

Chapter 2

Current Climate and Recent Trends

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2.1 Introduction

While the paleoclimatic record is based on indirect measurements—for example, biological and geological indicators—more recent climate history is defined largely by in situ observations over the last 100–150 years. This chapter centers on what these observations and theoretical understanding reveal about the climate of North America. The focus is on: the current climate and the physical and geographical features that regulate it; natural climate variability on time scales of weeks to many decades; and recent trends in key climate variables such as temperature and precipitation, snow cover, and sea ice.

2.2 The Climate of North America

2.2.1 *Physical Influences*

The continent of North America encompasses approximately 24.3 million km², or about 4.8 % of the Earth's surface and 16.5 % of its land area, and extends from about 7 to 83°N in latitude. Thus, parts of the continent experience tropical, desert, temperate, and polar climates. The latitudinal and seasonal distribution of solar energy exerts the predominant control on North American climate. Along a south-to-north latitudinal transect a variety of climates occur, all influenced by different components of the general circulation of the atmosphere including subtropical easterlies, mid-latitude westerlies, and a more mixed wind regime in the far north.

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The Inter-Tropical Convergence Zone (ITCZ) varies by season in this sector between 5° and 10°N, and thus extends northward into the southernmost portion of North America in summer (Hastenrath 2002). North of the ITCZ, the descending branch of the Hadley circulation limits precipitation and results in generally arid conditions from near southern Mexico to southern Nevada. The driest areas are between 25° and 35°N. Arid conditions extend into southern Canada in the western portion of the continent as a consequence of rain shadows that occur in the lee of prominent north–south oriented mountains. Mid-latitude westerlies consisting of progressive frontal cyclone waves and quasi-stationary continental-scale waves influence most of the continental United States and Canada. These waves result in often dramatic changes in energy and moisture regimes within a season and their mean track moves with the annual cycle, northward during the summer and southward during the winter. The northern edge of the boreal forest extends from southern Hudson Bay (54°N) in the east to nearly the Arctic Circle (67°N) in the west.

East–west differences across the continent are present at almost all latitudes, a consequence of two main influences. The first influence consists of differing oceanic influences on the western and eastern coastlines. Along the western continental US coast, currents differ between winter and summer. During summer, south of about Vancouver Island, the Pacific Subtropical High and associated clockwise ocean gyre lead to the cold southward-moving California Current, with strong upwelling a few tens of kilometers west of the shoreline, which cools the ocean and the adjoining land. The California Current extends south to about the Baja Peninsula. In winter, the narrower and weaker Davidson Current flows north along the US West Coast. In summer and winter, North of Vancouver Island the Alaska Current flows counterclockwise around the gyre as a northward flow along the continental shoreline. By contrast, along the eastern coast of North America, the northward-moving Gulf Stream brings warm water and moist tropical air into the mid-latitudes throughout the year.

The second influence consists of a nearly unbroken combination of mountains and high plateaus that stretch from southern Mexico to nearly the Arctic Ocean along the western side of the continent. These mountains cause substantially more precipitation to fall on their windward than their leeward sides. On windy days compressional heating of the descending air on the leeward side of the mountain increases temperature and decreases relative humidity, further reducing cloud cover and increasing aridity. Mountain ranges help to geographically define and steer air masses. Maritime air masses tend to influence the windward side of mountain ranges providing a moderating influence on seasonal and diurnal temperature variations, whereas continental air masses tend to exist east of the Rockies and undergo larger amplitude diurnal and seasonal changes in temperature. Mountain ranges also modify progressive mid-latitude weather systems and enhance the cyclonic (counter-clockwise) spin in the atmosphere on the leeward side of topography, particularly in the presence of warm and moist air. Such conditions can result in lee side cyclone formation in the Great Plains and facilitate conditions conducive to severe weather from the mountains eastward to the plains

In addition to these large scale patterns, several factors lead to considerable variations in the detailed spatial “texture” of climate. Foremost among these is the widespread presence of complex topography, followed by coastal effects of oceans and lakes, and increasingly, human alterations of land cover properties. The human population now numbers about 530 million (USCB 2011). As will be elaborated, topographic complexity affects not only the fine-scale structure of basic climatology, but also can lead to fine-scale structure in the temporal characteristics of climate variability, including climate change. Vegetation is an important factor in determining the communication between atmosphere and substrate, via vertical fluxes of energy and water, and until the arrival of humans was mostly controlled by climate (and somewhat by geology as well).

2.2.2 Data Sources

This chapter draws upon a wide variety of material and data to describe the present climate, much of which has been well understood and available for at least half a century (e.g., Bryson and Hare 1974). For discussions of variability and change, source material consists of in situ data from individual stations and from gridded surface data sets. The latter are intended to produce spatially and temporally complete representations of fields of climate data (at the scale of the gridding).

The Climatic Research Unit (CRU) at the University of East Anglia produces numerous global climate datasets that cover the modern climate record (Brohan et al. 2006). For our purposes, we draw upon a high spatial resolution product of monthly temperature and precipitation database (CRU TS 3.1) on a 0.5° grid from 1901 to 2009 compiled from nearly 4,000 weather stations globally. Data coverage in North America is good in the continental US, southern Canada, and parts of Mexico, and is less dense in the Arctic and central Mexico.

Surface based observations are supplemented with three-dimensional “reanalysis” data. Reanalysis is a process to derive spatially and temporally interpolated fields of common climate descriptors (temperature, humidity, wind speed and direction, solar radiation, and pressure) by assimilating available observations from radiosondes, aircraft, surface barometers, satellites, and special research activities, with a numerical model to create an internally consistent state of the atmosphere. In general, reanalysis does not assimilate measurements of surface temperature or precipitation, but rather derives these through numerical model output. Therefore, reanalysis values provide a set of observations nearly independent of surface based observations. Operational models evolve over time, and thus the manner in which they process observational input to produce model output varies from one generation to the next. To avoid spurious changes introduced exclusively by advances in models and data processing, a selected model is “frozen” in development, and initialized with the entire history of input, which has been saved in raw form (see Kalnay et al. 1996; Mesinger et al. 2006). That is, the entire history is “re-analyzed” with a modern model. Several such reanalysis

efforts have been undertaken around the world. The National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis product has been run in a fixed mode every 6 h from 1948-present and was selected based on its longevity over other reanalyses that span a much shorter time period. Reanalysis output can suffer from changes in observational systems, and is often too coarse to capture local-scale features.

2.2.3 Description of Current Climate

In one sense, climate can be considered to consist of the complete set of statistical characteristics of the atmosphere, including the entire probability distribution with its extremes, autocorrelations in time and spatial pattern correlations, and the timing and sequencing of events, at all time and space scales, and the seasonal cycle of all of these characteristics. In general we have patchy and piecewise knowledge of this full matrix of desired information. We here briefly touch on long term means and how those vary in space and time and through the year and the day.

In broad terms temperature tracks the supply of net radiation (the combined shortwave/longwave budget). Annual mean temperature is generally highest in the lowest latitudes and at lower elevations. On monthly scales the highest temperatures are thus found in the subtropical deserts in summer and the lowest temperatures in winter at high latitudes or high elevations. The diurnal (day-night) difference in temperature is greatest away from coasts and at higher elevations where downward longwave radiation (which impedes nocturnal cooling) is less.

Unlike temperature, precipitation is discontinuous in space and time. In most locations a single season or portion of the year dominates. In some areas, such as the northeastern United States (notably Central Park in New York), monthly average precipitation exhibits very little seasonality. However, precipitation seasonality in most regions is unimodal and varies gradually across space. For example the Great Plains has a summer maximum (convective heating and northward advection of moisture from the Gulf of Mexico), and the West Coast south of the Alaska Archipelago has a winter maximum (storm track shifts south). Other locations have two or more peaks within the year. For example, the desert Southwest of the United States receives frontal precipitation in winter during the maximum southern excursion of the jet stream; in summer a second season of precipitation is associated with continental heating and the North American Monsoon. Within Colorado, with its complex topography, and geographic setting permitting multiple trajectories for moisture inflow, the climatologically wettest month is spread among at least 10 separate calendar months (Redmond 2003). In mountainous terrain, seasonality can be very different from valley to summit over very short horizontal distances (Farnes 1995). The confluence of the general circulation and complex topography produce a very wide range of mean annual precipitation across North America, most notably in the western half of the

continent. Precipitation in North America varies from about 50 mm in Death Valley to about 8000 mm in the Olympic Mountains of northwest Washington, and up to 11000 mm near Glacier Bay Alaska (PRISM estimates, Chris Daly pers. communication).

2.3 Climate Variability

Fluctuation is inherent to climate. Atmospheric fluctuations occur on all time scales, and extend from turbulence to “weather” time frames to geological eras. Climate statistics do not appear to exhibit statistical stationarity on any time scales (Bryson 1997; Milly et al. 2008). Bryson stated this as an axiom: “The history of climate is a non-stationary time series.” The term “climate variability” is generally reserved for temporal, rather than spatial, usage, a convention followed here. Climate variability has sources in external drivers (e.g., solar output, volcanic eruptions, orbital mechanics, atmospheric composition, ocean-land configurations, and land surface alterations) and in internal dynamics associated with a multitude of feedbacks and subsystem interactions not forced by aforementioned external drivers. The term “natural variability” is sometimes employed to describe unforced internal dynamics. The interactions among climate subsystems can occur locally or act at a distance (as “teleconnections”) via the agency of Earth’s two main fluids, air and water. Some internal climate modes are temporally regular, even oscillatory, in nature, and others are chaotic, highly non-linear, and almost completely unpredictable. Temporally lagged behavior and interactions are also common. The climate system and its many subsystems are highly interconnected, in space and time, and it is essentially impossible to completely understand the variability at a chosen point even with a lengthy and accurate climate history from only that point.

Climate can be alternatively thought of as both an “enabler” of weather, and as a consequence of weather. Climate and weather are inseparable, and neither can be understood without reference to the other. Traditionally, in mathematical terms weather is often considered to be an initial value problem, and climate a boundary value problem. However, boundary values on some time scales are initial values on other time scales. Similarly “change” in climate is merely slow variability on longer time scales, reflected in “change” in statistical descriptors of weather. This is why it is important to understand sources of “natural” multi-year and decadal variability, to form a perspective from which human and other external effects can be viewed, and hopefully disentangled.

This variability through time has an infinitude of small and large consequences to the biosphere and ecological systems. As such it is of primary interest to a wide array of resource managers. This is especially true in the large parts of North America that are under public stewardship, such as the western US (Mote and Redmond 2011), where half the land area is in such status.

2.3.1 Spatial Scale Considerations

Central tendencies of climate (e.g., means or totals) vary with spatial scale. A general expectation among climatologists is that spatial scales of temporal variability will be on the order of tens, hundreds, and even thousands of kilometers. Climate “anomalies”, defined as a departure from long-term means, should show broad coherence over the scale of the driving mechanisms. These are often associated with teleconnection patterns that are continental in scale, reflecting fluctuations in the position and strength of the jet stream and subsequent continental-scale wave patterns. However, fine scale spatial structure in climate variability, even at longer time scales, is also quite plausible and even likely in the presence of topographic diversity, coastlines, glacial features, and urban settings. Furthermore, the strength of spatial correlations may differ between day and night. The atmosphere just above the surface tends to become decoupled at night, when vertical stability is greater, and nighttime minimums are more likely controlled by local factors. Fine scale structure in variability properties is not emphasized further in this chapter, but should be closely borne in mind because individual station measurements used to assess broad patterns may be subject to very local influences on the scale of tens to hundreds of meters. The determination of variability characteristics requires long (several decades) time series encompassing several “cycles” of the variability of interest; unfortunately, dense long-term homogeneous observational data sets in complex terrain are exceedingly scarce.

2.3.2 Temporal Scales of Variability

As noted earlier, the climate system exhibits variability over a very large range of time scales. Without records covering very long (geologic) periods of time at high temporal resolution, we cannot definitively characterize the complete spatio-temporal variability properties of climate. For this chapter we are primarily concerned with variability on time scales spanning a few weeks or intraseasonal (fluctuate within a season), interannual (fluctuate year-to-year) and multidecadal (fluctuate over a 2–3 decade time period) intervals. Tectonic and orbital mechanisms change little over these time frames.

