Temperature

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Temperature measures the intensity of heat within the physical world and thereby determines, influences, or limits most geographic features and processes. As primary examples, temperature (1) determines the solid, liquid, or vapor state of water, (2) influences heating and cooling needs within the built environment, and (3) limits the geographical range and distribution of many plant and animal species. Temperature also is a critical variable in analyzing climatic variability and change.

While heat refers to the total reservoir of kinetic energy of a substance, temperature refers to the average kinetic energy of a substance. As a comparative example, the water in a small lake in northern Wisconsin during summer will have a higher temperature than that of the water in nearby Lake Superior, but Lake Superior has a far greater overall heat content.

In many cases, what is referred to as "temperature" is the *air temperature*, usually measured at a standard height of about 1.5 m above the ground. Many other measures of temperature are routinely made and these include (1) profiles of air temperature from the near-surface to the upper atmosphere (made with instruments on weather balloons known as radiosondes), (2) sea-surface temperature (measured either by satellite sensors or by "in situ" sampling), and (3) soil or deep-earth borehole temperature profiles. The focus here primarily will be on near-surface air temperature; however, many of the concepts can be extended to other measures of environmental temperature.

Controls on temperature

Within the natural environment, air temperature is most closely related to the availability of solar radiation. Therefore, temperature has strong relationships with variables that influence solar radiation, such as latitude, cloud cover, aerosols, and surface albedo. The annual and daily cycles of solar radiation availability create corresponding cycles of temperature, although the peaks of the annual and daily temperature cycles lag the solar radiation cycles, as temperature is more closely related to net radiation than solar radiation alone.

The initial heating or cooling of air temperature near the Earth's surface occurs via conduction with the surface below. While incoming solar radiation (insolation) is the primary heat source for the atmosphere, reflection and longwave radiation emission by the surface offset solar input. These factors added together are known as net radiation. Net radiation indicates the overall amount of radiative heat gained or lost by the surface. A net radiation gain generally means warming and net radiation loss generally means cooling of the surface and, subsequently, of the atmosphere that is in contact with the surface.

The thermal climate of the Earth over recent millennia has varied only slightly, primarily because the amount of solar radiation reaching the Earth is relatively stable. In addition, the incoming and outgoing radiation from the Earth's surface as a whole are nearly in balance.

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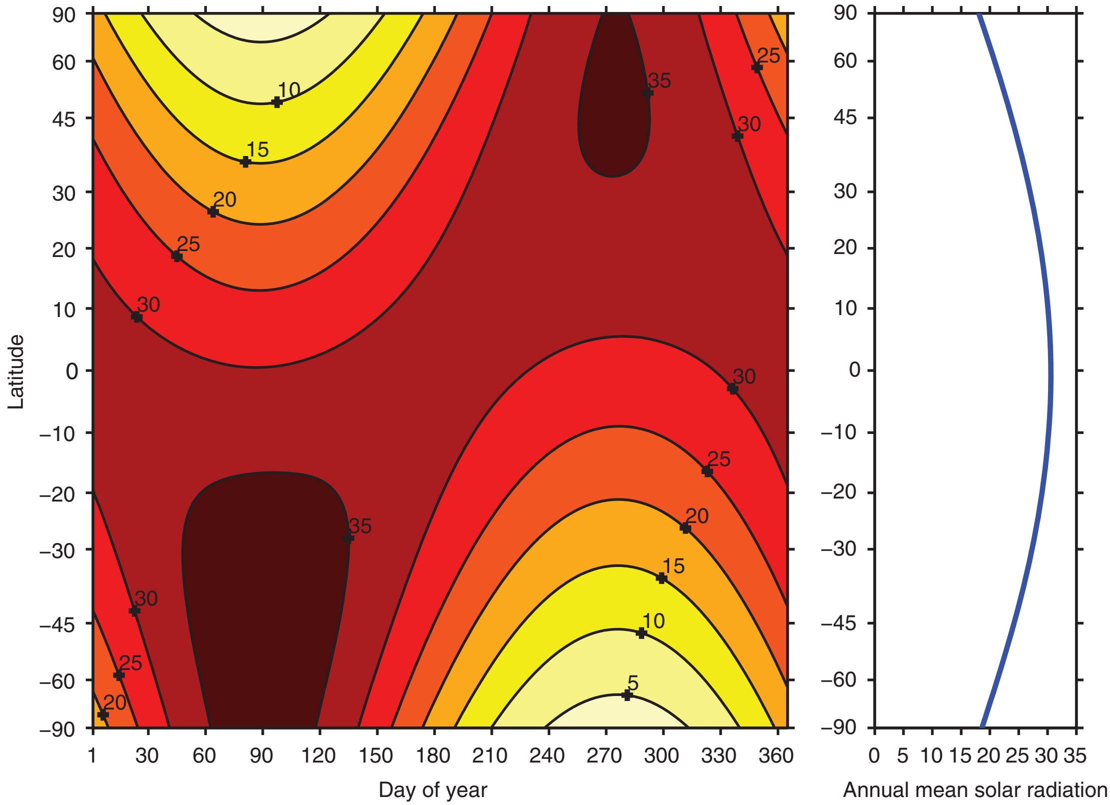

