in *American Academic Encyclopedia*. Princeton, NJ: Arete.

Mather, J.R., 1985. The water budget and the distribution of climates, vegetation and soils. *Publications in Climatology*, Laboratory of Climatology, Vol. XXXVIII, No. 2.

Mather, J.R., and Sanderson, M., 1991. *The Genius of C. Warren Thornthwaite: climatologist–geographer*, Tucson, AZ: University of Arizona Press.

Mather, J.R., 1966. Water budget climatology. In Hanson, S., ed., *Geographic Ideas That Have Changed the World*. New Brunswick: Rutgers University Press.

MAUNDER, EDWARD WALTER (1851–1928), AND MAUNDER MINIMUM

Maunder was the son of a Wesleyan minister, best remembered for two things: his solar research, and as the founder of the British Astronomical Association. He attended King's College in London, and later secured a post as photographic and spectroscopic assistant at the Royal Greenwich Observatory; he initiated a long series of daily photographic sunspot records there. Eventually he was able to prepare a famous diagram, known as Maunder's "butterfly diagram", linking the latitudes of sunspot groups with the state of the 11-year solar cycle. He also carried out spectroscopic work. He publicized the apparent dearth of sunspots between the years 1645 and 1715, coinciding with a very cold spell in Europe; this is now formally known as the "Maunder minimum". His photographic records of sunspots were meticulous, and are still of great value today.

The sun was by no means his only interest. He observed the planets, and in a famous experiment demonstrated that the "canals" on Mars, claimed by Lowell and other observers, were optical illusions. He also made careful studies of the zodiacal light. He took part in various eclipse expeditions, notably the Royal Greenwich Observatory expedition to Mauritius in 1901, during which he obtained an excellent series of photographs. On other expeditions in 1896 and 1898 the weather was unkind. Between 1877 and 1887 Maunder was coeditor of the *Observatory Magazine*; for many years he served on the Council of the Royal Astronomical Society, and was joint Honorary Secretary from 1892 to 1897. In 1890 he founded the British Astronomical Association, and subsequently served as President. He officially retired from Greenwich in 1913, but was recalled during World War I to carry on the sunspot record.

His first wife died in 1888. He then married Miss Annie Russell, herself a brilliant mathematician and an astronomer on the staff of the Royal Greenwich Observatory, who collaborated with him in much of his later work in connection with the sun. He wrote several books, including a history of the Royal Observatory (1900), *Astronomy Without a Telescope* (1901) and *Astronomy of the Bible* (1908).

Maunder minimum

The Maunder minimum refers to a period from about 1645 to 1715 when sunspot activity virtually disappeared (Eddy, 1976; Schove, 1983).The Maunder minimum is one of a number of protracted periods of low sunspot activity; among others identified are the Oort minimum (1010–1050), the Wolf minimum (1280–1340) and the Spörer minimum (1420–1530). It was following the work of Gustav Spörer that Maunder first called attention to the interruption of the normal course of sunspot activity during the second half of the seventeenth and first part of the eighteenth centuries. Maunder published his findings in

1894 but the research appeared to have created little academic interest. In 1922 Maunder published another paper with the same title as that produced earlier (A prolonged sunspot minimum), but it remained for more recent astronomers and astrophysicists to verify and discuss it. Eddy (1976) noted that the sunspot dearth, 1645–1715, is supported by direct accounts in contemporary literature of the day. That the low sunspot activity was real, and not caused by a limitation in observing capability, is illustrated by drawings made of the sun at the time. Additionally, a tree-ring anomaly spanning the same period shows evidence of a concurrent drought in the US Southwest (Douglass, 1928) and when a ^{14}C lag time is assumed, there is a high degree of agreement in time between major excursions in world temperatures and excursions of solar behavior in the records of ^{14}C. The coincidence of Maunder's prolonged sunspot minimum with the coldest excursion of the Little Ice Age has been noted by many researchers examining possible relations between solar activity and terrestrial climate (e.g. Schneider and Mass, 1975; Fairbridge, 1987; Grove, 1988). Given the time relationship between the Little Ice Age and the Maunder minimum, the role of solar variability and sunspot activity has also been examined in terms of recent and future climatic change (e.g. Kelly and Wigley, 1990) and as it may be related to possible changes in the sun's brightness (Baliunas and Jastrow, 1990).

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Bibliography

Anon, 1928. E.W. Maunder. *Monthly Notes of the Royal Astronomical Society*, **89**: 318.
Anon, 1928. E.W. Maunder. *Journal of the British Astronomical Association*, 38, 231.
Anon, 1926/7. A.S.D. Russell (Mrs Maunder). *Journal of the British Astronomical Association*, **57**: 238.
Baliunas, S., and Jastrow, R., 1990. Evidence for long-term brightness changes in solar-type stars. *Nature*, **348**: 520–523.
Douglass, A.E., 1928. *Climatic Cycles and Tree Growth*, vol. 2. Washington, DC: Carnegie Institute.
Eddy, J.A., 1976. The Maunder minimum. *Science*, **192**: 1189–1202.
Fairbridge, R.W., 1987. Little Ice Age. In Oliver, J.E., and Fairbridge, R.W., eds., *The Encyclopedia of Climatology*. New York: Van Nostrand Reinhold, pp. 547–551.
Grove, J.M., 1988. *The Little Ice Age*. London: Methuen.
Kelly, P.M., and Wigley, T.L.M., 1990 The influence of solar forcing trends on global mean temperature since 1861. *Nature*, **347**: 460–462.
Maunder, E.W., 1894. A prolonged sunspot minimum. *Knowledge*, **17**: 173.
Maunder, E.W., 1922. A prolonged sunspot minimum. *Journal of the British Astronomical Association*, **32**: 140.
Schneider, S.A., and Mass, C., 1975. Volcanic dust, sunspots and temperature trends. *Science*, **190**: 741–746.
Schove, D.J., 1983. *Sunspot Cycles* (*Benchmark Papers in Geology*, vol. 68). Stroudsburg: Hutchinson Ross.
Warner, D.J., 1974. Maunder, Edward Walter. *Dictionary of Scientific Biography*, vol. 9, pp. 183–185.

Cross-references

Little Ice Age
Sunspots

MAURY, MATTHEW FONTAINE (1806–1873)

Matthew Fontaine Maury was born in Spottsylvania County, Virginia in 1806 and died in Lexington, Virginia in 1873. Maury worked his way from midshipman to Lieutenant in the US Navy. From 1825 to 1834 his voyages took him to Europe, the Pacific coast of South America, and around the world. He showed special interest in navigation, publishing a *Treatise on Navigation* in 1836. Although his sea career was ended by a leg injury in 1839, Maury was appointed as superintendent of the Depot of Charts and Instruments in 1842. His job title changed to director of the US Naval Observatory and Hydrographical Office in 1844, when the Depot was moved and renamed. In this post Maury collected both ocean current and meteorological data, recorded in specially designed logbooks, from sea captains. He began publishing *Wind and Current* pilot charts of the North Atlantic in 1847, helping to reduce sailing times. He also proposed cooperative weather observation stations on land. Maury used two research vessels after 1849 to collect ocean temperature data and sea-floor samples. Data collected led to his publication of a bathymetrical map of the North Atlantic Basin in 1854, which depicted the sea floor between Yucatan and Cape Verde.

Maury wrote *The Physical Geography of the Sea* in 1855, recognized as one of the earliest oceanographic books, if not the earliest. Maury is widely considered to be "the father of oceanography". Several editions of *The Physical Geography of the Sea* were published, with later iterations taking on the title *The Physical Geography of the Sea, and its Meteorology*. With the exception of polar movement, Maury's 1855 depiction of global circulation appears to be remarkably close to current understandings. As important as his contributions were, however, many of his ideas rightly received criticisms on scientific grounds (e.g. tradewind control by magnetic forces).

Maury resigned and became commander of the Confederate Navy during the Civil War, then Imperial Commissioner for Immigration under Emperor Maximilian of Mexico. With the collapse of the Mexican Empire he moved to England, returning to the US in 1868 as professor of meteorology at Virginia Military Institute. His professorship there lasted until his death in 1873.

Lisa M. Butler Harrington

Bibliography

Daintith, J., and Gjertsen, D., eds., 1999. *A Dictionary of Scientists*. New York: Oxford University Press, pp. 365–366.
Maury, M.F. (edited by John Leighly), 1963. *The Physical Geography of the Sea and its Meteorology.* (Based on the 8th edn, 1861.) Cambridge, MA: Belknap Press.
Oliver, J., 1996. Climatic zones. In Schneider, S.H., chief editor. *Encyclopedia of Climate and Weather*, Vol. 1. New York: Oxford University Press, pp. 141–145.
Stringer, E.T., 1972. *Foundations of Climatology: an introduction to physical, dynamic, synoptic, and geographical climatology*. San Francisco, CA: W.H. Freeman, p. 5.

MEDIEVAL WARM PERIOD OR "LITTLE CLIMATIC OPTIMUM"

Histories in western Europe have long recognized a "post-Carolingian" climatic amelioration, approximately since the death of Charlemagne (AD 814), the first of the "Holy Roman Emperors" of the Middle Ages. Long-term indicators (sea level, glaciers, etc.) suggest it may have begun around AD 550, i.e. the "post-Roman Rise". The expression "Little Climatic Optimum" (LCO), a term believed to have been coined by Bryson, is essentially synonymous with "Early Medieval Warm Epoch" (Lamb, 1977), our "MWP", and also with the "Sub-Atlantic" (SA-2) Biozone of some palynologists.

Lamb (1977, p. 35) identifies a broad long-term "summer wetness index" (rising) and compares it with a "winter severity index" (falling), based on 50-year means. Both display a systematic lengthening of the upper westerlies' wavelength across northern Europe, reaching its peak around AD 1100. A comparable warming swing marked the twentieth century, but is now complicated by anthropogenic pollution. The overall cycle would appear to be about 900 years. The two were separated by the "Little Ice Age" (LIA) from 1300 to 1750. The warmest interval of the MWP in Europe was 1150 to 1300. A study of Canadian borehole temperatures (mentioned by Lamb, 1977, pp. 104–105) suggests that the mean MWP was up to 1.6°C warmer than the average for the last millennium. An interesting humanistic comparison of the MWP and the LIA is made by Fagan (2000).

Sea-surface temperatures in the high latitudes were higher, and sea-ice coverage was appreciably reduced, thus introducing a positive feedback with the decreasing albedo of open water. These factors played an important role in the occupation of Iceland by Viking (Norse) colonists in the ninth century (AD 860). In the North Atlantic there were frequent summers with calm seas and easterly winds that greatly favored their particular type of ships (shallow draft and square sails). Eventually there came the explorations by Eric the Red and Lief Ericsson, the colonization of southern Greenland by the Norsemen (Dansgaard et al., 1975), and their brief settlement (AD 986) at Anse aux Meadows in Newfoundland ("Vinland"), the first European foothold in North America.

In the same MWP Arctic warming came a widespread development of Eskimo cultures, reaching as far north as Ellesmere Land (AD 900). From the Bering Sea they spread out to northern Alaska and in Siberia, as far north as the New Siberian Islands. In the eleventh century a new wave of migrants brought the "Thule Culture" that spread from Alaska to northern Greenland.

In North America native peoples spread northward up the valleys of the Mississippi and Missouri, bringing an agricultural economy into Wisconsin and Minnesota, where pollen studies have shown there were reliable summer rains (Griffin, 1961). Eventually they reached the northern Rockies and even Utah. The Hohokam people were meanwhile developing settled agriculturally oriented communities in New Mexico and Arizona.

In Asia the behavior of the Caspian Sea provides hydrographic evidence of the MWP climate. From the ninth to fourteenth centuries the rise of the Caspian by 18 m flooded vast areas of the

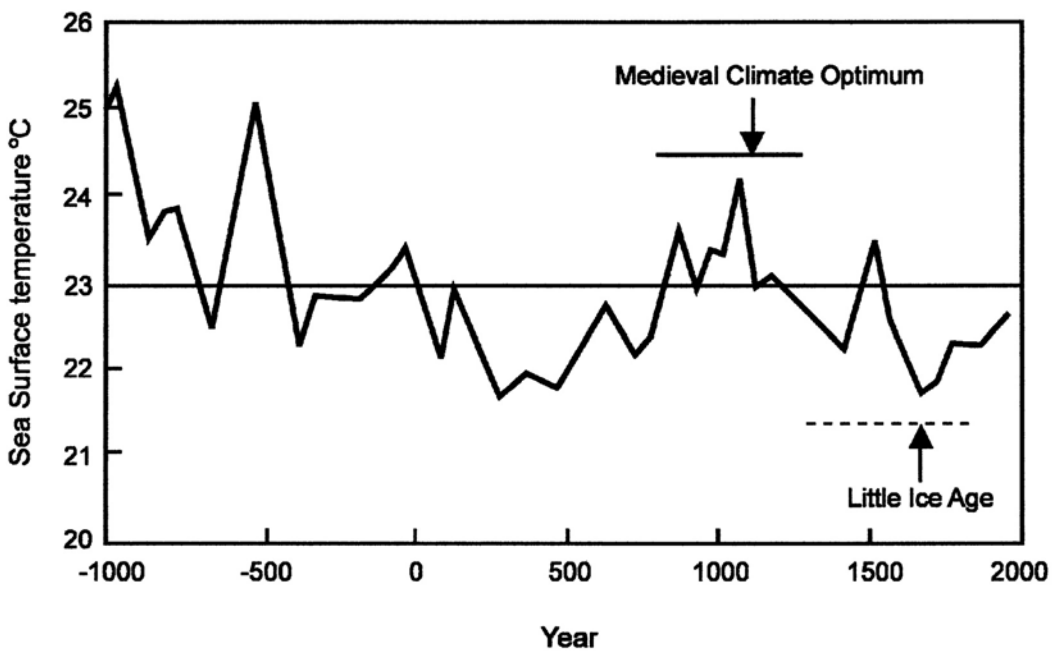

Figure M1 Trace of Sargasso Sea temperature, over 3000 years, based on benthic foraminifera oxygen-18 isotopes (adapted from Keigwin et al., 1994). The positions of the Medieval Warm Period and the Little ice Age have been expanded by the writer (R.W.F.) to conform to the other climatic proxies identified in the text.

Volga delta, reflecting a general increase in precipitation over northern Russia. Clearly this also reflects increased evaporation over the warming Gulf Stream and North Atlantic in general. The populations of central Asia showed a general expansion, but this reached a crescendo in the early thirteenth century, which launched the celebrated conquests of Genghis Khan and his "Mongol Hordes" (AD 1205–1225).

It was the exploration of central Asia in the first decades of the twentieth century that provided the database for Ellsworth Huntington's concept of climatic determinism (Huntington, 1924) that set up a great deal of controversy at a time when climate was assumed (by the "Establishment") to be a changeless aspect of the environment. A long sediment core obtained from Lake Saki (Crimea) illustrates the higher precipitation levels of the LCD also in southern Russia. The thickness of clay varves suggests a heavier rainfall at this time than for 2300 years before (Lamb, 1977, p. 408; Xanthakis et al., 1995).

The MWP warm cycle was registered also in Scandinavia and the Alps by glacier retreats (Röthlisberger, 1986; Karlén et al., 1995). Particularly in Sweden such glacier melts have been dated most precisely by sedimentation measurements in the corresponding glacial lakes. Worldwide response was marked by eustatic sea-level indicators including higher-than-normal beach ridges in many different regions, ranging from the Arctic coast of Alaska, to the coasts of the Gulf of California, and southwestern Florida (Fairbridge, 1992). The first radiocarbon dates of this high sea-level stage were obtained in Western Australia, on Rottnest Island, and became thus the "Rottnest Stage" (Fairbridge, 1961). In Scandinavia it is often referred to as the "Viking Stage" from the abundant relics displayed by the isostatic emergence around Stockholm and Uppsala. In northern Germany, the Netherlands and Belgium (Bennema, 1954), it is partly drowned by tectonic subsidence, but after correction its level was about 0.5 m above present MSL.

Inundation of low-lying coastal areas completely changed the geography in some areas. There was frantic dike building in northern Germany, the Netherlands, Belgium, France and England. The critical moment when man-made dikes were overtopped in particular storms is recorded by historic disasters when tens of thousands of people and livestock were drowned while farms and whole villages were destroyed (Bakker, 1957). In the lower Rhine valley river floods amplified the oceanic factors (Berendsen, 1995). Outstanding events included the flooding of the lower Ijssel valley in the Netherlands with the creation of the Zuider Zee (1250/1251, 1287), and in Britain with the creation of the Norfolk Broads (Lamb, 1977, p. 433). Other areas were affected including western France, the Rhone and Po deltas.

The MWP ("Rottnest") rise of sea level had an important demographic effect inasmuch as it drove coastal dwellers inland, creating endless strife with the people (mostly farmers and pastoralists) already occupying the areas. In the southern Baltic (modern Poland and northeast Germany), southern Denmark and the Frisian belt in northwest Germany, postglacial tectonic subsidence amplified the eustatic rise of the MWP. An economic side-effect of the MWP sea-level rise was in the salt industry. During periods of low sea level, as during much of the Roman era, there were coastal salt pans in which sea-water ponds were organized in series to permit progressive evaporation. This was a major industry because common salt (NaCl) is an essential dietary condiment, as well as being in widespread demand for food preservation, leather tanning, etc. With the MWP rise, the salt pans were repeatedly inundated and in many cases could not be expanded inland because of low bluffs, dunes or cliffs. Industrial disaster was alleviated by a wholesale shift inland to "fossil" salt deposits, mostly of Triassic or Permian age that had been long exploited since the Bronze Age or earlier. These deposits are known in Britain, Spain, France, Germany, Austria, Poland and Russia (Moores and Fairbridge, 1997). Many geographic names in central Europe incorporate the roots Salz, Sal, etc., e.g. Salzburg, Salzkammergut, Salzach, etc.

Metallic ores were extensively mined in mountainous areas of Europe when glaciers receded and exposed dry outcrop areas. With the ensuing Little Ice Age (after 1300), there were neoglacial advances and flooding of mines, leading to closures of the high passes in the Alps (Lamb, 1977, pp. 273–274).

Lamb (1967, 1977, pp. 276f.) was fascinated by the history of grape vines and viticulture in Medieval Britain, a country not particularly celebrated today for its home-grown wines. There were widespread vineyards south of latitude 53°N in Medieval times, some 500 km north of present-day limits in France and Germany. To quote from William of Malmesbury writing in 1150 (translated from Latin) of the vale of Gloucester "here you may behold highways and public roads full of fruit-trees, not planted, but growing naturally.... No county in England has so many or so good vineyards as this, either for fertility or for sweetness of the grape...." The export of British wines to France and Germany was a major factor in international trade, and even led to diplomatic friction. It should be recalled that in 1066 William of Normandy had led an invasion of Saxon Britain, at that time considered a remarkable prize. But then came the Little Ice Age, and soon after 1300, within 10 years the entire wine industry collapsed, the vines surviving only in rare cases under the protection of walled gardens.

The eastern parts of North America and Greenland seem to have shared equally with the European experience of the MWP, as did New Zealand and other parts of the southern hemisphere. However, much of the Pacific region, China, Japan and the western part of North America experienced the exact opposite. In Japan the mean blooming dates of the cherry trees of the imperial gardens in Kyoto have long been monitored as the harbingers of spring and the occasion of joyful festivals. For the MWP, however, the mean blossoming date was a full 10 days later than the millennial average, or 11 days comparing the mean for the twelfth century with that for the fifteenth century. To Lamb (1977, p. 400) "this suggests an eccentric position of the circumpolar vortex over the northern hemisphere in Europe's High Middle Ages, with the climatic zones persistently displaced north over the Atlantic sector and south over the whole Pacific sector and the Far East".

Archeomagnetic studies employing the paleomagnetic orientation of baked bricks or fireplace clays indicate that the North Magnetic Pole at the time lay in the eastern hemisphere (Bucha, 1984, 1988). Naked-eye sunspot observations and auroras (closely correlated to solar activity) were widely reported from China, Korea and Japan, which were then much closer to the North Magnetic Pole (Schove, 1983).

In the Mediterranean latitudes there were commonly heavy summer rains during the MWP. In central America these had an important role in bringing about the downfall of the Mayan civilization in Yucatan and elsewhere, where the maize (sweet corn) economy requires less humid seasons (Brooks, 1949), which needed to be nicely adjusted to their rainy seasons. Suffering repeated droughts, the population in the central valley of Mexico reached a minimum around AD 1000–1100.

The same delicate adjustment of seasons may also help to explain the wonderful surge in culture and religious buildings at Angkor in Cambodia, which reached a crescendo around AD 1000. During the Little Ice Age these structures were seasonally inundated by the rise of the Mekong and backfilling of Tonle-Sap that eventually buried the entire complex in 10 m or more of sediment.

Basing his estimates mainly on botanical proxies (palynology), Lamb (1977, p. 404) presented a comparison between the major climatic intervals for Britain of the last 10 000 years. The MWP discloses the warmest annual mean temperatures since the "Atlantic Biozone" 6000 years ago and 0.8°C above the warmest decades of the twentieth century. It was also wetter, except for the two summer months, July and August, which were characteristically dry, and thus good for grain harvesting, haymaking, thatching and so on (besides viticulture, noted earlier).

A series of northern hemisphere-oriented maps is presented by Lamb (1977, pp. 444/5) for selected centuries depicting troughs of the upper westerlies across Europe and the typical frontal depressions. For summer months during the MWP climax (AD 1000–1099), the mean polar jet swung from southern Greenland across northern Scandinavia to the Arctic shore of Siberia. Meanwhile the major frontal depressions radiated from northern Greenland to east-coast North America, from near Spitsbergen to Sicily, from near Lake Baikal to western China, and from Alaska to about Wake Island. For the same areas and season during the LIA (1550–1599) the polar jet ran across Scotland and southern Sweden, while the principal European frontal depressions ran from western Spain across the British Isles and the west of Norway.

For the winter seasons the MWP polar jet crossed Quebec and swung north of Iceland to the Barents Sea while its southern branch crossed Spain and the Mediterranean to Syria. The latter brought plentiful winter rains and prosperity to the kingdom of Palmyra and the peoples of Palestine and Mesopotamia. Five centuries later in the LIA the winter jet unified and crossed Iceland to head for southern Russia, bringing frequent droughts to the Middle East.

These mean jet trajectories help explain the antiphase relationships between climate proxies in western Europe and the Far East. The inferred position of the polar vortex caused swings to and fro, to eastern or western hemisphere; this appears to be analogous, on a greatly amplified (century) scale, with the "Atlantic Oscillation" (AO) which operates on a decadal basis. Biological proxies are very interesting. In central Europe history reports incidents of locust (*Locusta migratoria*) plagues in the drought summers as follows: ninth century, 8 seasons; tenth to twelfth centuries, none or next to none; fourteenth century, 15 seasons; fifteenth century, none; sixteenth century, 6; seventeenth century, one only; eighteenth century, 6; nineteenth century, 4 (Lamb, 1977). From recent data in North Africa Landscheidt (1987) believed there was a correlation between locust plagues and solar radiation.

Summary

1. The MWP, or LCD (Little Climatic Optimum), is a roughly 450-year climatic cycle, about AD 850–1300. Based on historical documentation this was a warming cycle in Europe.
2. Based on (a) CO_2 and temperature indicators in the Greenland ice cores, and (b) ^{14}C flux (inverse) levels in the dendrochronology, this interval is confirmed as a warming cycle, at least 1°C on average above millennial means.
3. Based on the astrochronology (using the 45.4 year resonance interval of the three major planets, Jupiter, Saturn and Uranus), there were three warm peaks: "Post-Carolingian" (AD 852), "Dunkerque-3" (944), "Viking/Rottnest" (1125). Two cool peaks (marked by extended interruptions caused by low sunspots and reduced solar emissions) occurred at the "Normanian" (peak: AD 1034) and the "Ottoman" (AD 1307).
4. In general, precipitation rose in step with the mean temperature oscillations, and glaciers tended to retreat. However, in exceptional areas glaciers showed temporary advances.
5. Mean sea level rose and fell eustatically and in more or less stable crustal areas it caused coastal retreat matching eustatic rise and vice-versa. However, in areas of crustal subsidence or rapid compaction, this relationship was reversed.
6. General atmospheric circulation (prevailing westerlies, trade winds, etc.) slackened, but in the sub-Arctic latitudes the 45-year summer storminess cycle persisted. With weakening trade winds, El Niño cycles became more important (but detailed monitoring is not available). However, the associated warm coastal currents reached farther south in Peru, and farther north in California.
7. The Gulf Stream and Kuro Shio were shifted northwards, with their characteristic ameliorating effects in maritime areas (increased oceanicity).
8. Sea ice coverage in the North Atlantic and Barents Sea decreased. Due to the longer ice-free seasons in areas of abundant sediment supply the beach ridge buildup was in places anomalously high.
9. In sub-Arctic lakes and swamp deposits the level of pine (or conifer) pollen rose in contrast to that of tundra species. Farther south the mixed forest boundaries moved northwards.
10. Equatorial lakes displayed little change in level, as did rainfall, but savanna (monsoon) boundaries tended to shift poleward by some hundreds of kilometers.

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Bibliography

Bakker, J.P., 1957. Relative sea-level changes in northwest floods in the Netherlands in recorded time. *Deutsches Geographiches Verhandlung*, **31**: 233–237. (Translation in Leatherman, S.P., ed., *Overwash Processes: Benchmark*, vol. 58. Stroudsburg, PA: DHR, pp. 51-56).

Bennema, J., 1954. Holocene movements of land and sea-level in the coastal areas of the Netherlands. *Geologie en Mijnbouw*, **63**: 343–350.

Berendsen, H.J.A., 1995. Holocene fluvial cycles in the Rhine delta? *Journal of Coastal Research*, special issue 17, pp. 103–108.

Brooks, C.E.P., 1926/1949. *Climate Through the Ages.* London: Ernest Bonn; 2nd edn. (1949); New York: McGraw-Hill (Dover reprint, 1970).

Bucha, V., 1984. Mechanism for linking solar activity to weather-scale effects, climatic changes and glaciations in the northern hemisphere. In Moerner, N.-A., and Karlen, W., eds., *Climatic Changes on a Yearly to Millennial Basis*. Dordrecht: Reidel, pp. 415–448.

Bucha, V., 1988. Influence of solar activity on atmospheric circulation types. *Annals of Geophysics*, **6**(5): 513–524.

Dansgaard, W., et al., 1975. Climatic changes, Norsemen and modern man. *Nature*, **255**: 24–28.
Fagan, B., 2000. *The Little Ice Age*. New York: Basic Books.
Fairbridge, R.W., (ed.), 1961. Solar variations, climatic change, and related geophysical problems. *New York Acad Science, Annals*, **95**: 1–740.
Fairbridge, R.W., 1992. Holocene marine coastal evolution of the United States. *SEPM Special Publications*, **48**: 9–20.
Griffin, J.B., 1961. Some correlations of climate and cultural change in eastern American prehistory. *Annals of the New York Academy of Sciences*, **95**(1): 710–717.
Huntington, E., 1924. *Civilization and Climate*, 3rd edn. New Haven, CT: Yale University Press, (reprinted 1977; Hamden, CT: Archon Books).
Karlén, W., Bodin, A., Kuylenstierna. J., et al. 1995. Climate of northern Sweden during the Holocene. *Journal of Coastal Research*, special issue 17, pp. 49–54.
Keigwin, L.D., et al., 1994. The role of the deep ocean in North Atlantic climate change between 70 and 130 kyr ago. *Nature*, **371**: 323–326.
Lamb, H.H., 1967. Britain's changing climate. *Geographical Journal*, **133**(4): 445–468.
Lamb, H.H., 1977. *Climate History and the Future*. London: Methuen.
Landscheidt, T., 1987. Long-range forecasts of solar, cycles and climate change. In Rampino, M.R., et al., eds., *Climate: history, periodicity, and predictability*. New York: Van Nostrand, pp. 421–445.
Moores, E., and Fairbridge, R W., (eds.), 1997. *Encyclopedia of European and Asian Regional Geology*. London: Chapman & Hall.
Röthlisberger, F., 1986. *10 000 Jahre Gletchergeschichte der Erde*. Aarau: Verlag Sauerlaender.
Schove, D.J. (ed.), 1983. *Sunspot Cycles.* Stroudsberg: Hutchinson & Ross (Benchmark **68**).
Xanthakis, J., Liritzis, I., and Poulakos, C., 1995. Solar-climatic cycles in the 4190-year Lake Saki mud layer thickness record. *Journal of Coastal Research*, special issue 17, pp. 79–86.

MEDITERRANEAN CLIMATES

The Mediterranean climate is distinct for its sunny and dry, mild-to-hot summers, and mild, wetter winters. The archetypal example of this climate is found in the area adjacent to the Mediterranean Sea – hence the name Mediterranean climate. Because its summer season is markedly drier than the other seasons of the year, it is sometimes referred to as the summer-dry climate type. The Mediterranean climate is unique amongst Earth's climates for its pronounced winter maximum of precipitation.

