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Climate and Terrestrial Ecosystem Change in the U.S. Rocky Mountains and Upper Columbia Basin

Historical and Future Perspectives for Natural Resource Management

Natural Resource Report NPS/GRYN/NRR—2010/260

ON THE COVER

Grinnell Glacier, Glacier National Park
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Climate and Terrestrial Ecosystem Change in the U.S. Rocky Mountains and Upper Columbia Basin

Historical and Future Perspectives for Natural Resource Management

Natural Resource Report NPS/GRYN/NRR—2010/260

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Executive Summary

Paleoenvironmental records provide critical information on past climates and the response of ecosystems to climatic variability. Ecosystems have changed in a variety of ways as a result of past climate change, and they will continue to do so in the future. At large scales, climate governs the distribution of vegetation across the landscape and acts as a strong control of important biophysical conditions (e.g., extent of mountain glaciers) and ecosystem processes (e.g., area burned by fire). Paleoenvironmental and instrumental records from throughout the western United States suggest that vegetation response to climate change varies along a hierarchy of temporal and spatial scales, and the responses range from wholesale shifts in biomes to small adjustments in forest density or structure. Anticipating how ecosystems may respond to ongoing and future climate change requires an understanding of the climate–ecological linkages on all these scales as well as cross-scale interactions that lead to abrupt responses and regime shifts.

Vegetation changes occurring on millennial time scales are related to changes in the seasonal cycle of solar radiation and its attendant effect on atmospheric circulation patterns and surface energy balances. After the ice sheets and local glaciers retreated 17,000–12,000 cal yr BP (calibrated years before present, equal to the number of calendar years before 1950), paleoenvironmental records from the Pacific Northwest and Rocky Mountains reveal a sequence of vegetation changes as a result of increasing temperatures and effectively wetter conditions. Initially, deglaciated regions were colonized by tundra communities and the climate was colder and probably drier than at present. After 14,000 cal yr BP, warmer and wetter conditions allowed present-day conifer taxa to expand first in open parkland and later as closed forest communities. By 11,000 cal yr BP, closed subalpine forests were widespread. The early Holocene (11,000–7000 cal yr BP) was a time of warmer and drier summer conditions than at present. Warmer temperatures led to an upslope shift in conifer ranges and xerothermic shrub communities occupied valley bottoms. Summer drought led to higher fire frequencies than at present in many ecosystems. A gradual cooling and increase in effective moisture in the mid-Holocene (7000–4000 cal yr BP) was followed by relatively cool, moist conditions (4000 cal yr BP–present).

Embedded with these millennial scale changes are centennial climate variations such as the Medieval Climate Anomaly (ca. AD 950–1250, AD = Anno Domini) and the Little Ice Age (ca. AD 1400–1700). These variations had less dramatic impacts on vegetation, but records describe shifts in ecotone positions, including the upper and lower treeline, and disturbance regimes. Gridded tree-ring networks suggest that within these intervals were multidecadal “megadroughts” associated with tree mortality and fire. On annual to decadal scales, climate variations have led to disturbance events and shorter droughts. These events have shaped successional pathways, tree growth and mortality, and community structure. Case studies investigating ecological response to these changes provide important lessons for understanding how ecosystems may respond to ongoing and future environmental change:

- The last century is an inadequate reference period for considering future climate change because it does not capture the range of natural climate variability that vegetation responds to or the magnitude of climate change projected for the near future. For example, managers often rely on the last several decades of fire occurrence as a baseline for managing different ecosystems in the West even though, because of fire suppression policies and fire elimination, fire activity of the last century is atypical of long-term historical patterns in much of the West and is unlikely to represent future conditions. Many vegetation types have evolved under a wide range of fire frequencies and intensities, calling into question the value of a static view of a fire regime, e.g., characterized by a mean fire return interval. To understand the full range of conditions that may be important for sustaining ecosystems in the future, long-term records of fire (e.g., centuries to millennia) provided by tree-ring and lake sediment data are essential.

- Rapid climate transitions have occurred in the past and will likely occur in the future. In the past, the response of vegetation has been highly variable, suggesting an equally complex response to future climate change. Among the likely outcomes will be a highly individualistic response by different species, a reorganization of plant communities, and the likelihood of differential lags in the ability of species to stay in equilibrium with climate change.

- Climate variability at large scales is often expressed in complex and asynchronous patterns across the U.S. Rocky Mountains and Upper Columbia Basin, largely because of interactions with topography and other sources of spatial heterogeneity. This can result in nearby communities showing different directions of change (e.g., precipitation regimes changing in opposite directions at different elevations).

- Many western ecosystems represent assemblages that formed as a result of a specific sequence of climate conditions during the last several thousand years. In particular, many middle and late-successional communities were established during the colder, wetter Little Ice Age and would be unlikely to form under current climate and land use conditions. Likewise, restoration to historical baselines is at best challenging and in most cases impossible.

Temperatures are projected to warm 1–5°C for much of the West by 2100, accompanied by declines in snowpack, earlier spring snowmelt, and reduced late-summer flows. While projections for future precipitation are less certain, increased precipitation is unlikely to offset increased evapotranspiration associated with even modest warming (e.g., 1–2°C), particularly during the summer. Consequently, drought is projected to increase in frequency and intensity over the next several decades, particularly in the Southwest and southern U.S. Rockies.

Investigations of the past suggest that we should expect dynamic and rapid ecosystem response to changing climate conditions, but in ways that may be difficult to predict. Paleoenvironmental records illustrate that while existing biomes have experienced distribution shifts, they have been resilient to climatic change across multiple time scales (e.g., decadal to millennial). They also suggest, however, that we should anticipate increased extreme and unforeseen disturbance synergisms; increased tree mortality, shifts in treeline position, and non-native plant invasions; and ultimately changes in plant community composition, structure, and function that may constitute novel vegetation assemblages. This poses significant challenges for developing management plans, but should not deter an adaptive management approach that allows for reassessment and modification of management strategies in the coming decades. By using scenario planning, managers can consider a wide range of possible future conditions to examine potential trajectories of ecological change. Managing for future conditions will, at the least, involve a continuation of well-established resource management and conservation practices.

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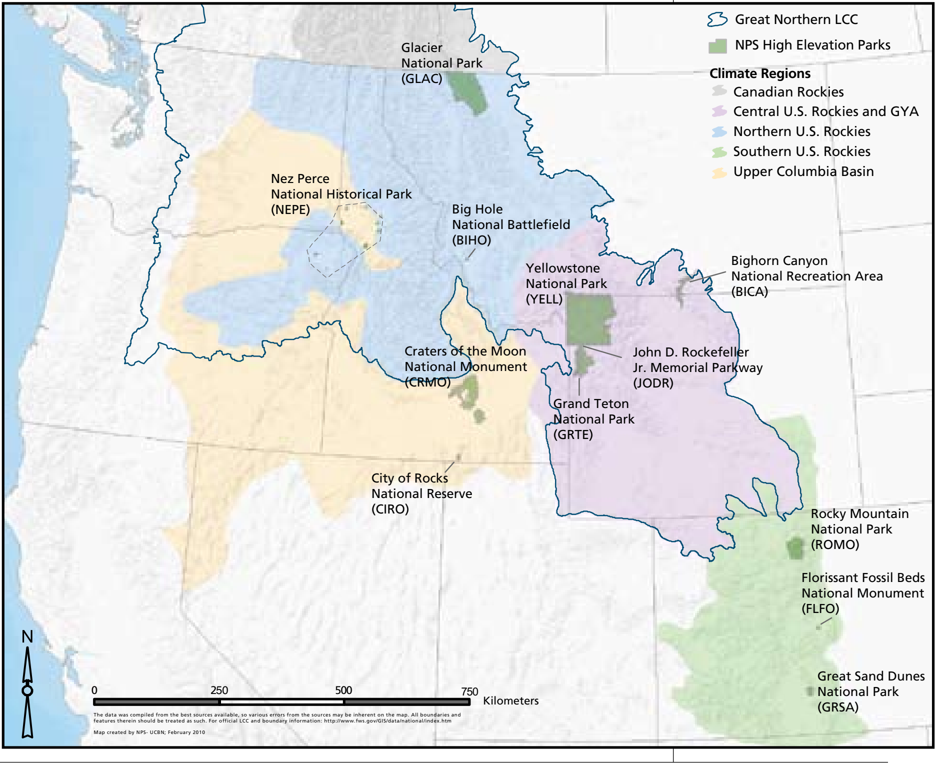
Chapter 1: Introduction

At large spatial and temporal scales, climatic conditions act as primary controls shaping the structure and distribution of ecosystems and the species they support. Changes in climate have dramatically altered ecosystem dynamics by shifting plant communities, creating opportunities for recruitment of new species, and restructuring land-surface processes and nutrient cycles (Solomon et al. 2007). The controls of climate vary at different time scales (i.e., millennial, centennial, multidecadal, decadal, annual and interannual), and the ecological response to climate change varies accordingly. Paleoecological data show that ecosystems in the West have undergone significant and sometimes rapid changes since the Last Glacial Maximum (ca. 20,000 years ago), and the biotic assemblages observed today are relatively recent phenomena (Thompson et al. 1993; Whitlock and Brunelle 2007; Jackson et al. 2009a). The future is also likely to be characterized by rapid biotic adjustment, including the possibility of novel assemblages as species respond individualistically to climate change (MacDonald et al. 2008; Williams et al. 2007). Understanding the drivers and rates of past climate change and the sensitivity of ecosystems to such changes provides critical insight for assessing how ecological communities and individual species will respond to future climate change (MacDonald et al. 2008; Shafer et al. 2005; Whitlock et al. 2003; Overpeck et al. 2003; Swetnam and Betancourt 1999).

The purpose of this report is to provide land and natural resource managers with a foundation of both climate and ecosystem response information that underpins management-relevant biophysical relationships likely to play an important role over the coming decades. We begin by synthesizing the climate and vegetation history over the last 20,000 years following the retreat of late Pleistocene glaciers. This time span provides examples of ecosystem responses to long-term (e.g., millennial) climate warming as well as several well-known periods of rapid climate change (i.e., substantial decadal to centennial scale climate perturbations). To contextualize past climate and ecosystem changes, and to provide a best estimate of future climate conditions, we also report on the most current statistically and dynamically downscaled Global Climate Model (GCM) projections of future changes in key climate variables (e.g., precipitation, temperature, snow water equivalent [CIG 2010]). Overall, our objective is to use the past to highlight a range of climate-driven biophysical responses to illustrate potential system trajectories and associated uncertainties under future climate conditions.

To meet the requested needs of the National Park Service (NPS), the U.S. Forest Service and the U.S. Fish and Wildlife Service, the geographic scope of this report encompasses the core regions of the Great Northern Landscape Conservation Cooperatives and the NPS high-elevation parks of the Rocky Mountains. For organizational purposes, the report divides the study area into four regions: the northern U.S. Rocky Mountains, the central U.S. Rocky Mountains and Greater Yellowstone Area, the southern U.S. Rocky Mountains, and the Upper Columbia Basin (fig. 1).

Figure 1. Location of study area. Climate region boundaries modified from Littell et al. (2009a), Kittell et al. 2002, and Bailey's (1995) ecoprovince boundaries. Climate regions represent coarse aggregations of biophysical constraints on modern ecological assemblages and the interaction between climate, substrate, elevation, and other conditions. (Figure created by M. Lonneker, NPS, 2010.)


Great Northern LCC
NPS High Elevation Parks
Climate Regions
Canadian Rockies
Central U.S. Rockies and GYA
Northern U.S. Rockies
Southern U.S. Rockies
Upper Columbia Basin
Glacier National Park (GLAC)
Nez Perce National Historical Park (NEPE)
Big Hole National Battlefield (BIHO)
Yellowstone National Park (YELL)
Bighorn Canyon National Recreation Area (BICA)
Craters of the Moon National Monument (CRMO)
John D. Rockefeller Jr. Memorial Parkway (JODR)
Grand Teton National Park (GRTE)
City of Rocks National Reserve (CIRO)
Rocky Mountain National Park (ROMO)
Florissant Fossil Beds National Monument (FLFO)
Great Sand Dunes National Park (GRSA)
N
0
250
500
750
Kilometers

The data was compiled from the best sources available, so various errors from the sources may be inherent on the map. All boundaries and features therein should be treated as such. For official LCC and boundary information: http://www.fws.gov/GIS/data/national/index.htm

Map created by NPS- UCBN; February 2010

The synthesis is organized into four sections: (1) biophysical responses and drivers of climate changes occurring on multi-centennial to millennial scales during the last 20,000 years; (2) biophysical responses and drivers of climate change on annual to centennial scales over the last 2000 years; (3) the last century of climate and ecosystem change as observed by high-resolution instrumental records; and (4) the next century of likely future climate and ecosystem changes under a range of greenhouse gas emission scenarios. For each section, we present the large-scale and regional drivers of climate change as inferred from GCMs and paleoenvironmental data. We then detail the associated biophysical and ecological responses documented in both modern and paleoecological proxy data with a focus on the implications for maintaining key resources in the face of changing conditions. The synthesis concludes with a discussion of challenges in planning for future conditions where there is high uncertainty about climatic change and ecological response, and provides a planning approach designed to address a wide range of potential conditions. The purpose of this review is to highlight important climate and ecosystem linkages using records relevant to the regions of great conservation and natural resource management value shown in figure 1. Accordingly, this document does not necessarily provide a comprehensive review of the suite of available climate and ecosystem-related research available for the entire study region.

1.1 Climate controls and variability at different spatio-temporal scales

The influence of variation in climate on ecosystems changes at different spatial and temporal scales. Understanding the potential influence of climate on biophysical processes often requires local to synoptic (regional or larger) information on the type and magnitude of forcings (mechanisms driving changing conditions), e.g., relative humidity-related changes versus changes in solar radiative output, long-lived greenhouse gasses, and ocean-atmosphere interactions and local albedo (low albedo = low reflectance or dark ground surface such as pavement; high albedo = high reflectance and lighter surface such as snow) along with an understanding of the sensitivity of the ecosystem property that is being measured (Overpeck et al. 2003; Webb and Bartlein 1992) (Table 1). It is also important to recognize the hierarchical nature of climate variation and change, and that short-term (i.e., daily to interannual) events are superimposed on longer ones, amplifying or dampening the magnitude of the underlying physical controls that influence ecosystem dynamics.

At the longer temporal range of major changes in Earth's climate system, variations on scales of 10,000 to 100,000 years are attributed to slowly varying changes in Earth's orbit known as Milankovitch Cycles (i.e., changes in Earth's precession, tilt, and obliquity [Hays et al. 1976, Berger 1978, Berger

Table 1. Climatic variation at different time scales and biotic response (modified from Overpeck et al. 2003).

Frequency and Scale of Variation (years)	Kind of Variation	Forcing Mechanism	Biotic Response
Millennial (1000–10,000)	Deglacial and postglacial variations	Ice sheet size, insolation, trace gases, regional ocean-atmosphere-ice interactions	Species migration, range expansion and contraction; community reorganization and establishment, species extirpation/extinction
Interdecadal–centennial (10–100)	Decadal and centennial anomalies	Internal variations in the climate system, solar variability, volcanism	Shifts in relative abundance and composition of different taxa through recruitment, mortality, and succession
Annual–interannual (<10)	Storms, droughts, ENSO events	Internal variations in the climate system, solar variability, volcanism	Adjustments in physiology, life history strategy, and succession following disturbance

and Loutre 1991], fig. 2). The net result of slowly varying changes in Earth's orbit have included multiple glacial and inter-glacial cycles driven by changes in global average temperatures over the past several million years. These temperature changes are initiated by changes in the seasonal and annual cycle of insolation (incident solar radiation; the amount of solar radiation received on a given surface area during a given time) over the high latitudes of the Northern Hemisphere that result in substantial positive feedbacks from changing concentrations of atmospheric greenhouse gasses (Vettoretti

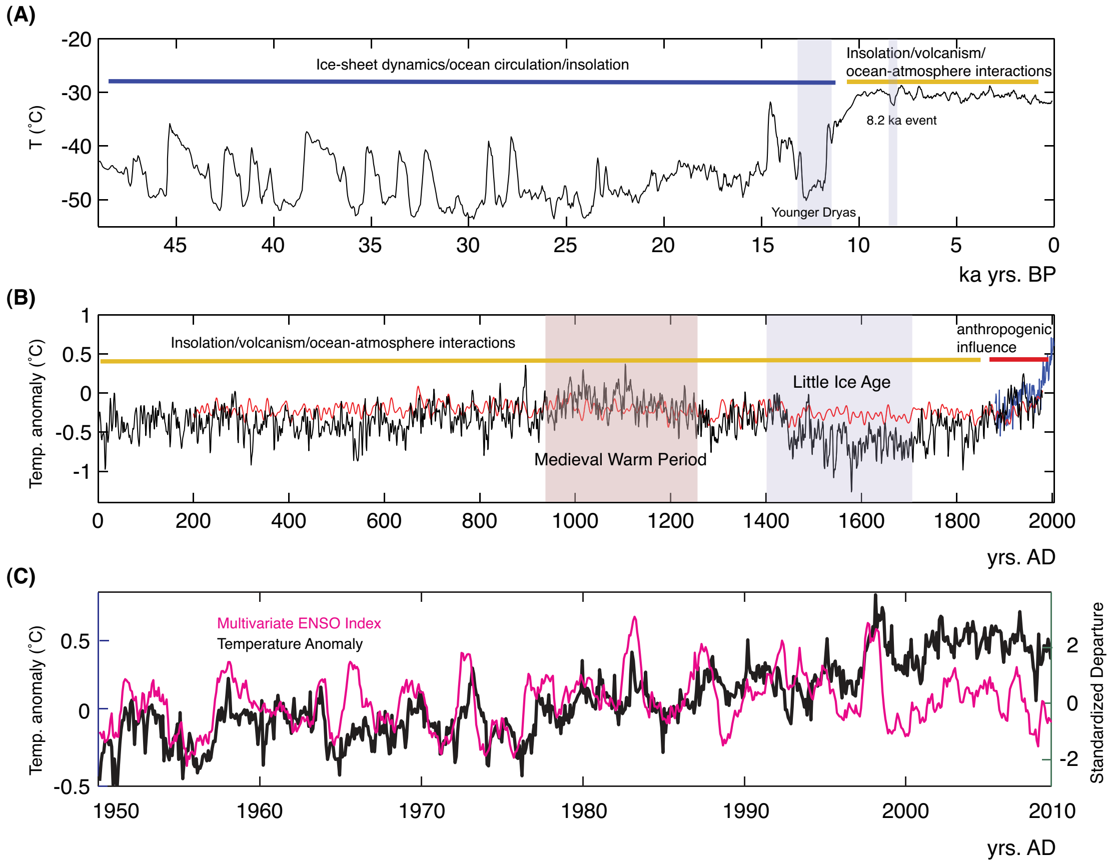


Figure 2. Primary drivers of climate and resulting climate variations at millennial, centennial, and interannual scales. (A) Temperature reconstruction from the central Greenland (GISP2) ice core record and the forcing mechanisms thought to influence variation during the glacial period (ca. 49,000–12,000 cal yr BP). The Laurentide ice sheet in North America began to recede and climate warming commenced ca. 17,000 cal yr BP. Gray bands indicate two abrupt climate changes: the Younger Dryas Chronozone ca. 12,900–11,600 BP, and the 8200 cal yr BP cool event (8.2 ka event); (B) Temperature anomalies for the Northern Hemisphere based on multiple proxy data (e.g., ice core, ice borehole, lake sediment, pollen, diatom, stalagmite, foraminifera, and tree-ring records) from Moberg 2005 (black line), Mann and Jones (2003, red line), and the instrumental record (blue) for the past 2000 years (Viau et al. 2006). Continental patterns of drought and interannual and decadal climate variability are associated with the Medieval Climate Anomaly (ca. 950–1250) and fewer fires during the Little Ice Age (ca. 1400–1700). (Time spans from Mann et al. 2009 but varies by region [Cook et al. 2004, MacDonald et al. 2008, Bradley et al. 2003, Carrara 1989]); (C) Recent global temperature anomalies (black line, based on 1900–2000 mean) from HadCRUT3v instrumental reconstruction (Brohan et al. 2006) and ocean-atmosphere variability. Magenta line represents the Multivariate El Niño–Southern Oscillation (ENSO) Index, which is based on six observed ocean-atmosphere variables. Positive values of the index depict El Niño events (Wolter and Timlin 1993, 1998). Source: NOAA Earth System Research Laboratory.

and Peltier 2004; Bond et al. 2001; Kutzbach and Guetter 1986; Kutzbach et al. 1998, 1993).