Figure 1 Time–latitude plot of top-of-the-atmosphere (extraterrestrial) solar radiation (left) and annual mean of top-of-the-atmosphere solar radiation (right), both expressed as a percentage of the solar constant. The tropics receive relatively uniform solar input, while many higher latitude regions have high solar inputs during summer months only when sun angles are higher and day length is particularly long. At high latitudes, substantially longer path lengths of radiation through the atmosphere (not shown) further reduce incoming solar radiation.

However, large variations in solar radiation occur daily and seasonally because of the geometry of the Earth and its orbital characteristics within the solar system. These characteristics – such as the Earth's orbit around the sun, axial tilt, and rotation, along with latitude – govern the angle at which solar radiation enters the Earth-atmosphere system. When totaled over the course of a day and expressed as a percentage of the solar constant, insolation at the top of the atmosphere at the poles varies from near zero during the winter solstice to over 30% at the summer solstices. Contrastingly, tropical areas receive nearly 30% of the solar constant at nearly all times of the year, resulting in year-round homogeneous temperatures (Figure 1).

By governing the amount of absorbed and reflected solar radiation and the amount of absorbed and emitted longwave radiation, characteristics of the Earth's surface and atmosphere exert a strong influence on temperature. The greatest influence of the surface on net radiation is albedo. Snow is the most effective natural reflector of sunlight, whereas most land surfaces and open water are efficient absorbers. Even small differences in land-surface albedo, however, can be great enough to cause large spatial variations in temperature during sunny days.

Emissivity is another important surface property because of its influence on longwave radiation. The amount of longwave radiation emitted by a surface (E) is dependent on both its temperature (T, in Kelvin) and its emissivity (ε) and is given according to the Stefan–Boltzmann law as $E = \varepsilon\sigma T^4$, where σ is a constant (5.67×10^{-8} W m^{-2} K^{-4}). Emissivity varies from 0 to 1, with natural surfaces generally having higher values than man-made surfaces (see Oke 1987, table 1.1). The effects of emissivity are at work 24 hours a day, whereas albedo is only a factor when sunlight is present. As a result, emission of longwave radiation plays a particularly important role in cooling processes at night but is overshadowed by the effects of solar radiation during the day. Higher nighttime temperatures often occur in areas with lower emissivity than adjacent areas where it is higher. Snow has a very high emissivity, which, combined with its high albedo and ability to insulate the atmosphere from ground heat below, makes it a very efficient cooling device in the climate system. Inversion of the Stefan–Boltzmann law to solve for temperature, $T = [E/(\varepsilon\sigma)]^{1/4}$, is used to estimate land- and sea-surface temperature from satellite- or aircraft-based sensors that measure longwave (thermal infrared) emission.