2.3.3 Hemispheric Patterns Relevant to North America

Several recurrent patterns of climate variability affect the North American continent (e.g., Barnston and Livezey 1987). Because of this recurrence, the temporal evolution of those patterns is often described as an “oscillation.” This term connotes a kind of regularity that is often not fully justified; a true oscillation typically carries within itself physical factors that in effect constitute the seeds of

Table 2.1 Climate oscillations, associated characteristic timescales, geographic foci, means through which they are quantified and approximate time horizon for skillful prediction

Climate mode	Characteristic time scale	Centers of action	Indices	Predictability
ENSO	2–7 years	Tropical eastern pacific and subtropical jet streams	SOI (atmosphere), NINO3.4 (ocean temp), MEI (combined)	<6 months
PNA	10–14 days	Quadrapole pattern over N. Pacific and N. American continent	PNA index (atmosphere)	7 days
AO/NAO	10–14 days	Dipole of pressure/height between Arctic and midlatitudes	AO/NAO indices (atmosphere)	7 days
PDO	40–70 years	North Pacific (midlatitude focus)	North Pacific (ocean temp)	N/A
AMO	40–70 years	North Atlantic (subtropical focus)	North Atlantic (ocean temp)	N/A
MJO	30–70 days	Tropical Indian and Pacific Oceans	Tropical convection and divergence (atm)	2 weeks

destruction for whatever phase it is in. However, the terminology is well established. Each pattern has different characteristics such as geographic extent, similarity of influence across seasons, and predominant expression at one or more characteristic time scales. Of importance to understanding and attribution, some of these time scales are long enough to appear as “trends” in observational records that only span a few decades. The temporal scale of these patterns can be traced back to the underlying mechanisms. For example, ocean–atmosphere patterns grow, evolve, and decay much more slowly than purely atmospheric patterns. In some cases these features and patterns appear to have predictive value (prognostic), and in other cases appear to have only descriptive value (diagnostic). These patterns are not mutually exclusive, but rather interact with one another, leading to reinforced or counteracting effects in specific locations and months. For long period oscillations (50–70 years), the observational record only covers about one and one-half oscillations, and further corroboration as to their continuous existence within the climate system must be obtained from indirect paleoclimate evidence.

Several major sources of variability that could affect North American climate (Table 2.1) are next discussed. Then, the attendant effects on climate are subsequently described where this has been at least partly established.

2.3.3.1 El Nino/Southern Oscillation

El Nino and La Nina are the warm and cool phases, respectively, of ocean surface temperatures on, and a few degrees of latitude either side of, the equator between

Peru and approximately the International Date Line. The terms, El Nino and La Nina, refer *exclusively* to ocean temperatures, in terms of departures from long term averages. The Southern Oscillation (SOI), by contrast, is defined *exclusively* as an atmospheric phenomenon. The SOI consists of a see-saw in sea level pressure departures from the long-term average between a wide area centered approximately on Tahiti and another area centered on northern Australia. Though the two measures are distinct, they are physically interlinked, and correlated in a manner that does slowly vary through time (McCabe and Dettinger 1999), so that the names are merged in the common acronym El Nino-Southern Oscillation (ENSO). Reflecting terminology origins, ENSO is said to be in the warm phase when positive ocean surface temperature departures are present in the ENSO region, and when the SOI is negative.

A third measure of ENSO status is in wide usage as well. The Multivariate ENSO Index (MEI) (Wolter and Timlin 1998), based on co-variability among six descriptors that correlate with El Nino and with each other, is derived from a principal components analysis of pressure, north-south and east-west wind, ocean and air temperatures, and cloudiness across the Pacific. The MEI is positive during El Nino. Figure 2.1 shows the time series of cool season MEI and associated temperature and precipitation patterns in North America.

ENSO is the largest source of interannual global climate variability and the most widely studied such phenomenon (e.g., Bjerknes 1969; Rasmussen and Carpenter 1982; Livezey et al. 1997). Tropical ocean surface temperatures and atmospheric pressure across the Pacific work together through positive feedbacks involving the east-west “Walker circulation” and tropical convection to create this coupled atmosphere-ocean pattern. Together, the two patterns vary with an irregular period of typically 2–7 years, and are often not predictable with current knowledge beyond around 6 months.

Though the primary centers of action for ENSO are in the tropical Pacific, its influence extends well beyond this area in longitude and in latitude. Alterations in the strength and position of the subtropical jet stream in the Pacific sector lead to large scale influences in climate patterns across the Northern Hemisphere. The influence of ENSO on North American climate is most prominent in the cool season, affecting the entire West Coast from Alaska to Baja, the US/Mexico border region eastward to Florida, and the Ohio River Valley (e.g., Redmond and Koch 1991; Livezey et al. 1997). There are also influences on tropical storm activity in the tropical Atlantic and eastern Pacific basins in the late summer and into autumn.

ENSO Effects The most pronounced is the north-south dipole pattern of temperature and precipitation anomalies centered over the western continental US in winter. This pattern of response begins in early October and lasts through the end of March. El Nino is preferentially associated with wet winters in the American Southwest (extending southward to near southern Baja Peninsula), and with dry winters in the Pacific Northwest and southwest Canada, essentially the drainage basin of the Columbia River. To the north, this pattern again switches sign around the Queen Charlotte Islands, with wet conditions seen along coastal Alaska from about Yakutat to Kodiak Island. The effects extend inland to the crest of the Sierra

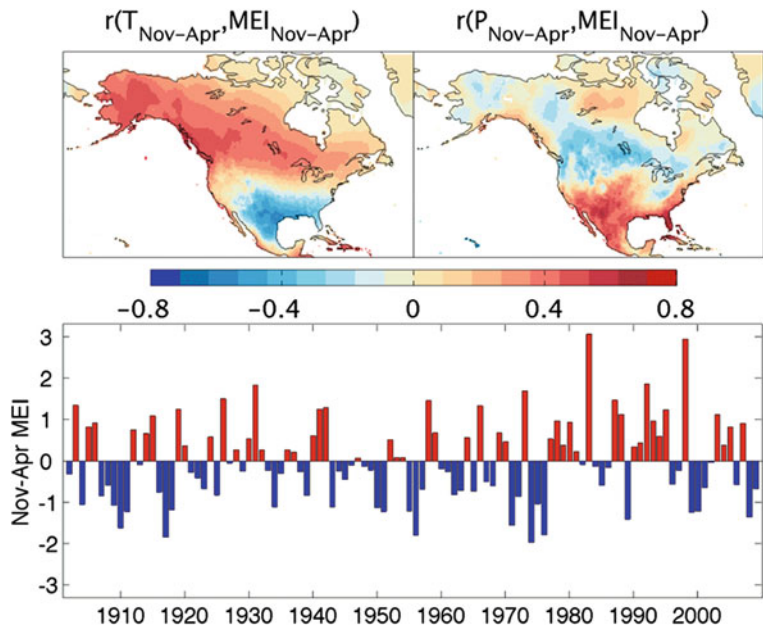


Fig. 2.1 *Top* Spatial correlation between Nov–Apr Multivariate ENSO Index (*MEI*) and average Nov–Apr temperature (*left*) and total Nov–Apr precipitation (*right*), 1901/02–2008/09. *Bottom* Time series of MEI for Oct–Feb from 1901/02 to 2008/09. Standardized units. Data from CRU TS 3.1

Madre and the Rockies, and in Alaska inland to the crest of the Alaska Range. Along the US–Mexican border, El Nino is associated with wet winters from southern California across to the San Juan Mountains of Colorado and eastward to Georgia and Florida. Effects in the US Southeast are nearly as pronounced as they are in the desert Southwest. The Ohio River Valley is usually drier during El Nino. Nearly the opposite effects are seen with La Nina. In the Southwest El Nino leads to an increase in the number of days with precipitation and in the average precipitation per wet day (Woolhiser et al. 1993), and vice versa for much of the Pacific Northwest.

The effects of ENSO just noted relate mostly to large scale winter storms and particularly influence water resources in the western United States because the majority of its annual precipitation occurs during the cool season. A few effects of ENSO can occur during the warm season. Summertime associations with La Nina have been noted in the upper Midwest United States (Kahya and Dracup 1993; Trenberth et al. 1988). El Nino tends to increase the upper level winds from the west over the tropical Atlantic and Caribbean Sea, and the resulting increased vertical wind shear helps suppress tropical storm development in summer and autumn. In the Eastern Pacific west of Mexico, normally nearly twice as active as the Atlantic, tropical storms become more numerous with El Nino, and a few more

of them recurve into the desert Southwest. Late summer in southern Mexico and Costa Rica generally experiences drier conditions with El Nino (Ropelewski and Halpert 1987, 1989; Giannini et al. 2000).

After the “1976 shift” in Pacific climate (Sect. 2.3.3.2), the frequencies of La Nina events (a decrease) and of El Nino (an increase) changed dramatically, from the preceding to the succeeding two decades. As a consequence, locations with a positive precipitation response to El Nino experienced an increase in precipitation. The Southwest United States saw its wettest two decades in the historical record. This “trend” decreased in magnitude during the 2000s when La Nina became more frequent. Whether such changes in ENSO return frequencies or ENSO locations (central versus eastern Pacific) are themselves perhaps a manifestation of climate change remains a matter of discussion (Ashok and Yamagata 2009; Yu et al. 2012).

Because global climate is a connected system, there are likely to be some linkages between the phenomena in this section. However, it does appear that to a first approximation, the relation between ENSO and North America is not greatly affected by any but the Pacific Decadal Oscillation.

2.3.3.2 Pacific Decadal Oscillation

First identified by Mantua et al. (1997), the Pacific Decadal Oscillation (PDO) is realized in sea surface temperatures of the North Pacific from about 20°N to the Gulf of Alaska. The pattern of ocean surface temperature anomalies consists of warmer than normal waters along the West Coast (approximately 20–60°N) concurrent with cooler than normal waters in the north central Pacific south of Anchorage and Southwest Alaska in its positive phase, and with opposite temperature departures in its negative phase. The characteristic time to complete a typical “cycle” (very irregular) is on the order of 50 years or so, spending 2–3 decades in each of its opposite modes. The pattern first came into prominence after the previously mentioned dramatic basin-wide “1976 shift” in Pacific climate (Ebbesmeyer et al. 1991; Trenberth and Hurrell 1994). The spatial pattern of the PDO components (sea surface temperature, wind, pressure) shows similarities to that of ENSO, but with much weaker signal in the tropics. Though visually similar, the two time series are only modestly correlated in time. The phase of the PDO has been shown to modify the typical effects of ENSO in some areas, such as the Pacific Northwest in winter (Hamlet and Lettenmaier 1999; Gershunov and Barnett 1998). Whether the PDO exists as an independent phenomenon, or instead represents an integrated sum of multiple nearby and remote forcings including ENSO in combination with annual reemergence of North Pacific ocean surface temperature anomalies (Newman et al. 2003), or has some other origin, has not been definitively settled.

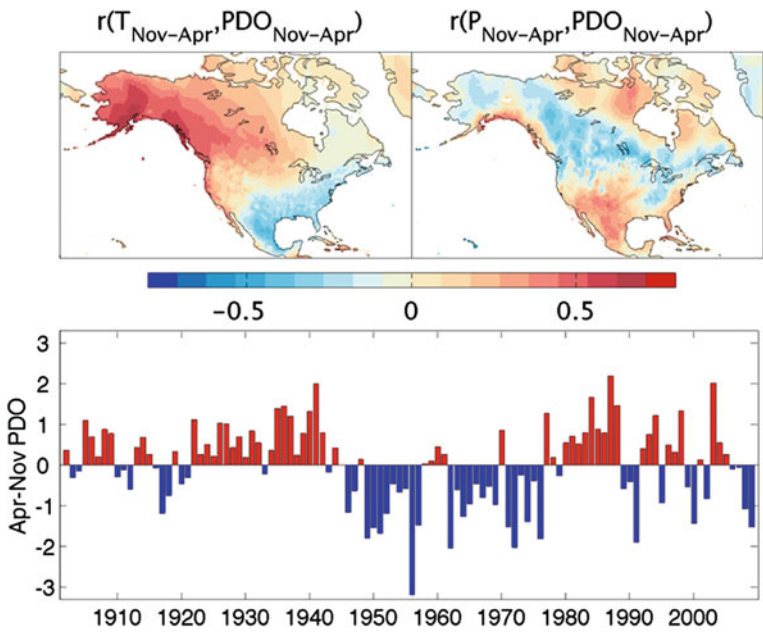


Fig. 2.2 *Top* Spatial correlation between Nov–Apr Pacific Decadal Oscillation and average Nov–Apr temperature (*left*) and total Nov–Apr precipitation (*right*), 1901/02–2008/09. *Bottom* Time series of Nov–Apr PDO from 1901/02 to 2008/09. Standardized units. Data from CRU TS 3.1

PDO Effects The phase of the PDO, sometimes in concert with other phenomena noted herein, does alter the probability distribution of seasonal precipitation, and to a lesser extent temperature, in different parts of North America, mostly in its western portions, over a broad latitudinal extent from northern Mexico to Alaska (Biondi et al. 2001; D’Arrigo et al. 2001). With its long time scale, PDO variations can masquerade as trends over periods of 20–40 years, as well as masking or amplifying observed trends resulting from anthropogenic factors. Unfortunately, a number of widely used data sets extend only from about World War II or afterward to present, or about this length of time. The PDO was negative for much of the 1940–1970s, and then reverted to a positive phase in the 1980–1990s with the 1976 shift. This oscillation coincides with strong increases in anthropogenic radiative forcing, resulting in challenges for the attribution of changes. Moreover, the small number of observed cycles of the PDO during which widespread climate observations are available is a limiting factor. This suggests that caution should be used when linking any observed phenomena to the PDO.