On multi-millennial scales (here specific to the last 20,000 years), the presence or absence of the large North American ice sheets, particularly the Laurentide ice sheet results in ocean-ice-atmosphere interactions that drive changes in atmospheric circulation patterns (i.e., the position of westerlies and preferential positioning of storm tracks [fig. 2a]), resulting in major changes in ecosystem distribution and structure. For example, as Northern Hemisphere summer insolation increased and the ice sheets and glaciers began to retreat, seasonal storm tracks shifted north and paleoecological records show widespread reorganization of plant communities throughout the West (e.g., MacDonald et al. 2008; Thompson et al. 1993; Bartlein et al. 1998; Whitlock and Brunelle 2007; Jackson et al. 2005; Betancourt et al. 1990). These large-scale and long-term changes in insolation and ice cover are important features of Earth's climate dynamics because they influence the persistence and strength of storm tracks, subtropical high-pressure systems, ocean-land temperature gradients, and consequently interannual to decadal scale drivers of climate variability such as the El Niño–Southern Oscillation (ENSO). For example, higher-than-present summer insolation in the Northern Hemisphere during the early Holocene (ca. 11,000–7000 cal yr BP) led directly to increased summer temperatures and indirectly to a strengthening of the Pacific subtropical high-pressure system off the northwestern United States, effectively intensifying summer drought in the region (Bartlein et al. 1998). Records of early Holocene glacier dynamics, lake levels, aeolian activity (blowing dust), vegetation, and fire show the ecological effects of this increased summer insolation at local to subcontinental scales (e.g., Whitlock and Brunelle 2007; Whitlock et al. 2008; Jackson et al. 2009a; Luckman and Kearney 1986; Osborne and Gerloff 1997; Rochefort et al. 1994; Graumlich et al. 2005; Fall 1997; Booth et al. 2005; Dean et al. 1996; Dean 1997).

On shorter and perhaps more management-relevant time scales, climate variations at interdecadal to centennial scales are related to changes in solar activity, volcanism, sea-surface temperature, and pressure anomalies in both the Atlantic and Pacific oceans. More recently, important contributions arise from rapidly increasing atmospheric greenhouse gas concentrations (Barnett et al. 2008; fig. 2b,c). The Medieval Climate Anomaly could be considered an example of centennial-scale climate variation due to the relatively warm and dry conditions that prevailed across the western United States from approximately 900 to 1300. The West experienced substantially reduced streamflows (Meko et al. 2007), shifts in the upper treeline (Rochefort et al. 1994; Fall 1997), and increased fire activity (Cook et el. 2004). Though the exact causes of the MCA are still debated, the prevailing evidence suggests that it was driven by changes in solar activity, volcanism, and perhaps sustained La Niña-like conditions in the tropical Pacific (Mann et al. 2009). At decadal to interdecadal scales, sustained sea surface temperature anomalies in the north Pacific and Atlantic oceans appear to be important drivers of climate variability across western North America (e.g., McCabe et al. 2004; Einfeld et al. 2001). The major indices that capture these modes of interdecadal variability include the Pacific Decadal Oscillation (PDO; see Mantua et al. 1997) and the Atlantic Multidecadal Oscillation (AMO; see Enfield et al. 2001, Regonda et al. 2005). Decadal climate shifts associated with changes in the PDO and AMO are well expressed in 20th century records of drought and winter precipitation (e.g., Cayan et al. 1998; McCabe et al. 2004), as well as in proxy-based reconstructions of precipitation and streamflow (fig. 2c; e.g., Gray et al. 2003). These events are often regional to subcontinental in scale and initiate and terminate within years, but often have widespread physical and ecological effects (e.g., Allen and Breshears 1998; Bitz and Battisti 1999; Pederson et al. 2004; Watson and Luckman 2004). Examples include widespread bark beetle outbreaks, increased forest fire activity and stress-related tree mortality, and rapid changes in glacier mass balance, snowpack, and streamflow.

The El Niño–Southern Oscillation (ENSO) is a major global control of both temperature and moisture patterns (fig. 2c). ENSO events are defined by changes in atmospheric pressure gradients across the tropical Pacific that are related to patterns of warming (El Niño) and cooling (La Niña) in the central and eastern equatorial Pacific which typically last 6–18 months and reoccur every 2–7 years (Bjerknes 1969; Cane and Zebiak 1985; Graham and White 1988; Horel and Wallace 1981; Philander 1990; Chang and Battisti 1998). The magnitude of these sea surface temperature anomalies varies, but they typically exert a substantial influence on regional temperature and precipitation patterns (McCabe et al. 2004; Einfeld et al. 2001). For example, El Niño events typically result in warm dry conditions across the northwestern United States and a southerly displacement of the winter storm track (i.e., the jet stream), resulting in cool wet conditions across the Southwest (McCabe et al. 2004; Enfield et al. 2001). The inverse is typically true for La Niña-like conditions, but in all cases this mode of climate variability appears to exert its strongest influence across the Southwest (e.g., Swetnam and Betancourt 1998), with an important but somewhat attenuated spatial signature across the Northwest (e.g., Dettinger and Ghil 1998).

In summary, climate-ecosystem linkages are evident across many time scales, from individual records as well as regional and global compilations. Shifts in species distributions and abundance are a response to climate variations occurring over seasons to millennia. Knowledge of climate drivers on all time scales is necessary to identify temporal and spatial dimensions of future changes and possible ecological responses to those changes.

1.2 How can understanding climatic variability inform management?

Knowledge of natural variations in climate at different scales provides a context for understanding how communities and individual species might respond to current and future rates of change (Table 2). For example, past changes in fire regimes have been largely driven by large-scale climate changes on millennial, centennial, decadal and annual scales (Whitlock et al. 2003; Kitzberger et al. 2007; Littell et al. 2009a; Westerling et al. 2006). In ecosystems where fire regimes are expected to change with future climate conditions, management efforts should focus on the ecological response to rapidly changing conditions as opposed to maintaining current or past conditions (Whitlock et al. 2010). Additionally, paleoenvironmental records showing evidence of rapid changes in climate and attendant ecological responses suggest that even small changes in climate can have large consequences and provide an important context for anticipating ecosystem response to future climate change (Whitlock and Brunelle 2007; Gray et al. 2006, 2004; MacDonald et al. 2008; Lyford et al. 2003).

Table 2. Examples of paleoenvironmental proxy data, spatio-temporal resolution, and range and type of reconstruction.

Proxy	Source	Temporal Resolution	Spatial Resolution	Temporal Range	Type of Reconstruction
Tree growth	Tree cores	High (seasonal/annual)	High	100–1000s yrs	Temperature, moisture
Charcoal	Lake/ peat sediments	High to moderate (annual–decadal)	High to moderate (1 to several km^2)	High (many millennia)	Fire
Pollen	Lake/peat sediments	Moderate (multidecadal–centennial)	Moderate (several km^2)	High (many millennia)	Vegetation
Oxygen Isotopes	Corals, tree, lake, ocean or ice cores	High to moderate (annual–decadal)	Moderate to low	High to very high (100,000+)	Temperature, ice volume

Reconstructing past environments using proxy data

Reconstructing past climatic conditions and the associated ecological response involves a number of direct and indirect measurements. Direct measurements include ground temperature variations, gas content in ice core air bubbles, ocean sediment pore-water change, and glacier extent changes. Indirect measurements or paleoclimate proxy typically come from organisms that respond to changes in climate through changes in their growth rate, abundance, or distribution as recorded in living or fossil specimens or assemblages of organisms. Each proxy indicates past change at different spatial and temporal scales and resolutions.

Obtaining lake sediment core, Rocky Mountain National Park. Photo courtesy P. Higuera.

Lake sediment, tree-ring cores, and packrat middens are three of the primary proxies used for reconstructing past conditions for the western United States (Whitlock and Larsen 2001; Fritts and Swetnam 1989; Betancourt 1990). Lake sediment cores often provide some of the longest records of vegetation and fire through analysis of pollen, plant macrofossils, and charcoal particles at intervals throughout the core. Most lakes in the Northwest were formed during deglaciation, and therefore provide a sedimentary record spanning the last 15,000 years or longer, depending on the time of ice retreat. Sediment cores are retrieved from modern lakes and wetlands using anchored platforms in summer or from the ice surface in winter. Samples for pollen and charcoal analyses are removed from the cores at intervals (e.g., every 0.5–1 cm) that depend on the detail and temporal resolution required. The pollen extracted from the sediment is chemically treated, identified under the microscope, and tallied for each sediment level sampled. Pollen counts are converted to percentages of terrestrial pollen and plotted as a diagram.

The reconstruction of past vegetation and climate from pollen percentages rests on the relationship between modern pollen rain and present-day vegetation and climate. Modern pollen samples have been collected at lakes throughout North America, and this information is calibrated to modern vegetation and climate. Past fire activity is inferred from the analysis of particulate charcoal, which is extracted and tallied from the sediment cores (Whitlock and Bartlein 2004). High-resolution charcoal analysis involves extraction of continuous samples from the core such that each sample spans a decade or less of sediment accumulation. These samples are washed through sieves and the charcoal residue is tallied under a microscope. Examining these relatively large particles enables a local fire reconstruction because large particles do not travel far from a fire. Charcoal counts are converted to charcoal concentration (particles/cm^3), which is then divided by the deposition time of each sample (yr/cm) to yield charcoal accumulation rates (particle/cm^2/yr). Detection of fire events involves identification of charcoal accumulation rates above background levels.

Tree rings, which provide records of past change at centennial and millennial scales, have several features that make them well suited for climate reconstruction, such as ease of replication, wide geographic availability, annual to seasonal resolution, and accurate, internally consistent dating.

Obtaining tree-ring core, Crazy Mountains, Montana. Photo D. McWethy.

Networks of tree-ring width and density chronologies are used to infer past temperature and moisture changes based on calibration with recent instrumental data, recording centennial to millennial change. Tree growth is highly sensitive to environmental changes and therefore tree-ring records are powerful tools for the investigation of annual to centennial variations. Tree-ring chronologies are used to reconstruct past growing season temperature and precipitation. The most sensitive trees are those growing in extreme environments where subtle variations in moisture or temperature can have a large impact on growth. For example, precipitation and/or drought reconstructions are often derived from extremely dry sites or sites at forest-grassland boundaries where moisture is the strongest limiting factor on growth. Similarly, sites at altitudinal and latitudinal treelines with ample moisture are often targeted for temperature-sensitive chronologies. The year-to-year variability in individual tree-ring width series (or other tree-ring parameters such as density) from long-lived stands of trees are combined to produce site histories or chronologies that span centuries or millennia. These chronologies contain considerable replication (e.g., two cores per tree, minimally 10–15 trees per site) and dating accuracy is rigorously verified by comparing ring-width patterns among trees. This cross-dating also allows tree-ring series from ancient dead wood (found in dwellings, lakes, sediments, and on the surface in cold, dry environments) to be combined with overlapping records from living trees, thereby extending records further back through time. Statistical relationships established between annual tree-ring width chronologies and instrumental climate records are used to hindcast estimates of precipitation and temperature.

Middens left by woodrats of the genus *Neotoma* also provide long-term records and are often found in arid environments where other approaches for reconstructing past environments are less viable. When packrats build nests, plant and animal remains often become crystallized and mummified in packrat urine, preserving rich deposits of macrofossils that can be used to reconstruct vegetation and climate. Middens located in caves or under rock ledges that provide protection from water are especially well preserved. The plant and animal parts from an excavated midden are dissected and identified, and then dated using radiocarbon techniques. A single midden typically represents a relatively discrete time interval when material was accumulated (one to several decades, [Finley 1990]), but a network of middens within one site can be stacked chronologically to provide a record of vegetation and climate change over a longer period. A reconstruction of vegetation typically includes the area within 30 to 100 meters (33–109 yd) surrounding a site (Betancourt et al. 1990).

Packrat midden. Photo courtesy USGS.

Chapter 2: The last 20,000 years of environmental change in the western United States

2.1 Drivers of millennial-scale climate variation

The climate variations of the last 20,000 years occurring on millennial scales are best understood through model-based simulations that look at the regional response to large-scale climate changes and paleoenvironmental data that measure specific components of climate change. Broad-scale climate variations were described by Bartlein et al. (1998) using the NCAR Community Climate Model (CCM1; 4.4º latitude by 7.5º longitude spatial resolution, mixed-layer ocean, crude depiction of western cordillera topography). The simulation provided estimates of climatic conditions over six discrete periods (21,000 cal yr BP=full glacial period with full-sized ice sheets; 16,000 to 14,000 cal yr BP =late Glacial period with shrinking ice sheets; 11,000 cal yr BP = early Holocene insolation maximum; 6000 cal yr BP= mid-Holocene transition; and present). Other recent regional-scale modeling studies have provided better temporal and spatial resolution with more realistic topography (Hostetler 2009; Hostetler et al. 2003). Results from these efforts are summarized below.

At the time of the Last Glacial Maximum (ca. 21,000 cal yr BP), the large Laurentide and Cordilleran ice sheets strongly influenced climatic conditions in the western United States (Bartlein et al. 1998; fig. 3), depressing temperatures approximately 10°C in areas south of the ice sheets and steepening the latitudinal temperature gradient. The presence of the large ice sheets also displaced the jet stream south of its present position, greatly reducing winter precipitation in the northwestern United States and Canada while increasing precipitation across the Southwest. Another element of the full-glacial climate was stronger than present surface easterlies related to a strong high-pressure system that persisted over the ice sheets. The presence of this strong high-pressure system steepened the latitudinal temperature gradient and weakened the westerly storm tracks, resulting in colder and effectively drier conditions across the Northwest and cold wet conditions in the Southwest (Hostetler et al. 2000; Bartlein et al. 1998; Thompson et al. 1993).

2.2 Glacial–Holocene transition

During the Glacial–Holocene transition (16,000–11,000 yr BP), solar insolation over high-latitude Northern Hemisphere landmasses increased, peaking ca. 11,000 yr BP when summer insolation was 8.5% higher and winter insolation was 10% lower than at present at 45ºN latitude. One consequence was a northward shift of winter storm tracks from their full-glacial position, bringing wetter winter conditions to the Northwest while the Southwest became increasingly dry (Bartlein et al. 1998). Increasing summer insolation resulted in warmer growing season temperatures, causing alpine glaciers and ice sheets to melt rapidly. (See fig. 4 for an example of modern air mass circulation.)

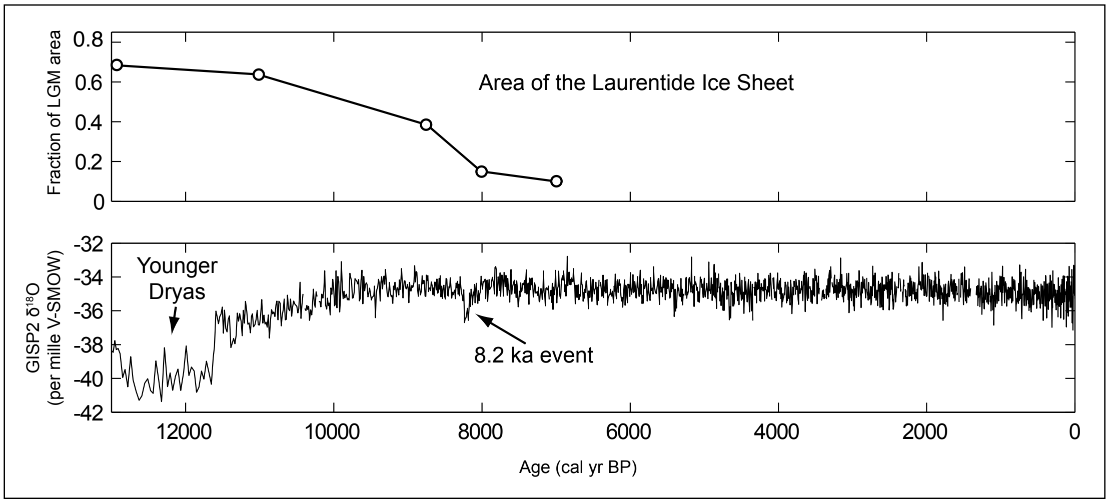


Figure 3. Area of the Laurentide ice sheet (top panel) and central Greenland temperature reconstruction (bottom panel). Ice sheet area estimated from Dyke and Prest (1987) and Barber et al. (1999); oxygen isotope record (bottom panel, black line) associated with variations in Northern Hemisphere temperature (higher isotope values represent warmer temperatures) from GISP2 (Stuiver et al. 1995). The Younger Dryas Chronozone was an abrupt cooling ca. 12,900–11,600 BP which represented a temporary reversal in warming during the Pleistocene–Holocene transition (Alley et al. 1993). (Figure modified from Shuman et al. 2002).

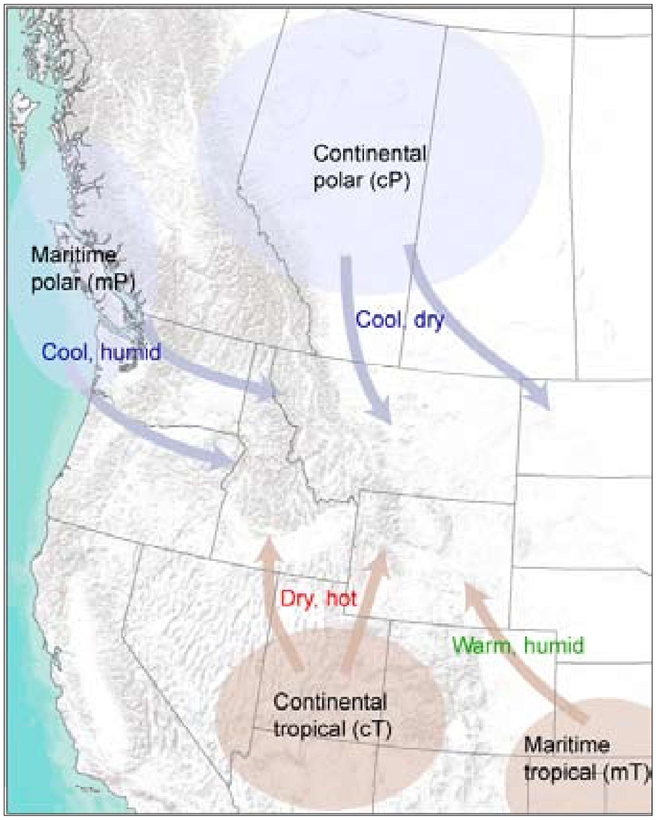


Figure 4. Primary air masses that influence the study area. Ice-sheet dynamics and ocean-atmosphere interactions have altered large-scale air mass circulation patterns throughout the late Glacial and Holocene, influencing temperature and precipitation. The U.S. Rocky Mountains lie in a transition zone where many of these air masses interact, strongly influencing the local and regional climate. Changes in the position and strength of these air masses, driven by large-scale changes in the climate system, are responsible for climatic variations on different time scales and act as an important control on vegetation. (Modified from Ahrens 2008 and Aguado and Burt 2010.)

At the end of the Pleistocene (the last 2.5 several million years of repeated glaciations prior to the Holocene), much of the Northern Hemisphere experienced an abrupt cooling known as the Younger Dryas Chronozone (YDC, ca. 12,900–11,600 BP) (Alley et al. 1993). This event is clearly registered in the North Atlantic region and across Europe, and is related to changes in ocean circulation during the melting of the Laurentide ice sheet. Evidence of the YDC is less obvious in the western United States, where most paleoenvironmental data show little or no response in terms of glacial activity (Licciardi et al. 2004; Heine 1998), vegetation change (Grigg and Whitlock 2002; Briles et al. 2005; Brunelle et al. 2005; Huerta et al. 2009), or shifts in fire activity (Marlon et al. 2009). In some areas, however, there is evidence of a re-advance of mountain glaciers (Osborn and Gerloff 1997; Reasoner and Huber 1999; Friele and Clague 2002; Menounos and Reasoner 1997) and vegetation changes that indicate cooling, including a lower treeline in central Colorado (Reasoner and Jodry 2000) and changes in the isotopic signature in speleothem data. Temperatures decreased as much as 5–10°C (9–18°F) in Greenland, but pollen data for the northwestern United States indicate much more moderate cooling (0.4–0.9°C [0.7–1.6°F]; Reasoner and Jodry 2000). The YDC ended rapidly with warming of ~7°C (13°F) in Greenland occurring within one to several decades (Alley 2000). Consequently, the period is closely scrutinized as an example of abrupt climate change (Alley et al. 2003).

A similar abrupt cooling occurred ca. 8200 cal yr BP (figs. 2–3) when a large influx of fresh water disrupted circulation in the North Atlantic, causing cooling that lasted several centuries in the North Atlantic region (Le Grande et al. 2006; Alley and Ágústsdóttir 2005). Like the YDC, the 8.2 ka event is not registered at many sites in the western United States, either because the signal is weak or the sampling resolution is inadequate to detect it. Geochemical proxies from lake sediments suggest that it has been associated with drier conditions at Bear Lake, Utah (Dean et al. 2006). The YDC and the 8.2 ka event illustrate how rapid climate changes due to ocean-atmosphere-ice interactions can occur. They also show a variable signal in regions distal to the North Atlantic origin, so that some sites in the western United States show a response while others do not.

Greater summer insolation (8% above present) and lower winter insolation (8% lower) in the early Holocene (11,000–7000 cal yr BP) profoundly affected the climate and ecosystems of the western United States. Increased summer insolation led to warmer temperatures throughout the region during the growing seasons, while winters were likely colder than at present. Model simulations show that increased insolation led indirectly to a strengthening of the eastern Pacific subtropical high-pressure system which suppressed summer precipitation in much of the Northwest. At the same time, it strengthened inflow of moisture from the Gulf of California to the Southwest and the southern and central Rocky Mountain region (Whitlock and Bartlein 1993), resulting in greater than present summer precipitation. East of the Rockies, increased summer precipitation was likely offset by increased temperatures and rates of evapotranspiration, making conditions effectively drier than at present, which is consistent with low lake levels and dune activation (Shuman et al. 2009; Stokes and Gaylord 1993).