Like the Earth's surface, the atmosphere has a variable albedo and emissivity. Clouds and other atmospheric constituents, such as small atmospheric particles or liquid droplets known as "aerosols," typically increase albedo and act as cooling processes. Conversely, greenhouse gases – such as water vapor, carbon dioxide, and methane – affect the absorptivity and emissivity of the atmosphere and act as warming processes. Even though greenhouse gas concentrations are relatively low, they have important direct radiative effects that are compounded by positive feedback mechanisms – processes that have acted to make Earth substantially warmer and more habitable than it would be without greenhouse gases. As a result, changing concentrations of greenhouse gases and atmospheric aerosols are very influential in regional and global climate change.

Water vapor is the most influential of the greenhouse gases in the atmosphere because of its high concentration compared to the others. Water vapor that is converted to liquid or solid form in the form of clouds, however, typically increases atmospheric albedo. The net effect of clouds on near-surface temperature is dependent on their optical thickness and their height. For instance, high cirrus clouds tend to lead to higher temperatures over clear-sky conditions because of their transparency to solar radiation and ability to absorb longwave radiation and re-emit it to the surface. Low clouds tend to be less transparent and have a higher albedo, leading to lower temperatures compared to clear-sky conditions. In addition, natural and anthropogenic aerosols and particles can linger for hours to days and lead to cooling. Volcanic eruptions can result in large-scale atmospheric albedo increases and temporary global cooling. The most influential of these volcanoes inject gases and particles into the stratosphere where they persist for months to several years.

The physical geography of the Earth's surface affects temperature spatially through topography and the distribution and configuration of land and water bodies. An important property of water is its role in dampening temperature variation. This is because of its high specific heat, which is about five times that of most land surfaces. Water absorbs energy when there is a surplus of net radiation and re-releases it when there is a deficit of net radiation, acting to decrease the rate of temperature change relative to surfaces with lower specific heats.

Water also plays a key role in modulating temperature via latent heat exchange. When water changes phase, an exchange of energy is involved that can offset background temperature tendencies. Phase changes from a lower energetic state to a higher one – such as melting or evaporating – consume latent heat energy so the warming that leads to the phase change is offset. Evaporative cooling is a good example: forests or other vegetated areas often have a lower temperature than adjacent nonvegetated areas because the transpiration process lowers the air temperature as water is evaporated. Conversely, phase changes from a higher energetic state to a lower one – such as condensing or freezing – release latent heat energy so that the cooling is offset. Citrus fruit growers make use of this process by spraying liquid water when temperatures approach the freezing point. The freezing process adds latent heat to the atmosphere and results in a local warming of the air that protects the fruit crops. Through these latent heat processes, there is no resultant change in total energy in the system, despite a change in temperature. As a result of these properties of water, locations that are near or downwind of water bodies, or those where atmospheric humidity is high from evaporation or transpiration from vegetation, have lower annual and diurnal variations in temperature. Semiarid and arid regions have a very high diurnal temperature range because of the lack of water vapor, vegetation, and transpiration.

Differential heating of the Earth's surface and the resulting temperature changes give rise to wind and air circulation from local to global scales. Cold air is denser than warmer air, which means cold air tends to displace warm air vertically when they are adjacent. At the global scale, transport of colder air from the poles equatorward and warmer air in the tropics poleward serves to alter temperature from that which net radiation alone would produce at the surface. Atmospheric circulation also influences and interacts with ocean currents, which exchange large amounts of heat latitudinally. In addition to the latitudinal gradients of ocean temperature, the orientation of the ocean basins and continents helps to direct ocean currents and produce the patterns that we see today. For instance, the Gulf Stream transports warm water from the tropics to Western Europe, helping to produce a climate that is much warmer than areas at similar latitudes in eastern Canada or eastern Russia where colder ocean currents exist (Figure 2).