Figure 2.2 shows the typical patterns of temperature and precipitation associated with the PDO during the cool season in North America, along with its time history for the last century. The spatial patterns exhibit considerable similarity to those of the higher frequency ENSO variations (correlation between PDO and MEI is 0.51).

2.3.3.3 Pacific North America

Western North America often experiences a large scale upper ridge at the same time the eastern half of the continent sees a deep trough. This common pattern in mid-level (3–10 km altitude) atmospheric pressure has two principal “centers of action” (with opposite signs) in the US Southeast and the Columbia River Basin, and two others (also of opposite signs) over the Gulf of Alaska and between Hawaii and the International Date Line (Wallace and Gutzler 1981; Feldstein 2002). This pattern may be partly caused by the east–west asymmetry in continental topography, the position of the jet stream, and tropical Pacific ocean surface temperatures. Its temporal variability is strongest at timescales of approximately two-weeks and can shift phase several times during a given season; this pattern is most pronounced during late autumn–spring. The PNA has somewhat similar influences on climate as the PDO and ENSO. Of particular note, the PNA has a strong influence on temperatures in northwestern North America during the cool season. In the Cascades of Oregon and Washington state, a 1 unit increase in the PNA (Abatzoglou 2011) equates to approximately a 400-m increase in the freezing level and plays a significant role in the proportion of precipitation that falls as snow. The gradual increase in the PNA from mid-twentieth century to around 2000 has been suggested to contribute to the widespread decline in mountain snowpack in western North America (Mote et al. 2005; Abatzoglou 2011). The PNA pattern does vary according to ENSO phase (Horel and Wallace 1981; Redmond and Koch 1991) as well as the phase of the PDO (positively correlated).

PNA Effects Figure 2.3 shows the correlation between the PNA index (defined as in Abatzoglou 2011) and temperature and precipitation fields in North America for the cool season, along with the associated winter time series. The spatial patterns do bear a resemblance to those associated with ENSO response over the continent.

2.3.3.4 Arctic Oscillation/North Atlantic Oscillation

The North Atlantic Oscillation (NAO) consists of an alternation between the strength of the subtropical high pressure system commonly located near the Azores Islands (Azores High), and the low-pressure system commonly located in the North Atlantic near Iceland (Iceland Low). The Arctic Oscillation (AO) exhibits a similar north–south seesaw in atmospheric pressure, but spans all longitudes and covers the polar cap north of about 60°N. The AO and NAO patterns are strongly correlated and considered by some to be part of the same phenomena, with extra emphasis of the NAO over the North Atlantic sector, north and east of Hudson Bay (Hurrell and van Loon 1997; Hurrell et al. 2003; Feldstein 2003) and the eastern United States (Joyce 2002).

Effects This north–south seesaw in atmospheric pressure displaces the westerly jet stream to the north or south, alters the position of the storm track and associated precipitation, and modulates the temperature of air masses in mid-latitudes and

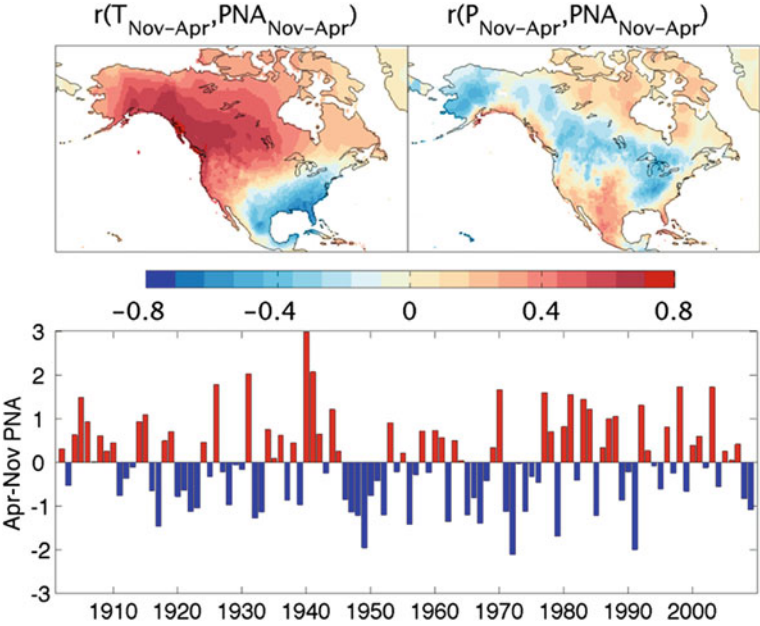


Fig. 2.3 *Top* Spatial correlation between Nov–Apr Pacific North America (PNA) index and average Nov–Apr temperature (*left*) and total Nov–Apr precipitation (*right*), 1901/02–2008/09. *Bottom* Time series of Nov–Apr PNA from 1901/02 to 2008/09. Standardized units. Data from CRU TS 3.1

over the poles. During the positive phase of the AO/NAO (subtropical pressure higher than usual), the sub-polar jet stream strengthens and moves northward, which allows for warmer air to move further north to mid-latitudes while the cold air is maintained over the pole. During the negative phase of the AO/NAO, the westerly jet stream weakens and migrates southward, which enables cooler Arctic air to dive southward into mid-latitudes, thus increasing the potential for cold air outbreaks and snow storms across southerly latitudes in eastern North America (Thompson and Wallace 1998).

Fundamentally, the AO and NAO are a dominant pattern of atmospheric variability and do not require coupled ocean–atmosphere interaction. This results in an intrinsic time scale of these patterns on the order of approximately 2 weeks (Feldstein 2003). However, observations show substantial interannual and decadal NAO variability, particularly on approximately a 20-year scale, increasing especially since about 1850 in paleoclimate coral records (Goodkin et al. 2008). The NAO underwent a significant increase in positive strength during the latter half of the twentieth century and is thought to have contributed up to half of the observed surface warming during winter across parts of northern Europe (Thompson et al. 2000a, b). By contrast, recent studies suggest that the amplified warming of the Arctic and associated loss of sea ice has decreased the latitudinal temperature

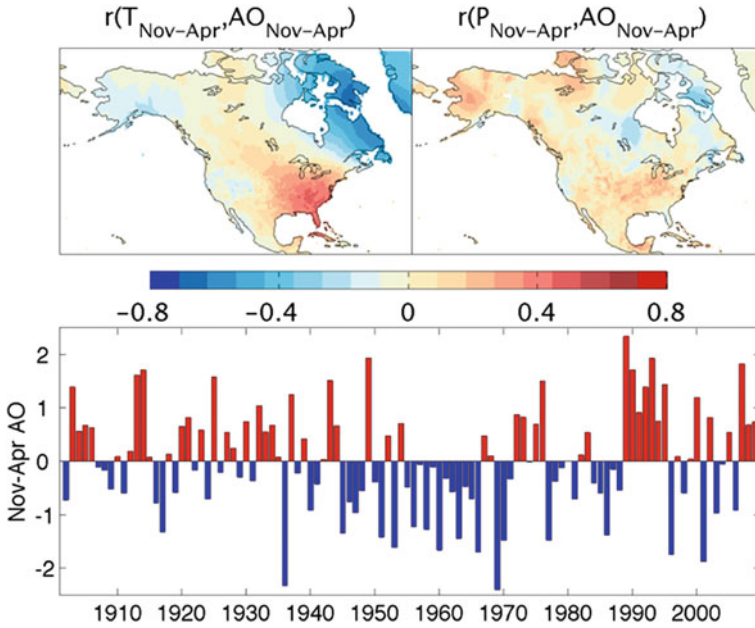


Fig. 2.4 *Top* Spatial correlation between Nov–Apr Arctic Oscillation (AO) and average Nov–Apr temperature (*left*) and total Nov–Apr precipitation (*right*), 1901/02–2008/09. *Bottom* Time series of Nov–Apr AO from 1901/02 to 2008/09. Standardized units. Data from CRU TS 3.1

gradient near the Arctic Circle, and the corresponding pressure gradient, weakening the barrier between cold air over the Arctic and warmer air in midlatitudes (Francis and Varvus 2012). In association, longer-term trends in circulation patterns do influence regional trends in temperature and precipitation and can obscure or amplify (or perhaps stem from) anthropogenically forced change (e.g., Abatzoglou and Redmond 2007).

Figure 2.4 shows that a positive value of the Arctic Oscillation during the cool season is associated with warm conditions in the eastern United States and Cuba, and very cool conditions in northeast Canada and western Greenland. Weak warmth extends north-northwestward to the MacKenzie River delta, and weak cool conditions occur in western sections of the continent.

2.3.4 Atlantic Multi-Decadal Oscillation

First noted by Enfield et al. (2001), the AMO refers to ocean surface temperatures in the North Atlantic poleward of 10°N. In its positive phase, this tripole pattern consists of higher ocean temperatures in the subtropics and the northern Atlantic, with somewhat less high ocean temperatures near Bermuda. Over the last century

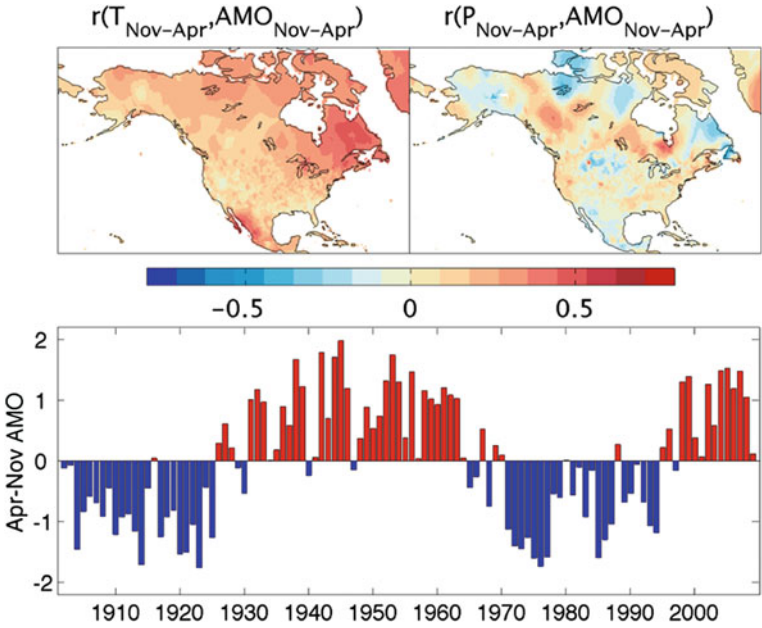


Fig. 2.5 *Top* Spatial correlation between Nov–Apr Atlantic Multidecadal Oscillation index and average Nov–Apr temperature (*left*) and total Nov–Apr precipitation (*right*), 1901/02–2008/09. *Bottom* Time series of Nov–Apr AMO from 1901/02 to 2008/09. Standardized units. Data from CRU TS 3.1

this pattern has shown four phase changes, at about 1900, 1930, 1970, and 1995, and thus an overall period of about 65 years during the one and one-half oscillations seen.

AMO Effects Effects are seen in the southeast (Enfield et al. 2001), but have been reported as far west as the Colorado Rockies (McCabe et al. 2007; Switanek and Troch 2011). With positive AMO in winter months, much of North America, especially eastern and northern Canada, tends to be warm (Fig. 2.5). The precipitation pattern is quite mixed and is not dominated by large spatial anomalies. The time series clearly shows multi-decade behavior.