Locations

Besides its occurrence in the Mediterranean Basin, it is found on the west coasts of the continental landmasses, between 25 and 40 degrees latitude. Its proximity to the ocean/sea has a significant influence on its temperature, precipitation, and humidity characteristics – particularly in near-coastal areas. The largest example – in geographic area – of the Mediterranean climate, occurs in the area adjacent to the Mediterranean Sea. Other major areas with this climate type include the west coast of North America, from the Mexico–United States border northward to the Pacific Northwest region of the United States, and southwestern and southern Australia. Smaller examples of Mediterranean climate occur in South Africa (Cape Province) and central Chile. In North America the north–south-trending Cascade and Sierra Nevada Mountains limit the eastern extent of Mediterranean climate. In total, the Mediterranean climate type covers only a very small portion – just 2% – of the continental landmass.

Climate characteristics

The semipermanent subtropical anticyclones in the Atlantic and Pacific oceans are largely responsible for the unique characteristics of the Mediterranean climate type. The strong subsidence and resultant adiabatic warming that occurs on the eastern limb of these anticyclones, produces the pronounced and distinct summer-dry period of the Mediterranean climate. The subtropical highs also act to keep summer cyclones poleward of the region, further reducing the opportunity for precipitation in the summer months.

As an example of just how dry the summer months of the Mediterranean climate can be, Santa Monica, California – located toward the equatorward margin of this climate type in North America – has never recorded more than a trace of precipitation during the months of June, July, and August (Oliver and Hidore, 2002). Winter is the distinctly wetter period for the Mediterrranean climate type because the subtropical anticyclones sag equatorward, allowing midlatitude cyclones embedded in the westerlies to bring rain to the area. In Perth, Australia, 85% of the annual precipitation comes during the winter half of the year. In Istanbul, Turkey, the 6 winter months of the year account for 70% of the precipitation that falls annually. San Diego, California, receives 90% of its total precipitation for the year during the 6 months from November through April (Oliver and Hidore, 2002).

Total annual precipitation amounts range between 40 and 80 cm, and characteristically increase poleward because of the diminished influence of the subtropical anticyclones, and thus greater frequency of penetration of midlatitude cyclones. Precipitation totals also increase in the interior highlands as orographic precipitation falls on the windward slopes of mountain ranges. In southern California, Los Angeles and San Diego average 37 and 26 cm of precipitation a year, respectively. However, in the adjacent mountains, annual precipitation totals are four to five times those amounts – with most of the precipitation falling as snow (Lydolph, 1985). Summers in the Mediterranean Basin are not as dry as they are along the west coast of North America because of the greater instability of the atmosphere in the area.

While the winter half of the year is the wetter and stormier half, the amount of precipitation that falls in any given year is quite variable. Moreover, precipitation events are generally shortlived and the period between events is characterized by fair weather. Thus, even during the winter months, pleasant conditions predominate. One caveat to this rule, however, is required along the California coast in El Niño years. During El Niño events the jet steam over the Pacific steers storms into California, bringing wetter than normal conditions to the region.

Along the west coast of North America, as winter progresses the polar front gradually moves southward, bringing precipitation to the more equatorward areas by late in the winter season. Maximum precipitation occurs during December in Seattle and Portland, in January in San Francisco, and in February in Los Angeles and San Diego (Lydolph, 1985). In the highland areas away from the coast, strong thunderstorms can occur and copious amounts of rain may fall in a short period of time – leading to flooding and hazardous mass wasting events.

Sea breezes and coastal upwelling moderate coastal temperatures in the summer. Daily maximum temperatures rarely get much higher than the low 20s (Celsius). In the winter, because of the moderating influence of the ocean, temperatures remain quite mild along coastal areas. Inland however, temperatures are warmer in the summer and colder in the winter. Sacramento, California, just 120 km to the northeast of San Francisco, has an average temperature for the month of July that is 9°C higher than that of San Francisco (Ahrens, 2002). Inland areas may even experience frost on occasion – which puts economically valuable citrus and other crops at risk.

Coastal areas adjacent to the Mediterranean Sea are warmer than their coastal counterpart in North America because upwelling does not occur in the Mediterranean Sea. Upwelling is quite strong, however, off the California coast during the summer months of the year. Air temperatures all along the length of the California coast, in July, are some 10–11°C below normal for their respective latitudes (Lydolph, 1985). Upwelling tends to be strongest just west of San Francisco Bay, making the area one of the coolest along the entire west coast of North America in the summer months. As moisture-laden air from the Pacific passes over the very cool water close to the coast in the San Francisco Bay area, it is cooled to its dewpoint and fog is produced. This fog is then advected inland. San Francisco, California, is renowned for its numerous days of fog each summer. Fog is common elsewhere along the California coast as well.

Inversions

The subsidence and adiabatic warming that occur on the southeastern edge of the Pacific subtropical high, in the summer, produce strong inversions in southern California's two largest cities, Los Angeles and San Diego. The base of the inversion typically lies only just a few hundred meters above sea level (Lydolph, 1985). Convection from surface heating in the moist marine layer below the inversion mixes emissions from stationary and mobile sources upward, and concentrates them just beneath the base of the inversion, where they collect and form a reddish-brown haze that is a signature of highly polluted urban atmospheres. Mountains to the north and east of the city do not allow for the horizontal dispersion of air, thus letting pollution build to very dangerous levels. Los Angeles, because of its geography and large number of automobiles, is infamous worldwide for its exceptionally high levels of air pollution.

Because inversions are such a common feature of Mediterranean climates, high levels of air pollution are characteristic of many large cities in Mediterranean climates. During the winter months the inversion is not as pronounced because of the equatorward movement of the Pacific High. Moreover, when migrating cyclones pass through the region they dramatically alter the thermal structure of the atmosphere and break down the inversions.

Summary

The abundant warmth, sunshine, high number of rainfree days, and cool sea breezes in coastal areas, make the Mediterranean climate type a favorite vacation destination. Its long, dry summers, however, do not allow for the growth of lush vegetation. Instead, scrubby, low-growing woody plants and trees predominate. In North America this type of vegetation is referred to as chaparral. In the Mediterranean basin such vegetation goes by the name *maquis*. Citrus, grapes (for wine-making), and other fruits and vegetables are grown in Mediterranean climates. Irrigation of these crops is required, however (Lydolph, 1985).

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Bibliography

Aguado, E., and Burt, J.E., 2001. *Understanding Weather and Climate*. Upper Saddle River, NJ: Prentice Hall.

Ahrens, C.D., 2003. *Meteorology Today: an introduction to weather, climate, and the environment*. Pacific Grove, CA: Brooks & Cole.

Lydolph, P.E., 1985. *Climates of the Earth*. Totowa, NJ: Rowan & Allanheld.

Moran, J.M., and Morgan, M.D., 1994. *Meteorology, the Atmosphere and the Science of Weather*. New York: Macmillian.

Oliver, J.E., and Hidore, J.J., 2002. *Climatology, an Atmospheric Science*. Upper Saddle River, NJ: Prentice Hall.

Cross-references

Art and Climate
Australia and New Zealand, Climate of
Europe, Climate of
Climate Classification
Precipitation Distribution

MICROCLIMATOLOGY

Microclimatology, the scientific study of microclimates, is concerned with the atmospheric layer that extends from the surface of the Earth to a height where the effects of the features of the underlying surface can no longer be distinguished from the local climate (American Meteorological Society, 2000). Microclimates vary in response to, and in turn are superimposed upon, larger-scale climates (the mesoclimate and macroclimate). Microclimates span a spatial scale (Figure M2) that includes the envelope of air that surrounds an insect or a spore on a plant leaf to the climatic conditions on a hillslope that permit or preclude the growth of specialty agricultural crops such as grapes for wine production. Hence, the horizontal extent of microclimates is often not rigidly defined, although several millimeters to 1 kilometer is often employed (Oke, 1987). Local climate or topoclimate terminology (Oliver and Hidore, 2002) is often employed between microclimates and mesoclimates to represent spatial scales between several hundred meters and 10 kilometers. Microscale atmospheric phenomena usually have temporal spans of less than a few minutes. Microclimate phenomena and their cumulative interactions, feedbacks and impacts may range from seconds to seasons. The thickness of the air layer of concern in microclimatology is also not rigidly defined (Figure M2). Sometimes it can extend from below the surface (the soil or substrate regime is often considered) to a height where recognizable microclimatic signatures in the atmosphere become essentially indistinguishable. During the night this might extend into the atmosphere for several hundred meters and, during the daytime, for over one kilometer. Usually, a much thinner air layer, in close proximity to the surface, receives particular attention. Another important aspect in the definition of microclimates and their relevance and importance is the perspective that is brought by those observing, studying or utilizing them. Human perspective and its acknowledged subjectivity often play a profound and fundamental role in delimiting the spatial and temporal scales of a microclimate as well as the scope of the controlling mechanisms.

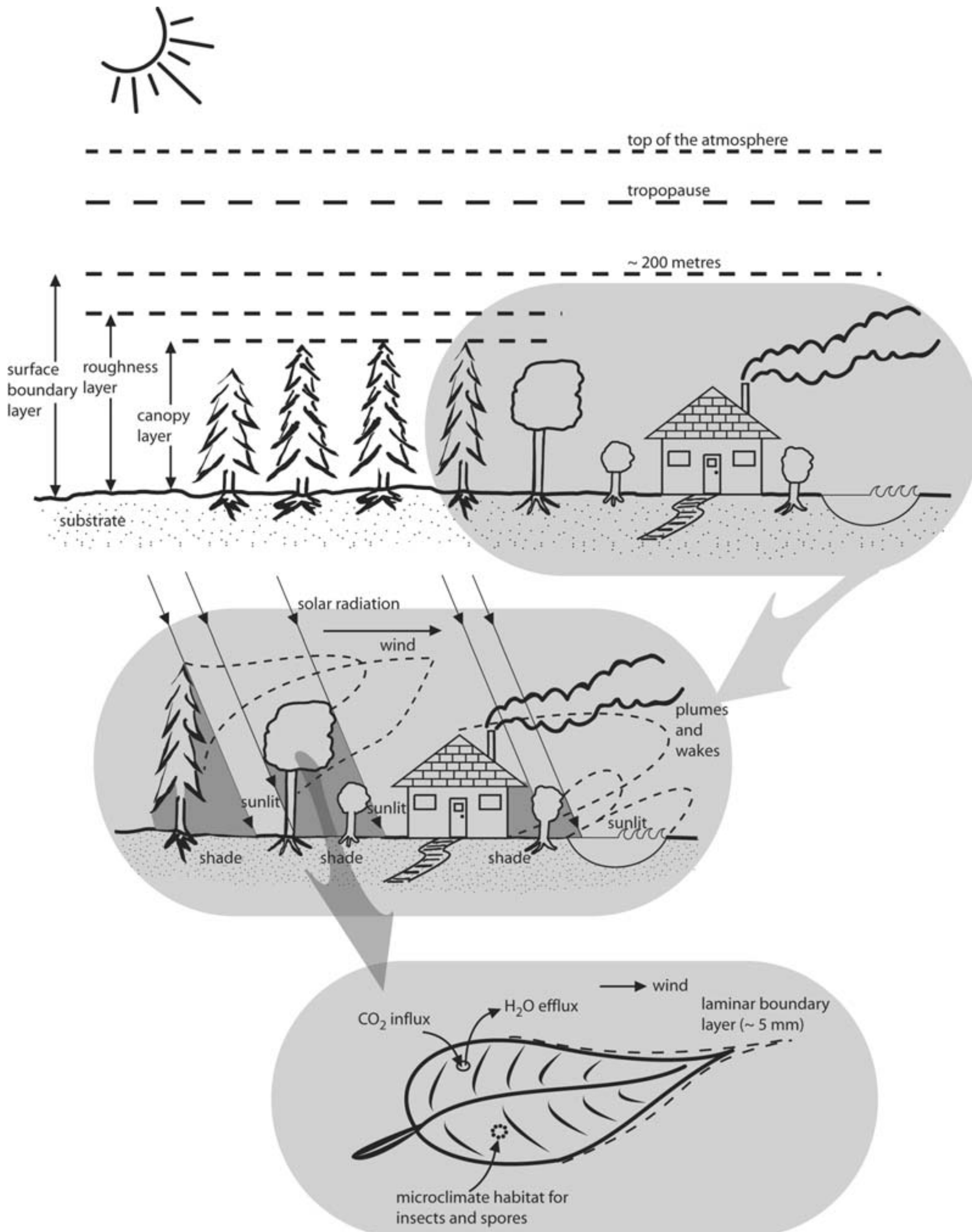


Figure M2 Schematic representation of a vertical cross-section of the Earth's atmosphere. Microclimates extend from the surface, or just below it, to a height where the effects of the surface features cannot be distinguished from the local climate. Numerous microclimates (forest, forest canopy, horticultural plants, building, lake edge, soil surface, insect habitat and others) can be readily identified.

Whereas microclimatology is a subdivision of climatology based primarily on spatial scale attributes, microclimates themselves are often subdivided to consider specific surfaces or habitats. Such subdividing, for example, gives rise to: the microclimatology of plant and animal environments in which the climatic resources for individuals and communities is examined (Bonan, 2002; Jones, 1992; Lowry and Lowry, 2001; Rosenberg et al., 1983; Stoutjesdijk and Barkman, 1992); agricultural microclimatology (Rose, 1966) and forest microclimatology (Lee, 1978) that focus more specialized treatments of particular surface types; forest canopy microclimatology that details biophysical and biochemical processes in foliage environments (Lowman and Nadkarni, 1995; Mynemi and Ross, 1991); urban microclimatology that focuses on human habitation and examines cities, building structures and materials (Brown and DeKay, 2001; Lowry, 1988), air quality (Bell and

Treshow, 2002) and landscape design (Brown and Gillespie, 1995). Additionally, microclimate investigations of many novel environments have been undertaken: cryptomicroclimatology which examines the microclimatology of confined spaces; the microclimatology of cultural heritage which examines the influence of atmospheric variables on cultural objects in both outdoor and museum environments (Camuffo, 1998; Thompson, 1990); animal microclimatology which examines the interrelationship between animals and microclimate variables in both natural and artificial environments (Albright, 1990; Johnson, 1987; Scientific Committee on Animal Health and Animal Welfare, 1999; Unwin and Corbet, 1991). Treatments on microclimatology can be found in general volumes on the atmosphere (for example Oliver and Hidore (2002) and many others) or for specific environments (i.e. mountains (Whiteman, 2000)), in specialized volumes that commence with first principles and cover both theory and application (Campbell and Norman, 1998; Geiger et al., 1995; Lowry and Lowry, 1989; Monteith and Unsworth, 1990; Oke, 1987; Rosenberg et al., 1983; Stoutjesdijk and Barkman, 1992) and in presentations for more focused readerships (i.e. microclimate and spray dispersion (Bache and Johnstone, 1992); extreme heat waves, societal response and human mortality (Klinenberg, 2002); and countless scientific journal publications). Additionally, other treatments have focused on the synthesis of microclimate information for various landscape types within political borders (Bailey et al., 1997).

Human recognition of, and influence by, climate phenomena predate recorded history (Calvin, 1991, 2002). Early human adaptation to the environment, migration and the utilization of resources for both survival and progress attest to an interrelationship with microclimates that parallels the long course of civilization and societal development (Brown, 2001; Butzer, 1971; Jones et al., 2001). Cumulative human impacts across a range of spatial scales have led to marked changes in the Earth's surface, atmosphere and in the concomitant atmosphere–surface interactions. Activities such as deforestation, agriculture, fossil fuel and biomass burning, industrialization and transportation, which occur at the microscale, have led to documented changes that are now being realized at local, regional and global scales (IPCC, 2001a,b,c). In turn, mesoscale and macroscale influences then impact the mosaic of microclimates that blanket the Earth's surface. Although human–climate interaction has been long recognized, the first major synthesis of microclimate research was not published until 1927. *Das Klima der bodennahen Luftschicht* by Rudolf Geiger (1894–1981) was a pioneering contribution, and four German editions appeared before *The Climate Near the Ground* was published in English (Geiger, 1966). It was the most exhaustive work on microclimatology until that time, and a modern edition (Geiger et al., 1995), to celebrate the 100th anniversary of Geiger's birth, now carries the volume's influence into the twenty-first century. Since the end of World War II, considerable advancements in the study of the Earth's environment, and associated with this microclimatology, have been achieved. A number of interrelated factors have played significant roles: the widespread quest for scientific understanding of the environment and the atmosphere has led to substantial increases in overall knowledge; the theoretical understanding of the air layer in contact with the surface, in terms of energy and mass exchange, has been greatly advanced; the development, refinement and extensive application of instrumentation, data recording and computational technology; the deployment of extensive spatial monitoring networks and the widespread availability of data (i.e. worldwide web and many other sources); the lure of and the opportunity to research unknown, challenging and remote locations on the Earth's surface; and the application of remote sensing technologies across a range of spatial and temporal scales for documenting patterns and trends (Kramer, 2002). Advancements have come from many persons in many countries, and contributions can be viewed as examples of interdisciplinary achievements with much microclimatic research being reported in agronomy, architecture, ecology, engineering, forestry, hydrology, physical geography, zoology and other communications.

The creation and control of microclimates

Microclimates arise in response to the external forcings of energy, precipitation and wind, and the magnitude of these forcings establishes the boundaries and character of the microclimates found. Transformations in microclimates arise from changes in the forcings and changes in the responses that occur at the surface to the forcings. Hence, a mosaic of microclimates is readily created over short distance scales.

The surface of the Earth is warmed by solar energy during daylight hours, swept by winds that are driven by interlocked large-scale circulation (macroclimates) and synoptic systems (mesoclimates), and is moistened by precipitation that arrives in either liquid (rain) or solid (snow) forms. Some of the inputs that create the climate of a surface are cyclical in character, for example the diurnal and seasonal variations in solar energy receipt and the associated thermal regimes. Others are quasi-periodic or perhaps close to random in character, for example the seasonal movement of storm systems, and the spatial and temporal occurrence of precipitation. For any geographical setting a general climatology will be created from the integrated influences of latitude, altitude, continentality and location in relation to mesoscale synoptic flows. Within this, an integrated blend of radiative, aerodynamic, thermal and moisture attributes establishes the microclimate for all surfaces (Table M3).

Table M3 Surface attributes important in the creation of microclimates. Adapted after Oke (1997)

Radiative	-Surface albedo, surface emissivity, surface temperature, geometric positioning of the surface and the surrounding environment that will influence radiant energy receipt and loss
Aerodynamic	-Surface roughness length, zero plane displacement, presence of elements upwind that obstruct or channel wind flow
Thermal	-Thermal conductivity, heat capacity, thermal diffusivity, thermal admittance
Moisture	-The surface character (vegetation, soil, etc.) that impacts plant transpiration and/or surface evaporation; the moisture status of the substrate and its availability for evaporation and/or transpiration

Solar radiation (in the wavelength range 0.15–4 micrometers) that reaches a surface has both direct beam and diffuse components. A portion reaching the surface is reflected, and the reflected solar radiation is dependent on the surface albedo. The result is net solar radiation (Figure M3), and this provides a direct contribution to surface heating. The albedo is an explicit surface control of the microclimate. Albedos vary from about 5–10% for water (midday), coniferous forest and dark soils, to

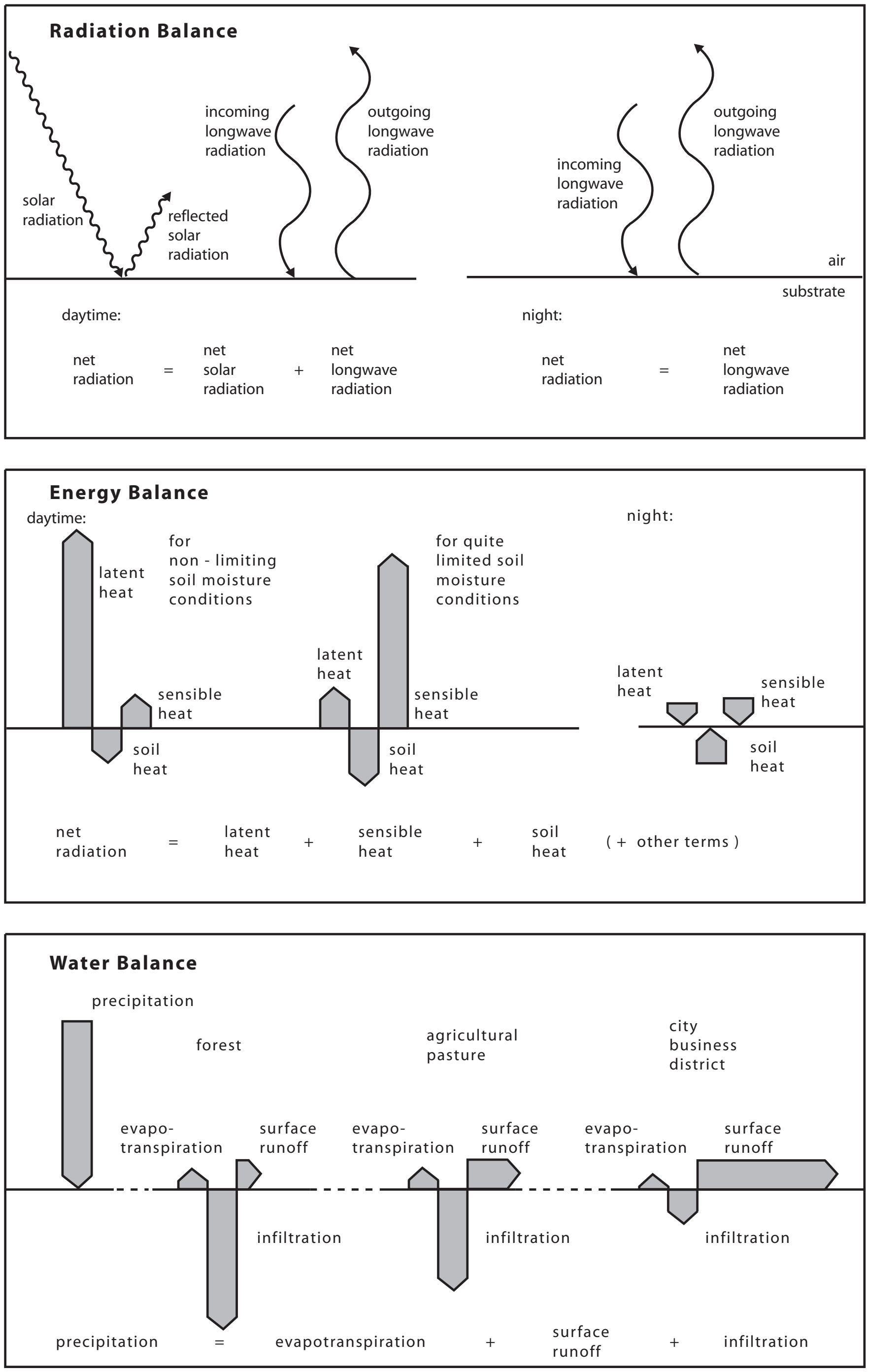


Figure M3 Schematic representation of the radiation, energy and water balances. The radiation balance is presented for both daytime and night periods. The energy balance is presented for both daytime and night periods, with the role of non-limiting and quite limiting surface moisture availability illustrated for daytime periods. For the water balance the response to precipitation input is illustrated for three distinct surfaces (forest, agricultural pasture and a city business district).

20–25% for agricultural crops and other vegetated surfaces, to 80–90% for fresh fallen snow. For three-dimensional surfaces, such as plant canopies, radiation is transmitted from the upper plant canopy toward the ground surface, with absorption and reflection of radiation by the foliage elements occurring. Solar radiation also penetrates into water bodies, snowpacks and glacier ice. Of the solar radiation reaching the Earth's surface, approximately 50% is in the visible spectral range. This also coincides with the approximate wavelength range that plants can biochemically harvest for photosynthesis. All elements in the Earth–atmosphere system also emit radiation, but at longer wavelengths (approximately 5–100 micrometers). The surface is both a recipient of the longwave radiation that is emitted by the atmosphere, as well as an emitter of longwave radiation to the atmosphere. Longwave radiation is governed by the Stefan–Boltzmann law, where the radiation is a function of the emitting surface's temperature raised to the fourth power and its emissivity. As Earth surface temperatures usually exceed the radiating temperature of the atmospheric column, the surface's net balance of longwave radiation is usually an energy loss. The net balance of solar and longwave radiation is known as net radiation (Figure M3). Daytime microclimates respond to the combined influences of solar and longwave radiation. At night the net radiation is solely a longwave radiation regime. The influence of diurnal and seasonal cycles in radiant energy input is pronounced in most microclimates (Figures M3 and M4).

The radiant energy surplus or deficit at the surface obeys the law of conservation of energy through the energy balance. In the most simple of treatments (Figure M3), net radiation is dissipated into latent heat, sensible heat and substrate heat. During the daytime, sensible and latent heat are energies employed in the convective transfer of heat and water into the air. Substrate heat is the energy lost by conduction into the underlying surface. The partitioning of available energy is primarily affected by surface moisture (Figure M3). At night, heating from the atmosphere and condensation often lead to surface-directed sensible and latent heat exchanges, although these are modest when compared to the magnitudes of the daytime regime. The substrate often serves as an energy source at night as conductive energy can flow toward the surface. Vertical profiles of temperature and humidity demonstrate the continuous coupling that exists between the surface and atmosphere in terms of energy and mass exchange (Figure M5). Simple treatments of the energy balance can be readily adapted to incorporate other factors. These could include the energy stored in surface vegetation, the energy used by plants in photosynthesis, the energy used in snow melt, the energy stored in the water bodies such as lakes, the energy contributed by anthropogenic sources, the horizontal transfer of sensible and latent energy from upwind environments, and the lateral transfer of energy to downwind environments. All of these can play significant roles in the creation and character of specific microclimates.

The input of precipitation to the surface and its routing is governed by the hydrological cycle. In the water balance the precipitation to the surface is lost to the atmosphere through evapotranspiration, to surface runoff by horizontal flow and to infiltration. The infiltrated water replenishes surface soil storage and thereafter the drainage flows horizontally or vertically into fluvial or groundwater systems. The character of the surface and substrate plays a pronounced role in the routing of precipitation (Figure M3). Precipitation input is not continuous, and the amount and timeliness is critical to the nature of the hydrological cycle across the range of spatial scales. Additionally,

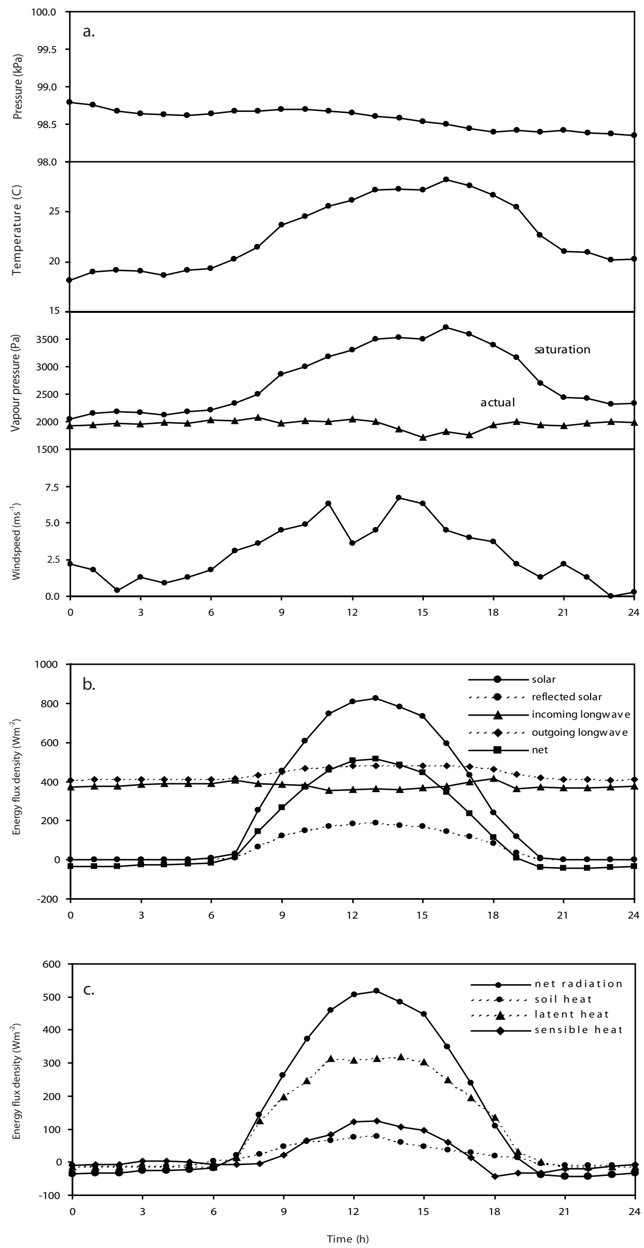


Figure M4 The diurnal trend of midsummer atmospheric variables monitored at a meteorological station and the radiation and energy balance components of a cropped surface located nearby. The day was almost cloudless and available moisture to the surface becomes slightly limiting during the afternoon period.

precipitation falling as snow may well introduce important temporal lags in the response of the hydrological system. The water balance is closely linked with the energy balance, as the transpiration from plant surfaces and evaporation from the soil are the water mass volumes represented in the energy balance by latent heat, the energy utilized in the change of phase of water from liquid to gaseous form. Analogous to the water balance, biogeochemical cycles can be defined and studied for various microclimates, with a prime example being carbon dioxide.