Topography has a strong influence on temperature because of the effects of altitude on pressure (and therefore temperature), but surface feature shapes and sizes also play a role. The slope and aspect of a surface can alter the incidence angle of solar radiation, such that slopes that face away from the sun are cooler than slopes facing toward the sun. This effect can be seen in mid-latitude regions where steeper slopes facing equatorward are warmer and often have vegetation indicative of a drier and warmer climate than slopes facing poleward, primarily because evaporation is directly related to temperature.

At higher elevations, the mean temperature is cooler because atmospheric pressure is lower, and therefore there is less molecular kinetic energy at these altitudes. However, because the atmosphere is thinner, greater insolation occurs at high elevations during the day as there is less atmosphere above to reflect or scatter insolation. On the other hand, there is a greater loss of longwave radiation at night because the thinner atmosphere does not absorb and return as much outgoing longwave radiation. These processes often result in a greater diurnal temperature range at high elevations. Factors like persistent clouds that limit net radiation or winds in exposed mountain

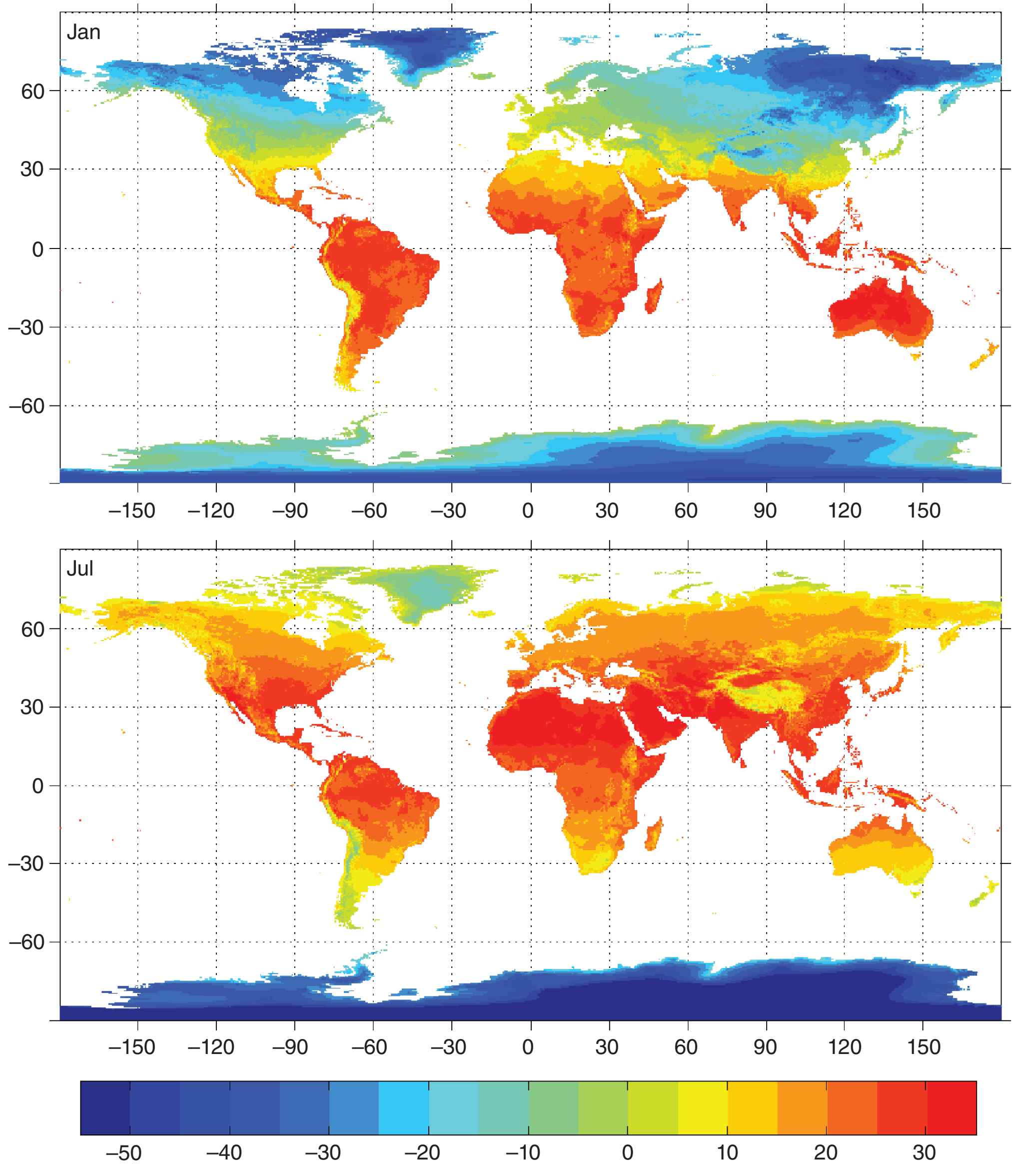

Figure 2 Maps of monthly mean air temperature over the terrestrial surface during January and July (°C). Data from Willmott and Matsuura (2009).

slopes that enhance atmospheric mixing, however, can reduce the diurnal temperature range.

Topography and elevation can work together to increase the diurnal temperature range. Valleys and enclosed basins tend to collect cool air at night, since it is denser than warmer air and thus drains to topographic lows. During the day, surface heating usually mixes this chilled air and then is further warmed from positive net radiation. These conditions require clear skies and dry air for a greater temperature variation with the presence of wind, clouds, or high water vapor concentrations dampening the effect. Valley environments can experience large seasonal temperature differences from surrounding areas, especially where winter snow or enhanced cloud cover occurs. The effects of snow and cloud on surface and atmospheric albedo give rise to the extreme annual temperature variation seen in nonglaciated cold environments, such as the interior regions of Asia and North America, where atmospheric humidity is normally low.

Global patterns of air temperature