2.3.4.1 Madden-Julian Oscillation

The Madden-Julian Oscillation (MJO; sometimes ISO, Intra-Seasonal Oscillation) is an eastward propagating tropical wave with a time scale of 40–70 days (Madden and Julian 1994; Zhang 2005). The time scale of 6–10 weeks is in the realm of short-term climate. The eastward propagating MJO “wave” features an extensive complex of thunderstorms and general upward motion, accompanied by an adjoining area of subsidence and downward motion that slowly migrates to the east

from the eastern Indian Ocean, across Indonesia, and into the equatorial western Pacific, decaying typically before reaching the International Dateline. The heat energy released by condensation in these tall convective towers can interact with the jet stream during its occasional southward excursions as it exits the Asian continent. This can in turn set up a wave train that propagates eastward across the Pacific to impinge on the North American west coast, bringing episodes of heavy precipitation (Mo and Higgins 1998; Higgins et al. 2000; Jones 2000). At times these lengthy episodes are accompanied by the “atmospheric rivers” identified by, e.g., Ralph et al. (2004, 2006) and Guan et al. (2012). The presence or absence of just a few significant storms can affect the character of a winter, especially at latitudes that are not normally continually affected by the jet stream (approximately 35°N southward). The phase of the MJO can be important in understanding variations in atmospheric regimes within a season and have been shown to influence the North American Monsoon (Lorenz and Hartmann 2006) and modulate tropical cyclone activity in the Atlantic and the Pacific depending on the longitude of convective enhancement (Maloney and Hartmann 2000a, b). There also appear to be interactions between MJO and ENSO and a current hypothesis is that the MJO can trigger and or modulate ENSO dynamics.

2.4 Recent History and Trends

Radiative forcing of the planetary energy budget from greenhouse gasses has been present since at least the beginning of the twentieth century, and in accelerated form over recent decades, where the response signal could be expected to be clearer. Especially since about 1940–1950, aerosol forcing from industrialization has likewise increased considerably (Hansen et al. 2005). Health consequences provide a strong incentive to reduce sulfates and particulates, so this forcing has leveled off globally (Hansen et al. 2005), allowing greenhouse gases to increasingly dominate. In addition, characteristics of observational data sets have changed over this interval.

Observed time series of climate elements averaged over a variety of spatial scales have shown rather different histories on monthly, seasonal, and annual bases. Because of differences in background state, there is no particular reason to expect spatial patterns of seasonal climate response to radiative forcing patterns to necessarily be the same during all months of the year. Seasonality in climate has very important consequences for water supply and demand, vegetative dormancy or growth, agriculture, ecological processes, and many other sectors.

For this reason, trends were computed for principal climate elements on an annual calendar year and seasonal (DJF, MAM, JJA, SON) basis, and for three different intervals of time. These were determined for all land points based on the CRU TS 3.1 data set. The three intervals and rationale for this choice are as follows:

(1) 1901–2009 This long period encompasses much of the era of rising carbon dioxide. The earliest years (1800s) with less dense data are avoided.

(2) 1948–2009 This encompasses the era since World War II covered by a number of more recent data sets, such as Global Reanalysis, and where more data are available.

(3) 1975–2009 Temperatures at a number of stations, and globally, have warmed at an accelerated rate in this 35 year period, and data coverage has gotten somewhat better at more isolated locations. A more robust climate response may thus be expected.

2.4.1 Temperature

Results for surface land temperature in North America are shown in Fig. 2.6, for all combinations of season and duration.

The most prominent feature is that temperature trends have accelerated since 1900, with the greatest rate of change since 1975 at most locations. All seasons and the year show large areas of increased warming rates during this period. Isolated areas show cooling, including during the past 35 years, mostly of limited geographic extent, with the exception of the spring cooling of south central Canada. Regional seasonal departures from spatial uniformity could reflect differences in regional radiative forcing, circulation shifts as a response to planetary-scale forcing, and the patterns of variability discussed in Sect. 2.3.3. The Arctic latitudes and the deserts of the American Southwest and the Mexican Northwest have shown the greatest recent warming.

The temporal histories of these regions and seasons almost without exception do not show slow monotonic increases, but rather are characterized by considerable variation from decade to decade, from region to region, and from season to season for a given region. These histories are shown in Fig. 2.7 for four regions (Alaska, Canada, continental US, Mexico) for the seasons and the year.

Figure 2.7 shows a general upward trend over the last 110 years in all four main regions, with the warmest years since about the mid-1990s. Temperatures generally leveled off or increased more slowly for the first decade of the new millennium. Temperatures were high during a previous warm period in the 1930–1940s, and for the period from 1940 to 1990 showed little overall net trends, with decreases to the mid-1970s and increases afterward. The interannual variability increases with latitude, so that lower latitudes have a somewhat higher signal to noise ratio. Thus while northern latitudes have warmed more in absolute degrees, lower latitudes have warmed more in normalized terms. Linear trends are listed to aid comparison, but human and natural impacts will actually consist of responses to the actual annual, seasonal, and daily time series.

Values for the four seasons are shown in Fig. 2.8 through Fig. 2.11. Winter (Fig. 2.8) has shown greater warming at higher latitudes (note scales vary by season). This trend is heavily influenced by the nature of the starting and ending

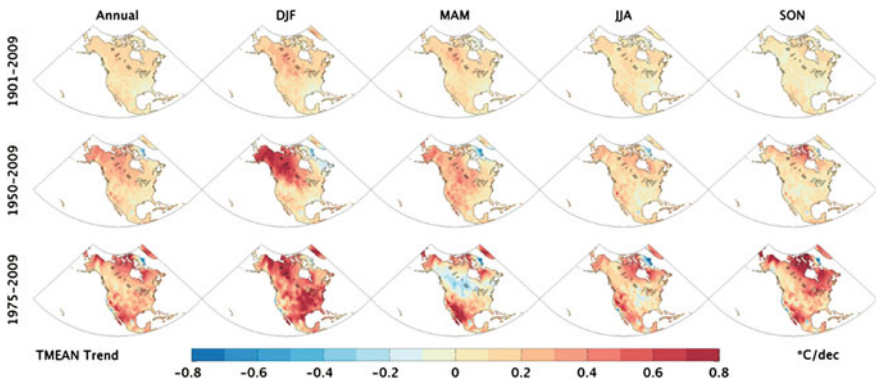


Fig. 2.6 Linear temperature trends in $^{\circ}\text{C}/\text{decade}$ for *five* seasons and *three* durations, based on CRU TS 3.1 data. *Top row* 1901–2009. *Middle row* 1950–2009. *Bottom row* 1975–2009. Columns, from *left to right* Annual, Winter (*DJF*), Spring (*MAM*), Summer (*JJA*), Autumn (*SON*)

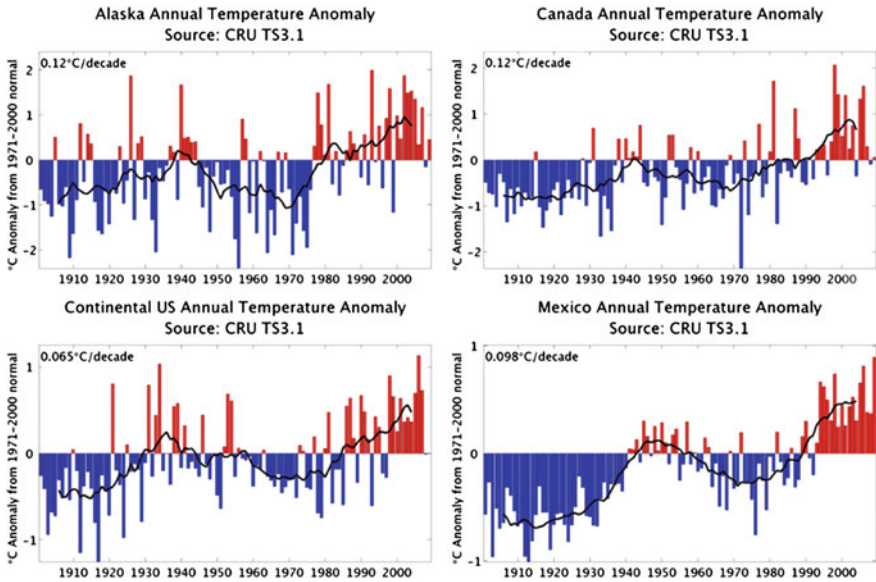


Fig. 2.7 Annual temperature anomalies (relative to 1971–2000 base period) for all CRU TS 3.1 grid points in (*upper left*) Alaska, (*upper right*) Canada, (*lower left*) continental US, and (*lower right*) Mexico. Data from 1901 to 2009. Note differences in scale between plots. Black line shows 10-year running mean

years. Much of the middle part of the twentieth century does not show strong warming or cooling in most seasons. Warmer winters have become more common since 1980 or 1990. Warm conditions were experienced in many parts of North America toward the end of the first third of the twentieth century.

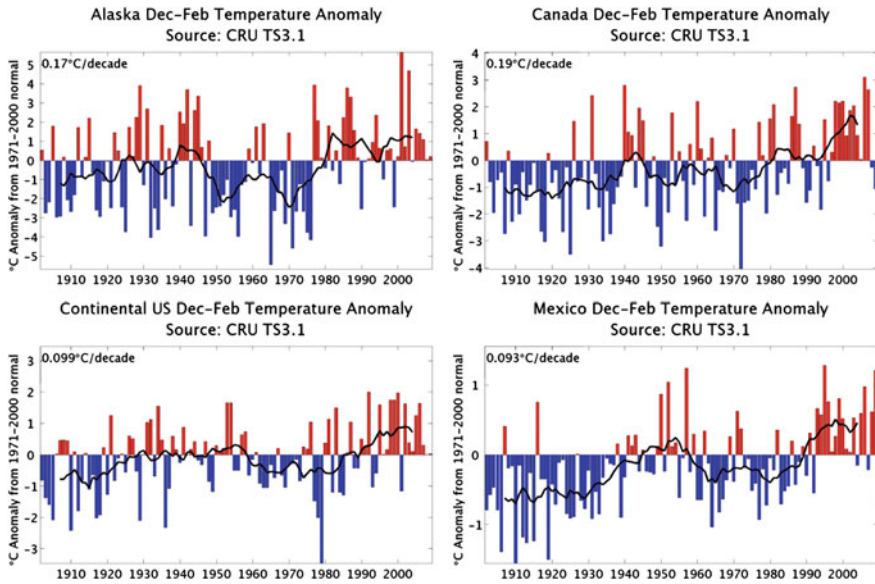


Fig. 2.8 Winter (Dec–Feb) temperature anomalies (relative to 1971–2000 base period) for all CRU TS 3.1 grid points in (*upper left*) Alaska, (*upper right*) Canada, (*lower left*) continental US, and (*lower right*) Mexico. Data from 1901/02 to 2008/9. Black line is 10-year running mean

Spring (Fig. 2.9) has shown rather widespread warming over the past 3–6 decades. The warm period shortly before the middle twentieth century is less prominent in these time series. Spring warming has considerable consequences to snow and snowmelt and consequently to hydrology in the western cordillera.

For much of North America, summer has shown little trend in temperature since 1900 (Fig. 2.10). The drought years of the “Dust Bowl” era were generally warmer in summer than the subsequent several decades. However, since about the late 1990s, summer has warmed very considerably and consistently. This rise in the final 10–15 years is responsible for much of the century long trend.

Temperature trends over North America north of Mexico during autumn (Fig. 2.11) have been generally flat or lackluster for most of the past century. As with summer, temperatures in autumn generally increased rapidly in the 1990s and have since remained on an approximate plateau.

As noted above, the NCEP/NCAR Global Reanalysis (NNGR) is based almost exclusively on upper air information (balloons, aircraft, satellites) over land, and so in many ways this widely used data set can be thought of as quasi-independent of the surface data represented by CRU TS 3.1. Surface values within NNGR are estimated from the vertically self-consistent assimilations used to derive the upper level data. Time series from this data set since 1948 are shown in Fig. 2.12 and can be compared with the portions of the time series since 1948 in Fig. 2.7.