Today the West is characterized by summer-dry areas under the influence of the subtropical high (i.e., the Northwest) and summer-wet areas where summer precipitation reflects monsoon activity (i.e., the Southwest). These two precipitation regimes are defined by topography (e.g., Yellowstone Plateau) and the boundary between them is relatively sharp (Whitlock and Bartlein 1993; Gray et al. 2004). The indirect effects of greater-than-present summer insolation strengthened both precipitation regimes, making summer-dry regions drier in the early Holocene and summer-wet regions wetter than at present (Thompson et al. 1993; Whitlock and Bartlein 1993; Bartlein et al. 1998). During the mid-Holocene (ca. 7000–4000 cal yr BP) and the late Holocene (4000 cal yr BP–present), summer insolation decreased and winter insolation increased gradually to present levels. Summer-dry regions became cooler and wetter, and summer-wet regions became cooler and drier than before.

2.3 Mid-Holocene transition

In the Pacific Northwest, including the Columbia Basin, the mid-Holocene was a transition period between the warm, dry, early Holocene and the cooler, wetter, late Holocene. In British Columbia, Hebda and Mathewes (1984) call this period the mesothermal and trace the expansion of hemlock (*Tsuga heterophylla*) and Douglas fir (*Pseudotsuga menziesii*) during this period. This signal is also evident in the northern Rockies and perhaps as far south as the Greater Yellowstone Area. In the southern and central Rockies, both summer-wet regions, paleoclimate data suggest several anomalously dry/wet periods during the Holocene (Shuman et al. 2009; Stone and Fritz 2006). Lake level data, for example, indicate that numerous sites throughout the Rockies experienced drier conditions during the mid-Holocene (ca. 6000 cal yr BP) than at present (Shuman et al. 2009), similar to climatic conditions across the Great Plains. Anomalously dry conditions for the central and southern Rockies also contrast with wetter-than-present conditions in the Southwest (Betancourt et al. 1990; Davis and Shafer 1992; Thompson et al. 1993; Fall 1997; Mock and Brunelle-Daines 1999; Harrison et al. 2003). Regional climate simulations from the Colorado Rockies shed some light on the drivers of mid-Holocene aridity; variations in the seasonal insolation cycle imposed local surface feedbacks (e.g., reduced snowpack and soil moisture) that were important drivers of submillennial-scale changes in precipitation and moisture (Shuman et al. 2009; Diffenbaugh et al. 2006).

2.3.1 Southern Canadian and Northern U.S. Rocky Mountains

Prior to 14,000 cal yr BP, much of the southern Canadian and northern U.S. Rocky Mountains were glaciated (fig. 5). The first plant communities to colonize

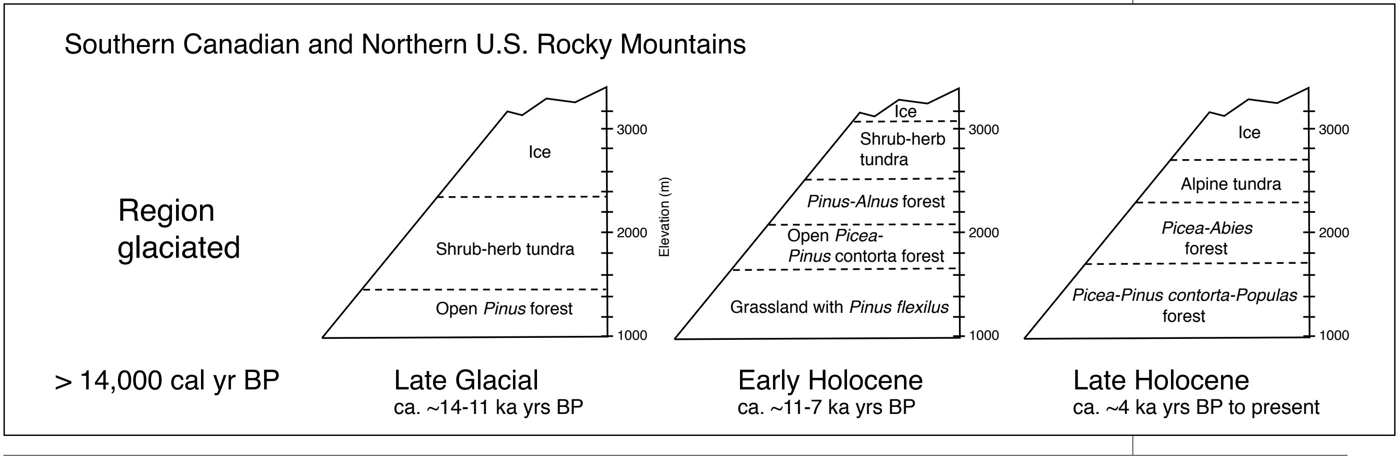

Southern Canadian and Northern U.S. Rocky Mountains
Region glaciated
> 14,000 cal yr BP
Ice
Shrub-herb tundra
Open *Pinus* forest
3000
2000
1000
Late Glacial
ca. ~14-11 ka yrs BP
Elevation (m)
Ice
Shrub-herb tundra
Pinus-Alnus forest
Open *Picea-Pinus* contorta forest
Grassland with *Pinus flexilus*
3000
2000
1000
Early Holocene
ca. ~11-7 ka yrs BP
Ice
Alpine tundra
Picea-Abies forest
Picea-Pinus contorta-Populas forest
3000
2000
1000
Late Holocene
ca. ~4 ka yrs BP to present

Figure 5. Ecological response to changing climatic conditions following glacial retreat in the southern Canadian and northern U.S. Rockies. Derived from MacDonald 1989 and Reasoner and Hickman 1989 (Whitlock unpublished).

deglaciated regions were alpine in character, composed of shrubs and herbs dominated by sagebrush, grasses, and alder (*Artemisia*, Gramineae, and *Alnus*) (Whitlock et al. 2002; MacDonald 1989; Reasoner and Hickman 1989). By the early Holocene, warmer-than-present conditions allowed shrub and herb communities to spread upslope to higher elevations than at present, and forests of pine and alder (*Pinus albicaulis/flexilis* and *Alnus*) developed at mid-elevations. The lesser abundances of spruce and pine (*Picea* and *Pinus contorta*) suggest warmer growing season conditions than at present. Low-elevation forests were characterized by spruce and lodgepole pine (*Pinus contorta*), and valley floors supported open grassland with limber pine. The upper treeline was at least 90 meters above present during the early- and mid-Holocene (ca. 8500–4000 cal yr BP). Low and middle elevations experienced the greatest fire activity during the early Holocene (ca. 9000–8000 cal yr BP) (MacDonald 1989) and forest composition resembled modern subalpine spruce-fir (*Picea-Abies*) forest. In the late Holocene, summer (growing) conditions became cooler, glacial activity increased, and the upper treeline decreased substantially (Reasoner and Hickman 1989). Cooler and moister conditions led to tundra at high elevations, spruce and fir forests at mid-elevations, and spruce, lodgepole pine, and aspen at low elevations.

Ecosystem response to centennial-scale climatic variations is evident from a 3800-year history of climate, vegetation, and ecosystem change inferred from pollen and charcoal concentrations in the lake sediment record from Foy Lake in northwestern Montana. Formed over 13,000 years ago as ice retreated from the Flathead Valley, the lake is situated at the eastern edge of the Salish Mountains, 3 kilometers southwest of Kalispell, Montana (Stevens et al. 2006). Several studies from the site provide historical reconstructions of climate and hydrologic variability and ecosystem response to climate change over the past several millennia (Stevens et al. 2006; Power et al. 2006; Shuman et al. 2009). Paleolimnologic and pollen data indicate that ca. 2700 cal yr BP, an abrupt rise in lake levels coincided with a transition from steppe and pine forest to pine forest-woodland to mixed conifer forest (Power et al. 2006), a transition linked to an increase in effective moisture (winter precipitation) shown in lake level records (Stevens et al. 2006; Shuman et al. 2009). Following the establishment of mixed conifer forests, lake levels decreased from 2200 to 1200 cal yr BP, and increases in grass, pine, and sagebrush and declines in Douglas fir and larch led to the development of a steppe/parkland/forest mosaic ca. 700 cal yr BP (Power et al. 2006; Stevens et al. 2006). Increases in grass and sagebrush in the late 19th and early 20th centuries coincided with human activities. Notable climatic events during this period include a long, intense drought ca. 1140 following a wetter period from 1050 to 1100 (Stevens et al. 2006).

These shifts in vegetation were accompanied by pronounced changes in fire patterns as evident in the charcoal record (fig. 6). Intervals dominated by forests coincide with high magnitude and frequent fires (e.g., stand replacing fires), periods dominated by steppe-parkland vegetation are associated with smaller and less frequent fires, and a decline in charcoal deposition in the last century likely reflects the impact of fire suppression (fig. 6). The Foy Lake record demonstrates the impacts of centennial-scale climate variations and their associated ecosystem response during the last two millennia. While relatively modest changes in vegetation cover occurred after the conifer forests were established ca. 2700 cal yr BP, multidecadal shifts in climate are evident in the fire reconstruction for the last several millennia.

2.3.2 Central U.S. Rocky Mountains and the Greater Yellowstone Area

Pollen records from Yellowstone and Grand Teton national parks show the nature of the biotic change that occurred in conjunction with broad climatic changes at different elevations (fig. 7, Whitlock 1993). Deglaciation (ca. 17,000–14,000 BP) was followed by colonization of tundra vegetation which, during the late Glacial and into the mid-Holocene, was replaced by subalpine communities of spruce, fir, and pine in many regions, first as an open parkland and then as a closed forest. With the warmer growing season conditions of the early Holocene, pine, juniper

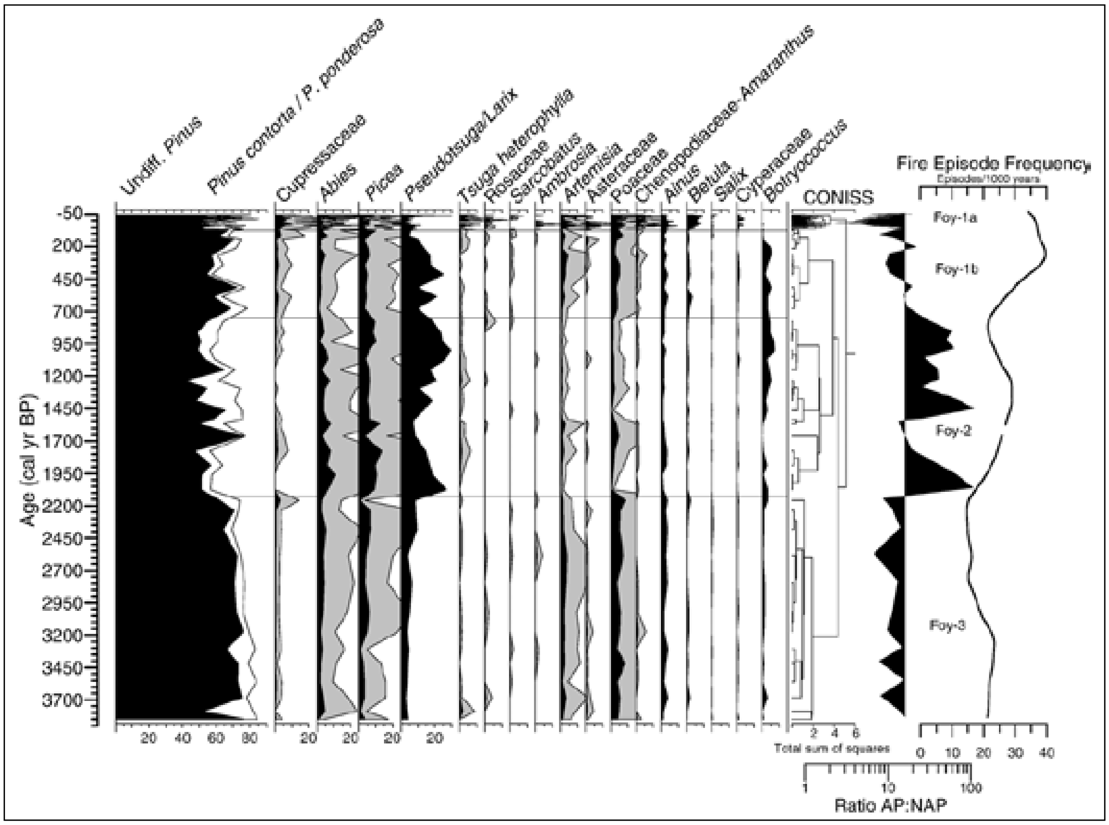


Figure 6. Changes in pollen percentages for different plant taxa during the last 3700 years. Black and gray silhouettes (5x exaggeration) represent changes in the proportion of pollen for individual taxa over time as a percentage of all terrestrial pollen. The arboreal pollen (AP) and non-arboreal pollen (NAP) ratio includes all upland tree pollen divided by all non-aquatic shrubs and herbs. Fire frequency is recorded as number of episodes per 1000 years. Examples of vegetation response to climate change include a distinct change from steppe vegetation to forest accompanied by increased fire frequency ca. 2125 cal yr BP and a return to more open vegetation ca. 750 cal yr BP. Source: Power et al. 2006; reprinted with permission.

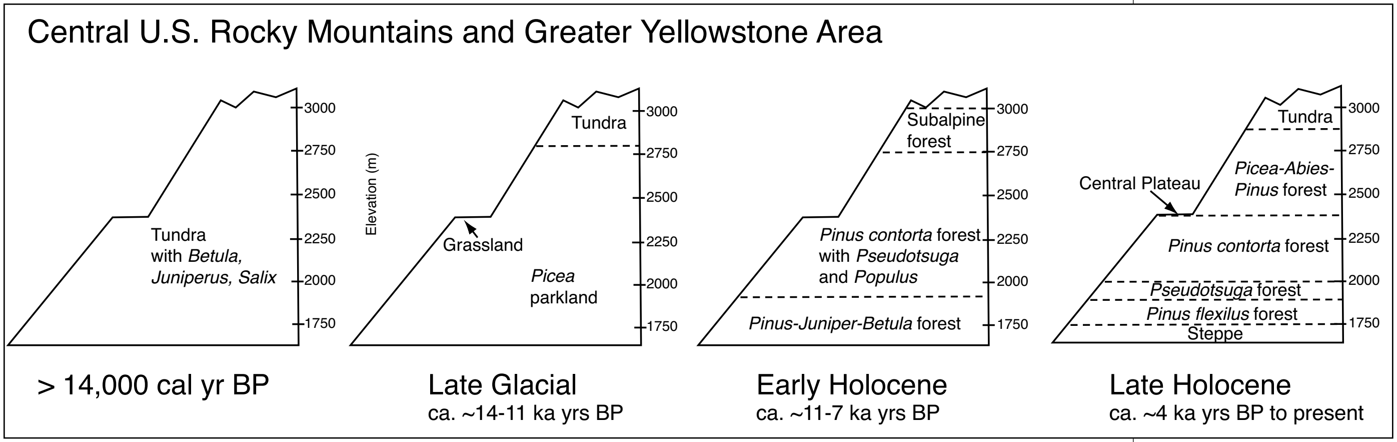

Central U.S. Rocky Mountains and Greater Yellowstone Area
Elevation (m)
3000
2750
2500
2250
2000
1750
Tundra with Betula, Juniperus, Salix
> 14,000 cal yr BP
Tundra
Grassland
Picea parkland
Late Glacial
ca. ~14-11 ka yrs BP
Subalpine forest
Pinus contorta forest with Pseudotsuga and Populus
Pinus-Juniper-Betula forest
Early Holocene
ca. ~11-7 ka yrs BP
Tundra
Central Plateau
Picea-Abies-Pinus forest
Pinus contorta forest
Pseudotsuga forest
Pinus flexilus forest
Steppe
Late Holocene
ca. ~4 ka yrs BP to present

Figure 7. Ecological response to changing climatic conditions following glacial retreat in the Central Rockies and the summer-dry region of the Greater Yellowstone Area (derived from Whitlock 1993). The figure illustrates that tundra occupied most of the area >14 ka BP, but is only present at high elevations in the late Glacial and late Holocene.

(*Juniperus*), and birch (*Betula*) were present at low elevations, whereas lodgepole pine, Douglas fir, and aspen (*Populus*) established and characterized mid-elevation forests. Subalpine forests (*Picea* and *Abies*) expanded their ranges to higher elevations, and the upper treeline was at an elevation similar to that of the present. Decreased summer insolation in the late Holocene (ca. 3000–4000 BP) led to cooler, wetter conditions. Sagebrush-steppe became present at low elevations and forests of limber pine (*Pinus flexilus*), Douglas fir, and lodgepole pine dominated mid-elevations (Whitlock et al. 1993). Subalpine communities were comprised of mixed spruce, fir, and pine forests, and the increasingly cooler conditions resulted in a lowering of the upper treeline to an elevation close to its present position.

2.3.3 Cygnet Lake

An examination of charcoal, pollen, and climate conditions from central Yellowstone provides an example of the linkage between fire and climate even in the absence of vegetation change (fig. 8; Whitlock 1993). Pollen from Cygnet Lake indicates the area was dominated by a tundra community of sagebrush and grass (Poaceae) when cool and wet conditions prevailed prior to 12,000 BP, after which the vegetation was dominated by lodgepole pine, which is favored in infertile rhyolite soils. The last 12,000 years reveal little change in the vegetation at Cygnet Lake compared to sites on non-rhyolite substrates that show strong responses to early, middle, and late Holocene climate forcings (Whitlock 1993). Despite the complacency of the vegetation, the Cygnet Lake fire record shows the effect of early Holocene drought on fire regimes; the charcoal data suggests rising fire activity in response to increasing summer insolation. Fire return intervals were 75–100 years between 11,000 and 7000 cal yr BP, and cooler wetter conditions after 7000 cal yr BP coincide with decreasing fire frequency. The present fire return interval of 200 to 400 years was reached in the late Holocene.

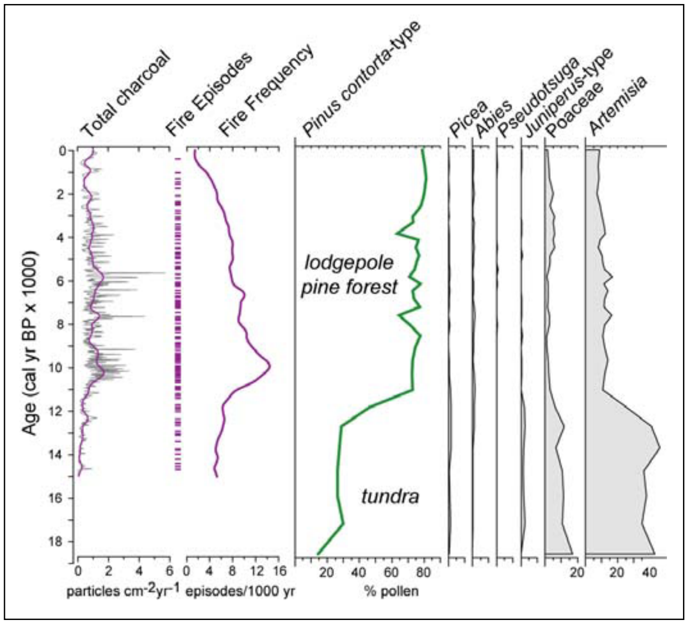


Figure 8. Ecological response to changing climatic conditions following glacial retreat in Yellowstone National Park. Pollen and charcoal diagram from Cygnet Lake, central Yellowstone (modified from Millspaugh et al. 2000).

2.3.4 Southern U.S. Rocky Mountains

A network of vegetation reconstructions from pollen and macrofossil data provides a history of climate and vegetation change in southwestern Colorado during the late Glacial and Holocene (Fall 1997). Prior to 14,000 cal yr BP, cooler and wetter conditions (2–5°C [4–9°F] cooler and 7–16 cm [3–6"] wetter than at present) supported tundra vegetation at high elevations (~300–700 m [984–2,297'] below the present treeline) and spruce parkland at low elevations (fig. 9, Fall 1997). During the late Glacial, fir (*Abies*) increased in abundance and subalpine forest was established at low and middle elevations (Fall 1997). Summer insolation increased during the early Holocene, and warmer temperatures allowed subalpine forests to expand above their present elevation. While the central and northern U.S. Rockies generally experienced warm dry conditions during the early Holocene, the southern Rockies experienced warmer, wetter growing seasons driven by a more intense North American monsoon (Thompson et al. 1993). Markgraf and Scott (1981) recorded an upslope advance of subalpine forests due to warmer conditions and an expansion of pine forests at both the lower and upper treeline facilitated by warmer and still wet conditions. Similarly, Fall (1997) found that from 9–4 ka BP, warm summers (mean temperatures

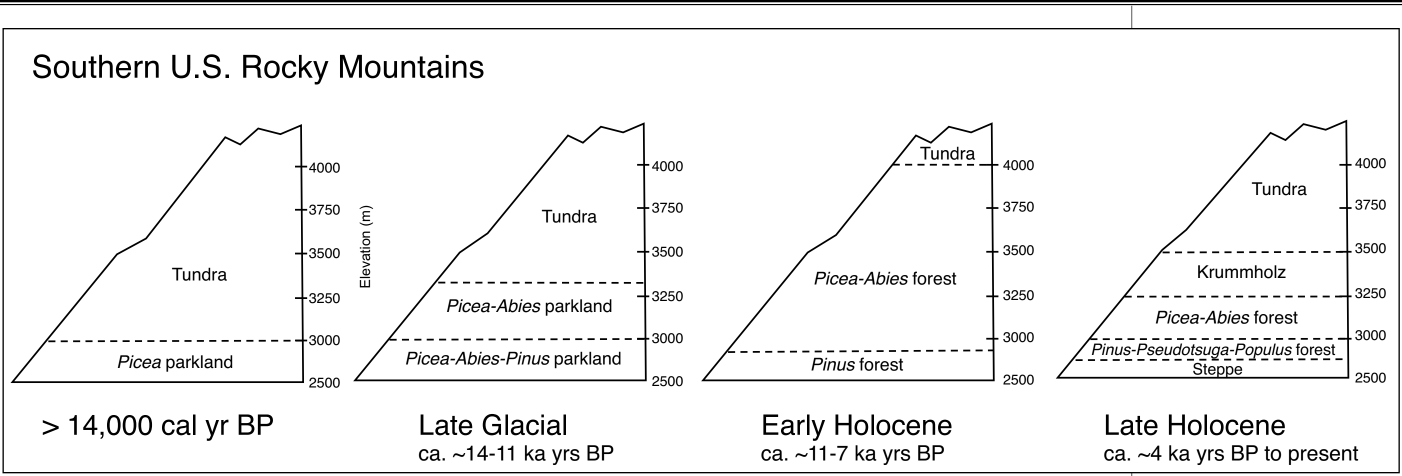

Southern U.S. Rocky Mountains
Elevation (m)
4000
3750
3500
3250
3000
2500
Tundra
Picea parkland
> 14,000 cal yr BP
Tundra
Picea-Abies parkland
Picea-Abies-Pinus parkland
Late Glacial
ca. ~14-11 ka yrs BP
Tundra
Picea-Abies forest
Pinus forest
Early Holocene
ca. ~11-7 ka yrs BP
Tundra
Krummholz
Picea-Abies forest
Pinus-Pseudotsuga-Populus forest
Steppe
Late Holocene
ca. ~4 ka yrs BP to present

Figure 9. Ecological response to changing climatic conditions following glacial retreat in the southern Rockies. Derived from Fall 1997.