Global-scale maps of monthly mean values of near-surface air temperature clearly show the influence of latitude, topography, and proximity to water bodies discussed above (Figure 2). In response to solar radiation variations, isotherms (lines of equal temperature) tend to be predominantly east–west. Major mountain ranges such as the Andes in South America, the Himalayas in Asia, and the Rockies in North America are very influential in global-scale patterns of temperature. Beyond the expectation of lower temperatures at high latitudes, the high elevations of the ice sheets on Greenland and Antarctica act to further reduce air temperature substantially. Some of the largest spatial gradients of air temperature occur in North America and eastern Asia during January – primarily because solar radiation gradients are largest in the winter hemisphere and both of these continents have vast interiors that are distant from oceanic sources of heat and water vapor. North America and eastern Asia also have substantial snow cover at higher latitudes, which further cools northern regions relative to southern ones. A quantitative example from North America shows that the rate of change of air temperature with latitude during January is about 1.0 to 1.5°C per degree of latitude, whereas July air temperatures decrease only 0.4 to 0.6°C per degree of latitude (Figure 3). These latitudinal rates of change of air temperature are the cause of variability in weather systems across the continent, but also are influenced by the same weather patterns, illustrating the nonlinear nature of the climate system.

Temperature and climate change

Changes and variability in near-surface air atmospheric temperature are among the most important and tangible indicators of climate change. Reliable liquid-in-glass thermometry developed in the mid-eighteenth century produced relatively consistent instrumentation during the historical record. High-quality data from microwave-based satellite sensing of temperature are available from about 1979 onward (Hurrell and Trenberth 1996). As temperatures have a wide range of variability, the analysis of temperature changes with time is simplified by averaging it over space and time. Global and hemispheric-scale averages of monthly or annual averages of temperature are typically used to monitor the energetic state of the near-surface environment. Often, these large-scale spatial averages are integrations of air temperature over land and sea-surface temperature over the oceans.