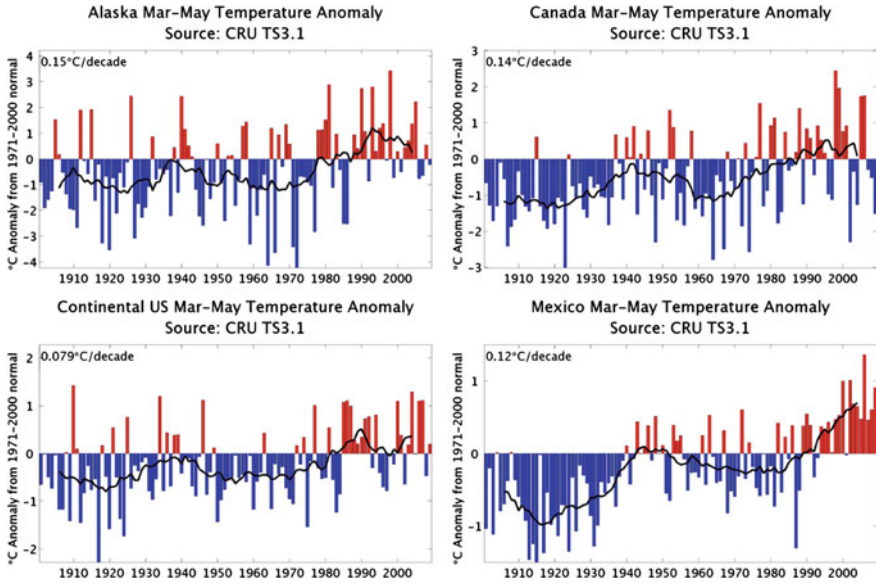


Fig. 2.9 Spring (Mar–May) temperature anomalies (relative to 1971–2000 base period) for all CRU TS 3.1 grid points in (*upper left*) Alaska, (*upper right*) Canada, (*lower left*) continental US, and (*lower right*) Mexico. Data from 1901 to 2009, plotted in ending year. *Black line* is 10-year running mean

In general, for the overlapping period from 1948 to 2009, the correspondence between surface based CRU TS 3.1 and upper air based NNGR is quite good in all four regions shown. These data sets do measure different things, however, so exact correspondence is not expected. In particular, the slightly warmer conditions of the 1940s and 1950s, cooling into the 1960s and 1970s, and warming thereafter, more pronounced in the 1990s, are generally replicated in both sets of data.

For each of these four geographic regions, annual and seasonal linear trends based on the CRU TS3.1 dataset for the three main time intervals discussed are summarized in Table 2.2. For the 109-year period, trends are statistically different from zero in all locations and seasons. Trends are generally larger over the last 60 years since 1950. In the 35 years since 1975, trends are generally larger than since 1950, except in Alaska. Annual trends in Canada, the continental US, and Mexico are significant on an annual basis over all three portions of the past 109 years. For the last 35 years, variability is large enough that some 95 % confidence bounds encompass zero trend. In most of North America, annual rates of increase appear to have generally accelerated, as have those in some but not all seasons. But, except for autumn, Alaska trends have not been greater over the past 35 years than those of the last 60 years. Mexico, closer to the tropics, has less inherent temperature variability, and thus its trends are less ambiguous even if smaller.

Large parts of North America are elevated, and mountain conditions are important for hydrologic, ecological, and cultural and human systems. In elevated

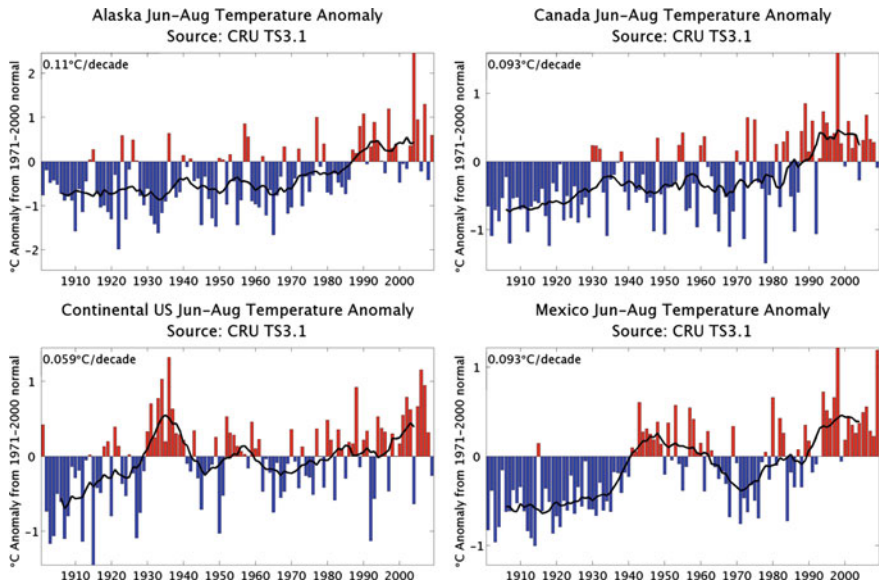


Fig. 2.10 Summer (Jun-Aug) temperature anomalies (relative to 1971–2000 base period) for all CRU TS 3.1 grid points in (upper left) Alaska, (upper right) Canada, (lower left) continental US, and (lower right) Mexico. Data from 1901 to 2009. Black line is 10-year running mean

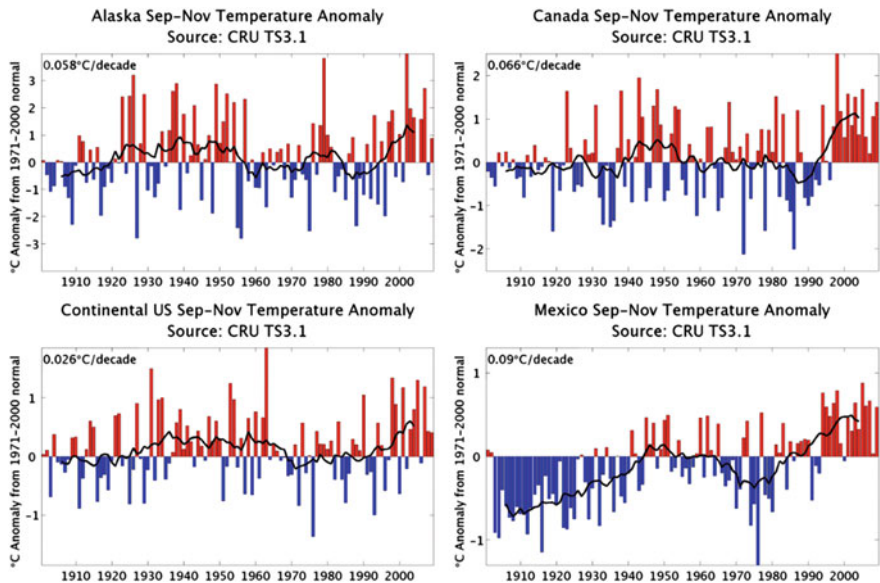


Fig. 2.11 Autumn (Sep-Nov) temperature anomalies (relative to 1971–2000 base period) for all CRU TS 3.1 grid points in (upper left) Alaska, (upper right) Canada, (lower left) continental US, and (lower right) Mexico. Data from 1901 to 2009. Black line shows 10-year running mean

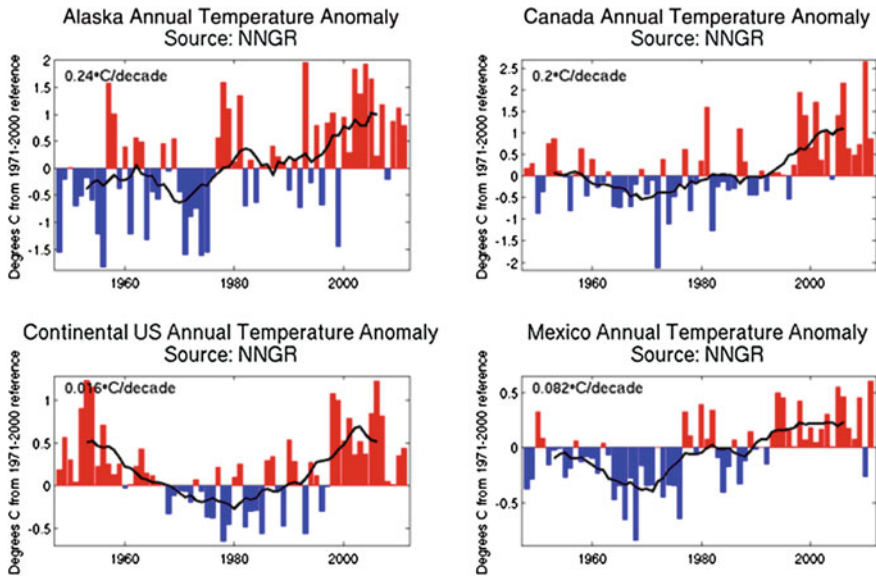


Fig. 2.12 Annual temperature anomalies (relative to 1971–2000 base period) for the surface layer from NCEP/NCAR Global Reanalysis for all grid points in (*upper left*) Alaska, (*upper right*) Canada, (*lower left*) continental US, and (*lower right*) Mexico. Data from 1948 to 2011. Black line shows 10-year running mean

Table 2.2 Temperature trends of four geographic regions, for three time intervals ending in 2009, for the four seasons (winter–DJF, spring–MAM, summer–JJA, autumn–SON) and annually

	Annual	DJF	MAM	JJA	SON
<i>1901–2009</i>					
Alaska	* 0.12 ± 0.06	* 0.17 ± 0.14	* 0.15 ± 0.09	* 0.11 ± 0.04	* 0.06 ± 0.09
Canada	* 0.12 ± 0.04	* 0.19 ± 0.09	* 0.14 ± 0.06	* 0.09 ± 0.03	* 0.07 ± 0.05
CONUS	* 0.07 ± 0.03	* 0.10 ± 0.06	* 0.08 ± 0.04	* 0.06 ± 0.03	0.03 ± 0.04
Mexico	* 0.10 ± 0.02	* 0.09 ± 0.03	* 0.12 ± 0.03	* 0.09 ± 0.02	* 0.09 ± 0.02
<i>1950–2009</i>					
Alaska	* 0.35 ± 0.14	* 0.62 ± 0.32	* 0.37 ± 0.25	* 0.23 ± 0.10	0.15 ± 0.23
Canada	* 0.21 ± 0.11	* 0.34 ± 0.22	* 0.20 ± 0.18	* 0.15 ± 0.09	* 0.16 ± 0.15
CONUS	* 0.11 ± 0.06	0.15 ± 0.17	* 0.19 ± 0.10	0.07 ± 0.07	0.05 ± 0.10
Mexico	* 0.12 ± 0.05	* 0.11 ± 0.08	* 0.16 ± 0.07	* 0.10 ± 0.06	* 0.11 ± 0.06
<i>1975–2009</i>					
Alaska	0.21 ± 0.29	0.20 ± 0.72	0.20 ± 0.54	* 0.25 ± 0.24	0.26 ± 0.54
Canada	* 0.28 ± 0.27	0.42 ± 0.51	0.02 ± 0.44	* 0.28 ± 0.20	* 0.45 ± 0.35
CONUS	* 0.26 ± 0.15	* 0.52 ± 0.42	0.12 ± 0.26	0.16 ± 0.18	* 0.26 ± 0.22
Mexico	* 0.33 ± 0.08	* 0.36 ± 0.16	* 0.41 ± 0.14	* 0.24 ± 0.13	* 0.30 ± 0.14

All trends positive. Values expressed in °C per decade. There is a 95 % chance that the actual trend is within the indicated bounds; asterisk (*) indicates different from zero

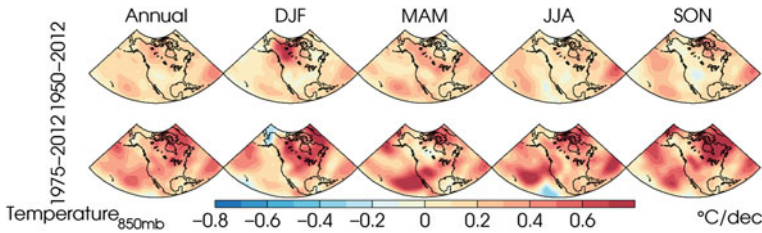


Fig. 2.13 Linear temperature trends at 850 mb (~ 1500 m elevation) in C/decade for five seasons and two durations, based on NNGR data. Top row 1950–2012. Bottom row 1975–2012. Columns, from left to right Annual (Jan–Dec), Winter (DJF), Spring (MAM), Summer (JJA), Autumn (SON)

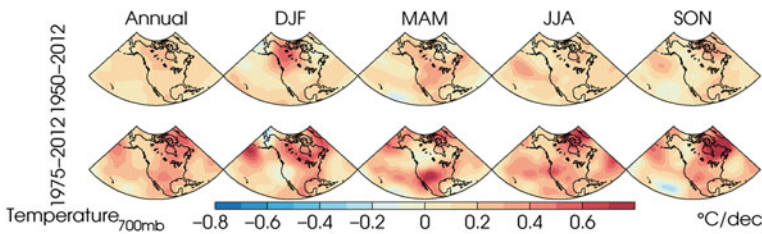


Fig. 2.14 Linear temperature trends at 700 mb (~ 3000 m elevation) in C/decade for five seasons and two durations, based on NNGR data. Top row 1950–2012. Bottom row 1975–2012. Columns, from left to right Annual (Jan–Dec), Winter (DJF), Spring (MAM), Summer (JJA), Autumn (SON)

areas, most rivers are fed by melting snow; for example, about 85 % of the flow of the Colorado River originates in the 15 % of the area above 2700 m (Christensen and Lettenmaier 2007). Because of the frequent presence of surface based temperature inversions, time series of temperature at elevated sites need not resemble those at sea level. For this reason, annual and seasonal temperature trends were derived for two standard pressure levels, one at 850 mb (~ 1500 m) and the other at 700 mb (~ 3000 m). The 850 mb level is near the upper part of large ice fields in Canada and Alaska, and the 700 mb level is near or slightly above the zone of snowpack accumulation and some glaciers in the western mountains of the conterminous United States.