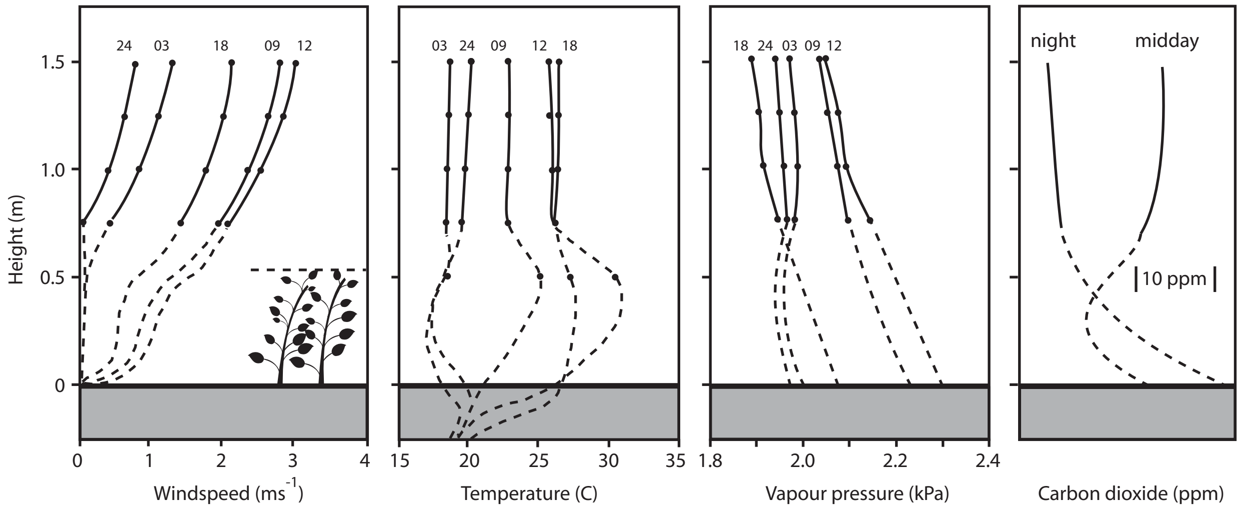


Figure M5 Hourly average vertical profiles of windspeed, temperature, vapor pressure and carbon dioxide for the near-surface layer for selected periods throughout the day (03, 09, 12, 18 and 24 hours). The profiles were measured on the day documented in Figure M5 (cropped surface in midsummer). Profiles within the plant canopy are denoted by dashed lines.

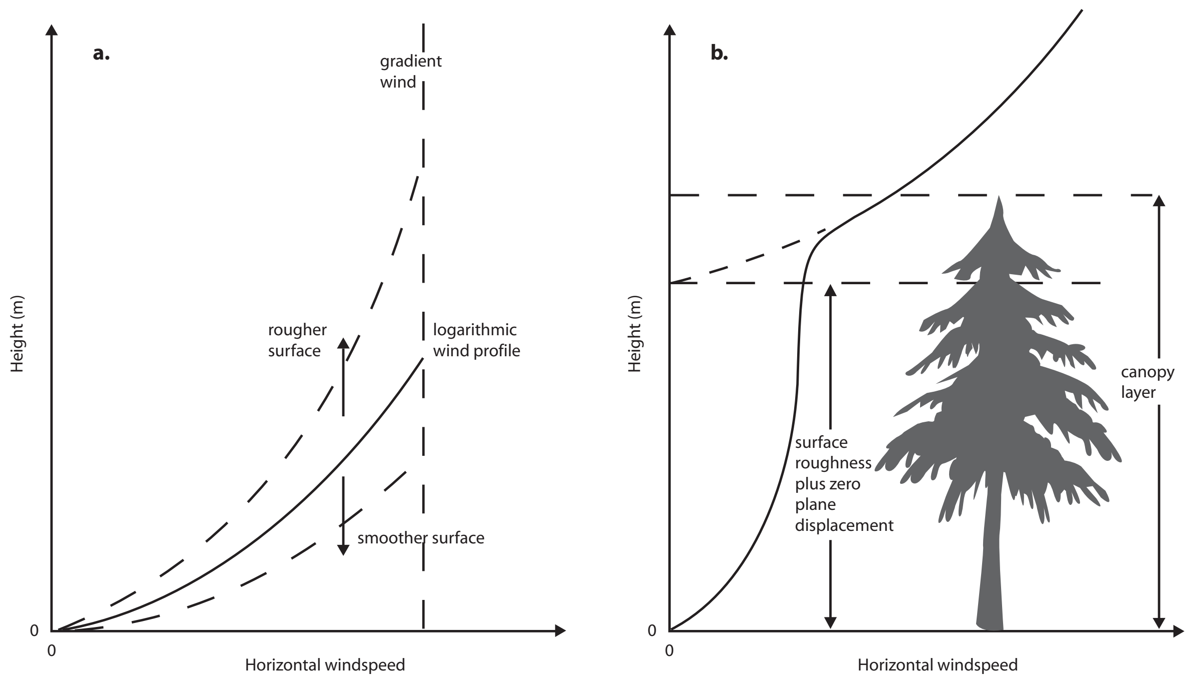


Figure M6 (**a**) The characteristic profile of windspeed near the Earth's surface when thermal enhancement or suppression is absent. Increasing surface roughness, from smooth (surfaces such as snow, calm water and desert sand) to rough (forests, cities), changes the profile of windspeed in the near surface layer. (**b**) The profile of windspeed through a plant canopy. For such three-dimensional surfaces a zero plane displacement must be subtracted from the height scale to preserve the characteristics of the logarithmic wind profile.

The flow of air above the surface, from regions of high pressure to low pressure, illustrates the combined influence of driving and steering forces. As the surface is approached, the wind velocity is reduced in response to friction effects. When no thermal gradients are present, the height variation of windspeed above the surface is logarithmic in character (Figure M6). Smooth surfaces such as water, snow and desert sand are relatively windy, but the effect of the surface extends only a short distance above the surface. Rough surfaces such as tall forests and cities generate more turbulence and affect a much deeper layer of air. The surface roughness parameter denotes the theoretical height where windspeed goes to zero, and it is typically 10% of the average surface element height (Figure M6). If a vegetation canopy covers the surface, the effective surface height is adjusted upward to account for this. This zero plane displacement is approximately 70% of vegetation height (Figure M6).

The fundamental property of flow for a surface is momentum. The momentum aloft is extracted from the flow by the friction effects of terrain roughness. Mixing of air by turbulence results in the creation of eddies. These parcels of air are of varying sizes and, at any instant, may exhibit almost random motion. Each eddy carries the characteristics of heat, mass substances and momentum. The vertical profiles of temperature, water vapor, other gases and windspeed exhibit fundamental linkages between the surface and the atmosphere (Figure M5). Turbulence in the lower atmosphere arises from buoyancy forces (free convection) and mechanical forces (forced convection). Departures from isothermal atmospheric conditions create changes in stability; these demonstrate profound roles in turbulent mixing. During daylight hours, unstable conditions predominate with strong lapse conditions existing (temperatures decreasing with height). Turbulence and, therefore, mixing is enhanced by buoyancy. At night, stable atmospheric conditions prevail with inversion conditions existing (temperatures increasing with height). Buoyancy is minimal and mixing is suppressed.

Few surfaces on the Earth are horizontal and smooth, and microclimate variation arising from topographic influence is commonplace. Topography transforms patterns of solar radiation, temperature, wind and precipitation. North, south, east and west slopes of hills or mountains, or the sides of valleys, can have markedly different microclimates (Figure M7), with solar energy receipt and wind patterns being major contributors to these differences. In northern midlatitudes, south-facing surfaces receive more solar radiation than do other orientations. North-facing slopes may receive no direct beam solar radiation for months, depending on latitude, declination of the sun, slope and orientation of the surface and slope position (valley bottom for example). In rugged terrain, adjacent topographic features may cast shade for portions of the day or year, and some places (valleys and steep poleward-facing slopes for example), may never receive direct beam solar radiation. As a result, each facet of the terrain has a unique radiation balance and a microclimate that differs from adjacent facets (Figure M7).

The contour of the landscape also creates winds that are superimposed on large-scale circulation patterns. Anabatic winds develop along slopes that have been heated by solar radiation. Associated cloud development throughout the daytime hours often provides a visible account of the intensity of the upslope flow. These winds are best developed during anticyclonic conditions, and when coupled with subsiding air in the valley bottom, a larger local circulation system is realized. Nighttime cooling results in an increase in the density of surface air on slopes and downslope katabatic (gravitational) flow results. Such topographically induced patterns of air flow can result in marked differences in near-surface temperature. In spring and fall, frost pockets can develop where cold air drainage is impeded by topographic features (valley bottoms, earthen berms, walls of buildings, rows of trees). In these pools of cold air, near-surface air temperatures at night may be many degrees less than temperatures a very short distance away. In valleys, nocturnal temperature regimes generally increase upslope as the impacts of the frost pockets diminish with elevation. Above the height of these zones of relative warmth, adiabatic cooling prevails.

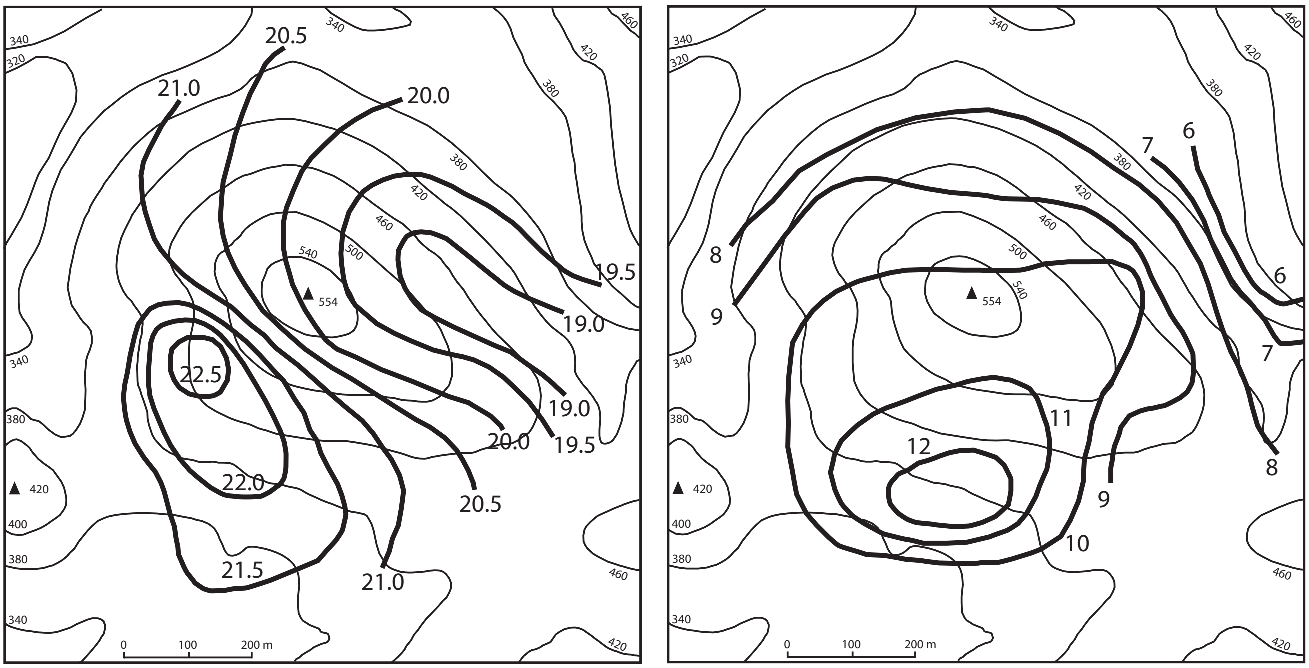
Precipitation onto any surface is determined by regional synoptic conditions, the slope and aspect of the surface, and the nature of the raindrops themselves. Changes in the trajectory of raindrops from the vertical results from wind effects, and small raindrops with lower terminal velocities are most affected. Locations with higher wind velocities have more perturbed precipitation patterns than protected sites. The intensity of precipitation at the surface depends on the angle of the wind-induced trajectories relative to the aspect and slope of the surface. A surface that is parallel to the trajectory of the raindrops, for example a wall during calm conditions, may receive no precipitation. Maximum intensities will be received on a surface that is perpendicular to the raindrop trajectory. Hence, in complex terrain and around architectural structures, significantly different precipitation amounts can occur on various slopes, and the differences can be accentuated or moderated over long periods. In winter, snow is subject to resuspension, transport and redistribution by wind. Snow accumulation zones do not necessarily coincide with the original pattern of the snowfall. Skillful design and positioning of snow fencing is a prime example of microclimate modification. The reduction of wind and turbulence around the barrier can enhance snow deposition and minimize the drifting across nearby transportation corridors.

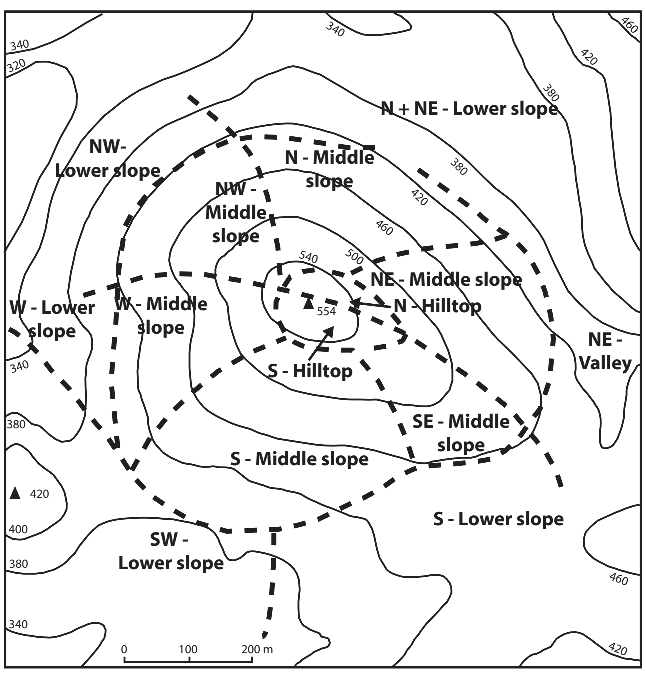
Examples of microclimates

Examples of microclimates are plentiful and contribute significantly to the makeup of the environments that surround us: the warming of one's face as it becomes aligned with the sun's beam on a cool morning; the seeking of shade by animals and people as a respite from the harsh afternoon sun; the distinct mosaic in surface temperatures experienced during summer outings to the beach or seaside; the creative use of sunlight and shade areas for ornamental plants; the cradling of snow by conifer branches in forests; and the fresh-fallen snow in open clearings that is swept into drift patterns that are repetitive from storm-to-storm and from year-to-year. Recent advancements in the study of microclimatology have sought not only to observe the features of microclimates, but also to investigate and explain the forcings and surface responses that give rise to the distinctiveness that is found over short distances.

With high solar energy, the microclimate energy forcing in tropical forest ecosystems (Sluiter and Smit, 2001) is considerable (Figure M8a). In such environments most of the solar radiation is absorbed by the upper levels of the forest canopy, and energy availability near the forest floor is very much reduced. In adjacent forest clearings (gaps), available energy is quite abundant in the near-surface layer. Diurnal temperature and windspeed observations document the distinct differences between these forest and gap environments (Figure M8b,c,d). Additionally, the size and nature of the gaps themselves play determinate roles in the microclimates created. An abundance of solar radiation and the availability of surface moisture provide microclimate resources for forest regrowth along the gap edges and within the gaps themselves. Ecological changes in response to microclimate resource availability will not only occur, but will in turn play an active role in modifying the surface microclimates.

The spatial pattern of surface thermal and moisture regimes is usually quite complex, and the transition across the boundary between different surfaces can display sharp discontinuities. Soil temperature and moisture transects across an open clear-cut in a midlatitude forest document both of these aspects (Figure M9). Variation within the forest and clear-cut surface regimes is considerable. Generally, soil temperatures are higher in the clear-cut than the forest, and differences are enhanced during cloudless conditions. The moisture of the uppermost surface materials is lower in the clear-cut due to the strong drying regimes present when compared to the shadier forest floor environment. Transitions across the borders of the clear-cut





Microclimate Summary:

Location	Solar Radiation	Temperature	Humidity
	[categories 1 (highest) to 5 (lowest)]		
Hilltop			
- South	1	3	4
- North	3	3	4
Middle Slope			
- South	1	1	1
- Southeast	1	2	2
- West	2	1	2
- Northwest	4	3	3
- Northeast	4	4	3
- North	5	5	4
Lower Slope			
- South	1	2	3
- Southwest	1	2	3
- West	3	2	3
- Northwest	5	4	4
- North + Northeast	5	5	5
Valley			
- Northeast	5	5	5

Figure M7 Topographic control of maximum and minimum near-surface air temperature for forested terrain in Germany. Isotherms are based on observations made during a series of sunny days in the month of June. Solar radiation, temperature and humidity regimes for each topographic zone are categorized from 1 (highest values) to 5 (lowest values). Adapted after Geiger (1961).

illustrate the role of south-facing and north-facing orientations in radiant energy receipt, with the latter demonstrating a sharper gradient of change.

Caves provide a means for both the assessment and interpretation of paleoclimates. Cave sediments have yielded significant archeological finds, and artistic renderings on the walls of caves provide profound insights into early humans and their activities. In a cave with a single opening (Figure M10), the cave entrance area can be sunlit and the air relatively hot. Upon entry, passageway narrowing diminishes the effectiveness of turbulent mixing, and the temperature cools and the air becomes moister. Further from the entrance, the temperature

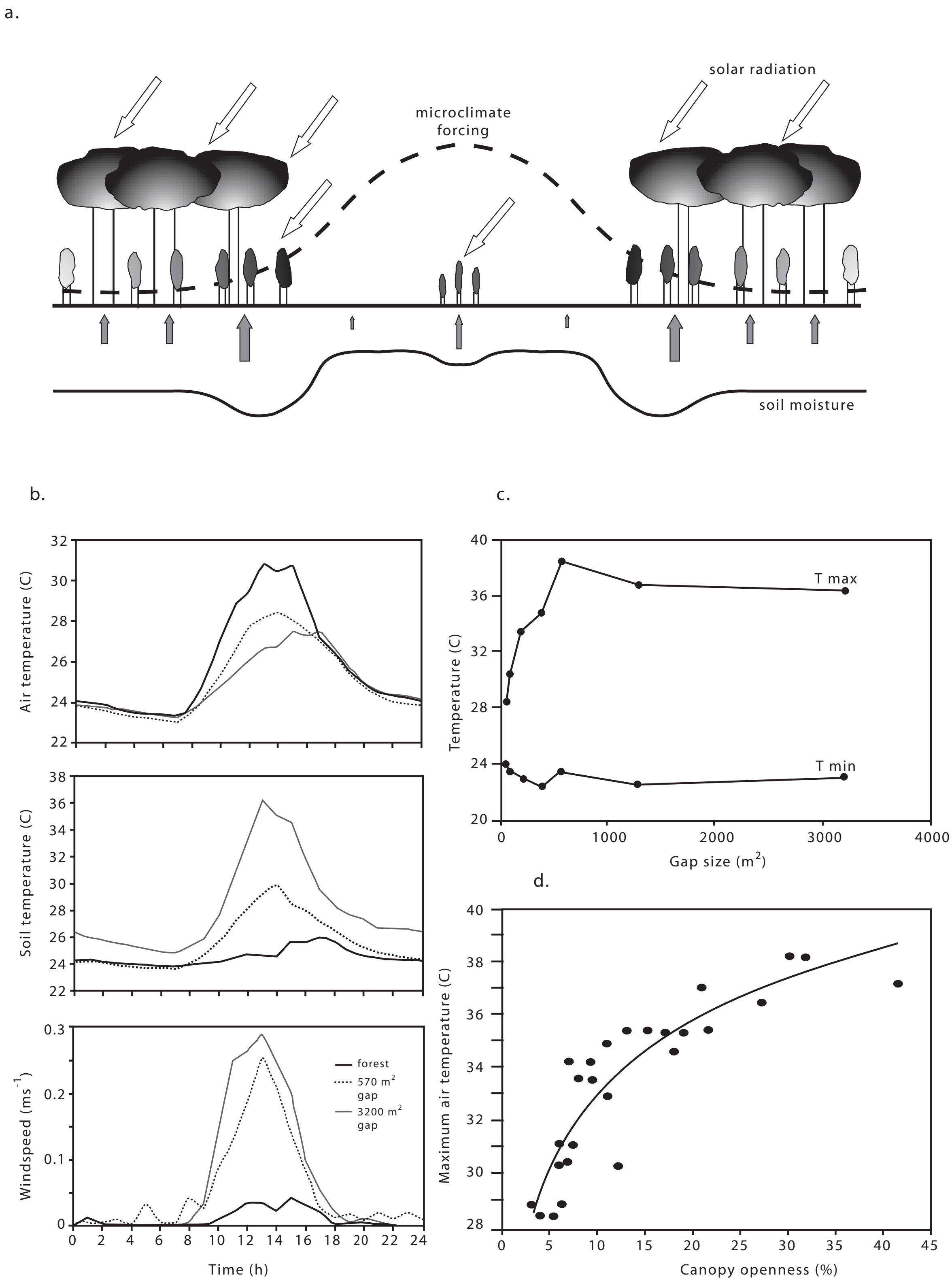


Figure M8 (**a**) A schematic representation of near-surface microclimate forcing in a tropical forest and clearing environment. The availability of soil moisture for the rooting zone is also indicated. (**b**) Diurnal air temperature, soil temperature and windspeed observations within a forest and in the center of two gaps with different areas in Guyana. (**c**) The influence of gap size on maximum and minimum air temperature. (**d**) The relationship between maximum air temperature and canopy openness for tropical forests. Adapted after Sluiter and Smit (2001).

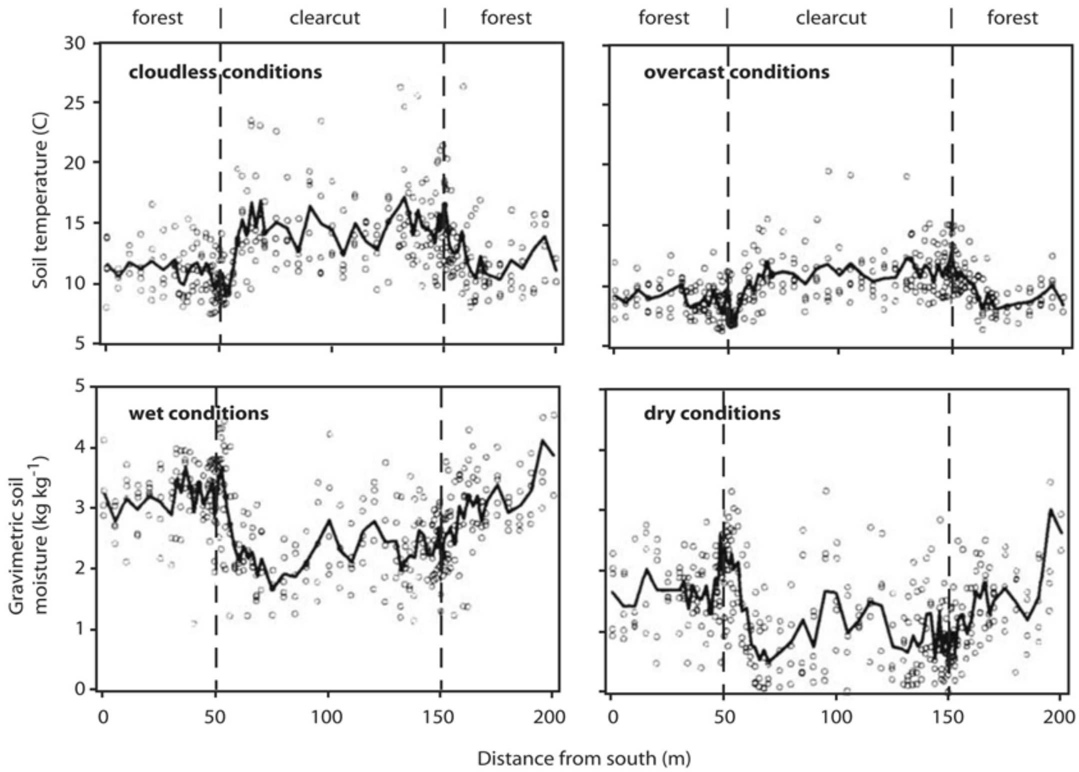


Figure M9 Transects of soil temperature and surface moisture across a 1 hectare clear-cut opening in a subalpine fir and spruce forest in British Columbia, Canada. Both average transect trends and individual site observations are presented. For soil temperature (20 mm below the surface), a comparison between cloudless and cloudy conditions is also displayed. For surface layer moisture a comparison between wet and dry conditions is exhibited. Adapted after Redding et al. (2003).

falls to near-constant values and the humidity approaches saturation. For caves with multiple openings, air circulation may develop on diurnal or seasonal timescales. Such unique crypto-microclimates have long been used by humans for shelter and security. Knowledge of such subterranean environments also has importance in understanding animal habitats, the characteristics of past and present human subsurface or hillslope dwellings, and the nature of the thermal and humidity regimes in below-surface storage facilities.

The survival and functioning of living organisms is dependent on an interrelationship between an organism and its surrounding environment. Organisms whose temperature is almost totally dictated by the surrounding environment are known as poikilotherms (so called “cold-blooded” animals such as fish, reptiles and insects). The body temperatures of poikilotherms exhibit a strong tracking and correlation with environmental temperature. Organisms that maintain near-constant deep-body temperatures are known as homeotherms (so called “warm-blooded animals” such as birds and mammals, including humans). Homeotherms exhibit near-constant body core temperatures over wide environmental temperature ranges. All animals respond to their microclimate, and heat production by animals can be separated into two distinct components: minimal heat production and regulatory heat production (Figure M11). With favorable environmental temperatures an animal will be in a thermoneutral zone and only minimal heat production will be required. Conditions will be near optimal, but perhaps with slight coolness or warmth. As environmental temperatures deviate from this zone, heat or cold stress dictates increases in regulatory heat production. If environmental conditions further deteriorate, and regulatory heat production is unable to stabilize the thermal regime of the organism, threshold levels can be exceeded. Mortality can then result. For some organisms daily existence and survival is dependent on such microclimate circumstances (Blumberg, 2002). Aspects of animal care, husbandry and transport (Albright, 1990; Johnson, 1987; Scientific Committee on Animal Health and Animal Welfare, 1999) are but a few of the practical acknowledgements of these principles.

Human activities that reshape the landscape give rise to many microclimate regimes. The thermal and moisture conditions in a cultivated field vary markedly from furrow crest to bottom. In many ways these are analogous to features found in larger landscapes (i.e. the topographic control exhibited in Figure M7). The excavation and landscaping of slopes or embankments with orientations that favor higher solar radiation receipt usually realize greater warmth and dryness (Brown and Gillespie, 1995). Humans have created built environments for shelter, security and a host of other reasons. The walls and roofs of buildings are extreme examples of the effects of orientation on solar radiation receipt. In the northern hemisphere, east walls receive direct beam solar radiation in the morning and west walls in the afternoon. Radiant energy receipt on north walls is minimal at most times of the

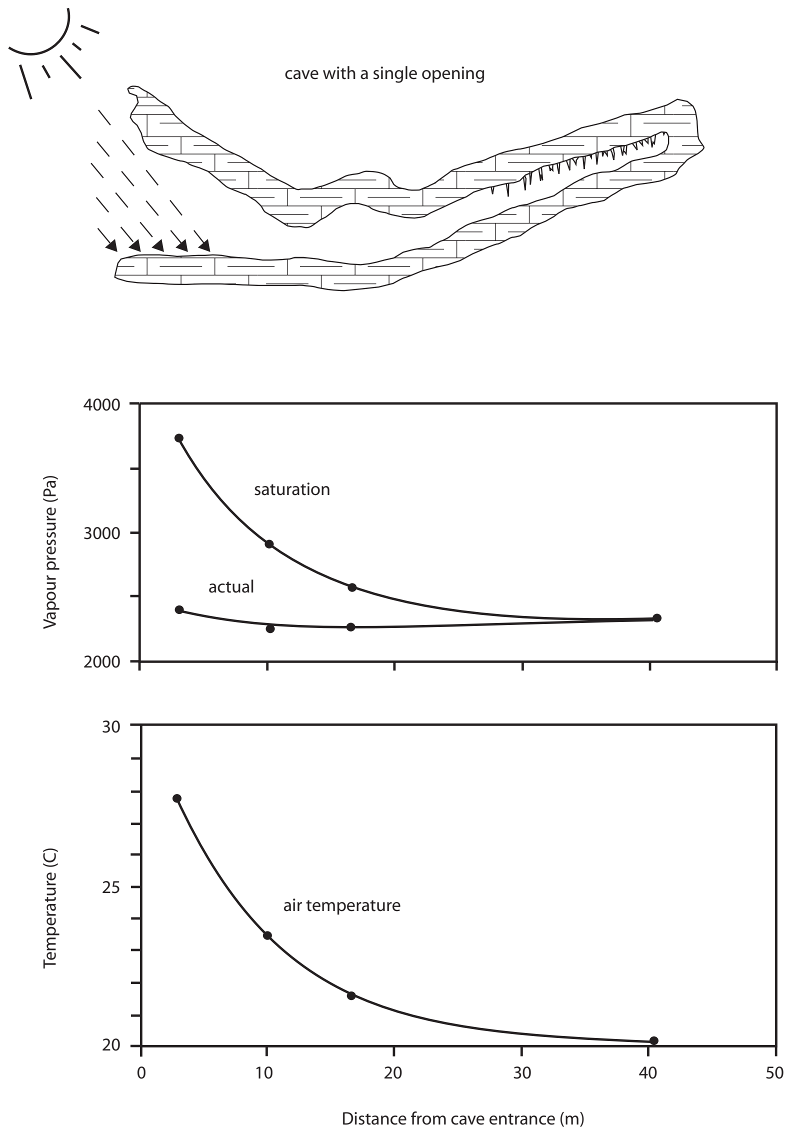


Figure M10 The temperature and humidity regime of a cave with a single opening in the eastern Mediterranean region. Adapted after Geiger (1961).

year and south walls may receive higher solar radiation during the late fall, winter and early spring. Microclimates within buildings may be linked to outside conditions (Brown and DeKay, 2001), or may have temperature, humidity and ventilation cycles that are artificially controlled. Concerns about indoor air quality and "sick buildings" are well documented, and provide challenges for architects and engineers. Humans themselves play distinct roles in the modification of the microclimates within buildings. The inhalation of oxygen and the exhalation of carbon dioxide and water vapor modify the atmospheric environment of rooms (Figure M12). The design of the best microclimate environment possible for buildings and cities, as well as a means for allowing people to modify the environment as needed, presents numerous and ongoing challenges.