Latitudinal and topographic influences on temperature are usually considered to be independent of climate change. As a result, to reduce the spatial variability associated with latitude and topography, air-temperature anomalies are created by removing a local mean from an air-temperature time series. Converting air temperatures to anomalies produces a smoother spatial field while preserving the temporal variability of each location. The temporal variability of nearby locations that are at very different elevations,

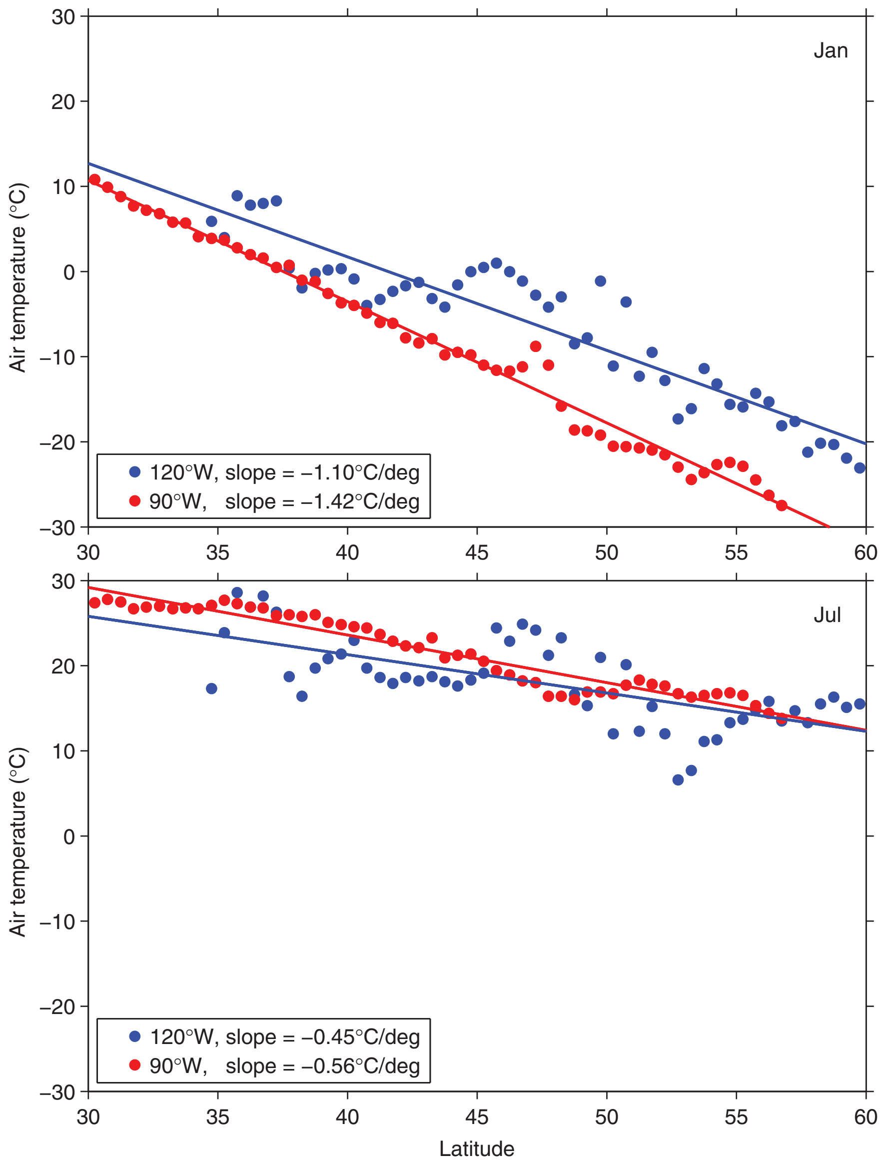

Figure 3 Latitudinal transects of monthly mean air temperatures during January and July at 120°W and 90°W in North America (same data as Figure 2). The gradient of air temperature with latitude (slope) in January is 2–3 times that of July. In both months, the more westerly transect (120°W) shows a weaker latitudinal gradient due to its closer proximity to the Pacific Ocean. Data from Willmott and Matsuura (2009).