Figure 2.13 shows temperature trends at 850 mb (~ 1500 m) from NNGR for two periods ending in 2012, on an annual and seasonal basis. The patterns are reminiscent of the surface based patterns, but in general show more regional coherence, and smaller areas of cooling, with almost no cooling seen over the land area of North America.

Similarly, temperature trends at 700 mb (~ 3000 m) are positive over nearly all of North America for the past 65 and 35 years (Fig. 2.14). Increased temperatures drive increased snowmelt, evaporation and transpiration from forested areas, and have similar consequences to a reduction in precipitation. Of much concern, forest

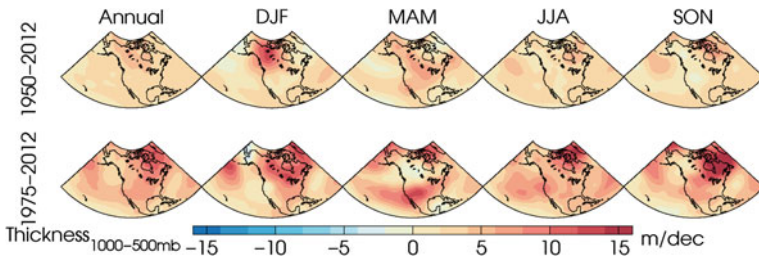


Fig. 2.15 Linear trends in 1000–500 mb “thickness” in m/decade for *five* seasons and *two* durations, based on NNGR data. 20 m is equivalent to about 1 °C. *Top row* 1950–2012. *Bottom row* 1975–2012. Columns, from *left to right* Annual (Jan–Dec), Winter (*DJF*), Spring (*MAM*), Summer (*JJA*), Autumn (*SON*)

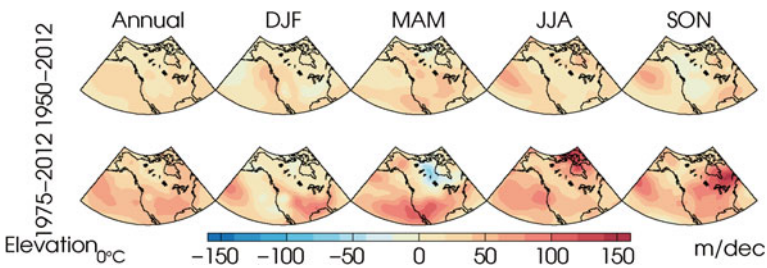


Fig. 2.16 Linear trends in elevation of the freezing level, expressed in m/decade for *five* seasons and *two* durations, based on NNGR data. *Top row* 1950–2012. *Bottom row* 1975–2012. Columns, from *left to right* Annual (Jan–Dec), Winter (*DJF*), Spring (*MAM*), Summer (*JJA*), Autumn (*SON*)

drying and drought are exacerbated by warm conditions, and very large insect infestations have developed in the Rocky Mountains and throughout the mountains in British Columbia over the past decade (Raffa et al. 2008). In addition, tree mortality, large fires, and fires at higher elevations have all become much more common since the late 1990s, and are currently of much concern to resource managers (Westerling et al. 2006; vanMangem et al. 2009).

A more integrated measure of the temperature of the lower atmosphere in very common use in meteorology is the vertical distance between two pressure levels, a quantity known as “thickness.” The thickness of the layer between 1000 mb (sea level) and 500 mb (~ 5500 m) encompasses half the mass of the atmosphere and all land surfaces except 3 peaks in Alaska and Canada. A column warming of 1 °C over this entire distance will increase the depth of this layer (its thickness) by about 20 m, because warm air expands. Figure 2.15 shows annual and seasonal trends in this vertically integrated temperature for the same two recent sets of decades. The warming during the past 65 years is seen to accelerate significantly starting about 1975.

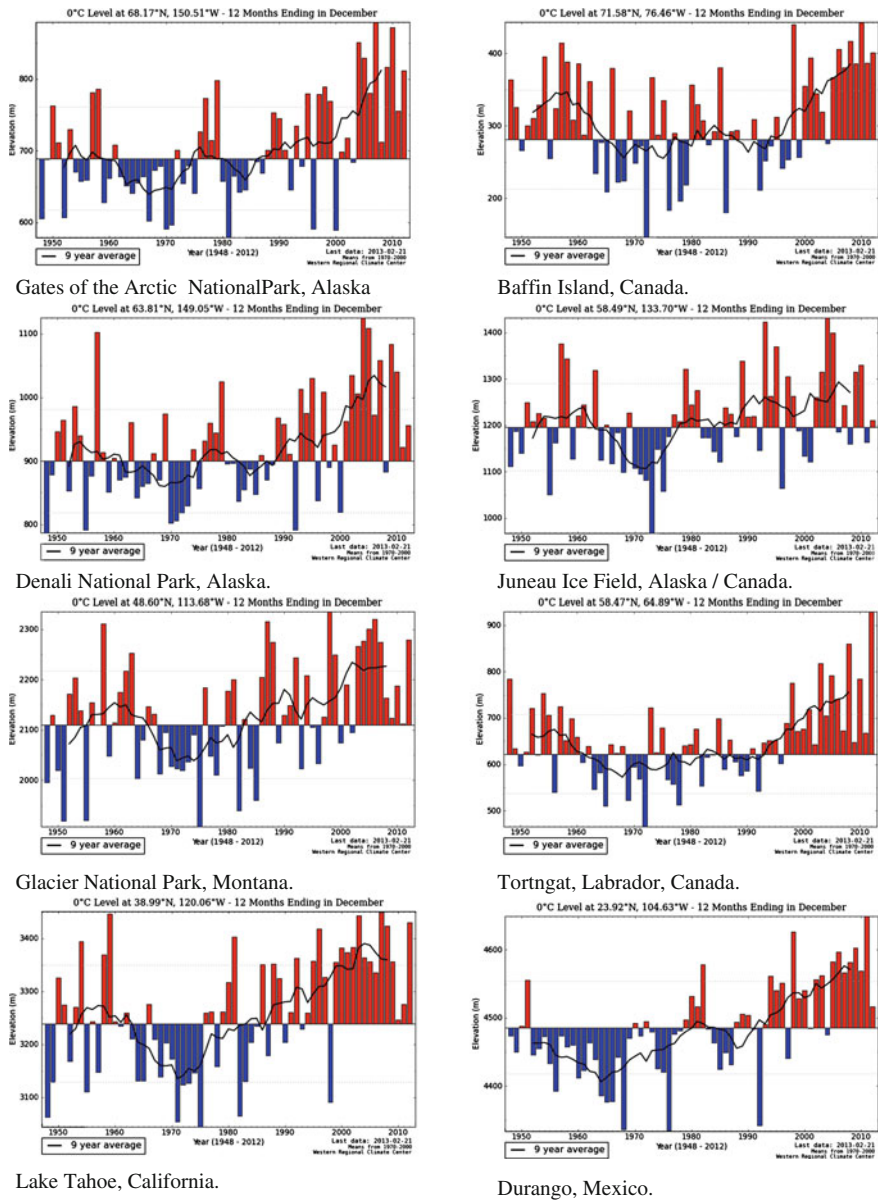


Fig. 2.17 Mean annual freezing level time series for selected Arctic and elevated regions of North America. *Black line* represents 9-year running mean. Based on NNGR, data from 1948 to 2012, from (NAFLT 2013), reference period is entire time series

Over snow covered areas, or at night, and in complex terrain, temperature does not always decrease with elevation because of decoupling of the surface layer from the overlying air. The height above sea level at which temperature reaches freezing

is very important to hydrology in mountainous regions, affecting whether precipitation falls as rain or snow, the accumulation of snow and the snowmelt season, and biological activity. A methodology to determine the upper freezing level has been used to determine the monthly and annual history of this level from NNGR over all of North America (NAFLT 2013). Figure 2.16 shows that the freezing level has increased annually and in all seasons.

Time histories of freezing level determined this way for a selection of sites in the Arctic and the western cordillera of North America are shown in Fig. 2.17. These can be readily extracted for any location in North America by means of an application that uses the NNGR data set (described at NAFLT 2013). A common feature of many of these important cold areas is a rise of over 100 m starting approximately around 2000. Such changes have many significant implications for hydrological and biological systems.

Most of northern North America is underlain by permafrost, including 42 % of Canada (Kettles et al. 1997). Warming of the overlying atmosphere, and changes in the energy budget caused by changes in snow cover and temperature, affect the depth to the permafrost level. Zhang et al. (2006) show that the depth from surface to permafrost level has increased throughout the twentieth century and into the twenty first century as a pulse of warmth penetrates downward. Thawing of permafrost is then expected to release carbon dioxide and methane into the atmosphere, a feedback loop that can accelerate warming (Lawrence et al. 2008; Cory et al. 2013).

Most of the emphasis here is on means, but extremes are more tangible and memorable, and act as disturbances with potential lasting consequences to human and natural systems. Meehl et al. (2009) noted that between 1 January 2000 and 30 September 2009, the US recorded 291,237 daily record high maximum temperatures and 142,420 daily record low minimum temperatures (ratio 2.05:1). Kunkel et al. (2009: their Fig. 10) show that heat waves (four consecutive days exceeding a once-in-five-year recurrence value) do not show an appreciable trend in the continental US. Similarly-defined cold waves likewise show a non-significant decrease, complicated by a large number of cold days in the 1980s. The decadal time series of heat spells is dominated by a large spike during the Dust Bowl era of the 1930s, with suggestions of a rising trend nationally and in some regions in the last 5–6 decades, and in Alaska (no 1930s data). When both are plotted together (Kunkel et al. 2012, their Fig. 1), there are strong suggestions of more heat/less cold extremes from the 1970s to 2000s. They further show (their Fig. 2) that the cold tails in the frequency distributions of maximum and especially of minimum temperatures have strongly warmed from 1950 to 2007. Figure 2.18 shows the trend in the coldest night per winter from 1920 to 2012 in the US. Extreme winter minimums are a primary cause of mortality for the landscape industry and ornamental gardens. These changes resemble those seen in updated cold hardiness maps for gardening (e.g., USDA 2012), although those maps are also affected to some extent by changes in methodology.