1.9°C [3.4°F] above present) facilitated the expansion of forests of spruce and fir (*Picea engelmannii* and *Abies lasiocarpa*) upward to an elevation of almost 4,000 meters (13,123'), at least 300 meters (948') higher than today (Fall 1997), and downward to elevations below 3,000 meters (9,483').

In the late Holocene, low-elevation montane forests mixed with steppe vegetation and low-elevation subalpine forests were defined by an increasingly open stand structure. Summer temperatures declined to pre-industrial levels ca. 1850, and spruce and fir dominated subalpine forests and krummholz vegetation. Montane taxa retreated upslope, sagebrush steppe expanded at lower elevations, and alpine tundra dominated a larger range of high elevation areas, suggesting that drier conditions increased for parts of the southern Rockies (Markgraf and Scott 1981; Fall 1997). Conditions similar to the present were established approximately two millennia ago, with modest treeline elevation fluctuations during the Medieval Climate Anomaly.

2.3.5 Upper Columbia Basin

Paleoenvironmental data are available from the Upper Columbia Basin (Barnosky 1985; Whitlock et al. 2000; Mack et al. 1978, 1976; Mehringer 1996; Blinnikov et al. 2002), the Snake River Plain (Davis 1986; Beiswenger 1991; Bright and Davis 1982), and the mountains of southern Idaho and Montana (Doerner and Carrara 2001; Whitlock et al. in review; Mumma 2010).

During glacial times, tundra-steppe communities dominated by *Artemisia* and Poaceae were widespread in the basins, reflecting cold dry conditions (figs. 10–11). As the climate warmed in the late Glacial period, pine, spruce, and fir parkland developed. The early Holocene period in the western Columbia Basin and Snake River Plain featured steppe vegetation, and records from adjacent mountains record an expansion of juniper, sagebrush, and Chenopodiaceae. Summer drought was more pronounced throughout much of the region during the early Holocene as the amplification of the seasonal insolation cycle resulted in warmer and

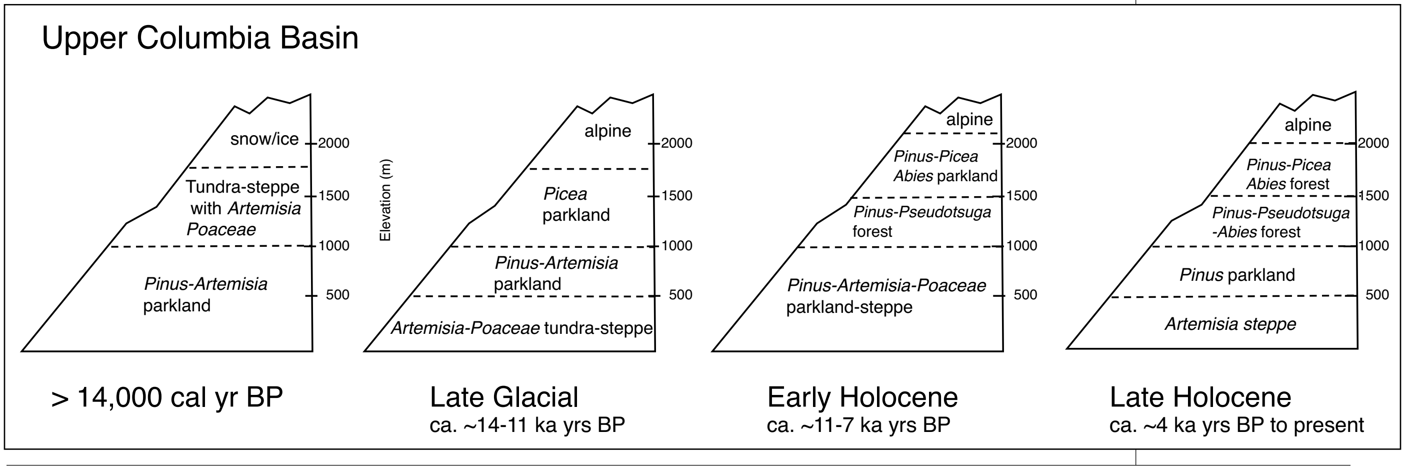

Upper Columbia Basin
Elevation (m)
2000
1500
1000
500
snow/ice
Tundra-steppe with Artemisia Poaceae
Pinus-Artemisia parkland
> 14,000 cal yr BP
alpine
Picea parkland
Pinus-Artemisia parkland
Artemisia-Poaceae tundra-steppe
Late Glacial
ca. ~14-11 ka yrs BP
alpine
Pinus-Picea Abies parkland
Pinus-Pseudotsuga forest
Pinus-Artemisia-Poaceae parkland-steppe
Early Holocene
ca. ~11-7 ka yrs BP
alpine
Pinus-Picea Abies forest
Pinus-Pseudotsuga -Abies forest
Pinus parkland
Artemisia steppe
Late Holocene
ca. ~4 ka yrs BP to present

Figure 10. Ecological response to changing climatic conditions in the western portion of the Upper Columbia Basin following deglaciation. Derived from Carp Lake, Washington (Whitlock et al. 2000).

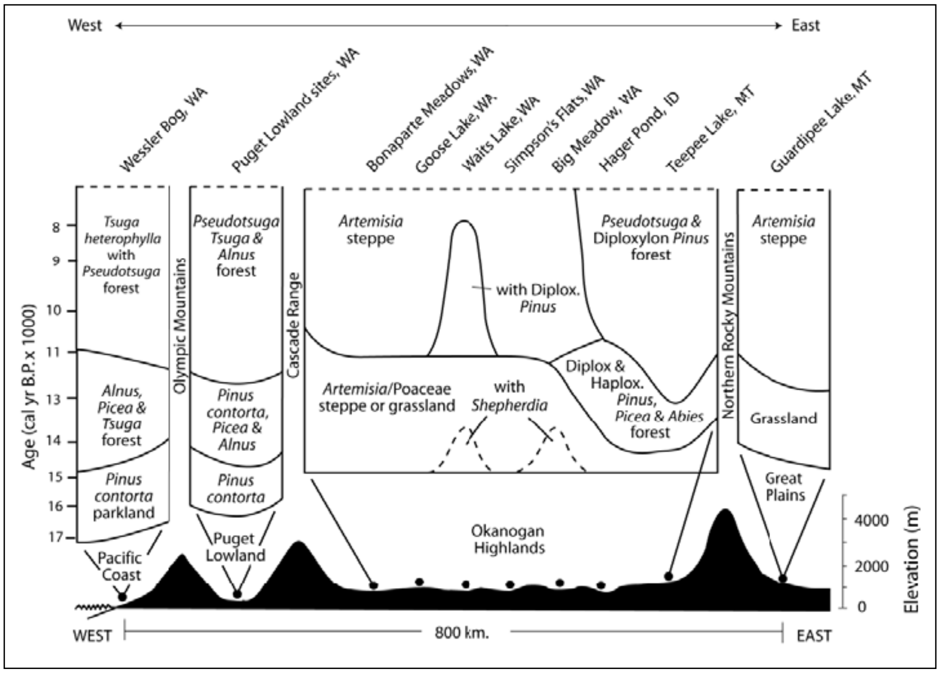

West
East
Wessler Bog, WA
Puget Lowland sites, WA
Bonaparte Meadows, WA
Goose Lake, WA
Waits Lake, WA
Simpson's Flats, WA
Big Meadow, WA
Hager Pond, ID
Teepee Lake, MT
Guardipee Lake, MT
Age (cal yr B.P. x 1000)
8
9
10
11
13
14
15
16
17
Tsuga heterophylla with Pseudotsuga forest
Alnus, Picea & Tsuga forest
Pinus contorta parkland
Olympic Mountains
Pseudotsuga Tsuga & Alnus forest
Pinus contorta, Picea & Alnus
Pinus contorta
Cascade Range
Artemisia steppe
with Diplox. Pinus
Artemisia/Poaceae steppe or grassland
with Shepherdia
Pseudotsuga & Diploxylon Pinus forest
Diplox & Haplox. Pinus, Picea & Abies forest
Northern Rocky Mountains
Artemisia steppe
Grassland
Great Plains
Pacific Coast
Puget Lowland
Okanogan Highlands
4000
2000
0
Elevation (m)
WEST
800 km.
EAST

Figure 11. Late Glacial and early Holocene vegetation history along the southern margin of the Cordilleran ice sheet, based on a transect of pollen records from western Washington to western Montana. *Abies*, fir; *Alnus*, alder; *Artemisia*, sagebrush; *Picea*, spruce; *Pinus contorta*, lodgepole pine; *Poaceae*, grass; *Pseudotsuga*, Douglas fir; *Shepherdia*, buffaloberry; *Tsuga heterophylla*, western hemlock. Source: Whitlock and Brunelle 2007; reprinted with permission.

effectively drier conditions at low and mid-elevations (Whitlock et al. 2000). Cool dry conditions in the Okanogan Highlands of northern Washington continued to support grasses and sagebrush, and the forest-steppe ecotone was north of its present location by as much as 100 kilometers (Mack et al. 1978). Pine, spruce, and fir were present in areas with greater precipitation (e.g., Waits Lake, eastern Washington) (Whitlock and Brunelle 2007, fig. 11). In the mid-Holocene, increased effective moisture allowed the establishment and expansion of pine woodland at middle and higher elevations, and the upper treeline was higher than at present. In the western Columbia Basin, the expansion of pine woodland (primarily *Pinus ponderosa*) was followed by an invasion of mixed forest in the late Holocene (Douglas fir, larch, fir, western hemlock, and oak) (Whitlock and Brunelle 2007). Some sites at higher elevations signal an interval of cooling during the late Holocene (ca. 1.7–3.5 ka BP) when the abundance of spruce and fir pollen increased (Whitlock and Brunelle 2007) and led to an expansion of mixed forests throughout the Okanogan highlands. Modern assemblages of Douglas fir, fir, western hemlock and spruce were established during this period (Whitlock and Brunelle 2007).

Chapter 3: What do paleoenvironmental records tell us about millennial scale climate variations?

The western United States has experienced large-scale changes in climate, vegetation, and disturbance regime since the last glaciation 20,000 years ago. With the initial recession of glacial ice more than 14,000 years ago, the climate was colder and generally drier than at present and most regions were colonized by tundra communities. As the climate warmed and precipitation increased from 14,000 to 11,000 years ago, these tundra communities were replaced by subalpine parkland and then closed subalpine forest. During the early Holocene (11,000–7,000 cal yr BP), the development of warmer and drier than present conditions led to more xerophytic vegetation and more fires in most areas. After 7000 cal yr BP, the climate became cooler and effectively wetter. In most regions, the modern vegetation and climate were established during the late Holocene (the last 4000 cal yrs). These changes highlight several important lessons for understanding the impacts of climate change on vegetation:

- Millennial-scale variations in climate over the last 20,000 years were caused by changes in the latitudinal and seasonal distribution of incoming solar radiation, the size and extent of the continental ice sheets, and attendant shifts in atmospheric circulation (e.g., southward displacement of the jet stream, the strength of the northeastern Pacific subtropical high-pressure system, and the intensity of monsoonal circulation). These slowly varying changes determined the distribution and composition of plant communities.

- Abrupt climate variation during the last 20,000 years led to rapid changes in the assemblages and distribution of vegetation across the U.S. Rocky Mountains and the Upper Columbia Basin and influenced ecosystem processes such as fire.

    - ✧ The Medieval Climate Anomaly (MCA, ca. 950–1250) and Little Ice Age (ca. 1400–1700) resulted in shifts in plant distributions and disturbance regimes in some locations, but were not uniformly manifested across the study area. For example, warm dry conditions in Yellowstone during the MCA led to increased fires in the summer-dry areas while the summer-wet areas of Yellowstone were effectively wetter (Whitlock and Bartlein 1993; Whitlock and et al. 2003).

    - ✧ The ecological impact of these shorter climatic events (lasting centuries) is variable either because the climate signal was regionally heterogeneous or because plant communities were not responsive to climate change on this relatively short time scale.

- The western United States is influenced by two precipitation regimes—a summer-dry area under the influence of the northeastern Pacific subtropical high-pressure system and a summer-wet region strongly influenced by summer monsoon circulation. In the northern U.S. Rocky Mountains, the location of these regimes is sharply delimited and constrained by topography. These two regimes were enhanced during the early Holocene, when summer solar radiation was higher than at present. As a result, summer-wet areas became wetter and experienced fewer fires than at present, and summer-dry areas became drier with more fires and xerophytic vegetation than at present (Whitlock and Bartlein 1993; Huerta et al. 2009; Millspaugh et al. 2004). The contrast between summer-wet and summer-dry regions was greatest in the early Holocene. This past response suggests that future changes in precipitation regimes will also likely be spatially heterogeneous, and that the boundary between precipitation regimes will likely be quite sharp in mountainous regions.

- Although climate exerts strong controls on the distribution of vegetation at large spatial and long temporal scales, edaphic factors can amplify or minimize the response at smaller scales, as illustrated by the persistence of lodgepole pine forests on rhyolitic soils in Yellowstone's Central Plateau throughout the Holocene (Whitlock et al. 1993; Millspaugh et al. 2000) and the influence of ultramafic soils on Holocene vegetation and fire regimes (Briles et al. in review).

- Climate change can influence the distribution of vegetation via direct climate constraints (e.g., temperature and precipitation) or indirectly by influencing key ecosystem processes such as fire and nutrient cycling. Feedbacks related to vegetation changes can also influence fire by changing fuel availability.

The superposition of climate changes occurring on multiple time scales means that no period in the last 20,000 years is an exact analogue for the future. Nonetheless, the paleoenvironmental records show the resilience of vegetation to periods of extreme drought, changes in disturbance regimes, and rapid climate change. These examples provide insights about the sensitivity and pathways by which ecosystems respond to climate changes of different duration and intensity.

Key findings:

1. Natural variations in climate and the accompanying ecological responses occur at multiple temporal and spatial scales, all of which must be understood to explain modern plant communities and their distributions. Paleoecological data provide evidence of a range of responses that are not adequately represented in the last two centuries. A baseline of natural variability for restoration efforts must therefore consider a longer time scale.

2. Many terrestrial ecosystems in the study region were established during the last 3000 to 4000 years in response to gradual cooling and increased effective moisture in the late Holocene. More subtle changes occurred during the Medieval Climate Anomaly and Little Ice Age (as discussed in the next section). The sequence of events that led to present vegetation is different from that projected for the future, which argues against strategies to restore to a reference condition. Instead, we need process-based approaches and flexible management responses.

3. Key ecosystem processes such as fire are driven by climate at large spatio-temporal scales. Patterns of the fire in the 20th century poorly represent the potential range of fire regimes that have occurred in the past and may occur in the future.

Chapter 4: The last 2000 years of environmental change

4.1 Primary drivers of change

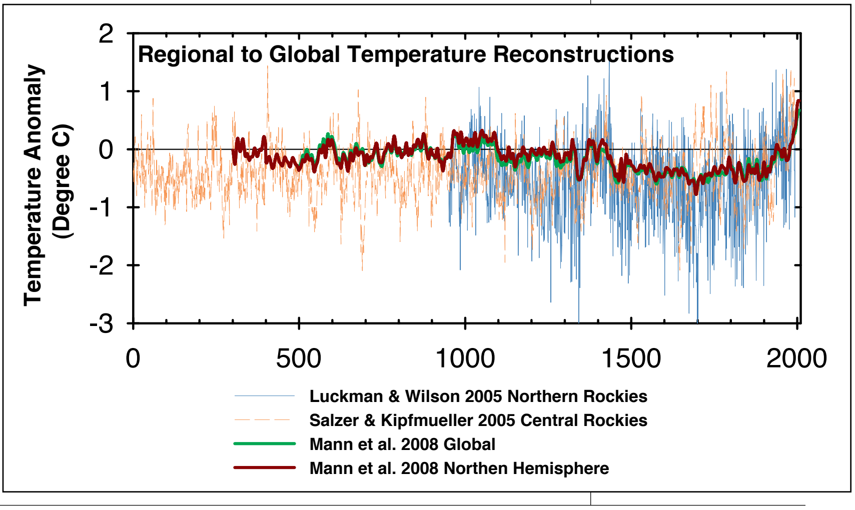
The primary drivers of climate during the last 2000 years include ocean-atmosphere interactions, volcanic eruptions, changes in incoming solar radiation, and increases in atmospheric greenhouse gases (GHGs) and aerosols due to human activities (fig. 2). Climate model simulations indicate that during the pre-industrial portion of the last 2000 years, solar fluctuations and volcanic eruptions were likely the most strongly varying forcings, and in combination with ocean-atmosphere interactions they likely resulted in periods of relative warmth and cold (Amman et al. 2007; Mann and Jones 2003; Mann 2007; Jones et al. 2009; Esper et al. 2002). GHGs and tropospheric aerosols varied little until around AD 1700 when they began to be significantly impacted by human activities (Solomon et al. 2007; Keeling 1976).The rapid rise in 20th century global temperatures is best explained by the combination of natural and anthropogenic GHG forcings, with GHGs playing an increasingly dominant role during recent decades.

4.2 Biophysical conditions

Compared to the conditions driving continental deglaciation and the Pleistocene/Holocene transition, orbital and radiative forcings over the last two millennia have remained relatively constant and can be considered more analogous to modern conditions. However, the major centennial-scale climate variations evident during this period can be linked to changes in solar output, volcanic forcing, and ocean-atmosphere interactions (Amman et al. 2007; Mann et al. 2009; Mann 2007; Jones et al. 2009; Esper et al. 2002). Because the rates of climatic change during the last two millennia were much smaller in magnitude than those associated with the late Glacial and early Holocene, the ecological response to climatic variation was generally less dramatic. For much of the study area considered in this synthesis, climatic variation during this period led to shifts in the extent and abundance of species found in modern vegetation assemblages but rarely to widespread changes in dominant vegetation. Vegetation types since the late Holocene are considered similar to present conditions and, overall, the magnitude and duration of the changes are not comparable to those of the Pleistocene–Holocene transition. Rather, the last 2000 years have had smaller magnitude and shorter duration (centennial, decadal, interannual) climatic controls on ecosystems that have nevertheless resulted in societally and ecologically relevant changes in both ecosystems and natural resources. At these shorter time scales, ocean-atmosphere interactions such as the Pacific Decadal Oscillation, the North Atlantic Oscillation, the Atlantic Multidecadal Oscillation, and the El Niño–Southern Oscillation interact to influence temperature, precipitation, and atmospheric circulation and help explain droughts and wet periods at interannual to interdecadal scales (e.g., Cook et al. 2004; Mann et al. 2009; Gray et al. 2004; Hidalgo 2004; Fye et al. 2003; McCabe et al. 2004; Graumlich et al. 2003; Enfield et al. 2001). Past records of temperature illustrate centennial, decadal, and interannual variation, providing a context for understanding ecosystem changes that occurred in different regions (fig. 12).

Tree-ring records from throughout the western United States show natural variation in temperature, precipitation, and available moisture during the last two millennia, some of which is synchronous across large areas of the four climate regions in this study, while other variations are more representative of local phenomena (figs. 12–13, Pederson et al. 2006; Cook et al. 2004). These records show that decadal and multidecadal fluctuations in

Figure 12. Comparison of regional and global temperature reconstructions. Derived from Luckman and Wilson 2005, Salzer and Kipfmueller 2005 (both based on tree-ring records), and Mann et al. 2008 (based on a multi-proxy reconstruction).



precipitation are a defining characteristic of the climate during past millennia and exert important controls on ecosystem processes and species distributions (e.g., Pederson et al. 2006; Cook et al. 2004, 2007). Regionally synchronous wet and dry intervals have been linked to low-frequency variations and state changes in sea surface temperature and pressure anomalies in both the Atlantic and Pacific oceans which are discussed in more detail later (McCabe et al. 2004; Gray et al. 2003; Cayan et al. 1998).