The microclimatology of cultural heritage represents a unique blend of seemingly independent disciplines (Camuffo, 1998). However, knowledge of macroclimate, mesoclimate and microclimate regimes is an important factor in the location and architectural design of museums and other facilities that house cultural heritage. Within museums the monitoring and control of microclimate variables (radiation, temperature, humidity and air quality) is fundamental to the preservation of the collections as well as for the comfort of visitors (Thompson, 1990). Outdoor statues and monuments face additional complications as they are continually subject to microclimate influences, for example diurnal and seasonal heating and cooling, and dry and wet deposition arising from pollution. Pisa's leaning tower has been the focus of tourism, public awareness, and scientific and engineering study for many hundreds of years. In addition to

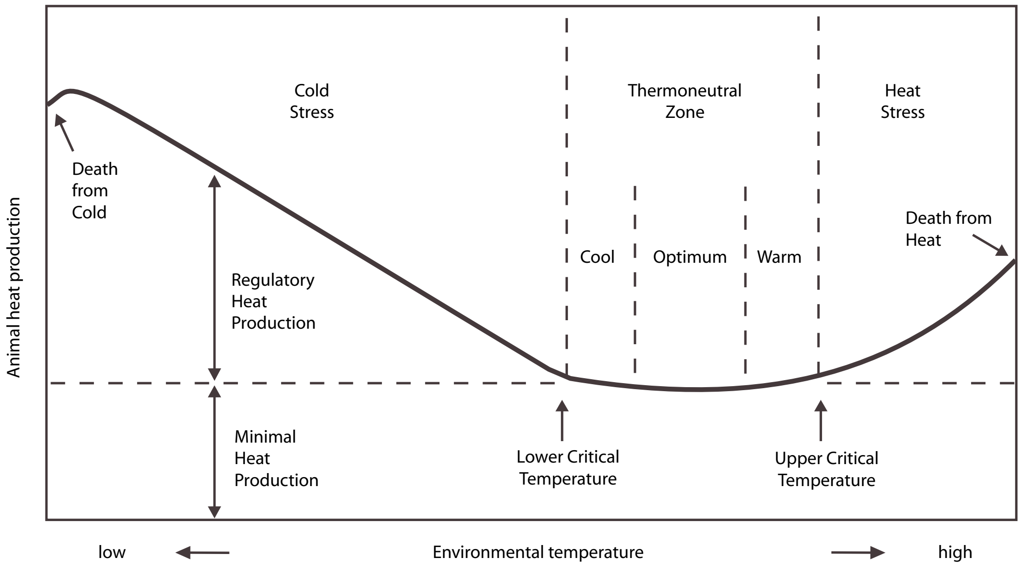


Figure M11 Schematic representation of heat production and the thermal zones of animals. Temperatures falling below the lower critical temperature results in cold stress (hypothermia), whereas temperatures rising above the upper critical temperature results in heat stress (hyperthermia). When critical thresholds are exceeded, mortality can result. Adapted after Scientific Committee on Animal Health and Animal Welfare (1999).

the tower's well-documented ballet with gravity, microclimate consequences are relentless (Camuffo et al., 1999) (Figure M13). The outer wall temperature regimes of La Tower Pendente di Pisa are closely linked to diurnal and seasonal trends in solar radiation. Throughout the day, the warmest wall temperatures move from the east-facing, to the south-facing, and finally to the west-facing walls. Expansion and contraction of the tower, and the associated diurnal twisting that is associated with this thermal warming of the stone, tracks the apparent motion of the sun in the sky. In summer, the warmed stone surfaces can exceed air temperatures by more than 10°C, and the north side of the tower is often colder than the surrounding air. The tower's splendor is continually disfigured by airborne particulate matter deposition, and a complex balance exists between deposition, the tilt and material of the tower, rainfall washout and surface runoff. Micropores on the stone surface play significant nanoclimate and picoclimate roles. The smallest micropores are always filled with water and constitute a reservoir for microbiological life, whereas condensation occurs only occasionally in the largest pores. Micropores between 4 and 5 nanometers undergo more than 250 condensation–evaporation cycles per year, and are influential in the dissolution of the material matrix of the stone and the migration of dissolved salts. Particles that are deposited on the tower can be washed away by rainfall. The washing is governed by the wind direction and the shape of the tower. Patterns in the surface washing demonstrate the direction of storms, the shielding provided by the tilt of the tower and the tower height. Rainfall washing is more effective at the top of the tower. Here airflow is less perturbed by the roughness of the surrounding urban landscape, and the trajectory of the raindrops has a greater horizontal component.

The Great Sphinx of Giza in Egypt is a globally recognized monument. Over past millennia it has been partially submerged in sand and subject to serious erosion about the neck and shoulder region. Excavation around the Sphinx has resulted in an advantageous microclimate modification as a result of altered airflow patterns (Figure M14). Wind that now sweeps across the Sphinx creates vortices on each flank of the monument, and the trajectory of abrasive granules has been altered with most falling into the excavation. As a consequence, one of the many threats to the preservation of the monument has been reduced (Camuffo, 1992, 1998).

New frontiers in the study of microclimatology are continuing to evolve. The study of the microclimatology of spacecraft, orbiting space stations and the surfaces of other planets has already commenced. In the latter, the integration of information gained from pioneer space probes, theoretical understanding and computer modeling holds the potential to yield remarkable advancements. The identification of severe microclimates on Earth, and their use as training areas for future planetary landings, is in its infancy. This attests to bold frontiers ahead for humans and their technology in the microclimates that will be visited far from Earth.

The challenges ahead

The impetus behind the advancements made in microclimatology knowledge during the twentieth century will continue into the twenty-first century. Knowledge about many environments exists, but only scant details are available for others. Demanding questions will continue to be posed, and the quest for insight and answers will persist. Global climate change is driven by processes that often commence with microscale

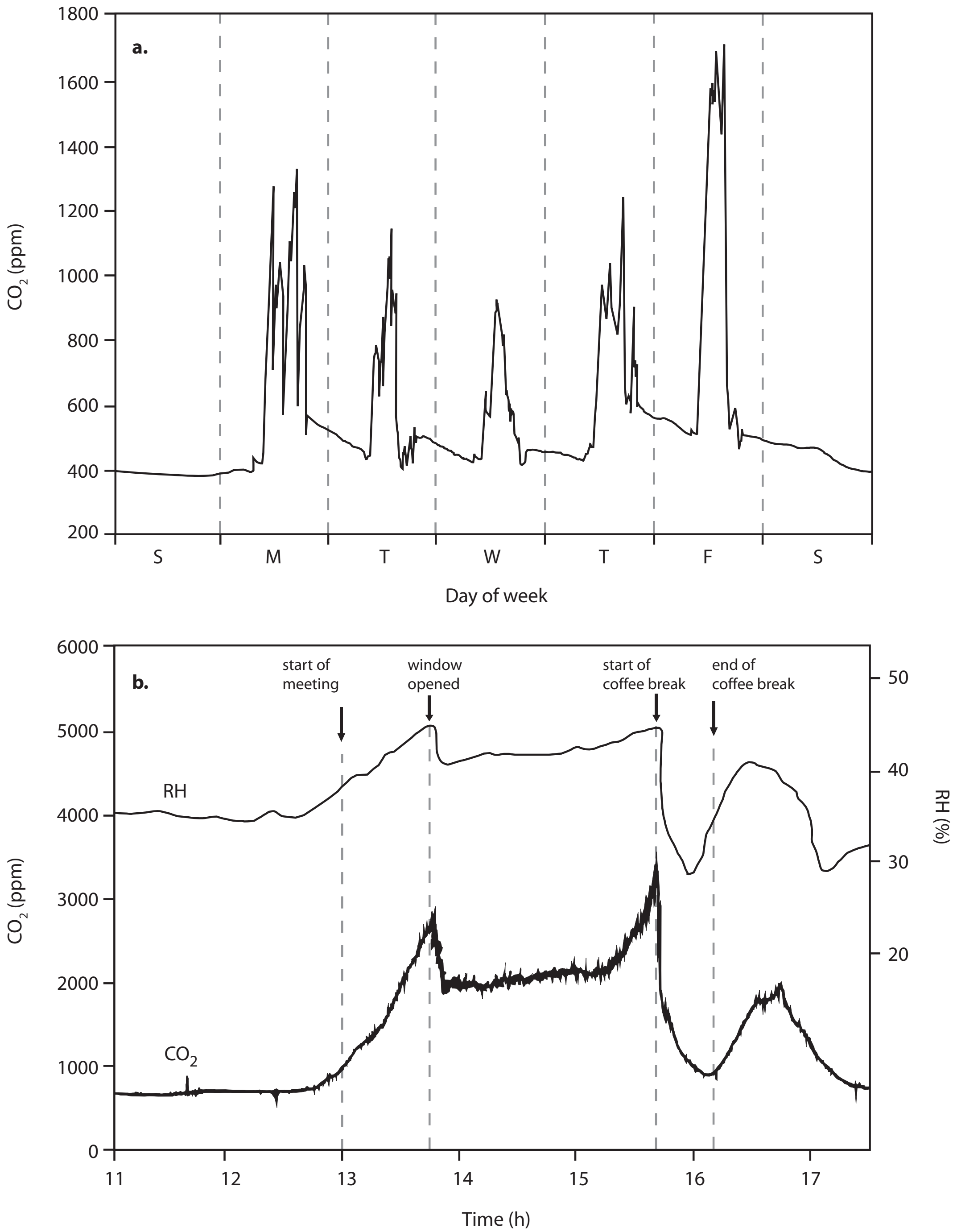


Figure M12 Variations in carbon dioxide in indoor environments. As CO_2 levels exceed 1000 ppm, tiredness and difficulties with concentration arise. Threshold values of CO_2 are approximately 5000 ppm. (**a**) The time-series of CO_2 in an office environment (maximum of three persons present) throughout a weekly period. Decreases in CO_2 occur during the daytime in response to windows being opened to allow the entry of fresh air. (**b**) The time-series of CO_2 and relative humidity during a meeting. Decreases in CO_2 and humidity occur due to window opening and the coffee break activities. Adapted after Endres (2001).

activities (IPCC, 2001c). Adequate representation and the scaling-up of important microclimate influences in numerical modeling approaches, for local and regional weather forecasting to climate change prediction and scenarios, will be an important contribution to the advancement of overall climate science (IPCC, 2001b). Forcings that are imbedded in natural and anthropogenic climate change will impact all microclimates. The response to these changes, in many environments,

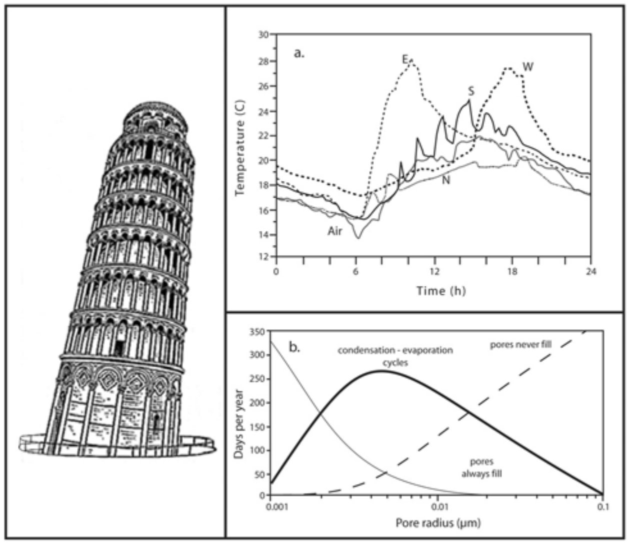


Figure M13 (**a**) The diurnal trend of surface temperatures on the outside wall surface of the central body of La Torre Pendente di Pisa (built from 1173 to 1370). Temperatures are given for positions facing the four cardinal directions. Air temperature was observed on the north side of the tower. (**b**) The influence of micropore size on the number of days each year in which condensation–evaporation cycles occur and the number of days that occur in which micropores are filled or never fill. Adapted after Camuffo et al. (1999).

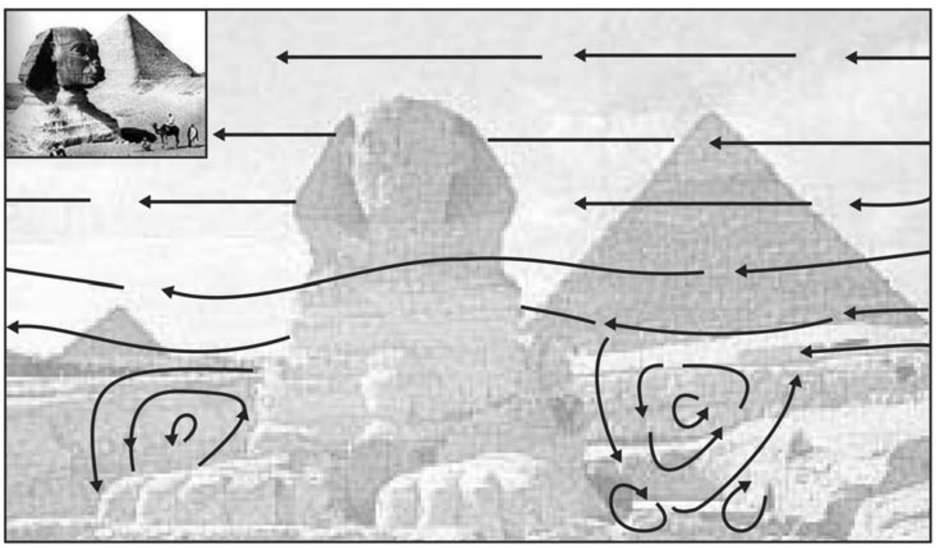


Figure M14 A schematic representation of airflow patterns across the Great Sphinx of Giza. Over the past millennia (insert), the Sphinx was often partially submerged in sand and subject to severe erosion in the neck and chest region. Excavation has resulted in the creation of a major vortex along each flank of the monument. The resultant trajectory of abrasive granules is such that most now fall into the excavation.

will have profound impacts on the organisms that inhabit them, perhaps their very survival (IPCC, 2001a).

Microclimatology is, and will remain, an inexact science. The profound challenges ahead will be the utilization of microclimate resources in the wisest and most prudent manner, and the minimization of deleterious modifications to microclimate regimes.

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Bibliography

Albright, L.D., 1990. *Environment Control for Animals and Plants*. St Joseph, MO: American Society of Agricultural Engineers.

American Meteorological Society, 2000. *Glossary of Meteorology*. Boston, MA: American Meteorological Society.

Bache, D.H., and Johnstone, D.R., 1992. *Microclimate and Spray Dispersion*. New York: Ellis Harwood.

Bailey, W.G., Oke, T.R., and Rouse, W.R., 1997. *The Surface Climates of Canada*. Montreal: McGill–Queen's University Press.

Bell, J.N.B., and Treshow, M., 2002. *Air Pollution and Plant Life*. New York: John Wiley.
Blumberg, M.S., 2002. Body Heat: temperature and life on Earth. Cambridge, MA: Harvard University Press.
Bonan, G.B., 2002. *Ecological Climatology: concepts and applications*. Cambridge: Cambridge University Press.
Brown, G.Z., and DeKay, M., 2001. *Sun, Wind and Light: architectural design strategies*. New York: Wiley.
Brown, N., 2001. *History and Climate Change*. London: Routledge.
Brown, R.D., and Gillespie, T.J., 1995. *Microclimate Landscape Design*. New York: John Wiley.
Butzer, K., 1971. *Environment and Archeology*. Chicago, IL: Aldine Atherton.
Calvin, W.H., 1991. *The Ascent of Mind: Ice Age climates and the evolution of intelligence*. New York: Bantam Books.
Calvin, W.H., 2002. *A Brain for All Seasons: human evolution and abrupt climate change*. Chicago, IL: University of Chicago Press.
Campbell, G.S., and Norman, J.M., 1998. *An Introduction to Environmental Biophysics*. Heidelberg: Springer-Verlag.
Camuffo, D., 1992. Controlling the aeolian erosion of the Great Sphinx. *Studies in Conservation*, **38**: 198–205.
Camuffo, D., 1998. *Microclimate for Cultural Heritage*. Amsterdam: Elsevier.
Camuffo, D., Sturaro, G., and Valentino, A., 1999. Urban climatology applied to the deterioration of the Pisa Leaning Tower, Italy. *Theoretical and Applied Climatology*, **63**: 223–231.
Endres, H.E., 2001. Air quality measurement and management. In Gassmaan, O., and Meixner, H., eds., *Sensors in Intelligent Buildings*. Weinheim: Wiley-VCH, pp. 85–102.
Geiger, R., 1961. *Das Klima der bodennahen Luftschicht*. Braunschweig: Vieweg.
Geiger, R., 1966. *The Climate Near the Ground*. Cambridge, MA: Harvard University Press.
Geiger, R., Aron, R.H., and Todhunter. P., 1995. *The Climate Near the Ground*. Braunschweig: Vieweg.
IPCC, 2001a. *Climate Change 2001: impacts, adaptation, and vulnerability*. Contribution of Working Group II to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge: Cambridge University Press.
IPCC, 2001b. *Climate Change 2001: synthesis report*. A Contribution of Working Groups I, II, and III to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge: Cambridge University Press.
IPCC, 2001c. *Climate Change 2001: the scientific basis*. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change. Cambridge: Cambridge University Press.
Johnson, H.D., 1987. *Bioclimatology and the Adaptation of Livestock*. Amsterdam: Elsevier.
Jones, H.G., 1992. *Plants and Microclimate*. Cambridge: Cambridge University Press.
Jones, P.D., Ogilvie, A.E.J., Davis, T.D., and Briffa, K.R., 2001. *History and Climate: memories of the future*. New York: Kluwer.
Klinenberg, E., 2002. *Heat Wave: a social autopsy of disaster in Chicago*. Chicago, IL: University of Chicago Press.
Kramer, H.J., 2002. *Observation of the Earth and its Environment. Survey of Missions and Sensors*. Berlin: Springer.
Lee, R., 1978. *Forest Microclimatology*. New York: Columbia University Press.
Lowman, M.D., and Nadkarni, N.M., 1995. *Forest Canopies*. San Diego, CA: Academic Press.
Lowry, W.P., 1988. *Atmospheric Ecology for Designers and Planners*. McMinnville, OR: Peavine.
Lowry, W.P., and Lowry, P.P., 1989. *Fundamentals of Biometeorology*, vol. I: *Interactions of Organisms and the Atmosphere*. McMinnville, OR: Peavine.
Lowry, W.P., and Lowry, P.P., 2001. *Fundamentals of Biometeorology*, vol. II: *The Biological Environment*. St. Louis, MO: Peavine.
Monteith, J.L., and Unsworth, M.H., 1990. *Principles of Environmental Physics*. London Edward Arnold.
Mynemi, R.B., and Ross, J., 1991. *Photon–Vegetation Interactions*. Berlin: Springer-Verlag.
Oke, T.R., 1987. *Boundary Layer Climates*. London: Routledge.
Oke, T.R., 1997. Surface climates processes. In Bailey, W.G., Oke, T.R., and Rouse, W.R., eds., *The Surface Climates of Canada*. Montreal: McGill–Queen's University Press, 21–43.
Oliver, J.E., and Hidore, J.J., 2002. *Climatology: an atmospheric science*. Upper Saddle River, NJ: Prentice-Hall.
Redding, T.E., Hope, G.D., Fortin, M.J., Schmidt, M.G., and Bailey, W.G., 2003. Spatial patterns of soil temperature and moisture across subalpine forest–clearcut edges in the southern interior of British Columbia. *Canadian Journal of Soil Science*, **83**: 121–130.
Rose, C.W., 1966. *Agricultural Physics*. New York: Pergamon Press.
Rosenberg, N.J., Blad, B.L., and Verma, S.B., 1983. *Microclimate: the biological environment*. New York: Wiley-Interscience.
Scientific Committee on Animal Health and Animal Welfare, 1999. *Standards for the Microclimate Inside Animal Transport Road Vehicles*. European Commission, Health and Consumer Protection Directorate-General, Directorate B, Scientific Health Opinions.
Sluiter, R., and Smit, N., 2001. Gap size effects on microclimate and soil moisture. In van Dam, O., ed., *Forest Filled with Gaps: effects of gap size on water and nutrient cycling in tropical rain forest*. Tropenbos-Guyana Series 10. Amsterdam: PrintPartners Ipskamp, pp. 49–66.
Stoutjesdijk, Ph., and Barkman, J.J., 1992. *Microclimate: vegetation and fauna*. Knivsta: Opulus.
Thompson, G., 1990. *The Museum Environment*. London: Butterworth–Heinemann.
Unwin, D.M., and Corbet, S.A., 1991. *Insects, Plants and Microclimate*. Slough: Richmond Publishers.
Whiteman, C.D., 2000. *Mountain Meteorology*. Oxford: Oxford University Press.

Cross-references

Albedo and Reflectivity
Bioclimatology
Climatology
Local Climatology
Scales of Climate
Vegetation and Climate

MIDDLE LATITUDE CLIMATES

The middle latitudes are regions of great atmospheric restlessness and variability, dominated at the surface and in the upper atmosphere by westerly winds (Hare, 1960). The climate is controlled at the surface by a succession of cyclones and anticyclones, normally moving from west to east, that are steered by the upper flow. The fall of temperature toward the poles occurs not in a uniform manner, but with the strong thermal gradients concentrated in one or more narrow latitudinal bands, or fronts. Baroclinic conditions lead to the development of jet streams just below the tropopause (Barry and Carleton, 2001).

The equatorward limit of middle latitude climates is often taken as the surface subtropical high-pressure belt. The poleward limit is more diffuse and variable, although it is often marked by the subpolar lows. The latitudinal extent of the climatic zone will vary from month to month and year to year, depending on changes in the position of the bordering centers of action, but most frequently it occupies the zone between 35 and 56 degrees N and S. As the overall equator-to-pole temperature contrast decreases from winter to summer, so does the strength of the westerlies. Day-to-day weather at a particular location is much affected by the latitude and strength of the westerly current. The upper westerlies vary between two extreme states. At times the upper flow is little deformed by waves, and the flow is parallel to the lines of latitude. At such times of high zonal index the westerlies blow strongly over large longitudinal zones and sweep a succession of depressions eastward at high speeds with frequent rain and gales in middle latitudes. At other times the Long or Rossby waves in the upper westerlies become greatly amplified, resulting in a meridional flow, a condition described as low zonal index. If the tips of the waves become cut off, leaving cold upper pools with associated low pressure at the surface in low latitudes and warm upper highs and associated surface anticyclones at high latitudes, blocking develops.

Blocks have preferred seasonal and geographical incidences, being most common in spring and in the eastern Atlantic and over eastern Asia. Having once formed, blocks frequently are very persistent and may dominate the circulation for several weeks and on occasion for a whole season, introducing large temperature and precipitation anomalies into middle latitudes. Evidence exists of annual variations in blocking frequency and persistent anomalous climatic conditions, e.g. the mid-1970s European drought could have been caused by persistent or repetitive blocking episodes.

The strength of the circulation in any season is often expressed in relation to oscillations in the strength and position of the main centers of action, e.g. the Icelandic Low, and the Azores High. Such pressure oscillations include the North Atlantic Oscillation (NAO) and North Pacific Oscillation (NPO). In recent years the NAO, in particular, has often been strongly positive, implying strong westerlies in the Atlantic-European sector and a succession of mild winters in much of Europe (Perry, 2000).

Great differences occur in the geographical extent of land and sea areas in the two hemispheres. In the northern hemisphere midlatitudes contain the large landmasses of North America and Eurasia that lead to prominent contrasts in regional climate with important longitudinal variations. West coast locations are exposed to the predominant westerly winds and have equable temperature regimes, often with abundant precipitation throughout the year. The interiors of the continents have a more continental climate with greater annual temperature ranges and lower average precipitation totals, most of which falls in the summer season as a result of convective processes. In the southern hemisphere the much greater extent of sea area ensures that the westerlies blow more consistently and strongly (being known by a variety of names, including the Roaring Forties), and the circulation is less disturbed by periods of blocking than in the northern hemisphere. The kinetic energy of the southern westerlies is about 60% larger than the northern hemisphere westerlies. Smith (1967) has suggested that continentality and oceanicity are the principal criteria in defining subdivisions of climate in middle latitudes with general temperature levels forming a secondary subdivision. The adjective “temperate” is often applied to midlatitudes or used as a proxy term (e.g. Trewartha, 1968) but is certainly not suited to all climates in these latitudes as Bailey (1964) has shown.

Marine midlatitude climates

Along the western margins of the continents in the middle latitudes of both hemispheres, the climate is under a considerable maritime influence. In Western Europe the topography allows penetration of maritime airmasses deep into the continent in contrast to the situation in both North America and South America. A distinction is normally made between warm temperate or Mediterranean and cool temperate climates.

Mediterranean type

The approximate latitudinal extent of this type is on the order of 5–10 degrees between latitudes 30/35–40. A large proportion of the total precipitation falls in the winter when day-to-day weather is controlled by the behavior of the westerlies (Perry, 1981). During the hot dry summers the subtropical high-pressure area and its attendant ridges take control. In the type area (the Mediterranean Basin), substantial winter precipitation totals occur during low zonal index phases, when a meandering jet stream with a major trough exists over the Mediterranean, favoring cyclogenesis. Lee depressions form in the Gulf of Genoa and south of the Atlas Mountains and move in a generally eastward direction through the basin, often becoming reinvigorated in the east near Cyprus. Whereas the sea acts as a heat sink in summer, in winter it represents a heat source and cold airmasses entering the basin quickly become unstable after passage over the relatively warm water. The high intensity of rainfall is reflected in the small number of rain-days, even in the wetter areas. Regional winds, e.g. the cold Mistral and Bora, are related to meteorological and topographic factors, whereas the persistent northerly winds of summer in the eastern Mediterranean known as the etesians give this area a distinct climatic subtype. High annual sunshine totals are a feature of Mediterranean climates, with totals exceeding 3000 hours being quite common. In the northern hemisphere Southern California has this type of climate, as does the southern hemisphere coast of Chile around Santiago, the West Australian coast around Perth, and the South African coast around Cape Town.

Cool temperate

Poleward of the Mediterranean climate, on the west side of the continents, the climate is changeable throughout the year with well-distributed precipitation, brought by a series of depressions and their associated fronts and a predominance of maritime air masses. In northwestern Europe and on the Oregon and Washington coasts of North America, large positive temperature anomalies occur in winter compared with the average for the latitude, as maritime airmasses cross the warm-water currents in the North Atlantic and North Pacific. Since summers are relatively cool for the latitude, annual temperature ranges of less than 20°C are common. Because depressions are deeper and more vigorous in winter near the coast, there is normally a fall or winter precipitation maximum, whereas spring and early summer have a minimum of precipitation, reflecting both the increased frequency of blocking anticyclones in these seasons and the lower moisture content of maritime airmasses at this period due to the lower sea temperatures. High precipitation totals are a feature of all mountain areas and large lapse rates give such areas a short growing season and a cloudy, damp and often raw climate. A high degree of changeability from day to day is a characteristic of these climates as rapid alternations of airmasses occur, although occasionally persistence of a particular synoptic situation leads to more settled conditions. The coldest winter weather and the warmest summer spells develop when continental airmasses replace, for a time, the more usual airflow from oceanic sources. In the southern hemisphere sizeable belts of this climatic type occur in Chile, Tasmania, and the South Island of New Zealand.

Continental midlatitude climates

In both North America and Scandinavia the transition from maritime to continental climates is rapid due to the barrier effect of the mountains imposed on the invasion of surface maritime airmasses, but across the European Plain the transition is much more gradual. Meridional airmass movement is a particular characteristic of the North American climate and rapid changes of temperature level can occur, especially in winter, as frontal depressions cross the continent. Very severe winters occur at times with disrupting snowstorms, and are the result of strong amplification of the long waves with ridging over western North America and a deep trough in the east. This allows deployment of Arctic airmasses into the eastern states, while often in the west weather is mild and dry. Blizzard conditions develop on the cold polar side of traveling lows. In summer severe heatwaves can develop when warm air from the Gulf of Mexico is advected northward into the central and eastern states.