for instance, becomes much more comparable after conversion to anomalies (Figure 4). Once temperatures are converted to anomalies, they can be interpolated to a regular grid and averaged accordingly. Analysis of global-scale air-temperature anomalies show that global

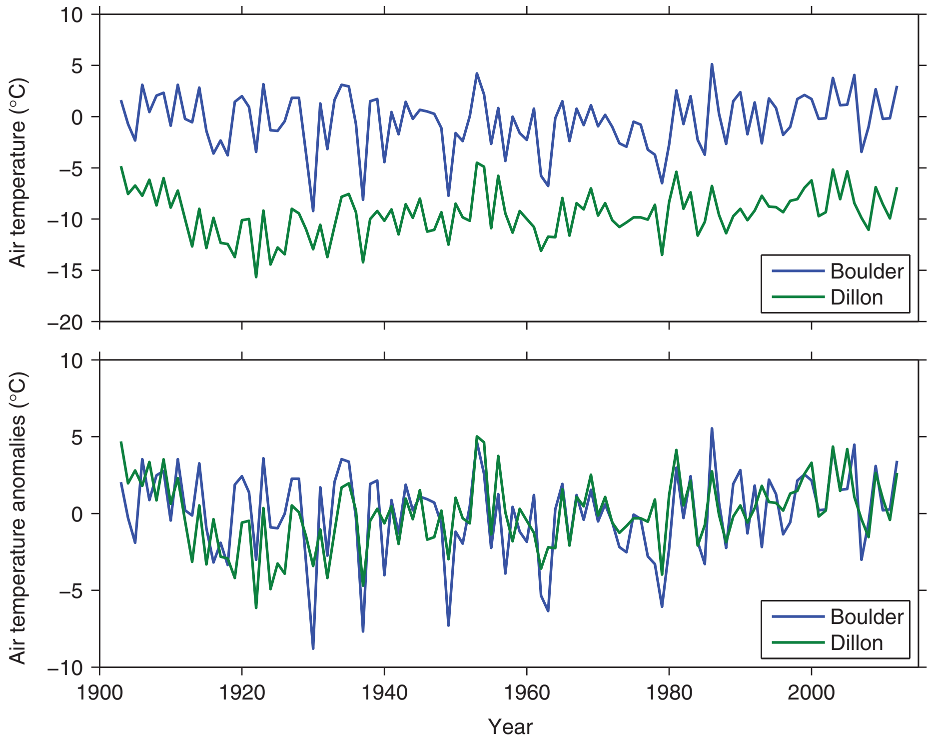

Figure 4 Time series of January mean air temperature (°C) at two locations in Colorado, USA, from 1903 to 2012 (data from Historical Climate Network, National Climatic Data Center). Boulder is located at an elevation of 1672 m while Dillon is at 2763 m. The two locations have January mean air temperatures that differ by approximately 9°C, but removing the long-term January mean produces anomaly time series that have similar interannual variability. As a result, air-temperature anomalies are used for comparative analysis and to estimate climatic change across regions that have wide ranges in elevation or latitude. Data from Historical Climate Network, National Climatic Data Center.

and hemispheric rates of recent warming range from 0.05 to 0.25°C/decade (Figure 5; Morice *et al*. 2012). Concentrations of greenhouse gases are increasing and they are mixed well in the atmosphere, but changes in temperature in the historical record have been larger during nights and winters, especially at high latitude where net radiation often is negative (Meehl *et al*. 2009; Robeson 2004; Seneviratne *et al*. 2012).