Two major, well-documented examples of centennial-scale climate change during the last two millennia are the Medieval Climate Anomaly (950–1250) and the Little Ice Age (1400–1700). As indicated by figure 13 (Cook et al. 2007), a number of anomalous warm dry periods and cool wet periods occurred (see Biondi et al. 1999; Meko et al. 2007; Salzer and Kipfmuller 2005; Cook et al. 2007), and resulted in extensive hydrological and ecological impacts. During the MCA, dry precipitation anomalies persisted across western North America (fig. 13). While the general climatic conditions during this time were defined by warmer, drier conditions, the local effects were highly variable. Elevated aridity and "mega-droughts" (lasting at least ~50 years) were common across the western United States, with more areas experiencing drought simultaneously than during the LIA or most of the 20th century (fig. 13, Cook et al. 2007).

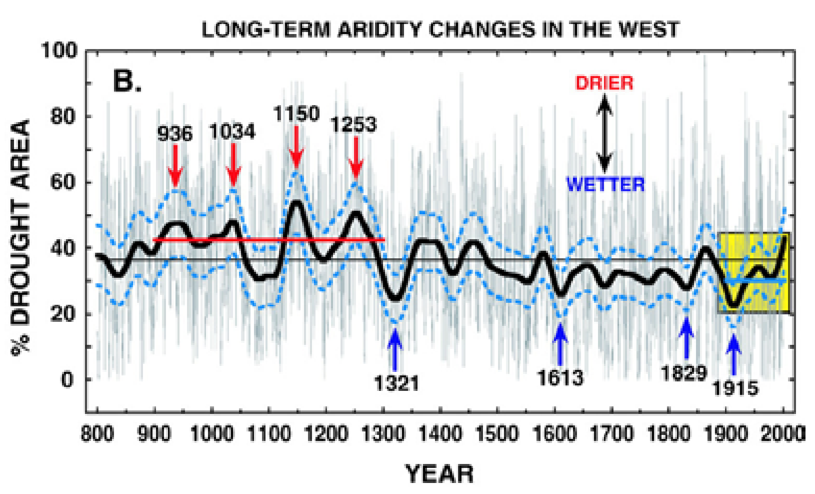


Figure 13. Long-term aridity changes in the West as measured by the percent area affected by drought (PDSIb_1, thick black line), 95% boostrapped confidence intervals (light-blue dotted lines) and the long-term mean (thin horizontal black line). The four most significant ($p<0.05$) dry and wet epochs since 800 are indicated by arrows. The 1900–2003 interval is highlighted by the yellow box. The average drought area during that period and for the 900–1300 interval is indicated by the thick blue and red lines, respectively. The difference between these two means is highly significant ($p<0.001$). Source: Cook et al. 2004, 2007; reprinted with permission.

Data from a number of sites suggest that regionally synchronous drought events occurred regularly during the MCA, with durations and extents unmatched in the late Holocene (Cook et al. 2007, 2004). Glacial retreat occurred in mountainous areas of Colorado, Wyoming, Montana, and the Cascades, with substantial reductions in streamflow (e.g., Meko et al. 2007; Gray et al. 2003) and lake levels throughout the study area (Millspaugh et al. 2000; Brunelle and Whitlock 2003). However, spatial and temporal variations in the generally warm dry conditions were widespread, as the central U.S. Rockies, Greater Yellowstone Area, and parts of the southern Rockies experienced increased summer moisture (Whitlock and Bartlein 1993). Upper treelines in some areas increased in elevation and areas now covered by krummholz were occupied by arborescent trees (Graumlich et al. 2005; Whitlock et al. 2002; Fall 1997; Rochefort et al. 1994; Winter 1984; Kearney and Luckman 1983). Additionally, alpine larch expanded 90 kilometers north of its current range (Reasoner and Huber 1999; Reasoner and Hickman 1989) ca. 950–1100 BP. The mechanisms and drivers leading to the MCA are still debated, but there is increasing evidence that low-frequency variation in ocean-atmosphere interactions was an important factor (Mann et. al 2009).

The LIA was a period of anomalous Northern Hemisphere cooling when mountain glaciers throughout the western United States expanded, many reaching their Holocene maximum (e.g., Pederson et al. 2004; Luckman 2000; Watson and Luckman 2006). While temperatures across the study area were persistently cooler than the long-term average (>1°C [2°F] cooler), some data suggest that the magnitude of the cooling decreased with latitude (Whitlock et al. 2002). In the northern Rockies where the most pronounced cooling occurred, conditions in the late LIA may have approached those of the late Pleistocene (Pederson et al. 2007; Luckman 2000; Clark and Gillespie 1997). Both tree ring and glacier data indicate sustained cool summer conditions and increased winter precipitation across the northern Rockies resulted in the major glacier advance during the LIA (Watson and Luckman 2004; Pederson et al. 2004). Advances of this magnitude did not occur in the southern and central Rockies, demonstrating the spatial variability in LIA climate anomalies (Clark and Gillespie 1997). As with the MCA, the drivers and mechanisms that influenced cooler conditions during the LIA are not well understood. Decreased sunspot activity during a period called the Maunder Minimum which led to decreased incoming solar radiation from ca. 1645 to 1715 is considered one of the main variables explaining cooler conditions for this interval of the LIA (Eddy 1976; Luckman and Wilson 2005).

The following case studies from the four climate regions highlight examples of biophysical and biotic response to climate change during the last two millennia and provide clues to the timing and extent of future biotic changes.

5.1 Northern U.S. Rocky Mountains

5.1.1 Drought variability and ecosystem dynamics in Glacier National Park

Records of past hydroclimatic changes, glacier dynamics, and fire activity in Glacier and Waterton national parks show how interdecadal and longer-term changes in climate interact to alter ecosystem processes in the northern Rockies (fig. 14). Long-term changes in regional temperature (e.g., the relatively cool conditions of the LIA in contrast to warmer temperatures in the 1350s to 1450s and especially over the last half of the 20th century, fig. 12) combined with persistent shifts in summer and winter moisture regimes over decadal to multidecadal time scales have a particularly strong impact on fire regimes in the Glacier National Park region (fig. 14a–c). Regional records spanning the last three centuries show that periods from the 1780s to the 1840s and the 1940s to the 1980s had generally cool wet summers coupled with high winter snowpack, resulting in extended (>20 years) burn regimes characterized by small infrequent fires with relatively little area burned. Conversely, decadal and longer combinations of low snowpack and warm dry summers resulted in burn regimes characterized by frequent fires and large total area burned (e.g., 1910–1940, 1980–present).

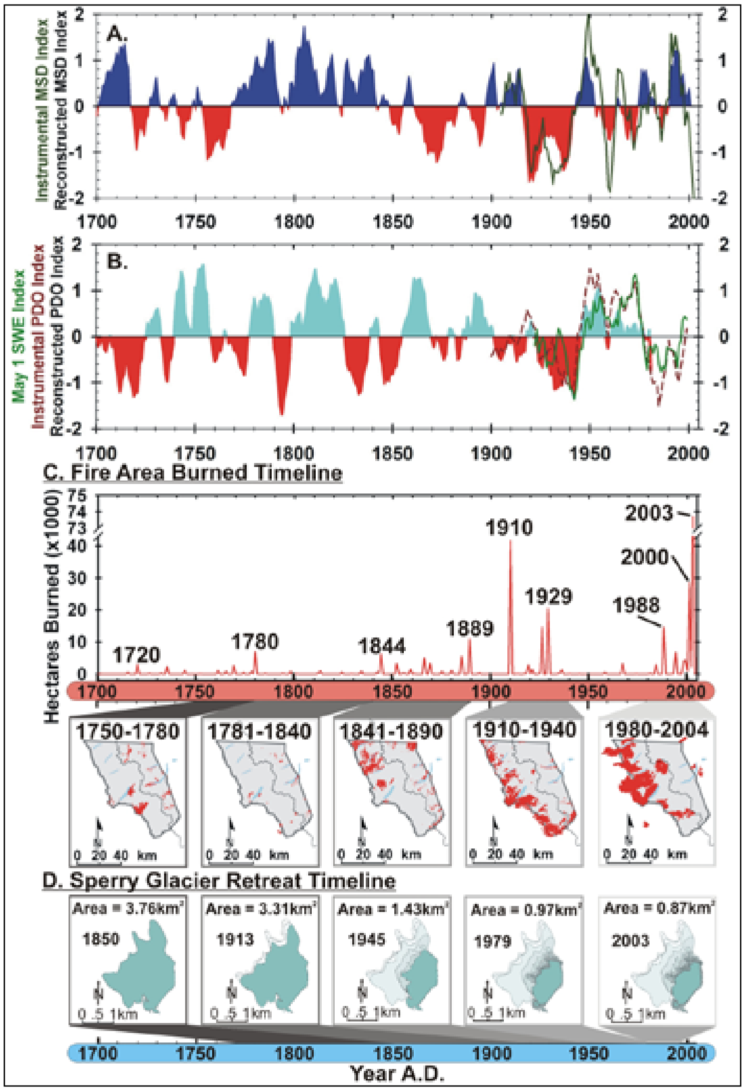


Figure 14. Relationship between Glacier National Park summer drought, inferred winter snowpack, fire area burned, and glacial recession 1700 to present. (a) Instrumental and reconstructed summer drought (MSD) normalized by converting to units of standard deviation and smoothed using a 5-year running mean. (b) Measured spring snowpack (May 1 SWE) anomalies (1922–present) and average annual instrumental and reconstructed PDO anomalies. Each time series was normalized and smoothed using a 5-year running mean. For ease of comparison, the instrumental and reconstructed PDO index was inverted due to the strong negative relationship between PDO anomalies and May 1 snowpack. (c) Fire area burned timeline for the GNP region. Timeline is presented with maps of fire activity during periods of interesting winter and summer precipitation regimes. (d) Maps showing the decrease in area of the Sperry Glacier at critical points from 1850 to 2003. The retreat patterns of the Sperry Glacier are representative of regional patterns of recession for glaciers sensitive to regional climate variability. Source: Pederson et al. 2006, American Meteorological Society; reprinted with permission.

Similarly, long-duration summer and winter temperature and moisture anomalies drive glacial dynamics in the northern Rockies (Watson and Luckman 2004; Pederson et al. 2004). Prior to the height of the LIA (ca. 1850), four centuries of generally cool summers prevailed (fig. 12). What drove the glaciers to reach their greatest extent since the Last Glacial Maximum was cool wet summers coupled with generally high snowpack conditions from 1770 to 1840 (Watson and Luckman 2004; Pederson et al. 2004). During subsequent periods when summer drought and snowpack were generally in opposing phases (e.g., 1850–1910) and summer temperatures remained relatively cool (fig. 14), glaciers experienced moderate retreat rates (1–7 m/yr [3–23']). From 1917 to 1941, however, sustained low snowpack, extreme summer drought conditions, and high summer temperatures drove rapid glacial retreat. The Sperry Glacier retreated at 15–22 meters (49–72') per year and lost approximately 68% of its area (fig. 14d). Other glaciers such as the Jackson and Agassiz glaciers at times retreated at rates ≥100 meters (328') per year (Carrara and McGimsey 1981; Key et al. 2002; Pederson et al. 2004). Climatic conditions from 1945 to 1977 became generally favorable (i.e., level to cooling summer temperatures, high snowpack with variable summer drought conditions) for stabilization and even accumulation of mass at some glaciers (fig. 14a,b). Since the late 1970s, however, exceptionally high summer temperatures (Pederson et al. 2010) combined with low winter snowpack and multiple periods of severe and sustained summer drought have resulted in a continuation of rapid glacier retreat.

As demonstrated by these records from Glacier National Park and surrounding regions, biophysical and ecosystem processes of the northern Rockies are strongly influenced by moisture and temperature variability at decadal and longer time scales. This relationship between biophysical and ecosystem processes with hydroclimatic changes that can appear to be static on shorter (e.g., decadal) scales poses challenges to management and sustainability in three ways. First, long-duration proxy reconstructions call into question the conventional strategy of defining reference conditions or management targets based solely on < 100-year records. For example, the use of 30-year climatology for the allocation of natural resources and development of resource management goals is flawed because a 30-year climatic mean may only capture a single mode of climate variability (e.g., an extended regime of wet or dry conditions). Second, abrupt, high-magnitude changes from one climate regime to the next can onset rapidly and have prolonged impacts on ecosystem processes. These persistent and frequent climate-related shifts may amplify or dampen the effects of management activities. Lastly, decadal and longer persistence of either deficits or abundances in climate-related biophysical processes can lead to management policies and economic strategies that, while appropriate during the current regime, may not be robust under subsequent climates. Overall, greater awareness of how ecosystems respond to climate change at longer temporal scales provides important context for future management.

5.2 Central U.S. Rocky Mountains and the Greater Yellowstone Area

5.2.1 Changing distributions of Utah juniper

A long-term reconstruction of changing distributions of Utah juniper (*Juniperus osteosperma*) in the CR-GYA provides an example of how climate variation, complex topography, and spatial distribution of suitable habitat and biotic factors interact to govern plant invasions. It also provides evidence contrary to the popular idea that plant invasions are characterized by a steady and continuous march across landscapes. The distribution of Utah juniper in the mountains of Wyoming, southern Montana, and Utah during the late Holocene has been tracked by radiocarbon-dating fossilized woodrat middens (Lyford et al. 2003). During a dry period in the mid-Holocene (ca. 7500–5400 BP), Utah juniper migrated north into the central Rockies via a series of long-distance dispersal events. Further range expansion and backfilling of suitable habitat was stalled during a wet period from 5400 to 2800 BP (Lyford et al. 2003). In response to warmer, drier conditions that developed after 2800 years BP, Utah juniper populations rapidly expanded within the Bighorn Basin, especially from 2800 to 1000 years BP. The notable absence of significant Utah juniper establishment and expansion during the MCA suggests that long-term climate variations determine the distributions of species with centennial-scale life expectancies. In the case of the Utah juniper, Lyford et al. (2003) noted that establishment rates are significantly more affected by adverse climatic conditions than by individual or population survival where the species is already established. This could explain, in part, the tendency for Utah juniper populations to remain static instead of contracting during the Holocene.

In general, Utah juniper range expanded during periods characterized by warmer, drier conditions, and expansion and establishment ceased during cool, wet periods (Lyford et al. 2003). Thus, the migration of Utah juniper into the central Rockies was at least partly controlled by millennial-scale climatic variations during the Holocene. Although Utah juniper distribution is severely limited by cool temperatures and high precipitation in higher elevations of the central Rockies, the species inhabits only a fraction of the suitable climate space in the region (fig. 15) because it is limited by suitable substrate; present distributions cover more than 90% of the substrate in the region deemed highly suitable for Utah juniper survival.

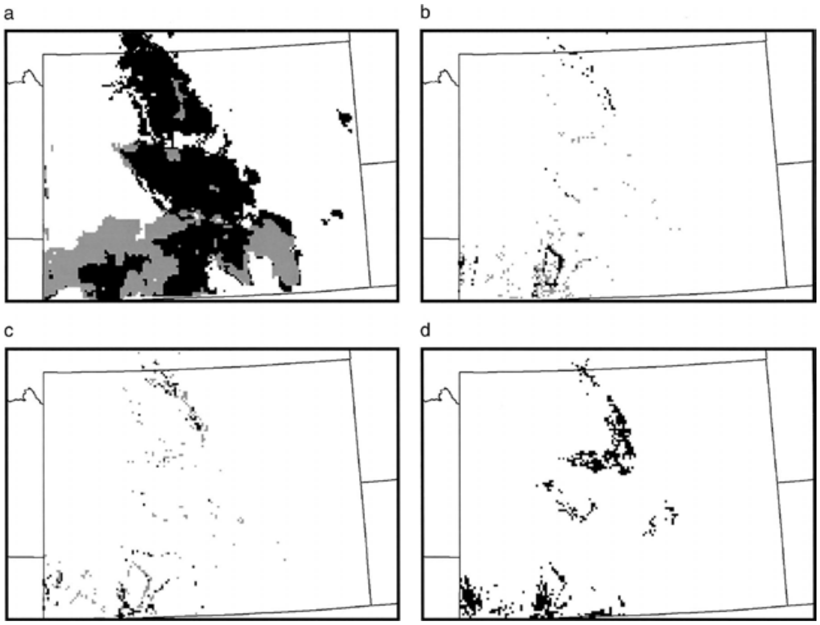


Figure 15. Habitats for Utah juniper in Wyoming and adjacent Montana. Black areas indicate extremely suitable habitats and gray areas moderately to highly suitable habitats. (a) Climate (ratio of growing season precipitation to growing season temperature). (b) Substrate, including soil, bedrock, and surficial-material type. (c) Climate and substrate combined. (d) Modern distribution (Knight et al. 1987, Driese et al. 1997). Note that climate (a) over-predicts the modern distribution, which is strongly constrained by substrate variables (b), and that favorable habitat is patchily distributed (c). Source: Lyford et al. 2003; reprinted with permission.

The case of the Utah juniper shows how climatic controls can influence species distribution, migration, and establishment in the central Rockies within the context of millennial-scale climate change. It also highlights the importance of recognizing other environmental factors that affect species distribution. While suitable climate can allow a species to become established, spatially variable factors such as substrate, dispersal, and competition influence how successfully it can disperse to and colonize areas with suitable climate. This provides an important example for considering how ecosystems and species will respond to changes in the spatial distribution of suitable habitat with changing climatic conditions. In this case, landscape structure and climate variability play key roles in governing the pattern and pace of natural invasions and will be important variables to consider when anticipating future changes in the distribution of plant species. The high temporal and spatial precision provided by this study illustrates that vegetation response to future conditions will be more nuanced than a steady march to newly suitable habitats, and better characterized by episodic long-distance colonization events, expansion, and backfilling (Lyford et al. 2003). This study suggests that models predicting plant invasions based on climate model projections may be oversimplified and encourages a more focused examination of how species dispersal will interact with the spatial distribution of suitable habitat and climate variability to govern future invasions.

5.2.2 Precipitation variability in Wyoming's Green River Basin

Tree-ring records from Wyoming's Green River basin provide a reconstruction of drought conditions for the last millennium and reveals how natural variations in dry and wet periods are a defining characteristic of the CR-GYA. Tree rings were used to develop a 1100-year record of the Palmer Drought Severity Index, a measure of drought that includes precipitation and temperature trends for southwestern Wyoming (Cook et al. 2004). This record is typical of many areas of the CR-GYA, showing above-average effective moisture in the early 20th century (Woodhouse et al. 2006; Gray et al. 2004, 2007; Meko et al. 2007) and the potential for severe, sustained droughts far outside the range of 20th century records, including several multidecadal droughts prior to 1300 (fig. 16).

Although using the last century as a reference for climate conditions would suggest that the Green River Basin is wet and relatively free of drought, this longer-term reconstruction indicates that some of the most severe droughts in the 1930s and 1950s were relatively minor compared to many dry periods in past millennia, and that the second half of the 20th century was relatively wet with no prolonged droughts (fig. 16). This study provides strong evidence that drought periods are a natural feature of the regional climate and that long-term records are critical for understanding natural variability of climate conditions in the western United States.

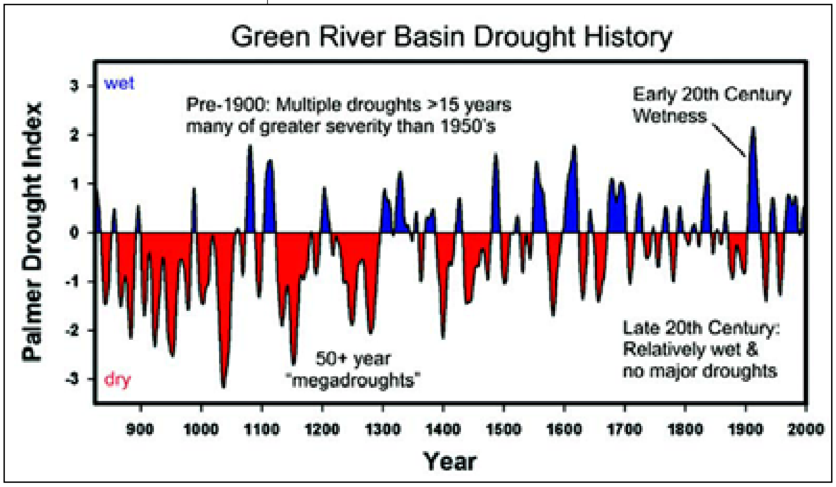


Figure 16. 1100 years of drought history in the Green River Basin region of southwest Wyoming (reconstructed from tree rings, Cook et al. 2004). The plot shows values for the Palmer Drought Severity Index (see also fig. 13). Positive values (blue) of the index indicate wet conditions and negative values (red) drought. Values are plotted so that each point represents mean conditions over a 25-year period. Source: Gray and Andersen 2010.