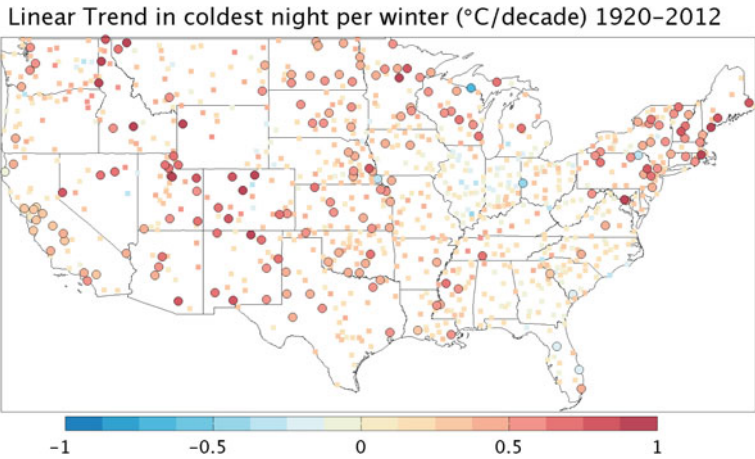


Fig. 2.18 Linear trend in coldest night per winter, 1920–2012, in $^{\circ}\text{C}/\text{decade}$. *Large dots* signify less than a 5 % chance that the trend is actually zero. Based on 1218 US Historical Climatology Network stations

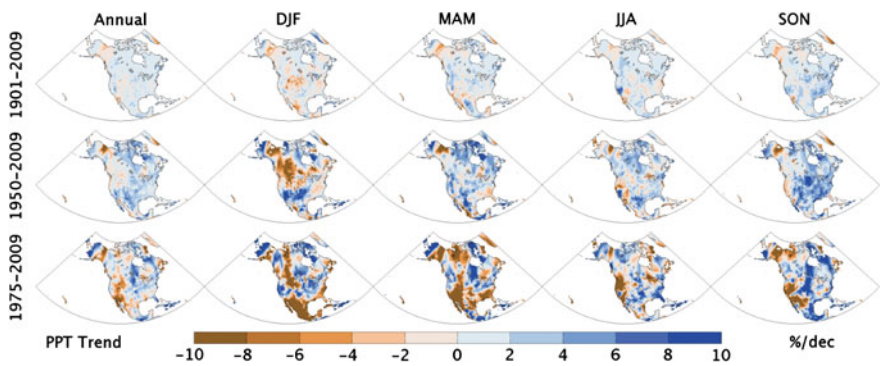


Fig. 2.19 Precipitation trends in percent/decade for *five* seasons and *three* durations, based on CRU TS 3.1 data. *Top row* 1901–2009. *Middle row* 1950–2009. *Bottom row* 1975–2009. Columns, from *left to right* Annual, Winter (*DJF*), Spring (*MAM*), Summer (*JJA*), Autumn (*SON*). Reference period is 1971–2000

2.4.2 Precipitation

Trends in precipitation derived from the CRU TS3.1 dataset are shown in Fig. 2.19 for the same seasons and years as for temperature in Fig. 2.6. Annual precipitation varies by 1–2 orders of magnitude within North America, and in complex terrain such variations can be seen within just a few kilometers. The month or season of maximum precipitation varies far more than for temperature, encompassing all months of the year. Furthermore, in mountain settings the season of maximum

Table 2.3 Precipitation trends in four geographic regions, for three time intervals ending in 2009, for the four seasons (winter-DJF, spring-MAM, summer-JJA, autumn-SON) and annually

	Annual	DJF	MAM	JJA	SON
<i>1901–2009</i>					
Alaska	−0.30 ± 0.58	−0.34 ± 1.15	−0.86 ± 0.96	−0.02 ± 0.86	*−0.35 ± 0.04
Canada	*+0.74 ± 0.19	+0.28 ± 0.52	*+0.85 ± 0.42	+0.86 ± 0.33	*+0.83 ± 0.39
CONUS	*+0.70 ± 0.43	−0.13 ± 0.80	+0.65 ± 0.74	+0.40 ± 0.56	*+1.82 ± 0.92
Mexico	*+1.01 ± 0.60	−0.63 ± 2.20	+0.75 ± 1.78	*+1.11 ± 0.76	*+1.37 ± 1.00
<i>1950–2009</i>					
Alaska	+0.76 ± 1.43	+2.32 ± 2.70	+0.43 ± 2.63	−0.31 ± 2.22	+1.61 ± 2.64
Canada	*+1.17 ± 0.40	*−1.94 ± 1.23	*+2.35 ± 1.07	*+1.62 ± 0.86	*+1.88 ± 0.90
CONUS	*+1.52 ± 1.04	−0.11 ± 2.10	+1.05 ± 1.71	*+1.28 ± 1.16	*+3.69 ± 2.20
Mexico	+1.32 ± 1.41	+1.10 ± 6.01	+1.97 ± 4.58	+0.72 ± 2.03	+2.08 ± 2.51
<i>1975–2009</i>					
Alaska	+0.33 ± 3.18	+0.07 ± 6.75	−1.80 ± 6.19	+2.85 ± 5.19	−1.19 ± 6.70
Canada	+0.65 ± 0.96	−0.63 ± 3.32	−0.05 ± 2.08	+0.38 ± 1.86	*+2.13 ± 2.11
CONUS	+1.12 ± 2.46	+1.31 ± 5.60	−1.19 ± 3.79	+2.43 ± 2.92	+1.56 ± 4.89
Mexico	+1.49 ± 3.22	*−16.14 ± 13.96	+0.07 ± 10.87	+0.59 ± 5.58	*+4.93 ± 5.80

Values expressed in percent per decade. There is a 95 % chance that the actual trend is within the indicated bounds; *asterisk* (*) indicates different from zero

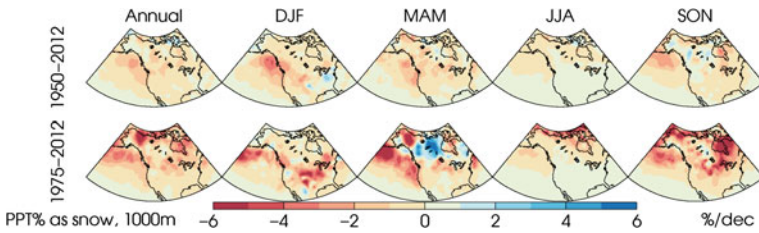


Fig. 2.20 Linear trends in fraction of precipitation as snow at 1000 m elevation, in percentage points per decade, for five seasons and two durations, based on NNGR data. *Top row* 1950–2012. *Bottom row* 1975–2012. Columns, from left to right Annual (Jan–Dec), Winter (DJF), Spring (MAM), Summer (JJA), Autumn (SON)

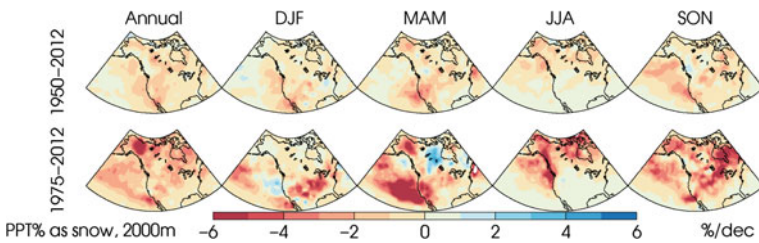


Fig. 2.21 Linear trends in fraction of precipitation as snow at 2000 m elevation, in percentage points per decade, for five seasons and two durations, based on NNGR data. *Top row* 1950–2012. *Bottom row* 1975–2012. Columns, from left to right Annual (Jan–Dec), Winter (DJF), Spring (MAM), Summer (JJA), Autumn (SON)

precipitation can similarly vary over tens of kilometers. For these and other reasons, trends in precipitation in absolute units are nearly meaningless, and are usually shown in percent per decade.

Precipitation is inherently highly variable in time, and trends in this element show much finer spatial structure than do temperature trends. Reflecting the multiple mechanisms that lead to precipitation, the interpretation of such precipitation trend maps is much more complicated than for temperature. On an annual basis the continent has become wetter over the past half century, more in the eastern than western United States, and generally in the Arctic. The expectation of increasingly wetter conditions in the northern latitudes and drier conditions in southern latitudes, projected to be just under way, is only somewhat evident. The desert region exhibits signs of drying annually and in all seasons. The unambiguous detection of trends in amount will require a longer passage of time than for temperature because of the large relative interannual variability. Table 2.3 gives information on precipitation trends for the four regions shown in Table 2.2.

Snow quantity, and its mere presence or absence, reflects joint occurrence of precipitation and temperatures below freezing. Thus an understanding of snow and its variation through time require consideration of both temperature and

precipitation. As a simple estimate of the relative contribution of liquid versus solid precipitation, the NNGR data set, which is available every 6 h since 1948, has been used to formulate an application accessible at NAFLT (2013). If precipitation occurs at an elevation of interest (e.g., 1000 m), this is called “snow” if the temperature is freezing or below, and rain if warmer. The fraction of total precipitation that falls as “rain” or “snow” can then be used to form a rough estimate of the rain/snow ratio. Except for very cold air, increases in temperature are expected to lead to more rain and less snow at given elevations.

Note also that precipitation from atmospheric Reanalysis is not “observed” but is rather the precipitation calculated by the assimilation model. Given the relatively coarse grid spacing of 2.5° of latitude and longitude, precipitation from large scale cyclonic systems, or organized systems of convection, is more likely to be accurately represented than is that from random convective elements such as summer air mass thunderstorms.

Trends in the fraction of precipitation falling as snow have been determined for all grid points in the North America sector of NNGR, and plotted for the past 65 and 35 years in Fig. 2.20 (1000 m elevation) and Fig. 2.21 (2000 m), typical elevations for mountain snowpack. At both elevations, the fraction of precipitation falling as snow, particularly in the cool season, has diminished by a few percent over the past 60 years (Abatzoglou 2011).

Societal infrastructure is highly susceptible to precipitation extremes, and extraordinary expense and attention are devoted to criteria for engineering design to avoid disruptive and often catastrophic consequences. The effects of anthropogenic climate changes on such extremes are now the subject of concerted investigation (e.g., Kunkel et al. 2013c). Across the United States there has been a general increase over the past several decades in extremely heavy precipitation events (Kunkel et al. 2013b). They found this increase to be primarily from the Great Plains eastward. Similarly, Bonnin et al. (2011) found little change in heavy precipitation frequency in the US Southwest, but a general increase in the eastern states. They showed, with high confidence, that for the mid-Atlantic and Ohio River Valley, the frequency of occurrence of historically defined once a year to once a century events, for event durations of 1–20 days, is increasing. They did note, however, that the uncertainty in the basic estimates of the values themselves is also comparably large. Another study along the West Coast of the United States and southwest Canada (Mass et al. 2011) found mixed evidence of temporal trends in the largest 2-day events over the period 1950–2009.

2.4.3 Snow

Snow has numerous hydrologic, biological, and energy budget influences. Snow characteristics of interest include depth of new snow, depth on the ground (or on floating ice), density and water content, albedo and age, chemical contaminants,

and duration of cover during the cold season. The measurement of snow is subject to subtle influences (Doesken and Judson 1996), and time series of observations are not necessarily homogeneous (Kunkel et al. 2007, 2009). An analysis by Kunkel et al. (2012) of the continental US shows that through the winter of 2010–2011, winters with large snow extent (>90th percentile) became slightly less common after the 1990s, but very little increase had been noted in low snow cover years (<10th percentile). Since 1900 there is relatively little trend in the frequency of either large or small snow-extent winters (Kunkel et al. 2012, their Fig. 7).

Analyses by Christy and Hnilo (2010) appear to show little trend in total seasonal accumulation in the southern Sierra Nevada of California over the twentieth century and into the first decade of the twenty first century. A subsequent analysis (Christy 2012) showed similar results for an expanded set of locations around California. These papers also emphasize and reiterate the significant effort required for quality control and data set rehabilitation before analysis is attempted.

Brown and Robinson (2011) examined the Northern Hemisphere snow-covered extent (SCE) from 1922 to 2010 and found that the last 40 years of this record showed a reduction of 7 % (March) and 11 % (April) from the earlier portion of the record. In March, most of this decrease was contributed by Eurasia, but in April both Eurasia and North America showed significant ($P < 0.05$) decreases in areal extent. Dyer and Mote (2006) concluded that downward trends in both coverage and depth were more apparent in late winter and spring than prior to January. Trends in Canadian snow cover are specifically highlighted by the government of Canada (Statistics Canada 2013) where they are available on an annual basis (their Chart 1). These downward trends are more prominent during the three main melt months of April, May and June (their Chart 3). All are statistically significant ($P < 0.05$). Additional details of snow cover variations over the last 40 years based on satellite observations are contained in Chap. 3 of this book.

2.4.4 Sea Ice

Sea ice in the Arctic Ocean generally reaches its maximum extent in February and decreases to its minimum extent in September, a quarter cycle after high sun. Long term measurements show an unsteady but continuing decline in ice-covered area. During September 2012 Arctic sea ice reached a record low individual daily extent of 3.41 million km², about 51 % of the 1979–2010 reference average (NSIDC 2012a). The monthly average Arctic areal ice extent in September also fell to its lowest value since the record began in 1979. The average September rate of decline for the entire 1979–2012 period has been 13 % per decade (NSIDC 2012b).