Human alterations of land cover also influence temperature, primarily by changing surface albedo, emissivity, and the availability of surface water for evaporation and transpiration (Pielke *et al*. 2011). For instance, many urban heat islands arise from the introduction of materials and urban geometry that result in increased absorption of solar radiation, decreased longwave radiation loss, and decreased evapotranspiration relative to natural landscapes. These processes, along with anthropogenic waste heat, frequently combine to increase air temperature over cities, with the effect usually being larger at night (Oke

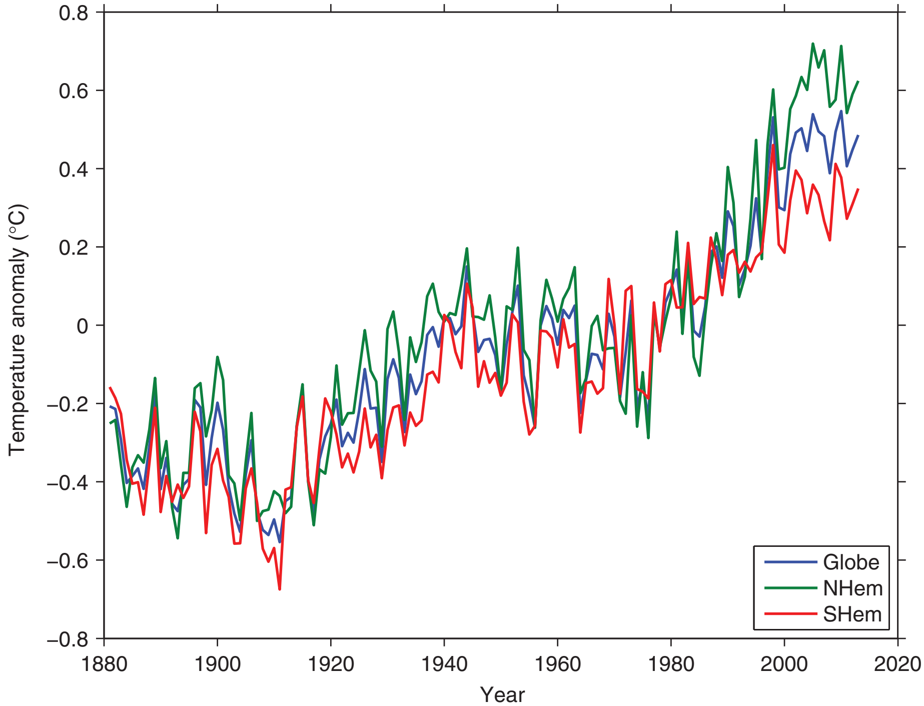

Figure 5 Time series of globally and hemispherically averaged temperature anomalies from 1881 to 2012 (HadCRUT4 data from Hadley Centre at UK Met Office and Climatic Research Unit at University of East Anglia). Both hemispheres have similar interannual variability, but the Northern Hemisphere has warmed slightly more rapidly in recent decades.

1987). Other larger-scale anthropogenic changes to the land surface include deforestation and agricultural activities. The direction in which temperature is forced is dependent on what landscape was altered because albedo, emissivity, and evapotranspiration can be either increased or decreased. For instance, irrigated farmland in a desert environment can lower the temperature through increased evapotranspiration. Deforestation can increase daytime temperature and decrease nighttime temperature through decrease in evapotranspiration and subsequent water vapor concentrations. Overgrazing removes vegetation and exposes soil, which reduces transpiration and water vapor while also lowering emissivity, with the net effect being higher daytime and nighttime temperatures.

SEE ALSO: Agroclimatology; Arid climates and desertification; Atmospheric aerosols; Atmospheric/general circulation; Climate change, concept of; Climate and societal impacts; Climatology; Earth's energy balance; Global climate change; Global dimming/brightening; Hydrologic cycle; Lake climates; Land-use/cover change and climate; Microclimatology; Mountain climatology; Oceans and climate; Paleoclimatology; Polar climates; Snow cover changes; Urban climatology; Water and climate change

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