Over the USSR the winter circulation is dominated by the intense Siberian anticyclone, although the position and the intensity of this cold anticyclone can cause considerable departures of temperature values from the normal in individual years. Mean January temperatures below −40°C occur in parts of Siberia, accompanied by dry sunny weather. During the short summers, temperatures can rise to 35°C on occasion, even as far north as the Arctic Circle. In these continental climates the transition seasons of spring and fall are very short. Although the seasonal variation of precipitation in the interior and eastern parts of both North America and Asia can be complex, there is normally a summer maximum brought on by instability showers and thunderstorms. Local variations in precipitation totals reflect such factors as the presence of large lakes, e.g. the Great Lakes of North America enhance snowfall totals on their eastern shores. In the United States the 51 cm annual isohyet follows approximately the 100° West Meridian, and in the dry area between the Rocky Mountains and this longitude occasional drought years occur, such as those resulting in the Dust Bowl conditions of the 1930s.

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Bibliography

Bailey, H.R., 1964. Towards a unified concept of the temperate climate. *Geographical Review*, **54**: 516–545.
Barry, R.G., and Chorley, R.J., 1998. *Atmosphere, Weather and Climate*, 7th edn. London: Methuen.
Barry, R.G., and Carleton, A.M., 2001. *Synoptic and Dynamic Climatology*. London: Routledge.
Hare, F.K., 1960. The westerlies. *Geographical Review*, **50**: 345–367.
Perry, A.H., 1981. Mediterranean climate: a synoptic reappraisal. *Progress in Physical Geography*, **5**: 107–113.
Perry, A.H., 2000. The North Atlantic Oscillation: an enigmatic see-saw. *Progress in Physical Geography*, **24**: 289–294.
Smith, D., 1967. Middle latitude climates. In Fairbridge, R.W., ed., *The Encyclopedia of Atmospheric Sciences and Astrogeology*. New York: Reinhold, pp. 604–605.
Trewartha, G.T., 1968. *An Introduction to Climate*, 4th edn., New York: McGraw-Hill, pp. 305–339.

Cross-references

Airmass Climatology
Climate Classification
Climatology
Continental Climate and Continentality
Europe, Climate of
Jet Streams
Mediterranean Climate
Maritime Climate
North America, Climate of

MILANKOVITCH, MILUTIN (1879–1958)

Milutin Milankovitch developed the theory of orbital control of terrestrial insolation and climate change. Milankovitch was of Serbian origin, born in Dalj (Slavonia) in what is now Croatia. He obtained his PhD in 1904 in Vienna and worked as a civil engineer for 5 years before going to the University of Belgrade as Professor of Applied Mathematics. Captured during World War I, but allowed to work at the Hungarian Academy of Sciences, Milankovitch completed the insolation theory for the Earth and had also worked out a climate history for Venus and Mars.

The "Milankovitch parameters" involved in his theory are: (1) the eccentricity (or ellipticity) of the orbit *e*, which measures the departure of the Earth's circumsolar orbit from a circle (with a period of about 90 000–100 000 years); (2) the obliquity of the ecliptic (or tilt), which is the angle between the equator and the plane of the orbit (principal period about 41 000 years); and (3) the precessional parameter, related to the longitude of the perihelion, which is conveniently expressed as the angular distance of the spring equinox point from the perihelion. It has two principal terms, about 19 000 and 23 000 years. This is related to the "general" *precession* that has been refined to 25 694 years (Berger, 1992).

After World War I, Wladimir Köppen, the famous climatologist, was working on a book with his son-in-law Alfred Wegener (of continental drift fame). This became a standard textbook on climate (1924) and carried the Milankovitch message. Unfortunately it only reached a German-speaking audience. However, a German geologist and archeologist from Breslau, Frederick Zeuner, applied the insolation theory to the central European geological record, and after moving to London in the 1930s Zeuner explained to a world audience the Milankovitch ideas (Zeuner, 1959). The "establishment" of the day responded to the Milankovitch theory with severe criticism, an outrage almost paralleled by the hostile reception of Wegener's continental drift theory. During World War II Milankovitch worked on a complete revision of his radiation theory, which was published (in German) by the Royal Serbian Academy of Sciences in 1941 as the *Kanon der Erdbestrahlung.*

In the post-World War II decades, however, dramatic discoveries were being made at sea thanks to sediment coring and deep-sea drilling. The sea-level record paralleled the last phase of the Milankovitch radiation curve for 65°N. During the next two decades, dating systems were expanded, deep-sea cores were obtained worldwide, and the results were conclusive: there was an exact match with the Milankovitch pattern (Fairbridge, 1967), and it was recognized as "the pacemaker of the ice ages" (Hays et al., 1976).

Milankovitch was destined to become one of those rare scientists who developed truly pivotal ideas. His inspirational conversion from engineering to orbital dynamics is elegantly told by Imbrie and Imbrie (1979), and may be followed with details and bibliography in an autobiographical work (1957). In 1979, to mark the 100th anniversary of the birth of Milankovitch, a symposium was organized in Belgrade by the Serbian Academy of Sciences, and a conference convened at the Lamont–Doherty Geological Observatory (Palisades, NY). The evidence was overwhelming (see symposium volume, edited by Berger et al., 1984).

The acceptance of the Milankovitch theory was far more than a quantification of dynamic change for the last ice age. Geologists are now applying it to the whole of Earth history. For climatologists, long-term modeling is now feasible, and for meteorologists it carries a crucial message: the terrestrial climate machine is neither chaotic nor unpredictable – it is forced by extraterrestrial agencies. The same message must apply also to other planets.

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Bibliography

Berger, A.L. (ed.), 1981. *Climatic Variations and Variability: facts and theories.* Dordrecht: Reidel.

Berger, A.L., 1992. Astronomical theory of paleoclimates and the last glacial–interglacial cycle. *Quaternary Science Review*, **11**: 571–81.

Berger, A.L., and Loutre, M.F., 1991. Insolation values for the climate of the last 10 million years. *Quaternary Science Review*, **10**: 297–317.

Berger, A., Imbrie, J., Hays, J. et al. (eds), 1984. *Milankovitch and Climate.* Dordrecht: Reidel.

Berger, A., Loutre, M.F., and Laskar, J., 1992. Stability of the astronomical frequencies over the Earth's history for paleoclimate studies. *Science*, **255**: 560–566.

Broecker, W.S., 1968. Milankovitch hypothesis supported by precise dating of coral reefs and deep sea sediments. *Science*, **159**: 297–300.

Emiliani, C., 1966. Isotope paleotemperatures. *Science*, **154**: 851–857.

Fairbridge, R.W., 1960. The changing level of the sea. *Scientific American*, **292**(5): 70–79.

Fairbridge, R.W., 1961. Eustatic changes in sea level. In Ahrens, L.H., et al., eds., *Physics and Chemistry of the Earth*, vol. 4. London: Pergamon Press, pp. 99–185.

Fairbridge, R.W., 1967. Ice-age theory. In Fairbridge, R.W., ed., *The Encyclopedia of Atmospheric Sciences and Astrogeology*. New York: Reinhold, pp. 462–467.

Hays, J.D., Imbrie, J., and Shackleton, N.J., 1976. Variations in the Earth's orbit: pacemaker of the ice ages. *Science*, **194**: 1221–1332.

Imbrie, J., and Imbrie, K.P., 1979. *Ice Ages: solving the mystery.* Short Hills, NJ: Enslow.

Köppen, W., and Wegener, A., 1924. *Die Klimate der Geologischen Vorzeit.* Berlin: Gebr. Borntraeger.

Kukla, G., and Gavin, J., 1992. Insolation regime of the warm to cold transition. In Kukla, G., and We, E., eds., *Start of a Glacial*. NATO ASI Series I, 3, pp. 307–339.

Kutzbach, J.E., 1985. Modeling of paleoclimates. *Advances in Geophysics*, **28A**: 159–196.

Liu, H.-S., 1995. A new view on the driving mechanisms of Milankovitch glaciation cycles. *Earth and Planetary Science Letters*, **131**: 17–26.

Milankovitch, M., 1941. Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem. Royal Serbian Academy, Special Publication 133. (English translation 1969 as *Canon of Insolation and the Ice-Age Problem*, by Israel Progr. Sci. Transl. Washington, DC: US Department of Commerce.)

Milankovitch, M., 1957. Astronomische Theorie der Klimaschwan-kungen: ihr Werdegang und Widerhall. *Serbian Academy of Sciences Monography*, **280**: 1–58 (incl. comprehensive bibliography).

Sanders, J.E., 1995. Astronomical forcing functions: from Hutton to Milankovitch and beyond. *Northeastern Geology and Environmental Sciences*, **17**(3): 306–347.

Zeuner, F.E., 1959. *The Pleistocene Period.* London: Hutchinson.

Cross-references

Climatic Variation: Historical Record
Paleoclimatology

MILITARY AFFAIRS AND CLIMATE

Today, the military is called upon for a variety of operations ranging from war fighting to humanitarian or peacekeeping missions (military operations other than war, MOOTW) to peacetime operations (for example training and research) similar to any other industry. Post-September 11 the military has also been asked to increase its role in domestic security issues and anti-terrorism special operations. All of these complex missions require timely weather and climate information. For the military, climate and weather are not distinct areas of study; rather, they flow along the entire knowledge continuum embracing strategy and tactics. Knowledge of the climatic conditions of an area is necessary in developing the strategic plans for operating in that area. Training as well as development, test, and evaluation of equipment also requires detailed climatic knowledge. On timescales up to a week, knowledge of the weather is critical in developing the tactics used during a mission. As demonstrated in Winters (1998), failure to effectively exploit this climatic knowledge, or a chance encounter with weather conditions representative of extreme departures from anticipated conditions, have resulted in disastrous defeats throughout military history. This essay will highlight many of the climatic elements that impact on the military as they perform their various missions. It will conclude with a brief look at "climates" of space and the oceans which also impact military operations. Readers interested in other areas of military geography should consult the numerous references in Collins (1998) and Plaka and Galgano (2000).

Impacts of climatic elements

A climatic element is some component of the climate system such as temperature or precipitation. Information on these elements is measured using conventional weather equipment. The (near) instantaneous values of these components is termed weather. The synthesis of these values, including not only average values but also their variability, is termed climate. Each has critical military significance. This section will relate the major climatic elements to some of their military impacts.

Temperature is the most obvious climatic element. Humans are homeotherms; they attempt to maintain a core body temperature within narrow limits. Extreme hot or cold temperatures impact the body thermoregulation system and degrade performance. In hot environments, armies will consume more fluids, and troops must be monitored to prevent heat illness. Extreme cold increases the logistical need for clothing and food to ensure good health. Equipment as well is impacted by extreme temperatures. Hot ambient conditions reduce the efficiency of engine radiators and may lead to overheating. Lubricants become less efficient at extreme temperatures causing increased wear or damage. In very cold conditions machinery may freeze if not continuously tended. In the free atmosphere temperature inversions may cause areas of anomalous propagation of electromagnetic radiation. The resulting degraded radar performance may prevent the acquisition or tracking of targets or threats. Jet engine efficiency also is a function of air temperature. Low-flying cruise missiles may have their effective range shortened in hot environments.

Humidity also impacts human comfort and performance. High humidity combined with high temperatures may produce dangerous heat stress conditions. At low humidity, drying effects can influence not only human comfort but fluid evaporation rates. A particular concern is the build-up in static charge which can make handling of fuels and munitions dangerous. Temperature and humidity determine air density, which impacts projectile performance. In tropical environments, high humidity contributes to the biofouling and deterioration of equipment. It is often necessary to artificially heat confined spaces to discourage the growth of mold or provide rust-resistant coatings to metal surfaces.

Precipitation has inspired many a soldier's lament. Heavy rains often result in trafficability problems as roads become impassable due to mud or debris flows. Floods are also a problem. Flash floods may imperil troops or equipment while riverine floods may cause delays in bridging operations. Snow or heavy rain restricts visibility and limits aircraft operations. The in-flight build-up of ice can affect aircraft control surfaces. Sensors or communication systems operating at microwave frequencies are particularly susceptible to interference from heavy precipitation.

Wind impacts include lowering of visibility due to blowing dust or sand. High winds may preclude the use of smoke or other obscurants to hide troop movements or impact the dispersal of chemical or biological weapons. Radar-absorbing chaff clouds, used as an electronic warfare countermeasure, may be quickly dispersed in high winds, thus reducing their effectiveness. High winds may also impact aircraft operations. Helicopter maneuvering in gusty winds is often problematic and launch and recovery of unmanned aerial drones is complicated by unsteady winds. Cross-winds and turbulence can impact aircraft providing low-level combat air support.

Many impacts on visibility have been listed under other climatic elements. Fog, blowing dust, and the smoke created by battle significantly impact low-level visibility. For aircraft, cloud layers may hinder target identification and weapons delivery. Many of these problems have technological solutions beyond the scope of this item. One historically important aspect of visibility has been all but eliminated by technology: night versus day. Night vision equipment and associated sensors which allow pilots and ground personnel to operate in near-total darkness have resulted in a radical change in battlefield tactics. War fighting is no longer a "sunrise to sunset" affair.

A final climatic element of military concern is atmospheric pressure. Aircraft must have oxygen systems for prolonged high-altitude flight. Troops operating in mountainous terrain need time to acclimatize to the lower pressures to avoid rapid fatigue, or in extreme cases altitude sickness. Even the often-vilified army cook must be aware of the impact of high elevation on cooking times and the boiling point of water!

Impacts of ocean and space "climates"

Most of this item has been directed toward the ground and lower troposphere. However, naval and marine units encounter another set of impacts peculiar to the ocean and the land/water interface. High winds and wave heights can make naval operations and shipkeeping extremely difficult to the point of structural damage. Personnel operating in and under the water must be protected from the effects of hypothermia. Amphibious landings must contend with tides, currents, and breaking waves. Submarine warfare is especially dependent on knowledge of sound-velocity profiles which are largely determined by seawater temperature. Fog and structural icing are also concerns.

With the development of space-based satellite observation and communication systems, the realm of "climate effects" has been extended to the exosphere. Periods of intense solar activity produce storms of charged particles which impact satellite operations. Multi- and hyper-spectral sensors are used to minimize the effects of clouds and water vapor on the satellite geometry.

Military weather and climate monitoring

The military maintains weather-monitoring stations at many of its facilities around the world. Part of their mission is to provide forecasting support to operational units. The US Air Force and Navy maintain detachments at the National Climatic Data Center. Here climatic data collected from weather stations around the world are used to produce climatic summaries for areas of interest to the military. During times of crisis these detachments can produce specific mission-directed products. Operational units are typically deployed with personnel trained in analyzing these weather and climate products.

Lessons learned from historical engagements are combined with advances in technology at a series of national laboratories operated by the Army, Air Force, and Navy. Research results and products from these laboratories are tested and evaluated at a series of field laboratories before being approved for use by operational forces. The testing must be done under similar climatic conditions to those in which the products will used; therefore field laboratories are located in tropical, temperate, and Arctic locations. In addition, troops must be trained to operate in a variety of climatic conditions prior to their deployment. The goal is to exercise personnel and machinery prior to employment to maximize effectiveness and minimize loss. Climate is one consideration in this important process.

Conclusion

Weather and climate have played an important role in historical war fighting. The role continues today along with increased responsibilities for MOOTW and peacetime operations. Climate impacts all these functions. From training to development of new weapon systems to operational deployment climate considerations abound. Strategic planners and logisticians make use of climate data in deciding what mix of forces will be used in a particular area and how they will be equipped and supplied. Operational commanders can prepare tactics based on climatic knowledge and be ready to adapt those tactics to take advantage of changing weather conditions. Climatologists work with engineers and other technicians to provide new equipment and climatological summaries relevant to the evolving missions of today's military.

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Bibliography

Collins, J., 1998. *Military Geography for Professionals and the Public*. Dulles, VA: Brassey's.

Plaka, E., and Galgano, F., 2000. *The Scope of Military Geography: across the spectrum from peacetime to war*. New York: McGraw-Hill.

Winters, H., 1998. *Battling the Elements: weather and terrain in the conduct of war*. Baltimore: Johns Hopkins University Press.

Cross-references

Climate Comfort Indices
Cultural Climatology
Human Health and Climate

MODELS, CLIMATIC

Climatic models are abstractions of climate that are used for one or more of three overlapping purposes: to explore and describe the interrelationships within and between the observed elements of climate, to predict important climatic or climate-forced events, or to explain the fundamental biophysical mechanisms that control climate. A widely accepted, comprehensive definition of climate and, subsequently, of climatic models has not been published although many climatologists would agree that climate is a synthesis of weather in which the time-period of integration often exceeds the length of a weather forecast. Climatic models, in fact, have been developed and used to investigate and describe climatic processes which are manifested over tens, hundreds and even thousands of years of Earth history as well as over but a few days. Equally variable is the geographic scale on which climate occurs and is modeled (Steyn et al., 1981).

Microclimatic models, for instance, may attempt to characterize climatic processes that function within only a few cubic meters, whereas macroclimatic models often strive to describe climate as it extends over hundreds or thousands of square kilometers and vertically from the surface to the bottom of the stratosphere and perhaps beyond (Schneider and Dickinson, 1974). From an anthropocentric vantage point, however – regardless of scale – all climatic processes and phenomena are manifested within the context of a four-dimensional (the three space dimensions plus time), circum-surface climatic environment (Figure M15). Models of climate, it follows, are chiefly formulated for and used within this context, even though application-specific requirements result in a wide variety of model forms.

A common purpose

A number of climatologists would qualify the above characterization by emphasizing that climatic models are distinct from meteorological models only when they attempt to couple terrestrial systems or oceanic processes to the lower atmosphere. Since the lower atmosphere and Earth are coupled across their contact surface, the energy and moisture crossing that interface control the equilibrium of atmosphere–Earth energy and mass exchange and, thus, have primary climatological importance. Climate at the surface also is primarily what humans experience, influencing a wide range of human activities. Climatic models, as a consequence, are mainly developed and used to explore, describe, predict or explain the integrated surface or primary interface (Figure M15) temperature and moisture states of the world's oceans, lakes, rivers, snowfields, soil, flora and fauna, as well as of people and their edifices.

As the integrated temperature and moisture states of any terrestrial or aquatic surface(s) of interest, that is climate, can only be evaluated by accounting for the fluxes of energy and mass (predominantly moisture) to and from the surface (i.e. by solving the surface "energy [and mass] budget"), the search for improved solutions to the budget – for the myriad of disparate space and time scales as well as environments – provides the *raison d'etre* for much of the development, refinement and use of climatic models. A primary reason for climatic modeling, in other words, is the establishment of good, quantitative solutions to all or parts of

$$Q(1-\alpha) + I{\downarrow} - I{\uparrow} = H + LE + G + F \qquad (1)$$

where Q is solar irradiance ($\mathrm{W\,m^{-2}}$), α is the integrated reflectivity of the surface to solar (shortwave) irradiance, $I{\downarrow}$ is longwave radiation from the surrounding environment and the atmosphere which is absorbed by the surface ($\mathrm{W\,m^{-2}}$), $I{\uparrow}$ is longwave radiation emitted by the surface ($\mathrm{W\,m^{-2}}$), H is the rate of sensible heat exchange between the surface and lower atmosphere ($\mathrm{W\,m^{-2}}$), L is the latent heat of vaporization ($\mathrm{J\,kg^{-1}}$), E is the rate of water vapor exchange between the surface and lower atmosphere ($\mathrm{kg\,m^{-2}\,s^{-1}}$), G is the rate of heat exchange by conduction between the surface and underlying ground or water ($\mathrm{W\,m^{-2}}$) and F is the heat convected by the ocean, lake or river from its surface ($\mathrm{W\,m^{-2}}$). Consistent with convention $[Q(1-\alpha)]$ and $I{\downarrow}$ are positive fluxes that contribute to surface warming whereas $I{\uparrow}$ is a positive, cooling term. On the other hand, H, LE, G and F are positive when the direction of transfer is away from the surface (cooling) and negative when they contribute to surface warming.

In order to solve the surface energy budget (equation 1), climate models must account for the ambient or representative energy and moisture states of the near-surface air, soil, snow, vegetation and so on, as well as for the states of the surface, because the magnitudes of most of the fluxes depend upon temperature, humidity and momentum gradients that exist between the surface, the atmosphere above and the terrestrial or aquatic environ below. That is, climatic models must provide supportive, quantitative characterizations of the near-surface state variables, such as air temperature, vapor pressure and soil moisture, before equation (1) can be adequately solved. Concern for the ambient states of the near-surface, in turn, leads to the related problem of evaluating those convective and radiative fluxes that take place across the climatic environment's secondary interfaces (Figure M15), for they – like the exchanges across the atmosphere–Earth interface – affect the states of the near-surface which then influence the surface energy budget. Thus, equation (1) is fundamentally nonlinear, in that the surface energy and moisture states are simultaneously required by the equation as input.

Evaluating the energy, mass and momentum fluxes into and out of the climatic environment entails modeling atmospheric dynamics and physics at the global scale. Over the last three decades, work on the development of global-scale climate models has increased dramatically, and become the main emphasis within the climate-modeling community. Our discussion, in turn, focuses on the history and status of global climate models. Although modern versions of these global climate models are both three-dimensional and time varying, important contributions to the history of global climate modeling and climatology have also been made with simpler models.

Global energy balance models (EBMs)

Global energy-balance models (EBMs) use the basic radiation laws and conservation of energy as their physical principles.

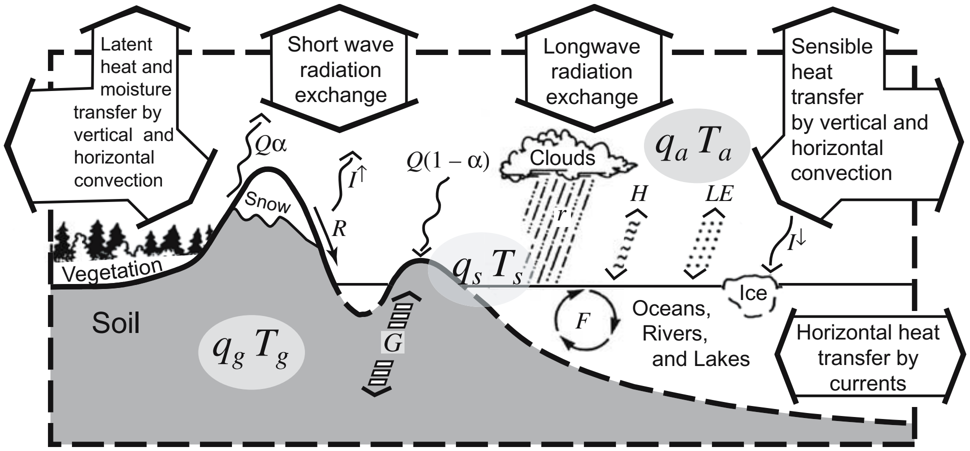

Figure M15 Schematic representation of the climatic environment represented in models. Primary fluxes occur to and from the surface (solid line), which also is referred to as the primary or atmosphere–Earth interface, and they are labeled with symbols for energy budget components (defined in text). Large symbols in oval represent temperature (*T*) and moisture (*q*) states of the atmosphere (q_a T_a), surface (q_s T_s), and near-surface ground (q_g T_g) respectively. Precipitation and total runoff are noted as *r* and *R*, respectively. Horizontal and vertical heat, moisture and momentum exchanges between the climatic environment and external environments are conceptually depicted by phrases within arrows crossing the secondary interfaces (dashed lines).

The balance implied is between income and outgo. On climatic timescales, whatever sunlight is intercepted by the Earth must be either reflected back to space or absorbed and reradiated as infrared radiation. A prototype for a zero-dimensional EBM is used in nearly every introductory weather and climate textbook in which the shortwave and longwave energy balances are presented for the Earth as a whole. In the textbook version, solar input is distributed between surface and atmospheric energy balances, with terms for latent heat, sensible heat, greenhouse gas warming, and albedo. All that is needed to make that into a climate model is to turn some of the ratios between energy flows into variables instead of constants, translate the percentage values into actual energy flux units, and constrain the model with the three energy balances it requires (surface, atmosphere, and space). Such a simple model can be used to provide heuristic calculations of the relative importance of the major climate forcing components, such as infrared radiation trapping (greenhouse effect) or changes in the vertical structure of the atmosphere.

Simple global EBMs reduce equation (1) to

$$Q[1-\alpha(T,f)] = I(T,f) \tag{2}$$

in which Q is the average solar flux at the top of the atmosphere, T is temperature, f is cloud-cover fraction, and α and I are functions representing the variation of planetary albedo and infrared output to space, respectively, with temperature and cloud-cover fraction. Simple linear relationships for α and I – derived from satellite data (Bintanja, 1996; Graves et al., 1993) – allow equation (2) to be inverted to solve for temperature.

In the most famous use of a zero-dimensional climate model, Arrhenius (1896) explored the role of carbon dioxide in the Earth's radiation balance. As early as the end of the nineteenth century, Arrhenius noted that carbon dioxide levels were rising from human use of fossil fuels, and he projected that a doubling of carbon dioxide levels in the atmosphere could produce a 1.5–3°C warming in the average surface temperature of the Earth. A century later, Arrhenius's calculation is still within the mainstream of climate model projections – neither alarmist nor contrarian.

Global energy-balance models have been usefully extended to higher dimensions, most notably to one-dimensional models. These zonal EBMs add a latitudinal component to the energy balance. In addition to the vertical energy balance implied by equation (2), a meridional transport term must be added, as in

$$\begin{aligned} &QS(\varphi)[1-\alpha(T(\varphi),f(\varphi),\varphi)] - I(T(\varphi),f(\varphi)) \\ &\quad = F(T(\varphi),f(\varphi),\varphi) \end{aligned} \tag{3}$$

in which φ is latitude, S is a distribution function for solar energy, and F is a function representing the net meridional flux of energy, and all the other functions now vary with latitude. A direct dependence on latitude is included for the albedo and meridional flux functions, in addition to their indirect dependence on latitude via the latitude-dependence of temperature and cloud cover. The direct dependence of albedo on latitude allows for variation of vegetative cover and continental extent with latitude. The direct dependence of horizontal flux convergence on latitude reflects the fact that models of this simplicity do not adequately simulate the high meridional transport in the Hadley circulation without some numerical assistance.

Two particular versions of equation (3), published nearly simultaneously, hold a place of high importance in the history of climate models. Budyko (1969) and Sellers (1969) independently presented simple versions of (3) with slightly different formulations for F and a similar formulation for $\alpha(T)$ in which the albedo of the land surface jumped sharply if the zonal temperature went below -10°C, representing the onset of a snowline. Both found that a small decrease in solar input, amounting to just a few percent, would raise the planetary albedo sufficiently to cause runaway cooling. This runaway positive feedback raised considerable interest in this kind of modeling through the 1970s. Through that decade, considerable knowledge of the Earth's stability, sensitivity, and feedback systems was gained via exploration of zonal EBMs (North et al., 1981).

Schneider and Dickinson (1974) reviewed a wide range of modeling activities and concluded that a hierarchy of climate models ranging from zero-dimensional, steady-state models through fully three-dimensional general circulation models had a role to play in understanding climate. Within a short time after that, however, zonal EBMs had been found to be too sensitive to describe global sensitivity to solar radiation variations, too insensitive to produce a realistic ice-age cycle from orbital parameter variations, too parameterized to produce a useful greenhouse-gas-change simulation, and generally not competitive for cutting-edge research. Today, zonal EBMs remain an important heuristic tool: they are well worth using as exploratory tools in an introductory course, and building a simple model can illustrate a variety of climate principles and numerical modeling techniques for an advanced student. A few research problems make good use of simple models when coupled with ocean models (Egger, 1999) or glacier models (Sakai and Peltier, 1999) for long-term climate experiments. Beyond these specialized uses, EBMs have mostly faded from view in research.

General circulation models (GCMs)

General circulation models (GCMs) share with global EBMs the idea that physical principles drive the system, rather than atmospheric data. Beyond that commonality, GCMs are both genetically and morphologically different from EBMs. The core of the modern GCM is the atmospheric GCM (AGCM).

AGCMs grew out of weather-forecasting models. The early history of numerical weather prediction has been told elsewhere (Thompson, 1978). The first useful weather simulations were developed in the years immediately after World War II as a project led by John von Neumann at Princeton, the inventor of the stored-program computer. The size and complexity of weather and climate models has been closely tied to improvements in computing technology ever since (Hack, 1992, figure 9.1). The transition from weather model to climate model occurs when a model's statistical behavior approaches the statistical behavior of weather over a long enough period that we may consider climate statistics from the model output to be comparable to climate statistics from weather data. Phillips (1956) ran a baroclinic model long enough to produce a climate simulation in what is often called the first general circulation model.

Six partial differential equations define an atmospheric general circulation model: three for conservation of momentum (one for each dimension of physical space), two for conservation of mass (one each for air and for water), and one for conservation of energy. The six independent field variables in this system are the three components of atmospheric velocity, temperature, specific humidity, and pressure. Thus, nominally, we have a system that can be solved. In practice, the effort required to create such a model is huge.