5.3 Southern U.S. Rocky Mountains

5.3.1 Changing pinyon pine distribution

Gray et al. (2006), who used woodrat midden and tree-ring data to track the spatial and temporal patterns of pinyon pine (*Pinus edulis* Engelm.) distribution in the Dutch John Mountains (DJM) of northeastern Utah, showed that the distribution during the Holocene has been strongly controlled by multidecadal precipitation patterns. The DJM population is an isolated northern outpost of pinyon pine that established ca. 1246. Similar to Utah juniper in the central Rockies, the pinyon pine probably reached the DJM via long-distance dispersal from the Colorado Plateau (Jackson et al. 2005) during the transition from the warmer, drier MCA to the cooler, wetter LIA (fig. 17). DJM pinyon pine expansion stalled in the late 1200s and significant recruitment did not resume until the 14th century pluvial, when regionally mesic conditions promoted establishment (fig. 17).

The case of DJM pinyon pine demonstrates the importance of episodic, multidecadal climatic variation in controlling rates of ecological change in the southern Rockies during past millennia. Records suggest that the development of the DJM population was not a steady movement associated with improving climate conditions but rather a markedly episodic invasion regulated by fluctuations in precipitation (Gray et al. 2006). In particular, this example highlights the consequences of having short-term, episodic climatic variation superimposed on centennial to millennial scale climate change, a pattern that can significantly affect species migration and establishment in ways that are more complex than a simple wave-like expansion.

As with previous studies of plant invasions at longer time scales (Lyford et al. 2003), this study suggests that climatic variation can amplify or dampen the probability of survival and reproduction after a species colonizes new areas. Climate can also modify the density and distribution of favorable habitats across the landscape and influence competitive interactions and disturbance processes (Gray et al. 2006). For example, different locations within a region may experience similar changes in average precipitation or temperature during a particular period, yet differences in the variability of precipitation could easily produce different disturbance dynamics and different end states. It is often assumed that vegetation responds to climate change with a steady wave-like movement to better growing conditions, yet the DJM example reveals that species are influenced by other factors, including the dynamics of long-distance dispersal and climate variability at different scales (Gray et al. 2006). Anticipating ecological response to climate change will require a better understanding of how natural climate variability regulates species migrations and invasions at smaller spatial and temporal scales.

5.4 Upper Columbia Basin

5.4.1 Climate variation and fire-related sedimentation

Fire-related sediment deposits in central Idaho reveal millennial to centennial scale climate variation and its control on fire regimes. Pierce et al. (2004) dated charcoal in alluvial fan deposits to reconstruct fire-related erosion events in dry forests dominated by ponderosa pine (*Pinus ponderosa*) and frequent, low-severity fires. They found that small sedimentation events occurred more frequently during the late Holocene, especially during the LIA, and suggested that these were associated with frequent fires of low to moderate severity. Large

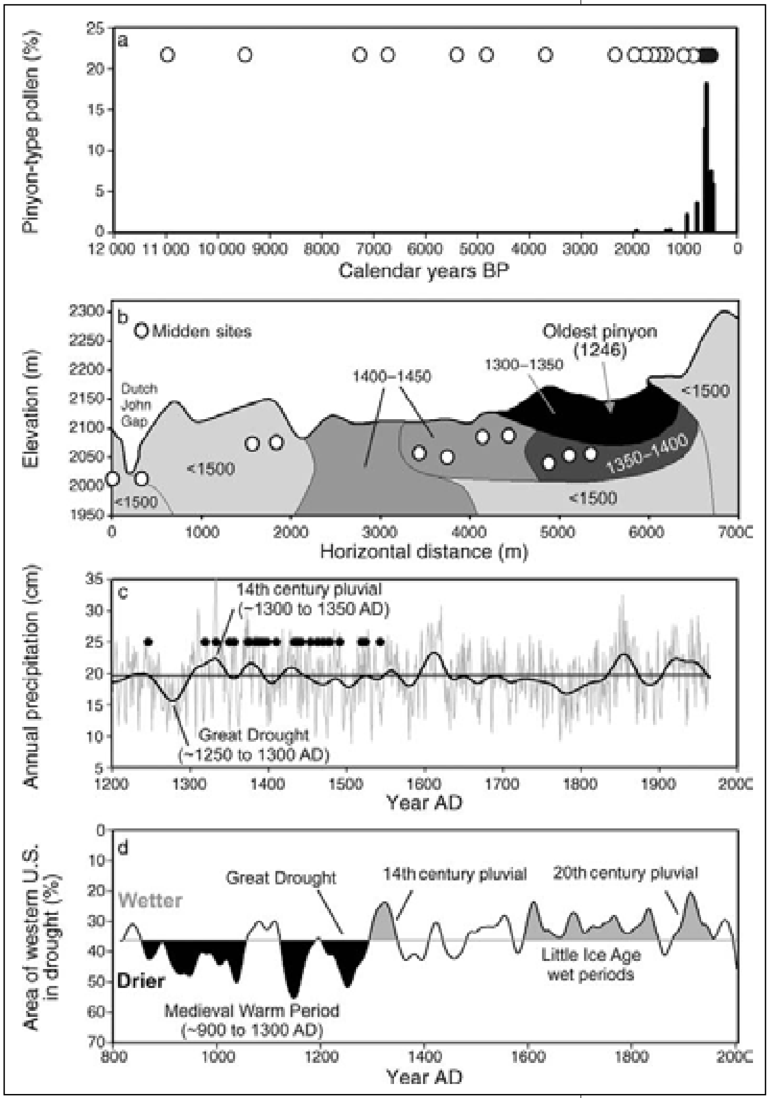


Figure 17. Establishment and presence/absence of pinyon pine in relation to climate variability and elevation during the Holocene. (a) Percentages of pinyon-type pollen (black vertical bars) and presence (solid circles) or absence (open circles) of pinyon pine macrofossils from 12,000 years of woodrat midden records collected in the DJM. (b) Map of the study area showing locations of midden sites (open circles) and sampling units used in the tree-ring age studies (shaded polygons). Each midden site is a cave or rock overhang where one or more of the 60 middens were collected. The estimated establishment dates are based on the average age of the four oldest pinyons found in each sampling unit. (c) Ages for the oldest pinyon on DJM and the four oldest pinyons in each of the eight sampling units (black dots) plotted against reconstructed annual (gray line) and 30-year smoothed (black line) precipitation values for the Uinta Basin Region. (d) Percentage of the western United States experiencing drought conditions during the last 1200 years as reconstructed from a large tree-ring network (Cook et al. 2004). Data are plotted as a 50-year moving average. The horizontal line at 37% (dark gray) shows the average or background level of drought through time. Significant multidecadal dry and wet periods identified by Cook et al. (2004) are shaded black and gray, respectively. Source: Gray et al. 2006; reprinted with permission.

sedimentation events were associated with prolonged periods of drought and severe fires that were most pronounced during the MCA (Pierce et al. 2004, fig. 18).

These results were compared with a similar record from northern Yellowstone National Park, where mixed conifer forests are associated with infrequent, stand-replacing fires (Meyer et al. 1995, fig. 18). Changes in inferred fire occurrence were synchronous between central Idaho and northern Yellowstone during the warmer, drier MCA. In central Idaho, longer intervals of warm dry conditions allowed for drying of large fuels to increase the frequency of large stand-replacing canopy fires and fire-related erosion events. Hence, strong climatic controls changed a fuel-limited, infrequent, low-severity fire regime to a fuel-rich, high-severity, stand-replacing fire regime. Large stand-replacing fires also increased in mixed conifer forests of northern Yellowstone during the MCA (fig. 18), coinciding with large pulses of fire-related sedimentation ca. AD 1150. Meyer et al. (1995) inferred that this activity resulted from increased intensity and interannual variability in summer precipitation. During the LIA, ca. AD 1400-1700, cooler wetter conditions in both northern Yellowstone and central Idaho are inferred to have maintained high-canopy moisture that inhibited canopy fires and facilitated the growth of understory grass and fuels required to sustain frequent low-severity fires in central Idaho (Pierce et al. 2004).

Paleoenvironmental investigations like these provide further evidence that millennial and centennial scale variations in temperature and precipitation have influenced biophysical conditions and important disturbance processes across the Upper Columbia Basin and CR-GYA. The results of these and similar studies (e.g., Whitlock et al. 2003) suggest that natural climatic variability acts as a primary control of ecosystem processes, which has important implications for management. Efforts to manage fuels in different stand types to restore specific fire regimes may be trumped by future climate variations, and actively managing for stand conditions that supported what are considered 20th century structural and disturbance characteristics may have limited impacts under future climatic conditions.

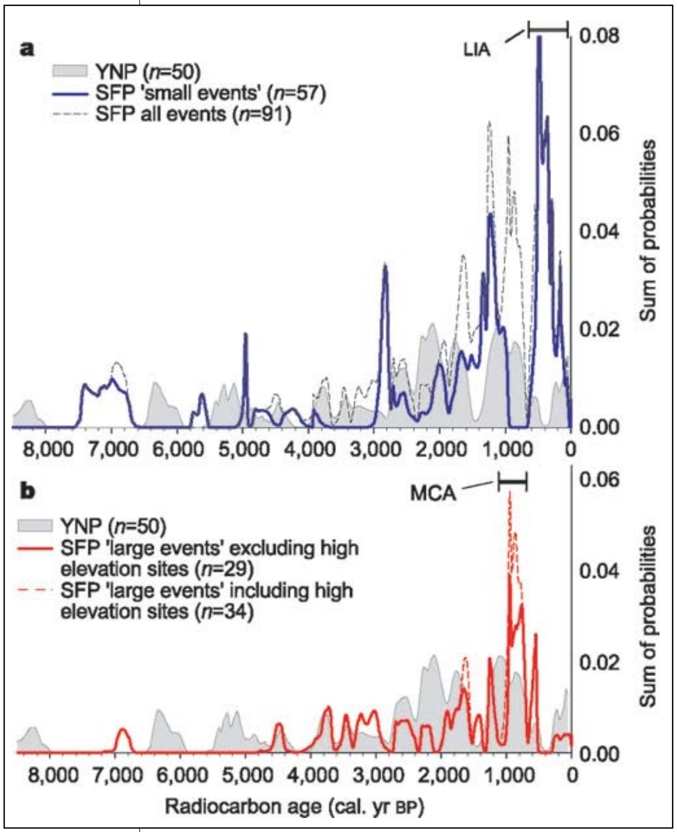


Figure 18. Fire-related sedimentation in the South Fork Payette River (SFP) area in central Idaho and in Yellowstone National Park (YNP), Wyoming. Probability distributions were smoothed using a 100-year running mean to reduce the influence of short-term variations in atmospheric 14C but retain peaks representing the most probable age ranges. The trend of decreasing probability probability before 4000 cal yr BP reflects fewer sites and decreased preservation of older deposits. (a) SFP small events are thin deposits probably related to low or moderate burns. Note correspondence of peaks with minima in YNP fire-related sedimentation and major peak during the LIA. Fewer near-surface deposits since 400 cal yr BP were selected for dating because of bioturbation and large uncertainties in 14C calibration. (b) SFP large events are major debris-flow deposits probably related to severe fires. Note correlation with the YNP record and peak in large-event probability during the MCA. Source: Pierce et al. 2004; reprinted with permission.

Chapter 6: What can we learn from the last 2000 years about decadal and centennial scale climate change?

Paleoenvironmental records of the last 2000 years provide information on decadal and centennial scale climate variability, the associated ecological responses, and the context for anticipating future change. These records suggest that ecosystems have responded to climatic variation primarily through shifts in ranges, ecotonal position, and community composition and structure. High-resolution records indicate that the rate, magnitude, and duration of climate change strongly govern the ecological response. Species respond to changing environmental conditions by moving up or downslope and by increasing or decreasing in density (e.g., treeline "infill") and abundance. In contrast to the widely held assumption that changes in plant distribution are characterized by steady advances across landscapes, paleoenvironmental evidence suggests that such range adjustments are episodic in response to climatic conditions, occurring rapidly when conditions are suitable and slowly or not at all otherwise (Gray et al. 2006: Lyford et al. 2003). Predicting plant response to future climate change will require consideration of the rate and magnitude of climate change, spatial heterogeneity in biophysical conditions, the catalytic and synergistic role of ecosystem responses (e.g., fire, nutrient cycling, insect outbreaks), and intrinsic biotic limitations (Whitlock and Brunelle 2007; Jackson et al. 2009a).

Paleoenvironmental records from throughout the western United States demonstrate that ecosystem processes are strongly influenced by climate variability occurring at decadal and longer timescales whereas management planning is often based on climate "regimes" that are determined using a few decades of climate data. For example, fuel management goals are often developed from fire patterns over the last century even though paleoenvironmental data show that this only captures a single mode of climate variability (i.e., an extended regime of warm and dry or cool and wet conditions, e.g., Pierce et al. 2004, Whitlock et al. 2003). Records of change in the U.S. Rocky Mountains and Upper Columbia Basin provide context for better understanding current changes and the character of ecosystem response we might expect in the future. In particular, these records highlight:

- Decadal to centennial scale climate variability are related to ocean-atmosphere interactions such as ENSO, volcanic eruptions, and solar variability. These events include the Medieval Climate Anomaly (ca. 950–1250), the Little Ice Age (ca. 1400–1700), and numerous decadal and multidecadal droughts and pluvials.
- Prolonged droughts in the last millennia, which rival 20th century droughts in duration and magnitude, influenced vegetation in many ways, including:
    - ✧ Changes in treeline position.
    - ✧ Increases and decreases in treeline density (e.g., the "infill" phenomenon experienced in some areas during the MCA).
    - ✧ Multidecadal to centennial scale climate variability that results in episodic species dispersal, colonization, and establishment.
- At small temporal (decadal to centennial) and spatial (m^2 to km^2) scales, local factors (e.g., substrate, disturbance, and human activities) interact with climatic variation to influence the distribution of vegetation.
- Human activity was superimposed on climatic variability and altered prehistoric fire regimes in some areas of the western United States, but the impacts were mostly local (Whitlock and Knox 2002). As evidenced by changes in fire regimes of the late 19th and early 20th centuries, human activities left a strong imprint on the landscape and can rapidly influence ecosystems and ecosystem processes such as fire.

Key findings:

1. Using the last century as a baseline for climate conditions does not capture important scales of natural

climate variability and is often an inadequate reference for considering future climate change.

2. Even large-scale climatic changes can have spatially variable impacts due to interactions among a number of controlling factors. For example, cooling during the Little Ice Age, evident at the scale of the Northern Hemisphere occurred asynchronously or may not have registered at some locations in the study area (fig. 14).

3. Rapid climate changes and associated ecosystem responses have occurred in the past and will likely occur in the future.

Throughout much of the western United States, the expression of natural variations in the climate system can differ greatly across elevational, latitudinal, and longitudinal gradients and at different spatial and temporal scales. Likewise, the footprint of broad-scale climate changes will vary across finer spatial and temporal scales, and from one area to another. As a direct measurement of climate conditions for the last century, instrumental observations provide the most accurate and reliable data available, but they are influenced by local biases (e.g., differing slopes or aspects) along with the signature of finer-scale processes. Thus, individual instrumental records should not be interpreted as representing the region as a whole, but rather as an indication of local conditions or, at most, conditions at similar locations (e.g., with comparable elevation and vegetation cover).

7.1.1 Temperature

For all climate regions in this study, 20th century climate change is characterized by high spatial and temporal variability. At broad spatial and temporal scales, however, it is possible to summarize trends in climatic conditions impacting large portions of the four climate regions. Since 1900, temperatures have increased in most areas of the western United States from 0.5°C to 2°C (Pederson et al. 2010; Mote 2003; Ray et al. 2008), although cooling has occurred in a few areas, e.g., southeastern Colorado (Ray et al. 2008) and individual sites in the Northwest (CIG 2010). Where temperatures are trending higher, the rate of change varies by location and elevation, but is typically 1°C since the early 20th century (Hamlet et al. 2007). In most of the northern portions of the study area, temperatures generally increased from 1900 to 1940, declined from 1940 to 1975, and have increased since then (Parson 2001). Similarly, in the southern Rockies, temperatures generally increased in the 1930s and 1950s, cooled in the 1960s and 1970s, and have consistently increased since then (Ray et al. 2008). The rate of increase for much of the study area doubled since the mid-20th century, with most of this warming occurring since 1975 (fig. 19).

Temperature increases are more pronounced during the cool season (Hamlet and Lettenmaier 2007). In the northern U.S. Rocky Mountains, annual rates of increase are roughly 2–3 times that of the global average (Vose et al. 2005; Bonfils et al. 2008; Pederson et al. 2010; Hall and Fagre 2003), a pattern that is evident at northern latitudes and higher elevation sites throughout the West (Diaz and Eischeid 2007). In addition, nighttime minimum temperatures are increasing faster than are daytime maximums, resulting in a decreased diurnal temperature range (Pederson et al. submitted). This has implications for species like the mountain pine beetle (*Dendroctonus ponderosae*), whose population dynamics are governed by minimum temperatures (Carroll et al. 2004). Mean regional spring and summer temperatures were 0.87°C higher in 1987 to 2003 than in 1970 to 1986, and were the warmest since 1895 (Westerling et al. 2006). Bonfils et al. (2008) and Barnett et al. (2008)

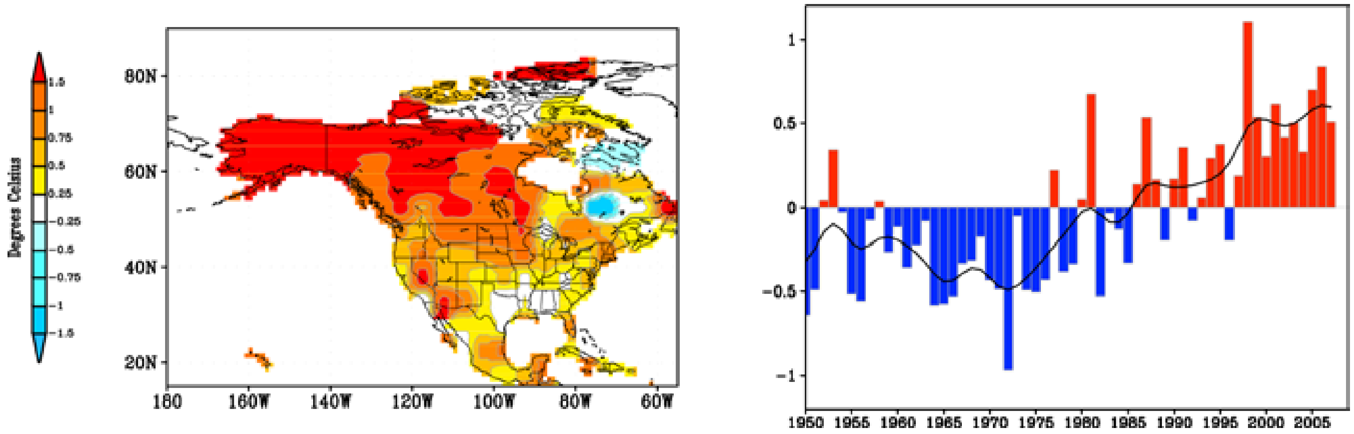


Figure 19. Observed annual average North American surface temperature, 1950–2007 (left) and the time series of the annual values of surface temperature averaged over North America (right). Annual anomalies are with respect to a 1971–2000 reference. The smoothed curve (black line) indicates low frequency variations. (Data source: UK Hadley Center's CRUv3 global monthly gridded temperatures.) Source: Ray et al. 2008; reprinted with permission.

found that the recent warming observed in mountainous areas across the West cannot be entirely explained by natural forcings (e.g., solar, volcanic, and ocean-atmosphere interactions); a major portion is attributable to human-influenced changes in greenhouse gas and aerosol concentrations.

7.1.2 Precipitation

Trends in precipitation for the study area are far less clear. Instrumental data from the last century show modest increases in precipitation for much of the northwestern United States (fig. 20; Mote et al. 1999, 2003, 2005), but no trends in parts of the southern Rockies (Ray et al. 2008). Natural variability in precipitation is evident in the instrumental record for all of the climate regions, and long-term drought conditions during the last century impacted large areas. Though 20th century droughts had substantial socioeconomic and ecosystem impacts, there is ample evidence they were not as severe in duration and magnitude as a number of drought events that occurred during the last millennium (Cook et al. 2007, 2004; Meko et al. 2007). For example, two droughts in the 1930s and 1950s impacted much of the study area. The 1930s drought was more widespread and pronounced in the northern and central climate regions while the 1950s drought was centered more on the south-central and southwestern United States (Cook et al. 2007; Gray et al. 2004; Fye et al. 2003). Research suggests that climatic conditions that influenced the nature and location of these droughts are likely linked to low-frequency oscillations in ocean-atmosphere interactions (McCabe et al. 2004; Gray et al. 2007; Hidalgo 2004; Graumlich et al. 2003), with evidence for substantial surface feedbacks during the 1930s “Dust Bowl” drought (Cook et al. 2009).

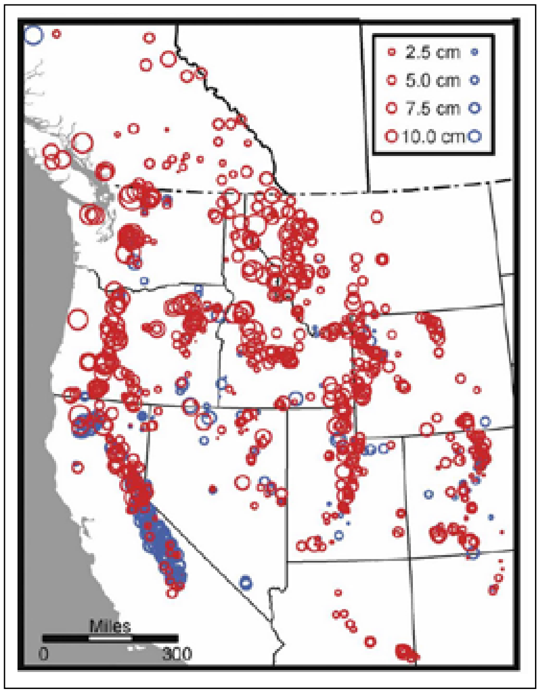


Figure 20. Trends in April 1 SWE from snow course observations, 1960–2002. SWE is an important indicator for forecasting summer water supplies. Positive trends are shown in blue and negative in red by the scale indicated in the legend. Source: Mote 2006, American Meteorological Society; reprinted with permission.