In addition, the age (and thus thickness and volume) of the ice has been decreasing over most of the North Polar cap, increasing ice susceptibility to sensible and radiative heating because less energy is needed to melt thinner ice. Wind from Arctic storms (e.g., 6 August 2012; NSIDC 2012c) can more readily

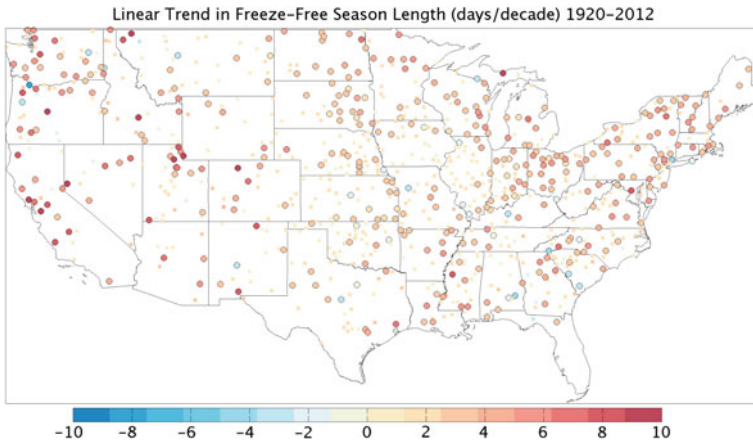


Fig. 2.22 Trend in length of season between last hard freeze ($< -2.2^{\circ}\text{C}$) in spring, and earliest in autumn, from 1948 to 2009. Trend in days per 62 years. *Large circle* indicates less than 5 % chance that trend is actually zero

push and mechanically reduce large areas of thinner ice, and can generate large waves over open water that readily destroy the edges of thinner pack ice. Areas of open water also allow waves to grow to sizes that eat away at shorelines, some with human villages immediately close at hand (ACIA 2004). Furthermore, elimination of the soil mechanical rigidity arising from permafrost melt allows wave action to more rapidly erode beaches and embankments. Additional discussion of sea ice variations over the last three decades is contained in the following chapter on satellite observations.

2.4.5 Trends from Other Indicators

Variations in climate can be detected by means of instruments and sensors. But climate also affects natural systems, and many of the impacts to such systems are noticed by human observers. They are often recorded in non-systematic or non-numerical form, such as entries in diaries or personal logs (Dupigny-Giroux and Mock 2009).

2.4.5.1 Biologically-Based

In many living organisms the rate of metabolism is proportional to temperature, because the chemical reactions that sustain life are themselves temperature dependent. In some cases, these processes cannot proceed unless above some critical threshold. The relation to temperature is often highly nonlinear. In other

cases (e.g., winter wheat) a certain value of accumulated cool temperature (chilling or vernalization, e.g., Chouard 1960) is needed before a plant can properly mature when spring temperatures permit growth. Insect activity and development are highly dependent on temperature. Therefore it is not surprising that a great variety of biological markers can serve as a proxy for temperature, or precipitation, or occasionally other climate elements.

However, these influences are selective, usually reflecting conditions during just one critical part of the year. For example, the growth rings of trees can reflect both precipitation and temperature during some few past months up to a year or two. However, for some species a tree growth ring may be influenced only by spring or summer temperature and be relatively insensitive to other seasons. Thus, biological measures (output signals) should be considered to be heavily filtered versions of the climate drivers (input signals). The properties of these plant and animal filters are complex, and depend particularly on the organism or species, so that every organism responds to the same climate and weather influences in different ways. The relations between such biological measures and climate must therefore usually be determined empirically via statistical methods.

Phenology is the study of how climate affects the timing of significant events in the life of an organism: first bud, first flower, wheat jointing, seed formation, egg hatch, migration of birds or land mammals, first and last fall freeze, life stages of insect pests, breeding conditions for plants and animals, etc. In recent years, systematic efforts have begun to record such events, using electronic recorders and web entry devices. The National Phenological Network (Betancourt et al. 2007) reflects the popularity of this approach, by entraining citizen volunteers.

Cayan et al. (2001) made use of the correspondence between spring blooming dates of lilacs and honeysuckles over a 40-year period, the first pulse of spring mountain snowmelt arising during brief warm spells, and conventional thermometer data. These showed surprisingly good correspondence, increasing the confidence in the quality and utility of each of the data sets employed.

Kunkel et al. (2013b) show that the summer freeze-free season has lengthened over the US by about 12 days from 1895 to 2011. This trend is more pronounced in the western than the eastern US, though all trends are statistically significant. Figure 2.22 shows trends in the date of last spring hard freeze (-2.2°C threshold) for the continental US over the period 1948–2009, based on surface data from the Historical Climatological Network. In much of the nation this date is trending earlier in the year, indicating warmer late winter and spring.

Xu et al. (2013) used satellite data from July 1981 through December 2011 to examine changes in vegetation amount and growth period length between 50 and 75°N latitude. Over these 30 years, the temperature of the summer photosynthetically active period has increased by $1\text{--}2^{\circ}\text{C}$ at all latitudes, the length of this period has increased by 4–5 days, and Normalized Difference Vegetation Index (NDVI) during summer growth season has increased over 32–41 % of the Arctic and boreal area. Vegetative browning (i.e., lower NDVI) occurred over just 4–5 % of the region. This vegetative greening was somewhat more prominent along the

coasts and on the eastern side of the continent. More detailed information from satellite measurements of vegetation changes is contained in [Chap. 3](#).

2.4.5.2 Lakes and Rivers

Trends in the temperature of lakes and rivers make a difference to organisms that depend on these waters. These could include migrating fish, birds needing open (versus ice-covered) water, water-borne micro-organisms and diseases, insects, amphibians, as well as local economies dependent on frozen water (snowmobiling, ice fishing) or liquid-water (fishing, boating, rafting) recreation. Some streams are subject to artificial influences (nuclear or coal power plant effluent, treated but warm sewage) or to the effects of reservoir management, where slow moving water can absorb sunlight, or cold water is released from the bottom of deep reservoirs. Nonetheless, many “natural” rivers and streams are available to serve as integrated measures of climate variability.

Saushal et al. (2010) found increases in river and stream temperatures across the United States ranging from 0.09 to 0.77 °C/decade. These usually reflected overlying air temperatures and were somewhat more prominent in urbanizing areas. Stream temperatures across numerous rivers and streams in the Northern Rockies of the United States have increased with the most acute changes during summer thereby reducing potential habitat for cold-water fisheries (Isaak et al. 2010). Coats (2010) shows an increase from 1970 to 2006 in depth-averaged annual temperature of about 0.13 °C/decade for Lake Tahoe, the 10th deepest lake in the world. At 400 m depth, temperatures warmed by about 0.4 °C in total during those 37 years. Dobiesz and Lester (2009) found a water temperature increase over the period 1968–2002 of 0.25 °C/decade ($P < 0.18$) of August surface water temperature in Lake Erie, 0.48 °C/decade ($P < 0.01$) in Lake Ontario, and 0.84 °C/decade ($P < 0.01$) in Lake Huron. In Lake Ontario most of this increase is in the upper 20 m of the water column, but temperature at 40–50 m has warmed by 0.18 °C/decade ($P < 0.05$) over these 35 years.

The number of days per winter that lakes are iced over has decreased by nearly a week per decade over the last 40 years in Vermont (Betts 2011). A long ice-cover record (1852–2012) from Lake Mendota in Madison WI shows latest ice-out years all occurring before 1900, and 5 of the earliest 10 ice-out years after 1995, with an average loss of 30 days of ice (Kunkel et al. 2013a; WSCO 2013). Wang et al. (2012) note a combined loss of Great Lakes ice coverage of 71 % from 1973 to 2010.

Duguay et al. (2006) analyzed autumn freeze-up and spring break-up dates of lake ice across Canada from 1951 to 2000. The most robust finding is a spatially extensive pattern of earlier spring break-up dates, with increasing trends in more recent years, and similar trends in 0 °C isotherm dates. Autumn freeze-up showed little overall change over this period. Satellite observations of lake ice are summarized in the next chapter.

2.4.5.3 Boreholes

A unique form of independent evidence is a result of numerous boreholes drilled into the solid earth around the globe, numbering in the tens of thousands. Many of these have very accurate (within 0.01 °C resolution) temperature profiles from the water that fills these holes, and that reflects temperatures in the solid rock at that particular depth. Slow variations in climate send pulses of temperature into the solid earth. A prolonged warming of 10–50 years would create a “wave” of warming above ambient levels slowly penetrating down into the solid earth. Using mathematical inversion techniques, the overall parameters (especially, filtered temporal history) of such waves can be determined from the precise instantaneous vertical temperature profile in such boreholes (e.g., Pollack and Huang 2000; Beltrami 2002). Such studies show that the upper portions of boreholes all around the earth are warmer than would be expected from a stationary unchanging climate, and contain the signal of large scale slow warming seen in thermometric time series from the overlying atmosphere. Estimates of warming are consistent with surface based measurements. Furthermore, these techniques can be used to recover approximate trends for the past 500–1000 years, though with decreasing time resolution as the time-scale becomes larger and the signal becomes smeared over time. For example, Minchin Lake in Canada shows a warming of about 1 °C over the past century, following on the heels of a period 150–300 years ago that was about 1.0–1.5 °C cooler than the most recent century (Beltrami et al. 2011).

2.4.5.4 Snowmelt Pulse Timing

In mountain regions, winter snow pack must warm throughout its depth to be on the verge of melting. At that time the “ripe” snow pack can react to warm spells of a few days duration and begin to melt rapidly, appearing almost as a step function or “spring pulse”. These dates are surprisingly susceptible to quantification, and show large scale spatial coherence stemming from the driving climate patterns (Cayan et al. 2001). Furthermore, the dates often agree to within a day or two and show uniformity across space. Stewart et al. (2004) were able to show that the date of this spring pulse has been occurring earlier in the year as the century advanced, and by as much as 2–3 weeks in parts of western North America.

This measure of climate consists solely of calendar dates from hydrologic measurements, and thus contains no explicit dependence on thermometers or the circumstances of their exposure. As such it forms an entirely independent measure of climate-related behavior. However, this measure only reflects temperatures in spring (especially) and winter (to a lesser extent), with very little influence of autumn or the previous summer (there is no autumn “pulse”). Thus, as with biological data, this kind of proxy data also involves a hydrologic “filter” applied to the driving climate time series. The approach similarly results in an inability to say anything definitive about certain parts of the year. As with all proxy data sets, we are forced to work with the signal that presents itself.

2.5 Key Findings

The main intent of this chapter is to describe the variability of climate in North America, including the slow variability often labeled as “trend.” Variability in the sign and strength of the characteristic patterns discussed in [Sect. 3.3.3](#) can account for some of the observed climatic variations on time scales from weeks to a few decades (e.g., the Pacific Decadal Oscillation). A discussion of potential causes (“attribution”) of climate change on decadal time scales is the focus of [Chap. 6](#).

The principal observational findings discussed in this chapter include the following:

- Temperature

- Annual and seasonal surface temperatures have been rising over all major portions of North America since 1901.
- Annual trends in temperature are positive and statistically different from zero for the last 109, 60 and 35 years in Canada, the continental US and Mexico, and have increased with time. For example, in the US, the trends for the three time periods are 0.07, 0.11, and 0.26 °C/decade.
- Seasonal temperature trends for all regions and the three time durations discussed are all positive, but not necessarily statistically significant. Within the year, for the entire 109 year period, the positive trends for all seasons and regions are statistically significant, except for the continental US in autumn.
- Shorter and essentially independent data sets covering the lower half of the atmospheric column over the past 60 years show increases in temperature similar to the surface data.

- Precipitation

- Seasonal and annual precipitation trends are mixed and vary geographically and over different time periods. The most robust trend is a general increase in precipitation in eastern portions of North America over the past full century (1–2 % per decade) and half century (2–3 % per decade). For North America as a whole, annual precipitation has increased by about 0.7 % per decade over the last century.
- Precipitation intensity has been increasing in eastern North America, less so in the northwest US, and very little in the US Southwest. The frequency of occurrence of heavy precipitation is increasing for event durations from 1 to 20 days, and return intervals for heavy precipitation are decreasing.
- Freezing levels have been rising. This has considerable implications for snowmelt driven stream flow regimes in mountain areas. Annual levels since 1995 have averaged about 100 m above the 65 year average for the Sierra Nevada.

- Snow cover
 - Spring snow cover for the entire Northern Hemisphere declined by 7 % in March and 11 % in April between the 1922 to 1970 period and the 1970–2010 interval. In March, most of this decrease was contributed by Eurasia, but in April both Eurasia and North America showed significant decreases in areal extent.
- Sea ice
 - Arctic sea ice volume and extent have decreased substantially, with the ice area in September (the month of annual minimum) shrinking by 13 % per decade from 1979 to 2012.
- Other key findings
 - Proxy indicators based on hydrologic or biological evidence widely indicate warming. These indicators often filter the climate effects of different seasons quite differently.
 - Patterns of ocean temperature and atmospheric circulation can vary on scales of a half century and complicate the interpretation of trends.
 - The climate system expression of reaction to changes in temperature or hydrology may well be in the form of preferred patterns. Such patterns are not spatially uniform, and do not occur with equal likelihood in all seasons.

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