Because of the complexity and scale of AGCM building, most state-of-the-art models bear institutional names, such as the NCAR CAM (the National Center for Atmospheric Research Community Atmospheric Model, Boulder, Colorado), the GFDL GCM (Geophysical Fluid Dynamics Laboratory, Princeton, New Jersey), or ECHAM model (European Center for Medium Range Weather Forecasting – Hamburg, Germany). Each of these has an evolving version history from over 30 years of continuous development involving hundreds of scientists and programmers. The most-developed models from large institutions coexist with a plethora of other models, but most share some common origins. What AGCMs have in common is the concept of simulating weather using a grid. The entire global atmosphere is broken up into boxes, both in the horizontal and vertical dimensions. Within each vertical column of boxes, all of the *subgrid-scale* processes must be evaluated. Between vertical columns of the model, most horizontal transfer processes proceed by advection, which requires a prediction of wind. For this prediction the equations of motion are solved approximately.

More difficult to understand, but currently more popular in use, are *spectral* methods (Bourke et al., 1977). In these methods, the fields that require horizontal derivatives are subjected to a generalized Fourier transform, so that they may be represented by a set of coefficients of the spherical harmonics. Some of the advantage of this method arises because the truncated Fourier series of spherical harmonics used to represent the horizontal fields is intrinsically smooth.

Grid scales have been decreased as computational power has increased. Up to a limiting point that we are far from approaching, decreasing the grid size so that fewer things need to be relegated to the subgrid-scale parameterizations will always be an improvement. However, halving the grid size requires an increase in computing power of more than a factor of 100, so grid sizes have not decreased at a rate that might be expected from a simple understanding of how fast computers have improved in recent decades. A common grid scale now would be roughly 3°latitude/longitude boxes, 20 layers in the vertical, and a 20-minute time step.

Climate system models (CSMs)

Climate system models (CSMs) are AGCMs within which boundary conditions at the land and ocean surfaces respond to the changing atmospheric conditions. In the earliest AGCM climate experiments a land model may have consisted of a simple parameterization of soil-moisture availability beneath each grid box and a specified vegetation albedo that could be raised if snowcover were predicted. A number of more realistic land-surface formulations have been developed since then, and they incorporate important vegetation and soil characteristics, as well as human-modified aspects of the land surface (Dickinson, 1984; Sellers et al., 1986). These land-surface models have been widely applied within CSMs, and they have improved significantly CSMs estimates of land-surface albedo, evapotranspiration and soil moisture. Dickinson (1984), in particular, raised the level of land modeling from a parameterization to a significant submodel with the Biosphere–Atmosphere Transfer Scheme (BATS). Continued development of BATS and similar models has led to land models which can be considered independent models in themselves, to be run either in conjunction with an AGCM or in standalone mode using atmospheric data (Bonan et al., 2002).

Just as land-surface boundary conditions have become more interactive and sophisticated, oceanic boundary conditions have matured. From "swamp" models in which the oceans were flat surfaces with unlimited thermodynamic capacity and moisture availability, we now recognize that for realistic long-term climate simulations, an oceanic general circulation model (OGCM) must be coupled to an AGCM. Such coupled experiments still have not been entirely successful in reconciling the energy and moisture fluxes between the atmospheric and oceanic components into a stable equilibrium. Usually, a flux correction field must be applied, a nonconservative difference between the energy leaving the top of the ocean and that entering the bottom of the atmosphere. Models have come close enough to equilibrium to avoid using the flux correction, but these still show long-term drift (Bryan, 1998).

Another boundary condition model required for coupling into a CSM is one for sea ice. Sea ice has an importance in climate systems well out of proportion to its mass, both because of its sharp albedo effect and because its role as an insulator with respect to both thermal and moisture fluxes between the oceans and the atmosphere. Although the thermal prediction of freezing or melting is fairly simple, and the simulation of the effects of sea ice on the atmosphere–ocean fluxes is also straightforward (Curry et al., 1995), sea ice moves in a complicated response to wind action, modulated by ocean currents. Errors in sea-ice prediction may be the largest identified, systematic problem with current CSM simulations.

Models of everything

Climate modelers have long dreamed of complete models of the climate system – models of "everything" which couple all of the related components of the climate system. The coupling of human activities to climate (Terjung, 1976), however – owing to its complexity, will continue to remain indirect for the foreseeable future. Nonetheless, all of the components of new CSM suites are available for simulating decadal to century-scale climatic variations. Specific improvements in nearly every aspect of atmospheric, oceanic, and land-surface models are being discussed widely and implemented. Even with dramatic improvements in these components, however, climate simulations will remain restricted to medium-term variations, perhaps up to the scale of a millennium, because the simulated climate system is still not complete.

The largest climatic changes of the last several million years have involved the advance and retreat of large ice sheets from North America and northern Eurasia, on time scales for which 1000 years is a high-speed fluctuation and 100 000 years appears to characterize a complete cycle. The pioneering ice-sheet model of Weertman (1976) demonstrated that such cycles could be controlled primarily by the lagged response within the ice sheet. The eventual model of everything will need to include an ice GCM for resolving problems on that scale. Ice-sheet models in current use are not quite ready for full coupling. The best continent-scale ice models are still map-plane models that have sub-grid scale weaknesses with important problems such as ice streams (Fastook and Prentice, 1994), whereas the fully three-dimensional ice sheet models that can simulate ice stream convergence cannot readily be scaled up to continental scales (Hanson, 1995; Blatter, 1995). Nevertheless, some interesting uncoupled simulations using ice-model-generated boundary conditions in an AGCM have been attempted (Bromwich et al., 2001). Full coupling of an ice-sheet model into a CSM is probably a few model generations away, but the path to that eventual coupling is at least being followed.

Another component that will eventually need to be dealt with is the vegetational response. When dealing with the last 20 000 years of climate, going from full glacial conditions to the present, we see significant vegetational responses. For example, the current Amazon rainforest was probably mostly savanna during the last ice age, with a few refugia of forests, from which the current rainforest was created by slow species migration as the climate changed. One of the first things specified into any paleoclimate simulation, immediately after the pattern of continents and oceans, is the vegetation types that will cover each grid box. No model currently presupposes the ability to modify those vegetation types in response to climate change, nor should models bother trying so long as a CSM is restricted to a relatively short simulation. We expect that such a vegetation model will be needed before the model of everything is truly complete, and we expect that building it will be a very difficult problem.

Outlook for the future

There is every reason to believe that the sophistication of our climate models will continue to increase over the next decades. Ongoing improvements in our computational resources will allow climatic models to incorporate increasingly realistic levels of detail at the same time that a burgeoning corps of climatologists is uncovering those essences of climate which will provide the bases for the models of the future. Vast and continuing increases in our stores of climatic and climate-related data – owing significantly to advances in remote sensing technology – will contribute many of the terrestrial and oceanic observations necessary to determine boundary and initial conditions for future models. With these advances in climatic theory and computational resources, as well as in the type and quantity of available data, climatologists will develop, refine and apply a hierarchy of CSMs. These CSMs will couple the oceans, vegetation, ice and sea ice to the atmosphere, and they will be used to describe, predict and explain climatic processes and phenomena over a wide range of geographic scales.

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Bibliography

Arrhenius, S., 1896. On the influence of carbonic acid in the air upon the temperature of the ground. *Philosophical Magazine, Series 5*, **41**(251): 237–276.

Bintanja, R., 1996. The parameterization of shortwave and longwave radiative fluxes for use in zonally averaged climate models. *Journal of Climate*, **9**: 439–454.

Blatter, H., 1995. Velocity and stress fields in grounded glaciers: a simple algorithm for including deviatoric stress gradients. *Journal of Glaciology*, **41**(138): 333–344.

Bonan, G.B., Oleson, K.W., Vertenstein, M., et al., 2002. The land surface climatology of the community land model coupled to the NCAR Community Climate Model. *Journal of Climate*, **15**(22): 3123–3149.

Bourke, W., McAvaney, B., Puri, K., and Thurling, R., 1977. Global modeling of atmospheric flow by spectral methods. In Chang, J., ed., *Methods in Computational Physics 17: General circulation models of the atmosphere*. New York: Academic Press, pp. 268–324.

Bromwich, D.H., Cassano, J.J., Klein, T., et al., 2001. Mesoscale modeling of katabatic winds over Greenland with the Polar MM5. *Monthly Weather Review*, **129**: 2290–2309.

Bryan, F., 1998. Climate drift in a multicentury integration of the NCAR Climate System Model. *Journal of Climate*, **4**(6): 1455–1471.

Budyko, M.I., 1969. The effect of solar radiation variations on the climate of the earth. *Tellus*, **21**: 611–619.

Crowley, T.J., and Baum, S.K., 1997. Effect of vegetation on an ice-age climate model simulation. *Journal of Geophysical Research*, **102**(D14): 16463–16480.
Curry, J.A., Schramm, J.L., and Ebert, E.E., 1995. Sea ice-albedo climate feedback mechanism. *Journal of Climate*, **8**: 240–247.
Dickinson, R.E., 1984. Modeling evapotranspiration for three-dimensional global climate models. In Hansen, J.E., and Takahashi, T., eds., *Climate Processes and Climate Sensitivity*. Washington, DC, American Geophysical Union Geophysical Monograph 29, pp. 58–72.
Egger, J., 1999. Internal fluctuations in an ocean–atmosphere box model with sea-ice. *Climate Dynamics*, **15**: 595–604.
Fastook, J.L., and Prentice, M., 1994. A finite-element model of Antarctica: sensitivity test for meteorological mass balance relationship. *Journal of Glaciology*, **40**(134): 167–175.
Graves, C.E., Lee, W.-H., and North, G.R., 1993. New parameterizations and sensitivities for simple climate models. *Journal of Geophysical Research*, **93**(D3): 5025–5036.
Hack, J.J., 1992. Climate system simulation. basic numerical and computational concepts. In Trenberth, K.E., ed., *Climate System Modeling*. Cambridge: Cambridge University Press, pp. 283–318.
Hanson, B., 1995. A fully three-dimensional finite-element model applied to velocities on Storglaciären, Sweden. *Journal of Glaciology*, **41**(137): 91–102.
North, G.R., Cahalan, R.F., and Coakley, J.A., Jr, 1981. Energy balance climate models. *Reviews of Geophysics and Space Physics*, **19**(1): 91–121.
Phillips, N., 1956. The general circulation of the atmosphere: a numerical experiment. *Quarterly Journal of the Royal Meteorological Society*, **82**: 123–164.
Sakai, K., and Peltier, W.R., 1999. A dynamical systems model of the Dansgaard-Oeschger oscillation and the origin of the Bond cycle. *Journal of Climate*, **12**(8): 2238–2255.
Schiller, A., Mikolajewicz, U., and Voss, R., 1997. The stability of the North Atlantic thermohaline circulation in a coupled ocean-atmosphere general circulation model. *Climate Dynamics*, **13**: 324–347.
Schneider, S.H., and Dickinson, R.E., 1974. Climate modeling. *Reviews of Geophysics and Space Physics*, **12**: 447–493.
Sellers, P.J., Mintz, Y., Sud, Y.C., and Dalcher, A., 1986. A simple biosphere model (SIB) for use within general-circulation models. *Journal of the Atmospheric Sciences*, **43**(6): 505–531.
Sellers, W.D., 1969. A global climate model based on the energy balance of the earth-atmosphere system. *Journal of Applied Meteorology*, **8**: 392–400.
Shukla. J., and Mintz, Y., 1982. Influence of land-surface evapotranspiration on the Earth's climate, *Science*, **215**: 1498–1501.
Steyn, D.G., Oke, T.R., Hay, J.E., and Knox, J.L., 1981. On scales in meteorology and climatology. *Climatological Bulletin*, **30**: 1–8.
Terjung, W.H., 1976. Climatology for geographers. *Annals of the Association of American Geographers*, **66**: 199–222.
Thompson, P.D., 1978. The mathematics of meteorology. In Steen, L.A., ed., *Mathematics Today: twelve informal essays*. New York: Springer-Verlag, pp. 127–152.
Weertman, J., 1976. Milankovitch solar radiation variations and ice age ice sheet sizes. *Nature*, **261**: 17–20.

Cross-references

Boundary Layer Climatology
Energy Budget Climatology
Water Budget Analysis
Statistical Climatology

MONSOONS AND MONSOON CLIMATE

Monsoon regions

The monsoon broadly refers to an atmospheric phenomenon in which the mean surface wind reverses its direction from summer to winter. However, the monsoon is popularly used to denote the rains without reference to the winds. The term "monsoon" has its origin from an Arabic word meaning season. In English, the original Arabic word is spelled in several ways, such as, "mausam", "mausem", "mausim", "mawsim" and "mausin". The term was used by seamen, several centuries ago, to describe southwesterly wind during summer, and northeasterly wind during winter over the Arabian sea. Along with progress in meteorological sciences, more regional circulations have been categorized as "monsoonal"; they are based on both wind and rainfall characteristics. Monsoonal regions over the globe are generally identified by certain characteristics of surface circulations in January and July as laid down by C.S. Ramage. These include a shift in wind direction by at least 120°, average frequency of prevailing wind directions exceeding 40%, mean wind strength in at least one of the months greater than 3 m/s and fewer than one cyclone–anticyclone alternation in either month in every 2 years. Although these monsoon criteria do not include the rainfall explicitly, the seasonality in rainfall is the most important manifestation of the monsoon circulation. Most of the countries lying between 35°N and 25°S and between 30°W and 170°E (Figures M16a,b) satisfy the criteria of monsoons and, hence, are widely accepted as a part of the single most important monsoon domain on Earth. The monsoon is primarily an Asian phenomenon. However, numerous studies have established that monsoon circulation occurs in some other parts of the world, such as western Africa, northern Australia and North America at varying intensities. In the monsoon regions, wind blows inland from the cooler oceans toward warm continents in summer and from cold continents toward the warm oceans in winter (Figures M16a,b). Thus, broadly speaking the summer monsoons of both the hemispheres are very wet (Figures M17a,b) and winter monsoons are dry. Comparison of seasonal wind directions in Figures M16a and M17b demonstrates that the monsoon conditions are best developed in east and south Asia, with winds from southwest in summer and from the northeast in winter. Accordingly, those are known as the southwest and northeast monsoons respectively. The Asian summer monsoon consists of the Indian monsoon, and the east Asian monsoon, both of which are responsible for abundant summer rainfall in the region shown in Figure M17a. The Indian monsoon is effectively separated from the east Asian monsoon by the massive Himalayan mountain range. About 80% of the annual rainfall over India occurs during the southwest monsoon from June to September. Consequently, the summer monsoon is very important for the economy of India, which is predominantly an agricultural country. Usually, in eastern Asia, the winter monsoon wind is stronger than the summer monsoon wind while the opposite happens in south Asia.

West Africa also experiences a wind reversal to some extent, from southwesterly in summer (Figure M16a) to northeasterly in winter (Figure M16b). In the west coast of Africa there is heavy rainfall from June to August, although the rains actually commence in March–April. In some parts of eastern Africa near the equator there are two rainy seasons, one during March to May, popularly know as the "long rains", and the other from October to November, which is termed as "short rains". These rainy periods fall between the two African monsoon circulations.

Over north Australia there is a northwesterly flow of humid maritime air in the southern hemisphere summer (Figure M16b) from December to February, and the southeast trades in the southern hemisphere winter (Figure M16a). The east Asian winter monsoon and the north Australian summer monsoon are

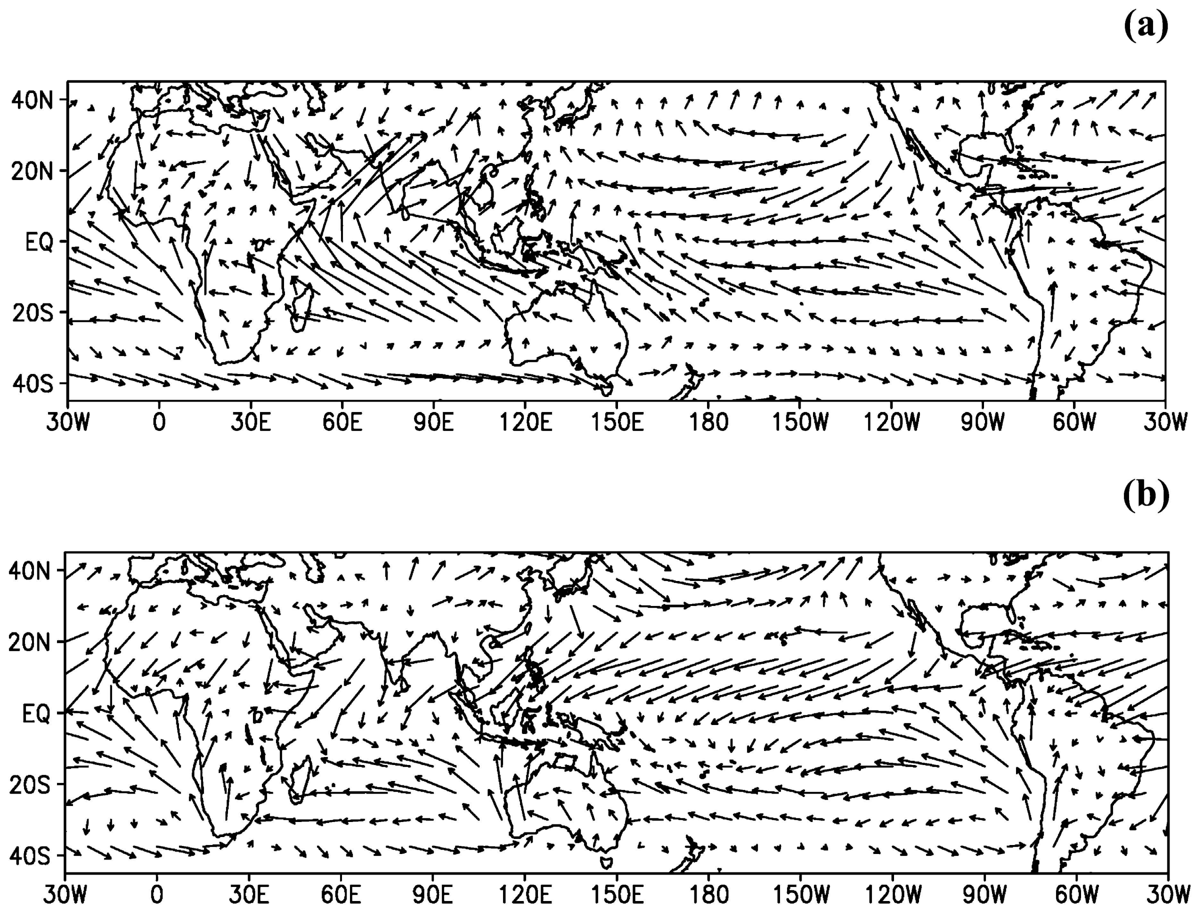

Figure M16 Surface wind (m/s) in the tropical belt (**a**) June, July and August average; (**b**) December, January and February average based on NCEP/NCAR reanalysis.

intermingled because the dry winter air of the northern hemisphere flows across the equator toward the southern hemisphere continents, picking up moisture from the warm tropical oceans to become the wet monsoon over north Australia.

The southwestern part of North America is also considered to be monsoonal. Here the surface zonal wind is observed to change from an easterly in January to westerly in July. There is a pronounced increase in rainfall over large areas of southwestern North America and southern Mexico from June to July, although the summer rains may last till September. This region also experiences significant winter rainfall.

Based on recent data, studies justify the existence of summer monsoon circulation over the subtropical south American highland. In southern hemisphere summer, distinct rainfall enhancement occurs over the central Andes and the southern parts and north coast of Brazil.

Important features of planetary scale monsoon

In 1686 Edmund Halley explained the Asiatic monsoon as resulting from thermal contrasts between the continent and oceans. In simple terms one may compare the monsoon circulation to a land–sea breeze occurring on larger spatial and temporal scales. In summer the continents surrounding the Arabian sea begin to receive large amount of heat, due both to solar insolation and the heat emitted from the Earth's surface; consequently, by the end of May, a trough of low pressure develops over a large area from Somalia to Pakistan and northwest India. This is usually known as the heat low. It is accompanied by subsidence and hence fine weather in general. The southerly wind, after crossing the equator, turns to the east under the influence of the Coriolis force and blows over the Indian subcontinent as a southwesterly monsoon wind (Figure M16a) in summer. This current appears to originate in the southeast trades. George Hadley in 1735 incorporated the concept of the Coriolis force, which arises due to the rotation of the Earth around its own axis from west to east.

In the northern hemisphere summer, prior to the arrival of monsoon, a low-pressure zone forms around 5°N and 5°S on either side of the equator. This is referred to as a near-equatorial double trough. During the Asian summer monsoon the trough is located north of 15°N and is associated with cyclonic vortices. Surface air converging into these troughs ascends, accompanied by moist convection, and gives rise to unsettled weather, mainly clouds and rain. Climatologically, the heat low and the near-equatorial trough in the northern hemisphere form a continuous low-pressure belt, although the two systems give rise to opposite weather. Hence, the location of the low-pressure

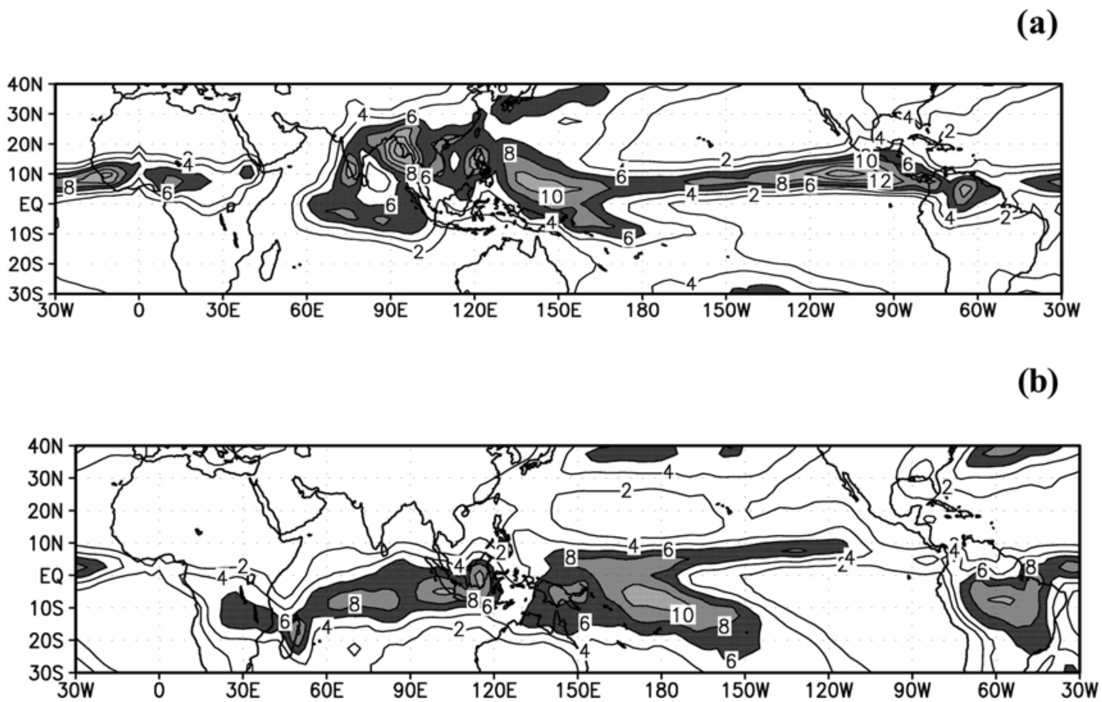

Figure M17 Accumulated rainfall (mm/day) in (**a**) northern (southern) hemisphere summer (winter) monsoon and (**b**) southern (northern) hemisphere summer (winter) monsoon based on Global Precipitation Climatology Project (GPCP) data updated by NCEP/NCAR.

belt over the Asian continent does not always coincide with maximum cloud and rains. Nevertheless, the huge land-mass of the Asian continent with wide east–west extension in the north and, primarily, an ocean to the south makes it geographically the most favorable for dominant summer monsoon flow compared to the other monsoons of the world.

The difference in the global distribution of heating gives rise to planetary-scale quasistationary waves in the tropics and subtropics, which form the most important component of both the summer and winter monsoons. The convective latent heat released forms the most important part of the differential heating, and globally the course of maximum rainfall follows that of convective heating. As shown in Figures M17a and M17b, the maximum precipitation bands occur at around 10°N and 10°S in the northern and southern hemisphere summers respectively. The monsoonal regions receive most of the rainfall due to the seasonal migration of the Intertropical Convergence Zone (ITCZ). This is a relatively narrow low-latitude zone in which the northeast and southeast trade winds originating in the northern and southern hemispheres respectively converge. The mean position of the ITCZ is somewhat north of the equator and it moves northward in the northern summer and southward in the southern summer. The ITCZ may not be the zone of organized moist convection always, especially over the Asian region. In this context it is convenient to use the term maximum cloud zone (MCZ) which is a bright cloud band of about 10° longitude width spreading along the latitude circles. There are two favorable locations of the MCZ; one within 15°N and 25°N latitudes over the land and another over the equatorial region (Figure M17a) at about 5°N have been identified. Successive generations of northward epochs of MCZ in the equatorial region pushes it to about 20°N in July–August from its mean position near 5°N. The northward progress of MCZ in surges is akin to that of monsoon rains over India.

Convection is best represented by low values of outgoing longwave radiation (OLR). A comparison of Figures M18a and M18b indicates the seasonal variation of low values of OLR, which is large over the monsoon regions. Specifically in the case of the Asian summer monsoon between the meridians 70°E and 110°E, the belt of low OLR extends over a large area from 5°S to 35°N, while for the rest of the monsoonal region it is confined to a narrow latitude belt (Figure M18a).

Each monsoon appears to have a dominant circulation either in a north–south or in the east–west direction, with a rising branch located near a heat source, and is associated with upper-level divergent circulation. Similarly, the descending branch lies over a heat sink. The north–south and east–west dominant circulations are popularly called the Hadley and Walker circulations respectively. Generally, the velocity potential field is used to depict divergence and convergence centers in the atmosphere. As shown in Figure M19a, in July the upper-level divergence circulation is prominent over the Asian monsoon region accompanied by convergence centers on both sides in the east–west direction as a part of the Walker circulation. This center of divergence is responsible for spillover of mass from the Asian summer monsoon region in all directions. Thus the Asian summer monsoon is supposed to have a profound impact on the global atmospheric circulation. In the northern hemisphere winter the center of divergence shifts to the south over the north Australian region, as shown in Figure M19b. In the northern hemisphere the Walker and Hadley type overturnings are observed to be prominent in summer (Figure M19a) and winter (Figure M19b) respectively.

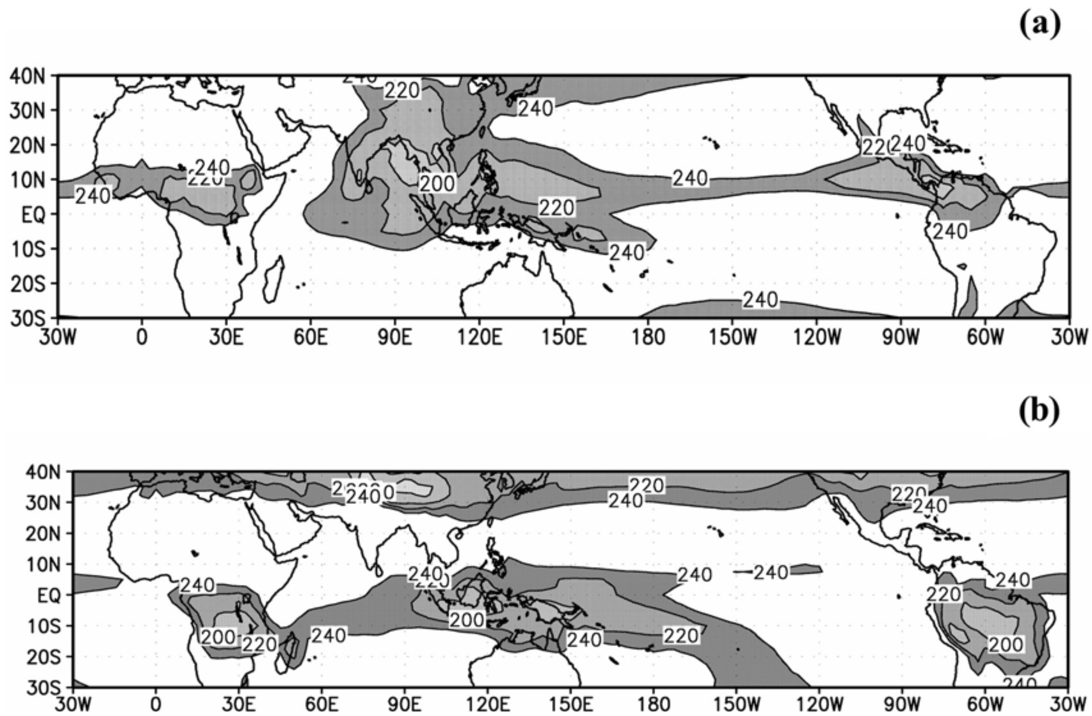

Figure M18 Same as in Figure M16 except for outgoing longwave radiation (W/m^2).