7.1.3 Surface hydrology

Generally speaking, snowmelt and peak runoff has tended to occur earlier since 1950, and river flows in many locations are decreasing during late summer (Pederson et al. 2010, 2009; Mote 2006; Barnett et al. 2008; Stewart et al. 2004; McCabe and Clark 2005). Recent impacts on snowpack and surface hydrology are strongly associated with more precipitation falling as rain than snow, and warming temperatures driving earlier snowmelt and snowmelt-driven runoff (fig. 21; Pederson et al. 2010, 2009; Mote 2006; Barnett et al. 2008; Stewart et al. 2004; McCabe and Clark 2005), leading to reduced surface storage of moisture and increasingly low baseflows during the dry summer months (Luce and Holden 2009). Examination of paleoclimate data and instrumental records (e.g., stream gages, snow course records, valley meteorological stations) suggests, however, that the total amount of cool season precipitation received across a particular region is more strongly associated with natural multidecadal, decadal, interannual, and intra-annual variability in ocean-atmosphere conditions (e.g., PDO, AMO, and ENSO).

Modest increases in precipitation have occurred in parts of the central and northern study area, but modest declines have taken place in parts of the southern Rockies, and significant declines in snowpack are evident throughout much of the Northwest. This decline is especially prevalent in the northern U.S. Rocky Mountains and parts of the Upper Columbia Basin (Hamlet and Lettenmaier 2007; Pederson et al. 2004, 2010; Selkowitz et al. 2002; Mote et al. 2005, 2008). Warmer temperatures and declining snowpack have, in turn, contributed to significant declines in the region’s glaciers. In Glacier National

Park, glaciers have decreased in area by more than 60% since 1900 (Hall and Fagre 2003; Key et al. 2002), and only 26 of the 150 glaciers and snow and ice fields present in 1910 remain (Pederson et al. 2010). Changes in glacier mass and area are emblematic of changes in surface hydrology across the West, with the recent substantial declines related to increases in greenhouse gases and aerosols (Barnett et al. 2008; Bonfils et al. 2008; Pierce et al. 2008; Hidalgo et al. 2009). During the last century, drought conditions have been increasing in the central and southern parts of the study area and decreasing in the northern parts (Andreadis and Lettenmaier 2006). Drought conditions are expected to increase for much of the study area over the coming decades (Hoerling and Eischeid 2007).

7.1.4 Ocean-atmosphere interactions

Ocean-atmosphere interactions are important drivers of interannual to multidecadal variability in temperature and precipitation (Pederson et al. 2010, 2009; Gray et al. 2007, 2004, 2003; McCabe et al. 2004; Hidalgo 2004; Cayan et al. 1998; Dettinger et al. 2000), but their impacts vary greatly across latitudinal, elevational, and longitudinal gradients. The El Niño–Southern Oscillation and Pacific Decadal Oscillation measures of high and low frequency variability in sea surface temperatures (SSTs) are, respectively, the predominant sources of interannual and interdecadal climate variability for much of the study area (Mantua et al. 2002, 1997). ENSO variations are commonly referred to as El Niño (the warm phase) or La Niña (the cool phase). An El Niño event is characterized by warmer than average sea surface temperatures in the central and eastern equatorial Pacific Ocean, reduced strength of the easterly trade winds in the tropical Pacific, and an eastward shift in the region of intense tropical rainfall (Bjerknes 1969; Cane and Zebiak 1985; Graham and White 1988; Horel and Wallace 1981; Philander 1990; Chang and Battisti 1998). A La Niña event is characterized by the opposite: cooler than average sea surface temperatures, stronger than normal easterly trade winds, and a westward shift in the region of intense tropical rainfall. Warm and cool phases typically alternate on 2 to 7 year cycles (Bjerknes 1969; Cane and Zebiak 1985; Graham and White 1988; Horel and Wallace 1981; Philander 1990; Chang and Battisti 1998).

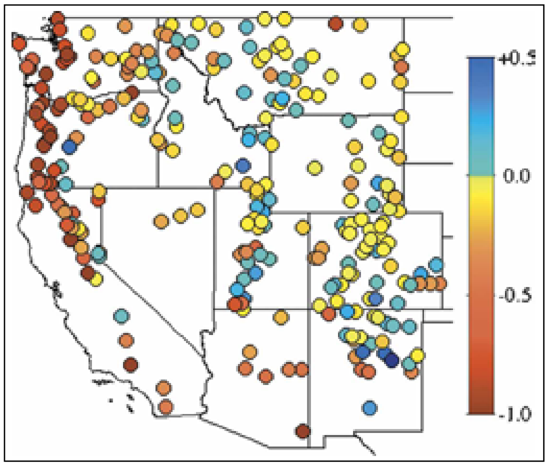


Figure 21. Trends in winter mean wet-day minimum temperatures, 1949–2004. Source: Knowles et al. 2006, American Meteorological Society; reprinted with permission.

The PDO exhibits alternate cool and warm phases with a spatial pattern similar to that of ENSO, but these phases typically last for 20 to 30 years (Mantua et al. 1997). Several switches occurred between warm and cool PDO phases during the 20th century and the magnitude of PDO phases increased in the latter half (McCabe et al. 2004; Mantua et al. 2002, 1997). The Atlantic Multidecadal Oscillation (AMO), representing low frequency (50–80 yr) oscillations in North Atlantic SSTs, has been linked to multidecadal variability in temperature and precipitation in the western United States through complex interactions with the PDO and ENSO, but the magnitude of the AMO influence is debated (Kerr 2000; McCabe et al. 2004; Enfield et al. 2001).

Changes in ENSO and PDO impact precipitation differently across the West, with winter precipitation in the Upper Columbia Basin and the northern Rockies being negatively correlated with warm conditions in the equatorial Pacific (i.e., during El Niños). Parts of the central and southern Rockies tend to be wetter than average during El Niños and dryer during La Niñas (Mote et al. 2005). Likewise, the Northwest generally receives more winter precipitation during the cool-phase PDO, and the Southwest often receives more precipitation during the warm phase. Further evidence for spatial heterogeneity in ENSO and PDO impacts can be seen in Greater Yellowstone, where windward aspects and the high mountains and plateaus to the west tend to follow more of an Upper Columbia Basin and northern Rockies pattern, while locations to the leeward and east often behave like the central and southern Rockies (Gray et al. 2004).

Multi-year droughts and extended dry regimes appear to be linked to complex interactions between the PDO, the AMO, and variations in ENSO. For example, the Dust Bowl drought, which was associated with a positive AMO and PDO, was centered primarily over the Southwest, whereas the 1950s drought (positive AMO and PDO) was centered more over Wyoming, Montana, and the Canadian Rockies (Gray et al. 2004; Hidalgo 2004; Fye et al. 2003). Drought conditions in the interior West are associated with low-frequency variations in the PDO and AMO (McCabe et al. 2004; Hidalgo 2004; Graumlich et al. 2003; Enfield et al. 2001) and these variations appear more pronounced in the northern and central Rockies than in parts of the Southwest (Hidalgo 2004). In the Southwest and coastal Northwest, variations in precipitation and warm-season water availability appear more sensitive to low-frequency ENSO variations than to PDO and AMO, although different combinations of these phases tend to amplify or dampen ENSO signals in climatic and hydrologic records (Gray et al. 2007, 2004; McCabe et al. 2004; Hidalgo 2004). While ocean-atmosphere interactions such as ENSO and PDO are partially responsible for variations in climatic conditions across this study area, research suggests that since the late 20th century increased greenhouse gas and aerosol concentrations have been amplifying, dampening, and, in some cases, overriding the influence of these phenomena (Barnett et al. 2008; Bonfils et al. 2008; McCabe et al. 2008; Gray et al. 2006, 2003).

7.1.5 Changes in storm track and circulation patterns

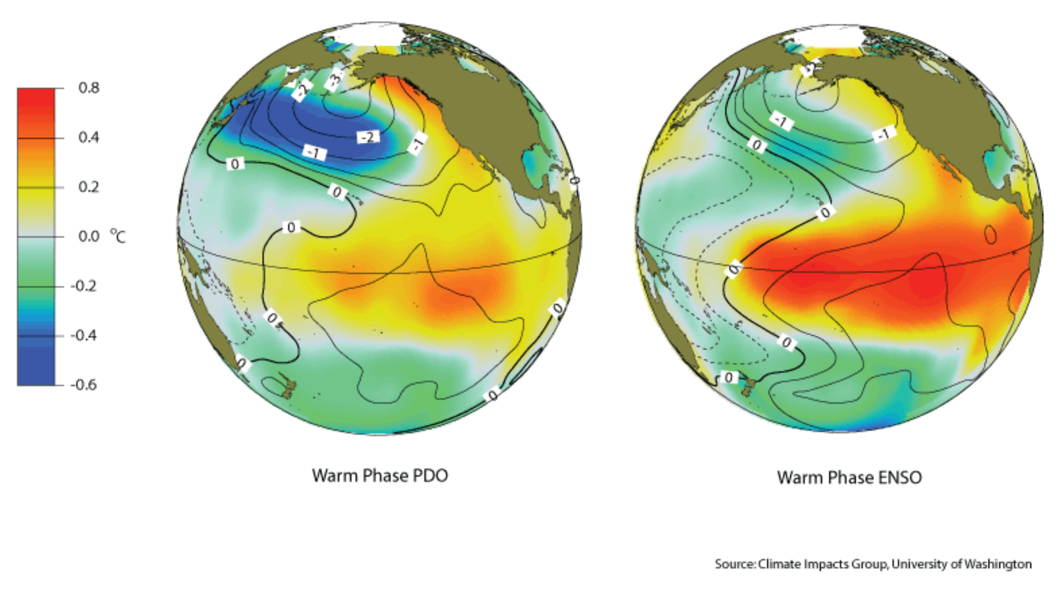
Simulations of 21st century climate suggest a northward movement of the storm tracks that influence precipitation over much of the western United States (Yin 2005; Lorenz and DeWeaver 2007), which could reduce precipitation for large parts of the study area. McAfee and Russell (2008) show that a strengthening of the Northern Annual Mode (an index of sea level pressure poleward of 20°N), which results in a poleward displacement of the Pacific Northwest storm track, increased west to east flow, reduced spring precipitation west of the Rockies, and increased spring precipitation east of the Rockies (McAfee and Russell 2008). This shift in the storm track is expected to persist well into the future and may reduce the length of the cool season, when circulation patterns provide the bulk of precipitation for large areas of the central and northern parts of the study area (McAfee and Russell 2008). If this becomes a more permanent shift in the storm-track position, it could lead to a longer duration of warm-season conditions (i.e., predominately warm and dry) for the Upper Columbia Basin, northern Rockies, and parts of the central Rockies. Changes in storm-track position and circulation patterns will be superimposed on a background of natural variability. Thus, various combinations of ENSO, PDO, and broad-scale trends could lead to local impacts that vary greatly in their magnitude over time.

7.1.6 Ecological impacts

Recent changes such as warming temperatures and associated declines in snowpack and surface-hydrology are already influencing ecosystem dynamics. Examples observed in the last century include: earlier spring blooming and leaf-out, forest infilling at and near the treeline, and increased severity of disturbances such as wildfire and insect outbreaks, all of which are likely to continue with additional warming. Spring blooming of a number of plant species has occurred earlier throughout much of the western United States, in some cases by as much as several weeks (Cayan et al. 2001; Schwartz and Reiter 2000; Schwartz et al. 2006). In the northern Rockies, increased density of trees at or near treeline has been observed at some sites (Butler et al. 2009). This "infill" phenomenon is not uncommon in the West and is predicted to continue where minimum temperatures rise, snowpack in high-snowfall areas decreases, and moisture is not limiting (Graumlich et al. 2005; Lloyd and Graumlich 1997; Rochefort et al. 1994; Millar et al. 2009). While evidence for infill is widespread, upslope movement in treeline

Ocean-atmosphere interactions: PDO and ENSO

The PDO has two phases: warm (positive index value) and cool (negative index value). Figure 1 shows the sea surface temperature anomalies associated with the warm phase of the PDO and ENSO, both of which favor anomalously warm sea surface temperatures near the equator and along the coast of North America, and anomalously cool sea surface temperatures in the central north Pacific. The cool phases for PDO and ENSO (not shown) have the opposite patterns: cool along the equator and the coast of North America and warm in the central north Pacific. Each PDO phase typically lasted for 20 to 30 years during the 20th century, and studies indicate that the PDO was in a cool phase from approximately 1890 to 1925 and 1945 to 1977 (Mantua 1997, 2002). Warm phase PDO regimes existed from 1925 to 1946 and from 1977 to at least 1998. Pacific climate changes in the late 1990s, in many respects, suggested another reversal from warm to cool phase and possibly back to warm.



Source: Climate Impacts Group, University of Washington

Figure 1. Warm phase PDO and ENSO. The spatial pattern of anomalies in sea surface temperature (shading, °C) and sea level pressure (contours) associated with the warm phase for 1900–1992. Note that the main center of action for the PDO (*left*) is in the north Pacific, and for the ENSO (*right*) in the equatorial Pacific. Contour interval is 1 millibar, with additional contours drawn for +0.25 and 0.5 mb. Positive contours are dashed; negative contours are solid. Source: Climate Impacts Group, University of Washington.

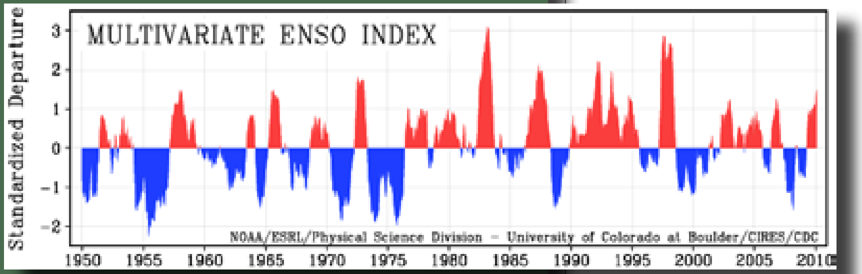


Figure 2. Multivariate ENSO index, 1950–2009. Positive (red) values indicate an El Niño event; negative (blue) values a La Niña event (Wolter and Timlin 1998, 1993).

Natural variation in the strength of PDO and ENSO events impacts climate regions in different ways. In the Northwest and parts of the central Rockies, warm-phase PDO and El Niño winters tend to be warmer and drier than average, with below normal snowpack and streamflow, whereas La Niña winters tend to be cooler and wetter than average, with above normal snowpack and streamflow (Graumlich et al. 2003; Cayan et al. 1998). In the southern Rockies and the Southwest, warm-phase PDO and El Niño winters tend to be wetter than average, with above normal snowpack and streamflow, and La Niña winters tend to be drier than average, with below normal snowpack and streamflow (Cayan et al. 1998; Swetnam and Betancourt 1998; Dettinger and Ghil 1998; Mote 2006).

position is much more variable, and research suggests that it will be characterized by a high degree of spatial heterogeneity in relation to other variables that control treeline position, e.g., aspect, soils, and micro-topography (Lloyd and Graumlich 1997; Graumlich et al. 2005; Bunn et al. 2005, 2007).

Changing climate conditions are also influencing disturbance processes that regulate ecosystem dynamics. Warming temperatures, earlier snowmelt, and increased evapotranspiration are increasing moisture stress on forest species and making them more susceptible to insect attack. An increase in the extent, intensity, and synchronicity of mountain pine beetle attacks in the western United States and Canada has been linked to forests stressed by drought, which makes trees less able to resist infestations (Nordhaus 2009; Hicke and Jenkins 2008; Romme et al. 2006; Logan et al. 2003; Carroll et al. 2004; Breshears et al. 2005). Warming temperatures have also influenced bark beetle population dynamics though reduced winter kill and by facilitating reproduction and dispersal (Carroll et al. 2004; Black et al. 2010). In some cases, past forest management (e.g., factors related to structural characteristics of host stands) may also facilitate beetle infestation (Nordhaus 2009; Black et al. 2010). The rate of fire disturbance is also increasing across the West, particularly in the northern Rockies (fig. 22, Westerling 2008).

The extent of the western United States burned in wildfires each year is strongly linked to interannual climate variability (Littell et al. 2008, 2009a; Morgan et al. 2008; Higuera et al. 2010). Changes in surface hydrology associated with reduced snowpack, earlier spring runoff and peak flows, diminished summer flows, and a lengthening fire season have all been linked to increased frequency of large fires, with the most evident impacts at mid-elevation forests in the northern Rockies since the mid-1980s (Westerling et al. 2006, fig. 22).

7.2 Northern U.S. Rocky Mountains

7.2.1 Temperature

Over the course of the 20th century, the instrumental record in the northern Rockies showed a significant increase in average seasonal, annual, minimum, and maximum temperatures (figs. 23, 24; Loehman and Anderson 2009; Pederson et al. 2010, submitted). Regional average annual temperatures increased 1–2°C (2–4°F) from 1900 to 2000 (Pederson et al. 2010). Seasonal and annual minimum temperatures are generally increasing much faster than maximum temperatures (Pederson et al. 2010). In particular, summer and winter seasonal average minimum temperatures are increasing significantly faster than the season's average maximum temperatures, causing a pronounced reduction in the seasonal diurnal temperature range (Pederson et al. 2010). The magnitude of minimum temperature increases also appears seasonally variable: in areas with mid-elevation snow telemetry (SNOTEL) sites, Pederson et al. (submitted) estimated minimum temperature increases since 1983 of 3.8 ± 1.72°C (6.8 ± 3.10°F) in winter, 2.5 ± 1.23°C (4.5 ± 2.21°F) in spring, and 3.5 ± 0.73°C (6.3 ± 1.31°F) annually (fig. 24). The magnitude of changes varies locally, but there are few exceptions to this general warming trend.

Temperature trends within the northern Rockies generally track Northern Hemisphere trends

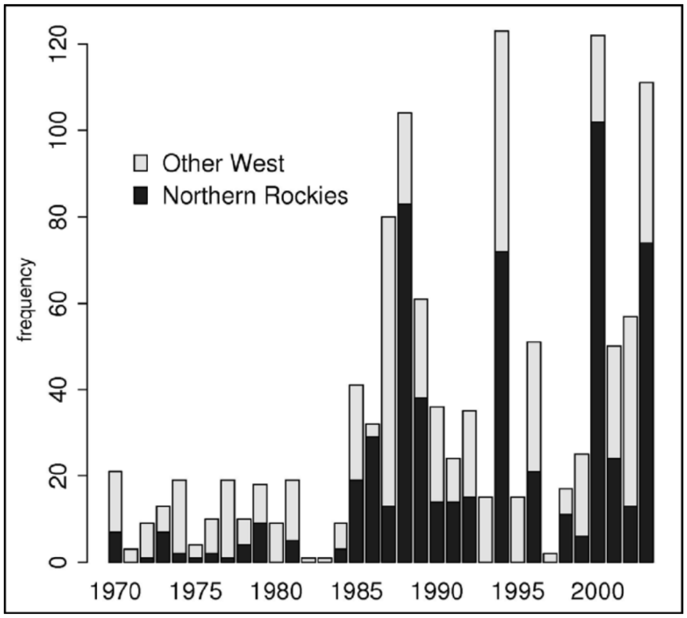


Figure 22. Annual number of forest fires >1000 acres (total column height) in the northern Rockies (black area) and other western states. Source: Westerling 2008; reprinted with permission.

across temporal scales (fig. 23). This similarity between regional and continental trends suggests that large-scale climate forcings such as greenhouse gases, sea surface temperature patterns, volcanic activity, and solar variability also drive regional temperatures (Pederson et al. 2010).

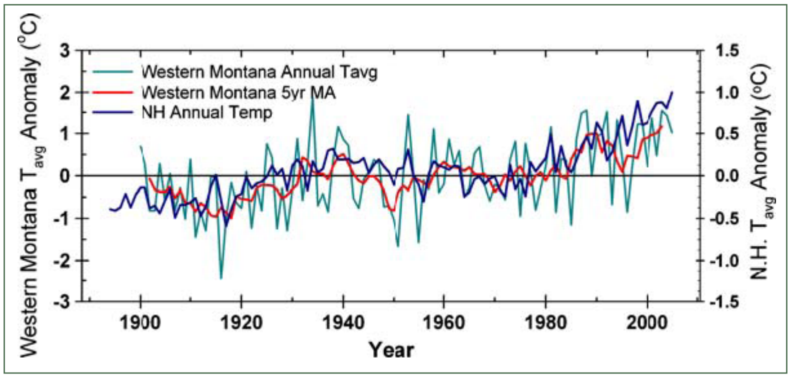


Figure 23. Comparison of variability and trends in western Montana (blue-green line) and Northern Hemisphere (dark blue line) annual average temperatures. A 5-year moving average (red line) highlights the similarity in trends and decadal variability between records. Source: Pederson et al. 2010; reprinted with permission.