Important features of regional monsoons

The principal components of the Indian summer monsoon are the monsoon trough, the cross-equatorial low-level jet over coastal Africa, the Mascarene High, the Tibetan anticyclone and the tropical easterly jet stream at the upper level, monsoon disturbances and the cloud cover which gives rise to rainfall. The Western Ghats and the Himalayan mountains help in the ascent of moist air, and this gives rise to heavy orographic rains. Around July the monsoon activity is most pronounced, and the low-pressure area is intense, extending from north Africa to northeast Siberia. At this time a trough lies over north India with its axis extending from a place in northwest India to the northern tip of the Bay of Bengal running almost parallel to the Himalayas. This is referred to as the *monsoon trough*. The pressure gradient is strong south of this trough. Simultaneously, in the southern hemisphere, off the coast of Madagascar, a region of high pressure persists giving rise to anticyclonic circulation. This is known as the *Mascarene High*. The *cross-equatorial low-level jet* is most pronounced at a height of about 1.0–1.5 km and its major part penetrates into east Africa during May. Subsequently, it crosses the Arabian Sea and reaches the Indian west coast. Due to significant boundary layer exchange processes between the ocean and the atmosphere, there is an abundant supply of moisture to the Indian summer monsoon. The Tibetan plateau acts as a source of heat in summer and as a sink in winter. The heating of this elevated landmass in summer leads to the development of an intense anticyclone in the upper troposphere, which is popularly known as the *Tibetan Anticyclone* with strong east-northeasterly flow over north India. Here deep convection exists along with strong vertical wind shear. Monsoonal circulation is also accompanied by a variety of cyclonic disturbances; especially over the Indian and adjoining oceanic region one observes the *onset vortex, mid-tropospheric cyclones, offshore vortices, lows, monsoon depressions and other cyclonic disturbances of higher intensity*. Observations show that most of the remnants of tropical cyclones from the South China Sea under favorable atmospheric conditions intensify to monsoon depressions. Studies also show that the horizontal and vertical wind shears of the mean monsoon flow in the presence of deep convection supply energy for the growth of monsoon disturbances.

The east Asian monsoon consists of a monsoon trough in the South China Sea and western Pacific, a cross-equatorial flow around 100°E, the cold anticyclone in Australia, the subtropical high in the western Pacific, the upper-level northeasterly flow, the convection along the monsoon trough and the midlatitude disturbances. In the middle of June the ITCZ is normally prominent over India, corresponding to an active phase of the Indian monsoon, and at the same time the subtropical convergence, which is called Mei-yu in China or the Baiu frontal zone in Japan, is most active, especially around Japan.

For the west African monsoon, the major components are the low-level southwesterly wind over the coastal areas of the Gulf of Guinea flanked by northeasterly winds to its north. The African easterly jet and tropical easterly jet are the dominant middle- and upper-level atmospheric flows in this region. With the change of season from summer to winter, there is no complete reversal of wind direction over the west coast of Africa (Figures M17a and M17b) as in the case of the Indian monsoon. The boundary between the southwesterly and northeasterly winds is known as the intertropical discontinuity in west Africa. The major rainfall in the east African monsoon is due to the north–south movement of the ITCZ.

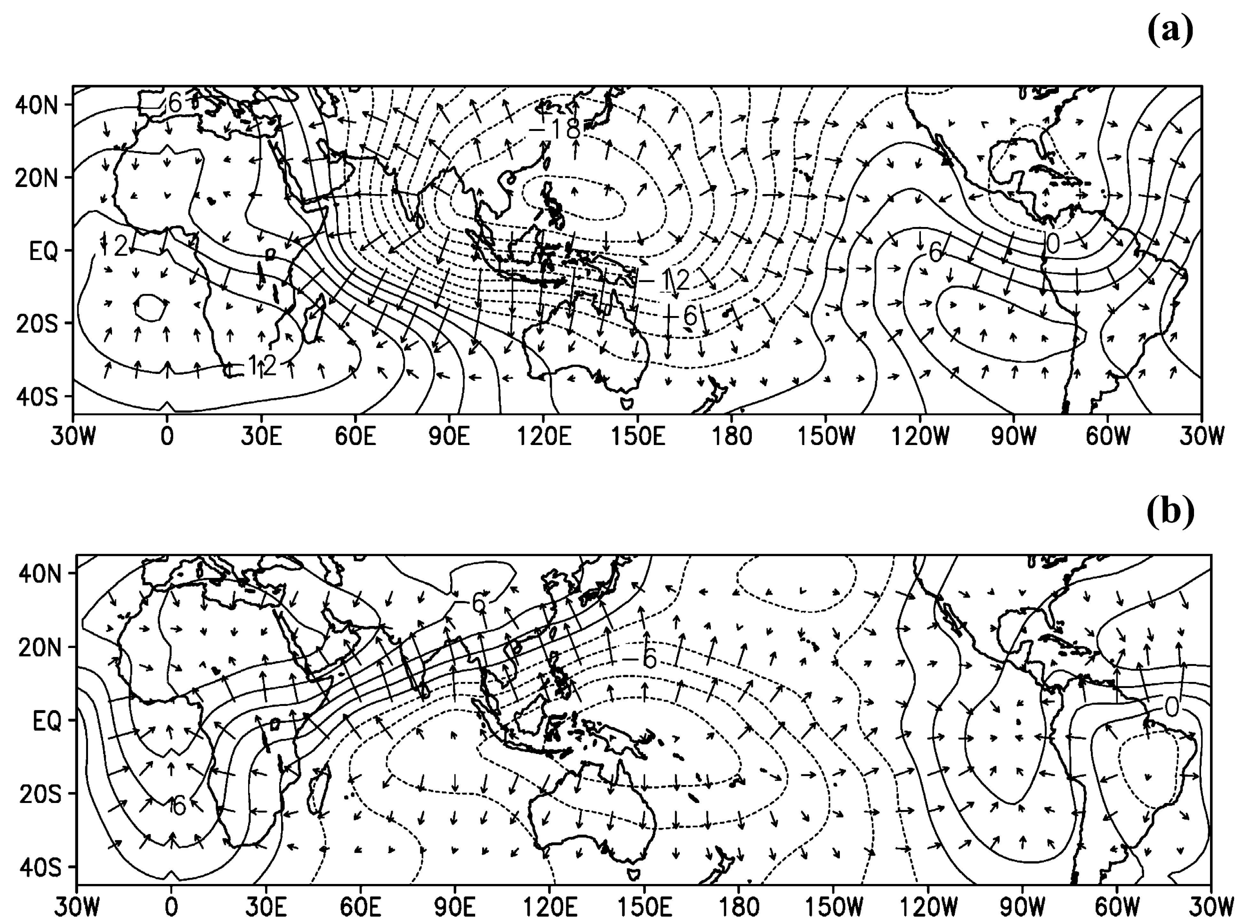

Figure M19 Divergent wind and contours of monthly mean velocity potential ($10^6\,m^2\,s^{-1}$) at 200 hPa (**a**) July and (**b**) January based be NCEP/NCAR reanalysis.

There are many similarities between the Asian summer monsoon and the northern Australian monsoon, although in general the latter is less intense. The ITCZ across northern Australia, near-equatorial upper-level easterlies and lower-level westerlies, cross-equatorial flow, tropical cyclones and monsoon depressions are the important components of the Australian monsoon. The southern hemisphere near-equatorial low lies to the south of the ITCZ over northern Australia. This low-pressure area is referred to as the monsoon trough and the corresponding westerlies are the north Australia monsoon winds.

The North American monsoon domain is considered to be large, as it spreads over much of the southwest United States, northwest Mexico and even Central America. The eastern north Pacific ITCZ between 120°W and 80°W migrates northwards (Figures M17a and M17b) in summer and the corresponding zonal wind becomes westerly at the lower level. At the upper level the zonal wind is easterly. The opposite happens in the winter. Over the North American Arctic region there is a seasonal reversal of wind direction and technically it may be considered to be monsoonal.

The occurrence of South American summer monsoon circulation can be shown by considering the difference between the prevailing surface wind and its annual mean. In the southern hemisphere summer this anomaly flow from the sub-Saharan region, after crossing the equator, becomes northwesterly along the eastern side of the Andes. In winter the anomaly flow reverses its direction. The seasonality of surface wind over this region is not normally noticed because of the fact that the easterly trade winds prevail over the tropical Atlantic throughout the year (Figures M16a and M16b).

Intraseasonal and interannual variabilities

The monsoon passes through different phases, such as its onset, active periods, breaks and withdrawal, which do not have any regularity. There are annual variations in the dates of onset and withdrawal of the monsoon at a place, and also in the duration and intensity of the active and break conditions of monsoon. The normal duration of the Indian summer monsoon is about 4 months starting from 1 June. It begins to withdraw from the northwest part of India by the middle of September. The extreme dates of onset of India summer monsoon is found to be as early as 11 May and as late as 15 June. Based on the temporal rainfall distribution at a particular place, there is a statistical method for delineating the normal dates for onset and withdrawal of monsoon at that place. The date of onset of monsoon over the southern tip of India every year is based on a

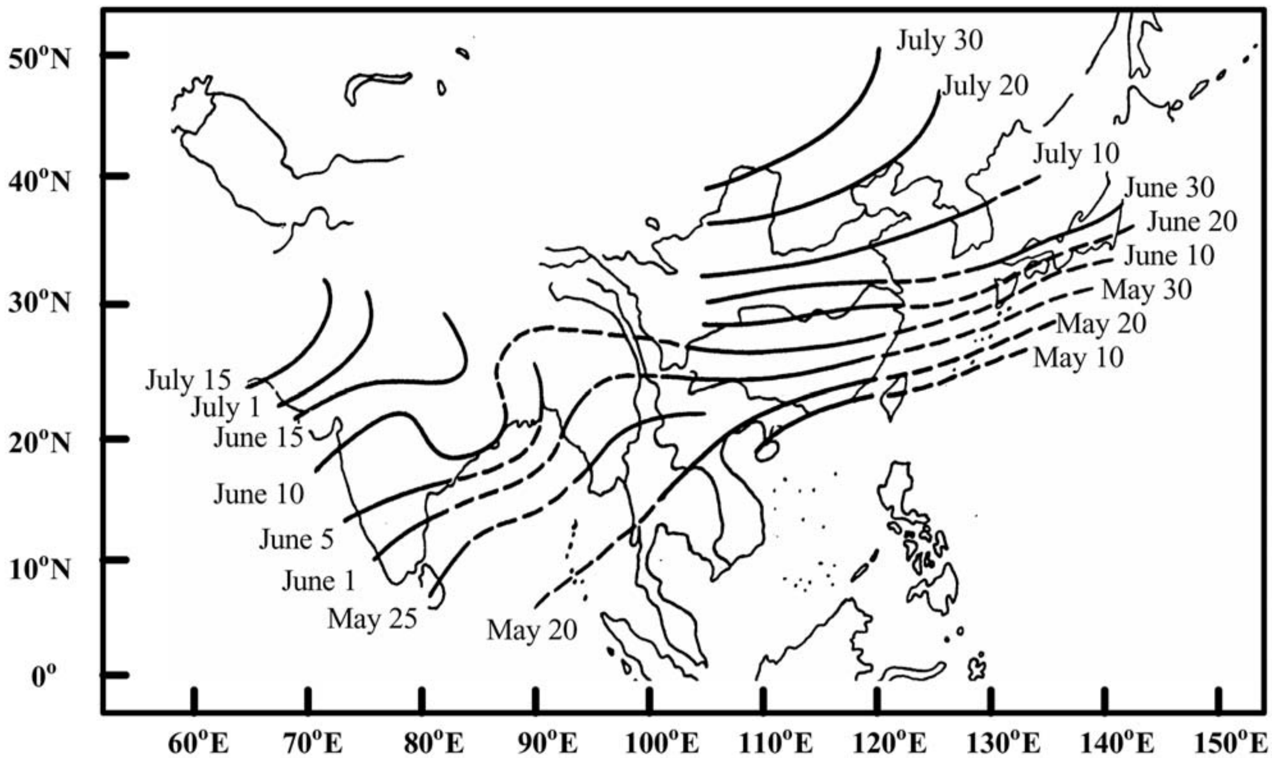

Figure M20 Mean dates of onset of Asian summer monsoon (based on Figures 3.1 and 3.9 in Rao (1976) and Chang and Krishnamurti (1987) respectively).

working rule by taking into account the continuously accumulated rainfall for certain days over certain meteorological stations in the west coast of southern India. Figure M20 shows the normal dates of onset of the summer monsoon at different places along with the northward progress of the rain belt over the Asian landmass. By mid-July the whole of India is observed to be covered by the summer monsoon. The withdrawal of the monsoon from India is slower than the onset; in fact it is difficult to know when southwest monsoon ends and the northeast monsoon begins over the extreme southern part of India.

The entire duration of the monsoon does not exhibit any uniformity or regularity in weather systems. During a season there are periods of active rainfall and also there are continuous days of no rainfall, called a "break". Normally, monsoon breaks are accompanied by cessation of rainfall in central and the adjoining parts of India, and an increase in rainfall over northeast India and foothills of the Himalayas. Observations show that, during breaks, the monsoon trough usually shifts northward to the foothills of the Himalayas and surface pressures are above normal over central parts of India. Prolonged monsoon breaks may lead to severe droughts and hence have disastrous effects. Thus, during a monsoon season, the characteristics and duration of weather systems and the amount and distribution of rainfall vary. Such intraseasonal variations in the Indian summer monsoon circulation and rainfall are the most important aspects of the monsoon.

The monsoon of a year is not identical with the monsoon of another year in all aspects. There are year-to-year variations in the strength of circulation and also in the associated rainfall at a place. It is observed that the mean rainfall over India during June to September in a year is about 88 cm with a coefficient of variation of 10%. Thus the amount of Indian summer monsoon rainfall (ISMR) equal to or more than 110% and less than 90% of the mean value are taken as excess and deficient rain respectively, and that equal to or more than 90% and less than 110% is normal. This annual variation in ISMR averaged over the whole of India may seem to be small, but its impact is high because of large spatial variations in the annual rainfall. The coefficient of variation of annual rainfall between different parts of India varies between 12% and 45%. There are occasions when some parts of the country experience deficient rain and drought, although the mean monsoon rainfall over the country as a whole is normal or in excess. The opposite also happens. Such temporal and spatial variations in the rainfall make the agriculture and the economy of India most vulnerable to the summer monsoon.

Variabilities of the Indian summer monsoon in the interannual and intraseasonal scales are scientifically intriguing. The exact reasons for such variations are not yet known. It is well known that changes in weather occur due to atmospheric instabilities. It is also known that the dynamic processes occurring in the atmosphere are nonlinear in nature, involving interactions between different spatial scales starting from few kilometers to hundreds of kilometers. Thus, "internal dynamics" play a very important role in the interannual variations of the seasonal mean circulation and rainfall. Secondly, the slowly varying "surface boundary conditions", such as the sea surface temperature (SST), soil moisture, sea ice, snow extent and depth, and land surface conditions play very important roles in the interannual variations of the ISMR. The exact physical processes responsible for the interannual variation of ISMR are not yet known. General circulation models (GCM) are increasingly used as convenient scientific tools to design and conduct sensitivity experiments on state-of-the-art computer systems for understanding the nonlinear processes leading to such variations. Large-scale field experiments using sophisticated sensors fitted to ships, aircraft, balloons and weather satellites yield good-quality meteorological data which help in understanding the evolution of weather systems. They are also needed for preparing reasonably accurate initial data for numerical weather prediction (NWP) models.

The Asian summer monsoon circulation is very different from other monsoons in its intensity, accompanying rainfall and its intraseasonal and interannual variabilities. A detailed analysis of satellite imagery indicates that the Asian summer monsoon is peculiar in having the double MCZ (one over the continents and the other over the seas) and also in the large meridional extent of low OLR spanning over 30° to 40° latitudes. It is believed that understanding the relationship between the continental MCZ and the oceanic MCZ may help a great deal in understanding the interannual and intraseasonal variations of ISMR.

Forecasting the Indian summer monsoon rainfall

Although the Indian summer monsoon is the most dominant feature of the atmospheric circulation, and although its annual appearance is certain, the most challenging task is forecasting its onset dates at different places, for active and break periods, withdrawal dates and above all the amount and spatial distribution of rainfall. Forecasts of rainfall at all time scales, such as short-range (up to 2 days), medium-range (between 3 and 10 days) and an extended range (a month to a season) are important. These involve the detailed understanding of the interaction between different physical processes, vast network for data collection, improved data assimilation procedures, sophisticated coupled ocean atmosphere models and the availability of immense computing power. The accuracy of the initial atmospheric variables is very important for the short-range forecasts, while both the initial atmospheric conditions and changes in the surface boundary conditions play very crucial roles in the forecasting of monsoons in the medium-range. In the long-range forecasts the variations in the slowly varying boundary conditions matter the most. Efforts are going on to forecast the Indian summer monsoon and its associated rainfall with reasonable accuracy by numerical weather prediction models at least a week in advance. As the mesoscale phenomena are very crucial, high-resolution regional models are nested to global models. They are best suited for forecasting monsoon rainfall up to few days with greater skill.

Forecasting the seasonal mean monsoon rainfall a couple of months ahead is very useful for farmers and planners as well. Based on the information of favorable and unfavorable predictors, a parametric model is generally used for a qualitative forecast of seasonal mean rainfall. Other statistical models, such as multiple-power regression models, dynamic stochastic transfer models, principal component regression models and models based on neural network technique are being tried for quantitative forecasting of ISMR for the whole of India, as well as for some of its homogeneous regions, such as northeast, northwest and peninsular India. These models depend on the long time relationship of some well-identified predictors such as the El Niño, Southern Oscillation Index, Eurasian and Himalayan snow covers, South Indian Ocean and Arabian Sea SST, Indian Ocean Equatorial Pressure, Central India temperature, northern hemisphere temperature and pressure, etc. with ISMR. The statistical models play a very important role in the long-range forecast of ISMR because the interannual variation of ISMR has not been successfully simulated by most of the NWP models. However, statistical models have their limitations, especially when the predictors are large in number as in case of ISMR. Changes in the global and regional circulation patterns may generate temporal variations of several predictors and hence affect their relationship with ISMR.

Because of the chaotic nature of the atmosphere there is a limitation of about 2 weeks in the deterministic prediction of weather. However, the seasonal mean monsoon circulation is potentially more predictable because of the slowly varying surface boundary conditions. The internal dynamics also affects the monsoon circulation. This makes the prediction of Indian summer monsoon very difficult. Intercomparison of the results of a number of GCM indicates that the simulation of ISMR is more difficult than the rainfall over the rest of the tropics. This is attributed to the occurrence of the double MCZ over the Indian summer monsoon region, one above the continent and the other over the Indian ocean. The successive fluctuations of the MCZ between these two favorable positions is found to be a major limitation for most of the GCM.

Oceans cover about three-fourths of the Earth's surface, and water has a much greater capacity for absorbing and storing heat energy than any other material on Earth. Thus the oceans have large memory. Hence, the interactions between the ocean and atmosphere have a crucial role in determining the state of the atmosphere. Results of sensitivity experiments show that the Pacific SST influences monsoon circulations more than other surface boundary conditions. Therefore, coupled ocean–atmosphere models are very essential for extended-range monsoon forecast by numerical methods. Such models also need proper parameterization of the land surface processes and cloud-radiation feedbacks. Ensemble simulations obtained from a combination of different GCM may help to yield reasonable forecasts of Indian summer monsoon.

The influence of global warming on the monsoons in general, and the Indian summer monsoon in particular, is not very clear, since the former occurs in the time-scale of a century. In general, global warming is expected to intensify the hydrological cycle and hence the monsoon rainfall. However, considering the limitations of GCM in simulating ISMR and the large time-scale of model integration, the different projections of Indian summer monsoon arising due to global climate change should be interpreted intelligently.

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Bibliography

Asnani, G.C., 1993. *Tropical Meteorology*. G.C. Asnani.

Das, P.K., 2002. *The Monsoons.* New Delhi: National Book Trust of India.

Das, P.K., 1986. *Monsoons*. World Meteorological Organization, WMO-No.613.

Chang, C.-P., and Krishnamurti, T.N. (ed.), 1987. *Monsoon Meteorology*. New York: Oxford University Press.

Fein, J.S., and Stephens, P.L. (ed.), 1987. *Monsoons*. New York: John Wiley & Sons.

Gadgil, S., Sajani, S., and participating AMIP Modelling Groups, 1998. *Monsoon precipitation in Atmospheric Model Intercomparison Project runs*. WCRP-100, WMO/TD-No.837.
Keshavamurty, R.N., and Sankar Rao, M., 1992. *The Physics of Monsoons*. New Delhi: Allied Publishers.
Krishnamurti, T.N. (ed.), 1978. *Monsoon Dynamics*. Basel: Birkhauser Verlag.
Lighthill, J., and Pearce, R.P. (ed.), 1981. *Monsoon Dynamics*. Cambridge: Cambridge University Press.
Ramage, C.S., 1971. *Monsoon Meteorology*. New York: Academic Press.
Rao, Y.P., 1976. *Southwest Monsoon*. India Meteorological Department.

Cross-references

Asia, Climate of South
Ocean–Atmosphere Interaction
Trade Winds and the Trade Wind Inversion
Tropical and Equatorial Climates
Winds and Wind Systems

MONTREAL PROTOCOL

An international agreement for the protection of the ozone layer was initiated in 1985 in Vienna, Austria. Details were defined in the *Montreal Protocol on Substances that Deplete the Ozone Layer* that was signed in 1987 and ratified in 1989.

The substances defined are CFC, HCFC, HBFC, halons, carbon tetrachloride, methyl chloroform, and methyl bromide. The phase-out time for these was defined in the Montreal Protocol, with appropriate changes made at Vienna in 1995, Copenhagen in 1996, and Beijing in 1999. In all cases different schedules were applied to developed and developing countries.

The agreement to reduce production of ozone-depleting substances has been successful. Public concerns, political action, and industrial innovation, including the identification and production of appropriate alternative chemicals, has, in most cases, seen the phase-out meet or exceed the provisions of the Montreal Protocol.

The complete text of the Protocol including adjustments and amendments is available at the UN Environment Program site listed below.

John E. Oliver

Bibliography

Editor's note: The most recent data and status concerning ozone depletion are available from websites including:
http://www.unep.org/ozone/Montreal-Protocol/MontrealProtocol2000.shtml
http://www.afeas.org.montreal_protocol.html
http://www.epa.gov.ebtpages.intestratosphericozone.html

Cross-references

Climate Change and Human Health
Global Environmental Change: Impacts
Kyoto Protocol
Ozone

MOUNTAIN AND VALLEY WINDS

There is a special category of local or tertiary winds that are directly related to the topography. Sir Harold Jeffreys called them *antitriptic winds* because of the dominance of friction in their thermodynamics. Known as moutain and valley winds or breezes (Figure M21), they are diurnal in character and most obviously operative in otherwise calm, clear weather.

At night on the high mountain slopes, the air cools rapidly by radiation and thus becomes denser; it flows downhill into the valleys, gaining velocity and momentum, as a *katabatic wind*. A similar cold wind often flows down the surface of a glacier. This nightly drainage of cool air from mountain slopes may lead to important temperature inversion in the high plains, e.g. of Colorado or central Switzerland, leading to extremely cold nights. In high plains of the central Rockies the basins at 2500–2800 m elevation are treeless and known as *parks*, although the actual tree line is much higher (3000–3400 m), and the slopes carry forests of conifers and aspen: the cold layer below the inversion predicates what might be called an *inverted tree line*.

In the daytime the mountain slopes are warmed by insolation and the breeze begins to flow upslope, usually beginning about half an hour after sunrise and continuing until half an hour before sunset. It may reach a maximum of 6 m/s (12 knots) up sunny slopes but much less on the northern slopes. In depth the air flow may exceed 150 m increasing uphill. This type of flow is called an *anabatic wind*.

According to Defant (1951), an additional factor was suggested by Wagner. A pressure gradient from the plain to the valley must exist during the day, whereas a reverse gradient must develop at night. Equalization of the pressure differences must

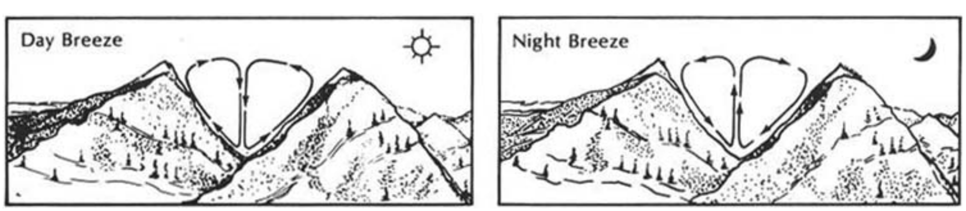

Figure M21 Schematic diagram illustrating moutain wind (breeze) and valley wind (breeze).

obtain at effective ridge altitudes, so that largest differences can be expected in valley floors. The pressure gradient is then due to the combined effects of the valley bottom slope and the effective ridge altitude; in practice one has two wind systems, a thermal slope circulation and a valley wind

The following sequence of mountain and valley winds may be identified:

1. Sunrise: onset of upslope winds; continuation of mountain wind. Valley cold, plains warm.
2. Forenoon (about 0900); strong slope winds, transition from mountain wind to valley wind. Valley temperature same as plains.
3. Noon and early afternoon; diminishing slope winds, fully developed valley wind. Valley warmer than plains.
4. Late afternoon; slope winds have ceased, valley wind continues. Valley continues warmer than plains.
5. Evening; onset of downslope winds, diminishing valley wind. Valley only slightly warmer than plains.
6. Early night; well-developed downslope winds, transition from valley wind to mountain wind. Valley and plains at same temperature.
7. Middle of night; downslope winds continue, mountain wind fully developed. Valley colder than plains.
8. Late night to morning; downslope winds have ceased, mountain wind fills valley. Valley colder than plains.

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Bibliography

Barry, R.G., and Chorley, R.J., 1998. *Atmosphere, Weather and Climate*, 7th edn. London: Routledge.

Defant, F., 1951. Local winds. In Malone, T. (ed.) *Compendium of Meteorology*. Boston, MA: American Meteorological Society, pp. 655–672.

Jeffreys, H., 1922. On the dynamics of wind. *Royal Meteorological Society Quarterly Journal*, **48**: 29–46.

Lutgens, F.K., and Tarbuck, F.J., 2002. *The Atmosphere*, 8th edn. Upper Saddle River, NJ: Prentice-Hall.

Cross-references

Local Winds
Microclimatology
Mountain Climates
Winds and Wind Systems

MOUNTAIN CLIMATES

Mountain climates show great variation in the values of climatic elements over a short distance due to elevational differences, complex topography, and the physical presence of the mountains themselves. Whereas mountain climates are usually characterized by their spatial variety, the same causes often lead to large temporal variation as well.

It has been estimated that mountains or high plateaus occupy 20.2% of the Earth's land surface (Louis, quoted by Barry, 1992) and influence the life-support systems of 400 million people (Ives, 1981). Messerli (personal communication, 2001) claimed that 10–26% of the world's population lived on or within 50 km of mountains. The most prominent mountain areas include the Cascade–Sierra Nevada and Rockies of North America, the Andes of South America, the Alpine system of Europe, the Himalayas and Tibetan Plateau and Caucasus of Asia, the east African Highland, and the mountain backbones of Central America, Borneo, New Guinea, and New Zealand. There are many less prominent highland areas that should be recognized, such as the ice plateau of Greenland and Antarctica and the mountain ranges in both, together with elevated regions of older geologic materials, exemplified by the Appalachians of North America and the uplands of Great Britain and Scandinavia.

The variety of climates in mountain areas leads to confusion in climatic classification, therefore such areas are usually either excluded or grouped together under a broad category such as "highland climate". Modern versions of the Köppen classification, for example, use the letter H for mountain climates (Christopherson, 2002); the H standing for the German *hochgebirge* (high mountain). No quantitative definition of mountain climates has ever been given but, if one were needed, it would include the characteristic of rapid spatial change of climate classes within complex terrain. Such a characteristic would include not only areas currently classified as H but also areas such as the central highlands of Madagascar, which sometimes are attributed to other classes in the Köppen system.

Although humans have noted some of the effects of mountain climates whenever they have had contact with elevated areas, the scientific study of them dates only from the eighteenth century. The investigations of H.B. De Saussure, started in 1787, assured him of recognition as the first mountain meteorologist (Barry, 1978). Numerous mountain meteorological observatories were established in the nineteenth century, many of which were later closed due to lack of funds or for other reasons. Barry (1992) lists 30 of the most important observing stations, many of which have long records and continue to collect data through to the present time. The continuing importance of mountain climates is demonstrated by the appearance of excellent reviews of the subject such as those of Barry (1992) and Whiteman (2000). Additionally, there are more frequent conferences on the subject both in Europe and in North America (American Meteorological Society, 2000), and a number of national and international research programs into mountain meteorology are currently operating or planned.

Variety in mountain climates

Geographic variety in mountain climates is seen both on the global scale and on meso- and local scales. On the global scale the nature of mountain climate is determined by the latitude of the site as well as other factors that influence most climates, such as proximity to the ocean or features of the general circulation of the atmosphere. Table M4 gives some idea of the range of temperature and precipitation values found at high-altitude stations at different latitudes. La Paz at 16°S has a very small annual temperature range, whereas the range at Lhasa (29°N) is rather larger and similar to the midlatitude station at Longs Peak, Colorado. The Antarctic station of Vostok (which might also be regarded as a high-latitude climate) has an even larger annual temperature range but is more noteworthy for its extremely low temperatures. The precipitation regime at La Paz is influenced by the southward movement of the Intertropical Convergence Zone giving a maximum of rainfall between December and February. The same phenomenon in conjunction with the Asian monsoon system is

Table M4 Monthly mean temperature (°C) and total precipitation (mm) at selected high-altitude stations

Station	Altitude	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	Year
La Paz, Bolivia 16°30′S 68°08′W	3658 m	11.7° 165.1 mm	11.7° 106.7 mm	11.7° 66.0 mm	11.4° 33.0 mm	10.3° 12.7 mm	8.9° 7.6 mm	8.6° 10.2 mm	9.5° 12.7 mm	10.6° 27.9 mm	11.7° 40.6 mm	12.5° 48.3 mm	12.0° 94.0 mm	10.8° 574.0 mm
Lhasa, Tibet 29°40′N 91°07′E	3685 m	−1.7° 2.5 mm	1.1° 12.7 mm	4.7° 7.6 mm	8.1° 5.1 mm	12.2° 25.4 mm	16.7° 63.5 mm	16.4° 121.9 mm	15.6° 88.9 mm	14.2° 66.0 mm	8.9° 12.7 mm	3.9° 2.5 mm	0.0° 0.0 mm	8.3° 406.4 mm
Longs Peak, Colorado 40°15′N 105°35′W	2729 m	−5.0° 17.8 mm	−5.0° 27.9 mm	−2.8° 50.8 mm	−1.1° 68.6 mm	5.0° 61.0 mm	10.0° 43.2 mm	12.8° 91.4 mm	12.8° 55.9 mm	8.9° 43.2 mm	3.9° 43.2 mm	−1.1° 22.9 mm	−5.0° 22.9 mm	2.8° 558.8 mm
Vostok, Antarctica 78°27′S 106°52′E	3420 m	−33.6°	−43.9°	−53.9°	−63.1°	−63.4° No precipitation data	−66.7°	−67.0°	−70.6°	−67.3°	−58.4°	−63.9°	−32.2°	−55.6°

Source: After Liljequist (1970), pp. 483–515 and Critchfield (1983), pp. 421–429.

responsible for the maximum of precipitation at Lhasa between June and September. Longs Peak shows a summer maximum of precipitation owing to the frequency of thunderstorms at this season. There is very little measurable precipitation in Antarctica where the annual value is believed to be less than 6 in (150 mm; Trewartha and Horn, 1980).

On smaller geographic scales mountain climates are affected by the rapid change of temperature with height, aspect, particular wind systems, and the barrier effect of mountains as they act with respect to precipitation values. The complexities of mountain climates are represented by surface energy budget values. All of these factors will be discussed below.

Characteristic features of mountain climates

Radiation

The amount of global solar radiation received at the Earth's surface increases with altitude due to the decreased radiation through the atmosphere and atmospheric water vapor that is concentrated at the lower part of the atmosphere. Sauberer and Dirmhirn (quoted by Barry and Van Wie, 1974) note an increase in global solar radiation in the Austrian Alps from 650 ft (200 m) to 9840 ft (3000 m) of 21% in June and 33% in December. The increase is approximately exponential and occurs in both cloudy and clear skies. However, varying cloud amounts in mountain areas can sometimes disturb this pattern (Greenland, 1978). The increase in radiation is most marked in the shorter wavelengths. Reiter and Munzert (1982) observed that total ultraviolet radiation could be increased 1.4 times in January and 1.5 times in June, accompanying an altitude increase from 2296 ft (700 m) to 9840 ft (3000 m), using a 5-year observing period in the northern European Alps. The large variety of slope orientations and gradients give rise to a whole spectrum of different aspects in mountain areas, and it is the varying degree of radiation receipt on these slopes that essentially leads to the importance of aspect in mountain climates. At latitude 50°, for example, there is a fourfold difference in the radiation receipt on 45° slopes facing north and south (Barry and Van Wie, 1974). Variation of radiation receipt due to aspect can overshadow that due to any other cause, and leads to variation in temperatures, evaporation rates, and processes of the soil and vegetation growth. The specific location of long-lasting snow and ice patches is also associated with aspect.

Temperature

Since the atmosphere is heated principally by infrared radiation from the Earth's surface, air temperature usually decreases with altitude. The decrease varies in the free atmosphere from about −2.7°F/1000 ft (−5°C/km) for saturated air to −5.5°F/1000 ft (−10°C/km) for dry air and averages about −3.2°F/1000 ft (−6°C/km). The air above mountain slopes is also influenced by latent and sensible turbulent heat flows and so the temperature changes with altitudes, or lapse rates, on slopes are often different from those in the free air (Coulter, 1967). Average lapse rates in mountain climates have also been shown to vary with season, global climatic zone and with airmass frequency (Barry, 1992).

The rapid change of temperature with height leads to a climatic zonation that, in some parts of the world, has been given a special nomenclature. Trewartha and Horn (1980) point out that in tropical Latin America four zones are recognized; these are Tierra Caliente (hot lands), Tierra Templada (temperate lands), Tierra Fria (cool lands), and Tierra Helada (land of frost). Each of these zones is associated with different types of vegetation or agricultural crops. Figure M22 displays a similar kind of zonation in the New Zealand Southern Alps. From the perpetually snow-covered slopes of Mount Tasman one descends through rainforest vegetation before reaching sea level.

Figure M22 The west coast of the South Island of New Zealand from the top of Mount Tasman. The snow-covered mountain peaks give way to subtropical rainforest before sea level is reached.

Another important feature of temperature in mountain climates is related to the sinking downslope of radiationally cooled, relatively dense cold air that accumulates at the bottom of valleys and in mountain basins. The colder air at the lower elevations, and the relatively warmer air above, result in increasing temperatures with height, a situation known as a temperature inversion. Inversions are commonly formed at night but can last through the whole day, especially in winter, when radiative heating is not strong enough to disperse them. Stagnating cold air in valley bottoms can have the effect of trapping air pollutants.

The frequency of temperature inversions and the consequent higher temperatures at some distance up the slope can be so great as to cause generally warmer areas in midslope. These are called thermal belts and have been used to advantage by agriculturalists in many parts of the world (Dunbar, 1966). The center of the thermal belt in the European Alps is found to be between 330 and 1310 ft (100–400 m) above the valley in areas of relief under 1650 ft (500 m). In higher mountains the thermal belt is centered 1150 ft (350 m) above the valley floor in summer and 2300 ft (700 m) above it in winter (Barry, 1992).

Precipitation

A general global pattern of precipitation change with altitude has been noted by Lauscher (quoted by Barry, 1992). There is a general decrease of precipitation with elevation in equatorial latitudes, which is also seen, to a lesser degree, at polar stations open to a maritime influence. In the tropics, however, there is a maximum of precipitation at about 3000 ft (900 m). In midlatitudes there is a general increase of precipitation up to the highest altitudes at which observations are made at about 12 000 ft (3660 m).

This global pattern is often superceded by the effect of mountain barriers on the vertical flow of air and subsequent increase or suppression of precipitation formation processes. When air is forced to higher altitudes in its passage over a mountain, the air is cooled and condensation of water vapor often gives rise to precipitation formation on the windward side. This is sometimes called orographic precipitation. Descending air on the leeward side of the barrier is accompanied by warming of the air, an increase in potential for evaporation, a decrease in relative humidity, and a decrease in the likelihood for precipitation formation. Dry areas on the lee side of mountain barriers are called rainshadow areas. These phenomena are found in many of the mountain areas of the world. The South Island of New Zealand is a good example of such orographic effects. On the windward side of the island at Milford Sound, annual precipitation is 245 in (6233 mm), whereas on the lee side in the interior plateau of central Otago the value decreases to 13 in (330 mm) at Alexandra (Garnier, 1958).

A complicating but important feature of mountain precipitation is the fact that much of it falls in the form of snow. The difficulty of measuring snowfall makes values of precipitation from the higher elevations of mountain climates less accurate than their lowland flat-terrain counterparts. In both environments it is hard to design a shield for precipitation gauges that suppresses local wind eddy currents and permits representative assessment of snowfall but, in addition, mountain observational sites are seldom representative of larger geographic areas. As a result it is often expedient to sample the water content of the mountain snowpack over snow courses set up in a systematic manner. As much accuracy in snowpack measurements as possible is necessary because many lowland areas depend on this water source for agricultural, industrial, hydroelectric energy, and domestic purposes. The frozen form of precipitation in mountain areas also adds immensely to their esthetic qualities in terms of such features as long-lasting snow cover and glaciers.

Wind

The decrease of the effect of friction between the Earth's surface and the movement of air in the free atmosphere causes wind velocities to increase with altitude. This effect is also noted from wind observations at high mountain stations although Wahl (quoted by Barry, 1992) found that for the European Alps mountain summit wind velocities are about half those at corresponding altitudes in the free atmosphere. Nevertheless, mountain areas exhibit some of the highest wind velocities anywhere on the surface of the planet – a gust of over 231 mph (103 m/s) having been recorded on Mount Washington, New Hampshire, where the annual average wind speed exceeds 31 mph (15.7 m/s: Smith, 1982).

More important than high wind velocities is the topographic influence of mountain barriers on winds. These barriers impose wavelike motion on airflow both in the horizontal and in the vertical planes. On the global scale, downwind of major mountain barriers, horizontal waves with wavelengths on the order of 3000 miles (4800 km) are established with their troughs to the lee of the mountain range. This is most pronounced in the northern hemisphere, resulting from the presence of the Rockies in North America and the Himalayas in Asia. A similar smaller-scale effect is noticed, however, east of the Andes in South America and in the New Zealand Southern Alps. In all cases the presence of the mountains has a marked effect on the climates downwind of them, particularly in causing rainshadow areas and, further downwind, regions of cyclogenesis or formation of cyclones. Waves are also imposed on airflow over mountains in the vertical plane. These vertical undulations or waves created by gravity acting on local variations in air density are sometimes called gravity waves (Whiteman, 2000). The effects of these waves can sometimes be noticed when lee wave, or

Figure M23 Lee wave clouds near Aviemore, South Island, New Zealand.

lenticular, clouds are formed on their crests, as illustrated in Figure M23. Wind phenomena in the lee of mountains are not limited to lee waves, however. Another interesting feature is a rotor effect that can give air movement at the surface back toward the mountain. This effect sometimes carries snow that can feed alpine glaciers and maintain their existence at latitudes that would otherwise be impossible (Johnson, 1980).

High-velocity winds in the lee of mountain ranges are also common features of mountain climates. These downslope winds may be warm or cold. The warm variety are called chinook winds in North America and föhn winds in Europe, and their warmth is due to compressional heating as the air moves downward to more dense parts of the atmosphere. "Chinook" is a North American Indian term meaning "snow eater", and signifies the large amounts of snow that can be melted by the wind, especially when it occurs in the spring (Brinkman, 1970). The wind storms at Boulder, Colorado, originally studied by Brinkman, have been found to be difficult to predict because of lack of upstream observational data but also because more than one synoptic-scale pressure situation can give rise to the storms (Leptuch et al., 2000). As in the case of the chinook winds, particular synoptic-scale pressure patterns are required for the establishment of the cold winds. These winds that have several local names are exemplified in Europe by the Bora winds of Croatia and Slovenia (parts of the former Yugoslavia). Besides a strong pressure gradient, their formation occurs when there is a damming up of cold air east of the mountains in the type area (Petkovsæk and Paradiû, quoted by Barry, 1992). Locally, under certain conditions, downslope winds can give rise to intense windstorms, causing much damage in populated areas (Miller et al., 1974). A windstorm in 1972 caused $2.5 million damage in Boulder, Colorado (Whiteman, 2000).

The modification of wind flow direction is another important aspect of wind in mountain climates. Modifications include the lateral movement of air around obstacles, the channeling of air through topographically formed tunnels, and the altering of wind direction between that found in valleys and that above the ridge line. The fact that it is possible to have different wind regimes in different parts of mountainous topography is another noteworthy feature of mountain winds. When two or more regimes influence each other they are said to be coupled, but when they act as separate entities they are said to be uncoupled.

Some of the most frequent kinds of mountain winds are thermally induced. Downslope movement of cold air at night is called katabatic flow, whereas upslope movement of a warm air during the day is termed anabatic flow. In the absence of strong synoptic-scale winds, mountain-valley wind systems can develop. The classic description of these involves the combined results of katabatic nighttime winds, giving rise to a downvalley wind best marked in the early morning. This is gradually stilled when radiative heating of slopes gives rise to anabatic winds. Air to feed these is drawn up the valley, leading to an upvalley wind system that is most noticeable in the early afternoon. More recent observational studies show the phenomenon to be very complex and involving compensatory winds at different levels and characteristic fluctuations in wind velocity (Barry, 1992). The theory of these wind systems has yet to be completely worked out but Whiteman (in American Meteorological Society, 1981) has made an important contribution to it in his study of the breakup of valley temperature inversions. Whiteman (2000) distinguishes four components of the overall mountain wind system; namely slope winds, along-valley winds, cross-valley winds, and mountain-plain winds. He comments that it is difficult to study the pure form of any one of these components because they interact with one another in various combinations and also with the winds of the larger-scale synoptic situation.

Surface energy budgets

The energy budget of the surface is an estimation of heat flow to and from the surface by radiation and by sensible and latent heat flows to and from the atmosphere, and by conduction of heat into and out of the submedium. The estimation of surface energy budgets often helps our understanding of a climate under consideration. Reiter (1982) has drawn attention to the importance on global climate particularly of sensible heat fluxes from the Tibetan Plateau, where a resulting large-scale wind circulation may be established. Reiter has described how anomalies of heat flowing from the Tibetan Plateau could set off significant interannual variability of monsoon circulation systems. Complete energy budgets of snow- and ice-free mountain area surfaces are difficult to estimate for any length of time and hence very few exist. Barry (1992), for example, while giving information on several others, tabulates only six studies, and all of these are for short periods. One of the few long-term studies (daily values for 1 year) for the New Zealand Southern Alps shows a positive net radiation to be used more or less in equal parts by sensible and latent heat flows, whereas ground heat flow is rather small (Greenland, 1973). These results, which relate to a shrub and grass surface, have been paralleled by shorter-term studies in similar environments such as Alaska (Wendler, 1971). However, the effect of aspect plays a large role in determining the radiation receipt, and therefore the energy budgets, of mountain areas, as has been demonstrated by Brazel and Outcalt (1973) for another Alaskan site. Cline (1997a) performed pioneering work on measuring the energy balances of melting snow surfaces at Niwot Ridge, Colorado. He found that, besides net radiation, surface flowing turbulent fluxes were an important source of energy for snow melt and could provide 25–54% of the energy needed. He has also developed a point energy and mass balance model of a snow cover (called SNTHERM) for use in alpine conditions (Cline, 1997b).

The energy budgets of snow and glacial surfaces of mountain areas are better known partly because their details are important for practical hydrological purposes. A summer study of the Peyto glacier in the Canadian Rockies indicates that heat flows are generally toward the ice surface. These heat flows are used in melting the ice; 71% of the ice melt was from net radiation, 23% from the downward transfer of sensible heat, and 6% was from the downward transfer of latent heat in the form of condensation (Munro, quoted by Oke, 1987). These values appear to be generally representative as judged by the reviews presented by Patterson (1994). The difficult problems of aspect and surface heterogeneity still remain, however, and because of this several investigators have turned their attention to attempting to model the energy budgets of such surfaces. Dozier and Outcalt (1979) performed pioneering work in this area. With the advent of satellite data the heterogeneity of mountain areas becomes more approachable. Duguay (1994) has used remote sensing data to map the growing season radiation of the Niwot Ridge alpine site in Colorado. Researchers at the Institute for Computational Earth System Science at the University of California, Santa Barbara, are leaders in handling the large datasets of energy budget values and other data that result from the remotely sensed products. In another advance, Kumar et al. (1997) have extended the process of modeling topographic variation in solar radiation for use within a GIS environment. It is with respect to their surface energy budgets that mountain climates show one of their greatest degrees of variability.

Applied aspects of mountain climates

The increasing use of mountain areas by humans has stimulated the field of applied climatology in mountain regions. Studies have taken a wide variety of forms. These forms include the study of climate change and tourism, hydro-electric power potential, and marginal agriculture (Beniston, 1994), investigations of weather-related mountain hazards such as avalanches (Mock and Birkeland, 2000) and lightning (Peterson, quoted by Barry, 1992), examination of forest fire weather and smoke management, air pollution dispersion, and aerial spraying (Whiteman, 2000).

One topic of increasing importance is the development of systems to interpolate and extrapolate values of climatic variables in mountain environments. Such an endeavor is extremely useful in almost all applied aspects related to mountain climates. This development is particularly important because there will never be enough actual meteorological observations in mountain areas. Indeed there are fears that increasingly mountain observatories are being closed down for budgetary reasons (Beniston et al., 1997). Pioneering work has been performed on interpolation and extrapolation by researchers at the University of Montana who first developed a methodology called MT-CLIM and later refined it into a system called Daymet. These methodologies permit the interpolation and extrapolation of such variables as air temperature, solar radiation, and surface humidity from sparsely distributed observation sites to other parts of the terrain (Running et al., 1987; Glassy and Running, 1994; Kimball et al., 1997; and Thornton and Running, 1999). This work has been complemented by the development of a methodology, called PRISM (Parameter-elevation Regressions on Independent Slopes Model). PRISM is an analytical tool that uses point data, a digital elevation model, and other spatial datasets to generate gridded estimates of monthly, yearly, and event-based climatic parameters, such as precipitation, temperature, and dewpoint for interpolating precipitation values in mountain (and other) terrain (Daly et al., 1997). Ollinger et al. (1998) have also used regression techniques to establish the spatial variation of temperature and precipitation in the complex terrain of the northeastern United States.

Research programs in mountain climatology

Our knowledge of the characteristics of mountain climates has been achieved by the painstaking research of many individuals. However, in recent decades, as in many branches of the atmospheric sciences, it has been realized that significant further progress could be made by employing large integrated research programs in addition to the work of individuals and small groups.

Some noteworthy early programs of the 1980s included ALPEX that was focused on the European Alps and ASCOT that was developed in the United States in order to investigate the potential of mountain areas for air pollution. In more recent years the Special Observation Period of the Mesoscale Alpine Programme (MAP) in the European Alps was completed in the fall, 1999. MAP was possibly the most comprehensive field program to date documenting the influence of mountains on the atmosphere. MAP is likely to provide data and the development of new theory in years to come. The following are the goals of MAP as outlined on the website of the programme

> (1a) To improve the understanding of orographically influenced precipitation events and related flooding episodes involving deep convection, frontal precipitation and runoff. (1b) To improve the numerical prediction of moist processes over and in the vicinity of complex topography, including interactions with land-surface processes. (2a) To improve the understanding and forecasting of the life-cycle of Föhn-related phenomena, including their three-dimensional structure and associated boundary layer processes. (2b) To improve the understanding of three-dimensional gravity wave breaking and associated wave drag in order to improve the parameterization of gravity wave drag effects in numerical weather prediction and climate models. (3) To provide data sets for the validation and improvement of high-resolution numerical weather prediction, hydrological and coupled models in mountainous terrain.

Another recent field program was the Intermountain Precipitation Experiment (IPEX). This program, focused in the US Rocky Mountains, also had detailed and complex goals as described by the program leaders.

> IPEX is a research program designed to improve the understanding, analysis, and prediction of precipitation over the complex orography of the Intermountain West of the United States. The goals of IPEX are to advance knowledge of the structure and dynamics of Great Salt Lake-effect and orographic precipitation, especially in and adjacent to the Wasatch Mountains of northern Utah; to better understand the relationships between orographic circulations and cloud microphysics; to verify and improve data assimilation, numerical weather prediction, and radar-derived quantitative precipitation estimates over the Intermountain West; to explore the electrical structure of continental winter storms; and to raise awareness of mountain meteorology and the associated scientific and forecasting challenges at the public, K-12, undergraduate, and graduate levels (Schultz et al., 2002).

Programs such as MAP and IPEX clearly demonstrate the areas of concern and depth of investigation in modern-day studies of mountain climatology.

The future outlook

In mountain climatology, as with many other fields, the more that is discovered, the more new questions arise. In addition, recent discoveries in mountain climatology have highlighted the increasing importance of the topic. For example, evidence suggests that late twentieth-century global-warming signals may be amplified in mountain climates. Temperature increases in parts of the European Alps of up to 2.0°C have been recorded. It has also been shown for the European Alps that temperature extremes shift by a factor of 1.5 for a unit shift in the global means of the twentieth century (Beniston et al., 1997). Beniston and colleagues have outlined the most pressing needs for further activities in the field. They advocate the encouragement of field studies including monitoring programs, an increase in the depth and breadth of the paleo-database, and more attention to the downscaling of climate models. The future appears to continue to be closely tied to the increasing human populations in and near mountains and the particular kinds of uses to which mountain lands are being put. The main activities include: (1) recreation; (2) timber production and watershed management; and (3) mining for mineral and energy production. Each of these activities carries with it particular, although sometimes interrelated, problems of mountain climatology.

Recreationists demand clean air and this, together with clean air legislation in some countries, provides the incentive for air-quality studies that examine pollutant transport, chemical reaction, and visibility studies. Avalanche potential prediction is another item of interest to recreationists. Mountain forest land management requires knowledge of mountain microclimate and climatic factors affecting growth rates and forest fire. Overall, whereas projection of future climates is a difficult task in itself, the task becomes even more complicated in mountain areas because of the complex nature of mountain climates. Increasing population growth, however, continues to raise the importance of understanding mountain climates.

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Bibliography

American Meteorological Society, 1981. *Second Conference on Mountain Meteorology*, 9–12 November. Steamboat Springs, Colorado. Preprints. Boston, MA: American Meteorological Society.

American Meteorological Society, 2000. *Ninth Conference on Mountain Meteorology*. Aspen, Colorado. Preprints. Boston, MA: American Meteorological Society.

Barry, R.G., 1978. DeSaussare H.B.: the first mountain meteorologist. *American Meteorological Society Bulletin*, **59**: 702–705.

Barry, R.G., 1992. *Mountain Weather and Climate*, 2dn edn. London: Methuen.

Barry, R.G., and Van Wie, C.C., 1974. Topo- and microclimatology. In Ives J.D., and Barry, R.G., eds., *Alpine Areas in Arctic and Alpine Environments*. London: Methuen, pp. 73–84.

Beniston, M. (ed). 1994. *Mountain Environments in Changing Climates*. London: Routledge.

Beniston, M., Diaz, H.F., and Bradley, R.S. (eds), 1997. Climatic changes at high elevation sites: an overview. In Diaz, H.F., Beniston, M., and Bradley, R.S., eds., *Climatic Changes at High Elevation Sites*. Dordrecht: Kluwer, pp. 1–19.

Brazel, A.J., and Outcalt, S.I., 1973. The observation and simulation of diurnal evaporation contrast in an Alaskan alpine pass. *Journal of Applied Meteorology*, **12**(7): 1134–1143.

Brinkmann, W.A., 1970. The chinook at Calgary (Canada). *Archiv für Meteorologie, Geophysik und Bioklimatologie*, **B18**: 269–278.

Cline, D.W., 1997a. Effect of seasonality of snow accumulation and melt on snow surface energy exchanges at a continental alpine site. *Journal of Applied Meteorology*, **36**(1): 22–41.

Cline, D.W., 1997b. Snow surface energy exchanges and snowmelt at a continental, midlatitude Alpine site. *Water Resources Research*, **33**(4): 689–701.

Christopherson, R.W., 2002. *Geosystems*. Upper Saddle River, NJ: Prentice-Hall.
Critchfield, H., 1983. *General Climatology*, 4th edn. Englewood Cliffs. NJ: Prentice-Hall.
Coulter, J.D., 1967. Mountain climate. *New Zealand Ecological Society Proceedings*, **14**: 40–57.
Daly, C., Taylor, G., and Gibson, W., 1997. The PRISM approach to mapping precipitation and temperature. *10th Conference on Applied Climatology*. Reno, NV: American Meteorological Society, pp. 10–12.
Dozier, J., and Outcalt, S.I., 1979. An approach toward energy balance simulation over rugged terrain. *Geographical Analysis*, **11**(1): 65–85.
Duguay, C.R., 1994. Remote sensing of the radiation balance during the growing season at the Niwot Ridge Long-Term Ecological Research site, Front range, Colorado, U.S.A. *Arctic and Alpine Research*, **26**(4): 393–402.
Dunbar, G.S., 1966. Thermal belts in North Carolina. *Geographical Review*, **56**(4): 516–526.
Garnier, B.J., 1958. *The Climate of New Zealand*. London: Edward Arnold.
Glassy, J.M., and Running, S.W., 1994. Validating diurnal climatology of the MT-CLIM model across a climatic gradient in Oregon. *Ecological Applications*, **4**(2): 248–257.
Greenland, D., 1973. An estimate of the heat balance of an alpine valley in the New Zealand Southern Alps. *Agricultural Meteorology*, **11**: 293–302.
Greenland, D., 1978. Spatial distribution of radiation in the Colorado Front Range. *Climatological Bulletin*, **14**: 1–14.
Ives, J.D., 1981. Editorial, *Mountain Research and Development*, **1**(1): 3–4.
Johnson, J.B., 1980. Mass balance studies on the Arikaree Glacier. In Ives, J.D., ed., Geoecology of the Colorado Front Range*. Boulder, Co: Westview Press,* pp. 209–213.
Kimball, J.S., Running, S.W., and Nemani, R., 1997. An improved method for estimating surface humidity from daily minimum temperature. *Agricultural and Forest Meteorology*, **85**: 87–98.
Kumar, L.A., Skidmore, K., and Knowles, E., 1997. Modelling topographic variation in solar radiation in a GIS environment. *International Journal for Geographical Information Science*. **11**(5): 475–497.
Leptuch, P., Brown, J.M., Bluestein, H.B., Thaler, E., and Richman, M.B., 2000. Forecasting downslope windstorms at Boulder, Colorado: the empirical–statistical approach revisited. *Preprints of Ninth Conference on Mountain Meteorology*. 7–11 August 2000, Aspen, Colorado. American Meteorological Society, pp. 105–108.
Liljequist, G.H., 1970. *Klimatologi*. Stockholm: Generalstabens Litografiska Anstalt.
Messerli, B., 2001. Opening remarks to International Year of Mountains 2002. http://www.iisd.ca/linkages/sd/mountains/
Miller, D.J., Brinkmann, W.A.R., and Barry, R.G., 1974. Windstorms: a case study of wind hazards for Boulder, Colorado. In White, G.F., ed., *Natural Hazards: local, global and national*. Oxford: Oxford University Press, pp. 80–85.
Mock, C., and Birkeland, K.W., 2000. Snow avalanche climatology of the western United States mountain ranges. *Bulletin of the American Meteorological Society*, **81**: 2367–2392.
Oke, T.R., 1987. *Boundary Layer Climates*, 2nd edn, London: Methuen.
Ollinger S.V., Aber, J.D., and Federer, C.A., 1998. Estimating regional forest productivity and water yield using an ecosystem model linked to a GIS. *Landscape Ecology*, **13**: 323–334.
Patterson, W.S.B., 1994. *The Physics of Glaciers*, 3rd edn, Oxford: Elsevier.
Reiter, E.R., 1982. Where we are and where we are going in mountain meteorology. *American Meteorological Society Bulletin*, **63**(10): 1114–1122.
Reiter, R., and Munzert, K., 1982. Values of UV and global radiation in the Northern Alps. *Archiv für Meteorologie, Geophysik und Bioklimatologie*, **B30**: 239–246.
Running, S.W., Nemani, R.R., and Hungerford, R.D., 1987. Extrapolation of synoptic meteorological data in mountainous terrain and its use for simulating forest evaporation and photosynthesis. *Canadian Journal of Forest Research*, **17**: 472–483.
Schultz, D.M., Steenburgh, W.J., Trapp, R.J., et al., 2002. Understanding Utah winter storms: The Intermountain Precipitation Experiment. *Bulletin of the American Meteorological Society*, **83**: 189–210.
Smith, A.A., 1982. The Mount Washington Observatory – 50 years old. *American Meteorological Society Bulletin*, **63**(9): 986–995.
Thornton, P.E., and Running, S.W., 1999. An improved algorithm for estimating incident daily solar radiation from measurements of temperature, humidity, and precipitation. *Agricultural and Forest Meteorology*, **93**: 211–228.
Trewartha, G.T., and Horn, L.H., 1980. *An Introduction to Climate*, 5th edn., New York: McGraw-Hill.
University Corporation for Atmospheric Research, 1980. ALPEX: upcoming GARP mountain subprogram. *UCAR Newsletter*, **4**(3): 1.
Wendler, G., 1971. An estimate of heat balance of a valley and hill station in central Alaska. *Journal of Applied Meteorology*, **10**: 684–693.
Whiteman, C.D., 2000. *Mountain Meteorology: fundamentals and applications*. New York: Oxford University Press.

Cross-references

Mountain and Valley Winds
Precipitation Distribution
Rainshadow