7.2.2 Precipitation

Throughout the West, high interannual, annual, and decadal variability in precipitation exceeds any century-long trends (Ray et al. 2008). General patterns throughout the latter part of the 20th century indicate that areas within the northern Rockies experienced modest but statistically insignificant decreases in annual precipitation (Mote et al. 2005; Knowles et al. 2006). Although few statistically significant trends are evident in regional 20th century precipitation, rising temperatures throughout the West have led to an increasing proportion of precipitation falling as rain rather than snow (Knowles et al. 2006). Winter temperatures well below 0°C make the northern Rockies less sensitive than other western regions where small temperature increases in temperature are affecting the number of freezing days (Knowles et al. 2006).

7.2.3 Surface hydrology

Like most of the western United States, the snow water equivalent (SWE) of winter snowpack largely controls surface runoff and hydrology in the northern Rockies for the water year (e.g., Pierce et al. 2008; Barnett et al. 2008; Stewart et al. 2005; Pederson et al. submitted). Studies have demonstrated a statistically significant decrease in winter snowpack SWE across the region during the second half of the 20th century (Barnett et al. 2008; Pederson et al. submitted).

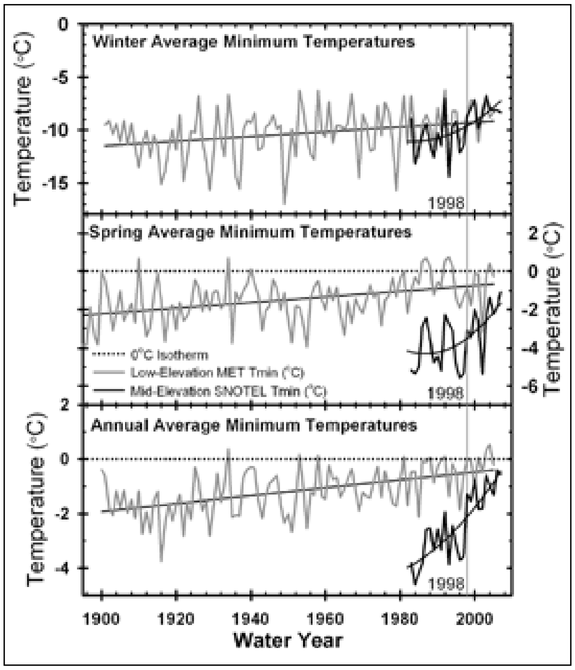


Figure 24. Average winter (Dec–Feb; top), spring (Mar–May; middle), and annual (bottom) minimum temperatures. Data from SNOTEL (water year Oct–Sep) and valley MET (Jan–Dec) stations. SNOTEL station Tmin (minimum temperature) records have been fit using a non-linear quadratic equation due to characteristics of these time series. All trends shown are significant ($p \leq 0.05$). Note the y-axis temperature scale changes for each panel. Source: Pederson et al. submitted.

7.2.4 Ocean-atmosphere interactions

The warm phase of the PDO was associated with reduced streamflow and snowpack in the northern Rockies during the 20th century, (Fagre et al. 2003), and the cool phase with increased streamflow and snowpack (Pederson et al. submitted). One ecological response to these shifts has been changes in the distribution of mountain hemlock (*Tsuga mertensiana*). At high elevations, where mountain hemlock growth is limited by snowpack-free days, a warm-phase PDO often results in decreased snowpack and increased mountain hemlock growth (Peterson and Peterson 2001). At low elevation sites where moisture is limiting, a warm-phase PDO commonly leads to less moisture and consequently decreased mountain hemlock growth and establishment (Fagre et al. 2003).

Pederson et al. (submitted) summarize how variation in Pacific SSTs, atmospheric circulation, and surface feedbacks influence climate conditions, snowpack, and streamflow in the northern Rockies. Winters with high

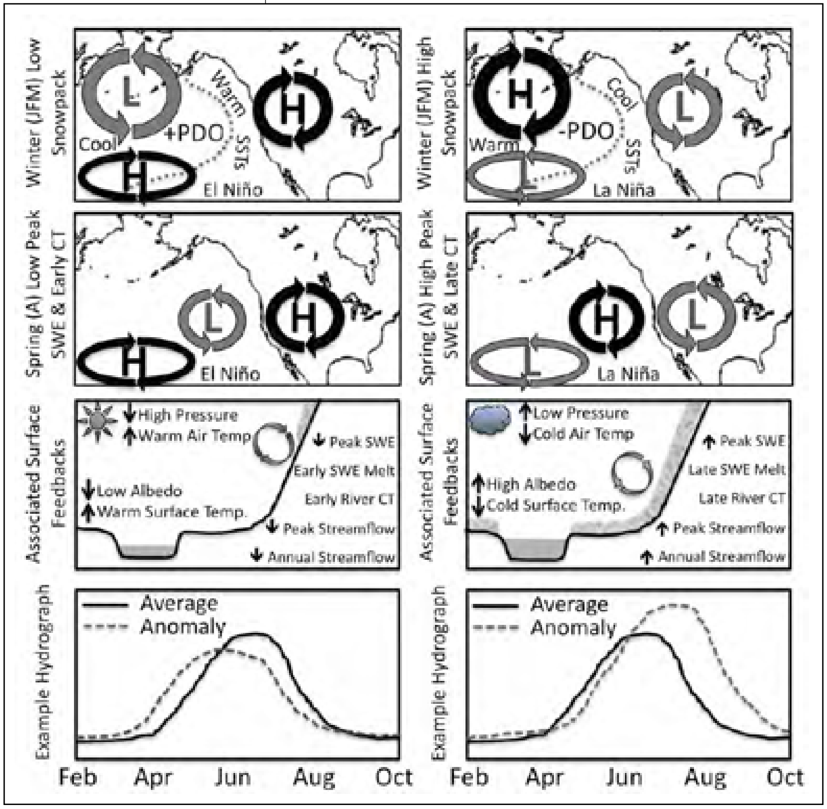

Winter (JFM) Low Snowpack
L
Warm
H
Cool
+PDO
SSTs
H
El Niño
Winter (JFM) High Snowpack
H
Cool
L
Warm
-PDO
SSTs
L
La Niña
Spring (A) Low Peak SWE & Early CT
L
H
H
El Niño
Spring (A) High Peak SWE & Late CT
H
L
L
La Niña
Associated Surface Feedbacks
↓High Pressure
↑Warm Air Temp
↓ Peak SWE
Early SWE Melt
Early River CT
↓Low Albedo
↑Warm Surface Temp.
↓ Peak Streamflow
↓ Annual Streamflow
Associated Surface Feedbacks
↑Low Pressure
↓Cold Air Temp
↑ Peak SWE
Late SWE Melt
Late River CT
↑High Albedo
↓Cold Surface Temp.
↑ Peak Streamflow
↑ Annual Streamflow
Example Hydrograph
Average
Anomaly
Example Hydrograph
Average
Anomaly
Feb
Apr
Jun
Aug
Oct
Feb
Apr
Jun
Aug
Oct

Figure 25. Idealized relationship between northern Rockies snowpack and streamflow anomalies with associated Pacific SSTs, atmospheric circulation, and surface feedbacks. Source: Pederson et al. submitted.

snowpack tend to be associated with the cool phase of the PDO, a weakened Aleutian Low, and low pressure centered poleward of 45°N across western North America (fig. 25). During years of high snowpack, mid-latitude cyclones tend to track from the Gulf of Alaska southeast through the Pacific Northwest and into the northern Rockies. The relatively persistent low-pressure anomaly centered over western North America is also conducive to more frequent Arctic-air outbreaks, resulting in colder winter temperatures.

The ENSO is an important driver of snowpack and streamflow at interannual scales, and the influence of related tropical Pacific atmospheric circulation anomalies persists well into the spring. Changes in spring (MAM) temperatures and precipitation are associated with changes in regional atmospheric circulation, and strongly influence the timing of streamflow in the northern Rockies (fig.25). Springtime geopotential heights over western North America influence the amount and more importantly the timing of snowmelt and streamflow across the northern Rockies. Specifically, high pressure anomalies centered over western North America are associated with higher spring temperatures and consequently an increasing number of snow-free days and earlier arrival of snow melt-out and peak streamflow. Atmospheric circulation changes in March and April can, in turn, initiate surface feedbacks that contribute to surface temperature and hydrograph anomalies (fig. 25). Hence, warming temperatures in the northern Rockies lead to earlier snowmelt and runoff and associated decreasing snowpack and streamflow, but these patterns can be partially attributed to seasonally-dependant teleconnections and atmospheric circulation patterns, as well as to surface-albedo feedbacks that interact with broad-scale controls on snowpack and runoff (Pederson et al. submitted).

7.3 Central U.S. Rocky Mountains and the Greater Yellowstone Area

7.3.1 Temperature

Temperatures for the CR-GYA increased 1–2°C (2–4°F) during the last century, with the greatest increases occurring in the second half of the 20th century (Vose et al. 2005; Bonfils et al. 2008; Mote 2006, 2003). This rate of increase is slightly higher than in the Southwest and slightly lower than in the northern Rockies, following a pattern of more pronounced temperature increases at higher latitudes (Cayan et al.2001; Ray et al. 2008). Increasing winter and spring temperatures have resulted in reduced snowpack, earlier spring snowmelt and peak flows, and, in some cases, lower summer flows for major basins (Mote 2006; McCabe and Clark 2005; Stewart et al. 2004; Hidalgo et al. 2008).

7.3.2 Precipitation

CR-GYA precipitation records show highly variable patterns across gradients in elevation, latitude, and longitude. No long-term trends are evident over the last century; reconstructions of hydrology from tree-ring records indicate interannual, decadal, and multidecadal variation (fig. 26; Watson et al. 2009; Graumlich et al. 2003). A greater proportion of precipitation is likely falling as rain rather than snow in this region but the impacts are less pronounced than in other parts of the western United States (Knowles et al. 2006). In many parts of the CR-GYA, the 1930s and 1950s were significantly drier than average and the 1940s wetter, although sub-regional variation is high, likely because the region is located between atmospheric circulation patterns, as discussed further below (Watson et al. 2009; Gray et al. 2007, 2004, 2003; Graumlich 2003).

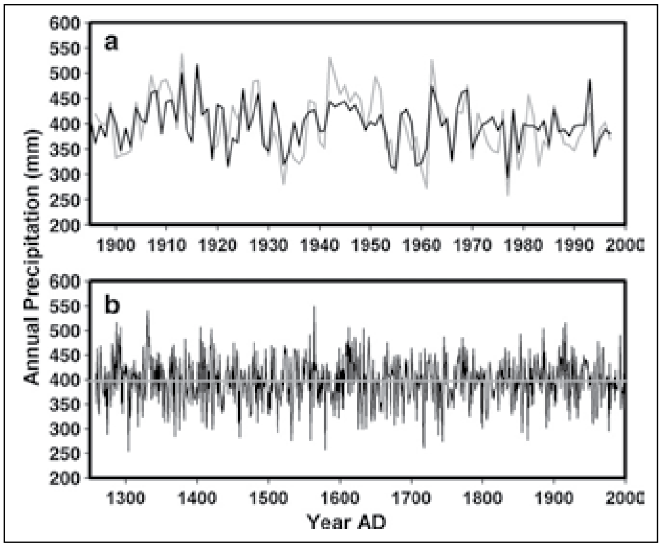


Figure 26. (a) Observed annual (previous July through current year June) precipitation for Wyoming Climate Division 1 (gray line) compared to precipitation estimates based on the stepwise regression model (black line). (b) The full stepwise version of reconstructed annual precipitation (black line) for 1173 to 1998. The horizontal line (solid gray) near 400 mm represents the series mean, and the vertical line (dotted gray) at 1258 divides the well-replicated portion of the record from reconstructed values in the earlier, less replicated years. Source: Gray et al. 2007; reprinted with permission.

7.3.3 *Ocean-atmosphere interactions*

The influence of ocean-atmosphere interactions on decadal, multidecadal, and interannual variation in climatic conditions is more spatially variable in the CR-GYA than in the other regions because it falls in a transition area between northwestern and southwestern U.S. circulation patterns (Gray et al. 2007, 2004; Graumlich et al. 2003), where variations in ocean-atmosphere interactions, topography, latitude, and longitude often result in opposite trends in climatic conditions at sites within the same region (Gray et al. 2004).

High-elevation snow basins within the western GYA typically respond to large-scale climate forcing in a manner similar to that of the Pacific Northwest, where the cool-phase PDO results in cool, wetter than average winters and the warm-phase PDO brings warmer and drier than average winters (Gray et al. 2007, 2004; Graumlich et al. 2003; Dettinger et al. 1998). Similar to the Pacific Northwest, these areas experience increased precipitation during La Niña events and decreased precipitation during El Niño events, and the ENSO seems to be linked to the magnitude of PDO anomalies, especially during the winter (Gray et al. 2007). Alternatively, lower elevation sites and eastern portions of the GYA respond more like the Southwest or show a variable response to ENSO that depends heavily on the strength of event and interactions with other climate drivers (Gray et al. 2004; McCabe et al. 2007). Years with strong El Niño SSTs have increased winter precipitation and La Niña events bring drier conditions. This difference between high and low elevation precipitation regimes is common throughout the central Rockies, complicating predictions of future precipitation in the region (McCabe et al. 2007).

7.4 Southern U.S. Rocky Mountains

7.4.1 *Temperature*

Temperatures have increased 0.5–1°C (0.9–1.8°F) throughout the southern Rockies during the last 30 years. The north central mountains of Colorado warmed the most (~1.35˚C, 2.43°F) and high elevations may be warming more quickly than lower elevations in some regions (Pepin and Lundquist 2008; Diaz and Eischeid 2007). Warming is evident at almost all locations, but temperatures have increased the most in the north central mountains and the least in the San Juan Mountains of southwestern Colorado (Ray et al. 2008). Only the Arkansas River Valley in southeastern Colorado shows a slight cooling trend during the 20th century; no trend is evident in this area for the second half of the century (Ray et al. 2008).

7.4.2 *Precipitation*

Precipitation records for the southern Rockies for the last century indicate highly variable annual amounts and no long-term trends (Ray et al. 2008; Dettinger 2005). Like elsewhere in the interior West, a greater proportion of precipitation is falling as rain rather than snow than in the past, but these changes are less pronounced than in the northern Rockies (Knowles et al. 2006). Decadal variability is evident in records of precipitation and surface flows and is linked to variability in ocean-atmosphere and

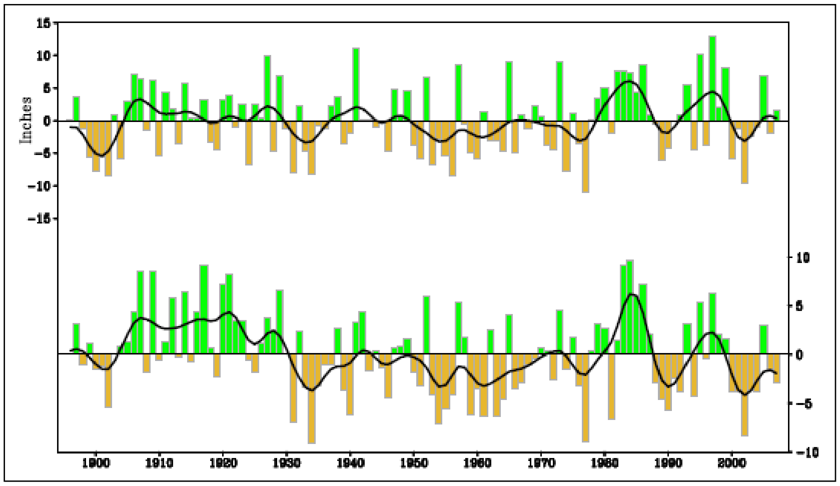


Figure 27. Observed time series (1895–2007) of annually averaged precipitation departures area-averaged over the Upper Colorado drainage basin (top) and annual Colorado River natural flow departures at Lees Ferry in million acre-feet (bottom). The precipitation data are based on 4-km gridded PRISM data. Colorado River natural flow data from the Bureau of Reclamation. Source: Ray et al. 2008; reprinted with permission.

land-surface interactions (fig. 27; Stewart et al. 2005).

7.4.3 Surface hydrology

Similar to trends evident throughout the interior West, more precipitation is falling as rain than snow in the southern Rockies, spring snowpack is decreasing, especially at elevations below 2500 meters, and peak streamflows are occurring earlier because of warmer spring temperatures (Knowles et al. 2006; Bales et al. 2006; Stewart et al. 2005; Hamlet et al. 2005; Clow 2007; Mote 2006, 2003). Summer flows are typically lower and annual flows show high variability but no significant trends in most locations (Ray et al. 2008).

7.4.4 Ocean-atmosphere interactions

Like the CR-GYA, the Colorado River Basin spans a transition area where the influence of Pacific Northwest and southwestern circulation patterns show opposite trends (Gray et al. 2007; Clark et al. 2001). During El Niño years, northern parts of this region experience drier than average conditions while the southern portions experience wetter than average conditions. The opposite conditions occur during La Niña years, and anomalies tend to be more pronounced in spring in southern portions. Long-term droughts are linked more closely to low-frequency oscillations in PDO and AMO, and are most commonly associated with the interaction between a cool-phase PDO and warm-phase AMO (McCabe et al. 2004).

7.5 Upper Columbia Basin

7.5.1 Temperature

For most of the Upper Columbia Basin, average annual temperatures increased 0.7–0.8°C from 1920 to 2003, and the warmest decade was the 1990s (fig. 28; Littell et al. 2009b). Average temperatures have increased as much as 2°C (4°F) in parts of the region, and increases have been more pronounced at higher elevations (Mote 2006, 2003). During the mid-20th century, average and daily minimum temperatures increased more in the winter and spring than in other seasons and more than maximum temperatures. During the second half of the 20th century, minimum and maximum temperatures increased at similar rates (Watson et al. 2009; Dettinger et al. 1994; Karl et al. 1993: Lettenmaier et al. 1994; Littell et al. 2009b).

7.5.2 Precipitation

Upper Columbia Basin precipitation trends are less clear than temperature trends, and observations indicate high decadal variability. Precipitation increased 14% for

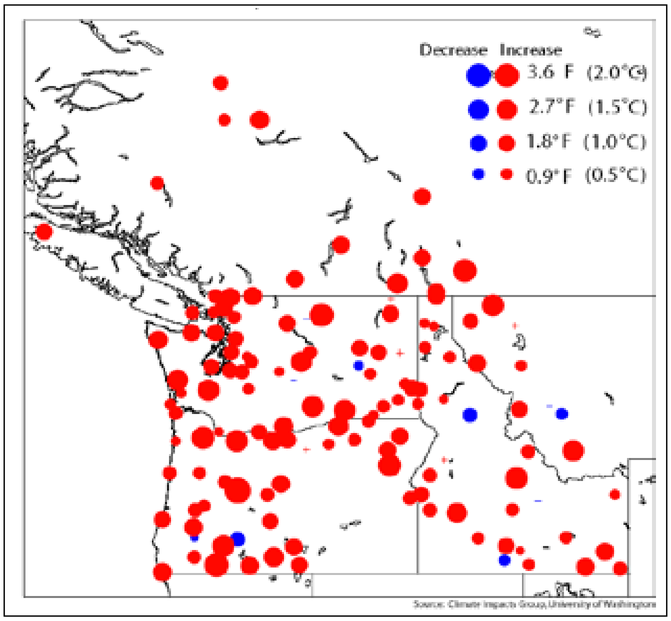


Figure 28. Trends in average annual Pacific Northwest temperature, 1920–2000. Increases (decreases) are indicated with red (blue) dots, and the size of the dot corresponds to the magnitude of the change. Pluses and minuses indicate increases or decreases, respectively, that are less than the given scale. Source: Climate Impacts Group, University of Washington.

the entire northwestern United States, (1930–1995), and increases ranged between 13% and 38% within the region (fig. 29; Mote 2003), but these trends are often not statistically significant, depending on the area and time interval measured. Similar to much of the interior West, variability in winter precipitation has increased since 1973 (Hamlet and Lettenmaier 2007).

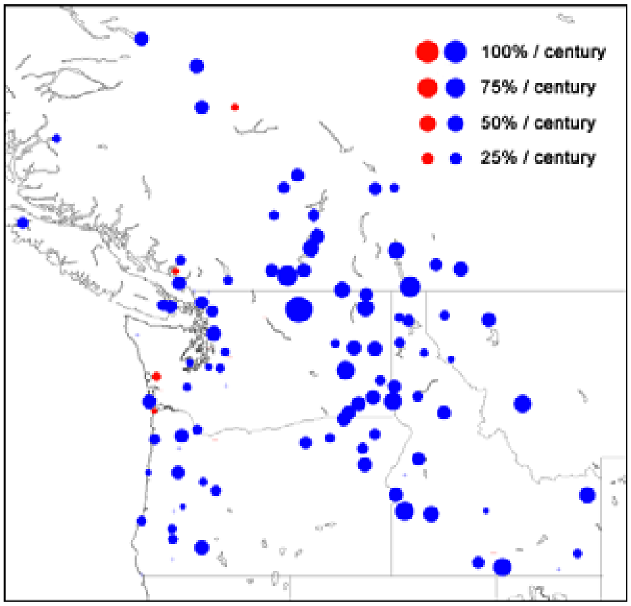
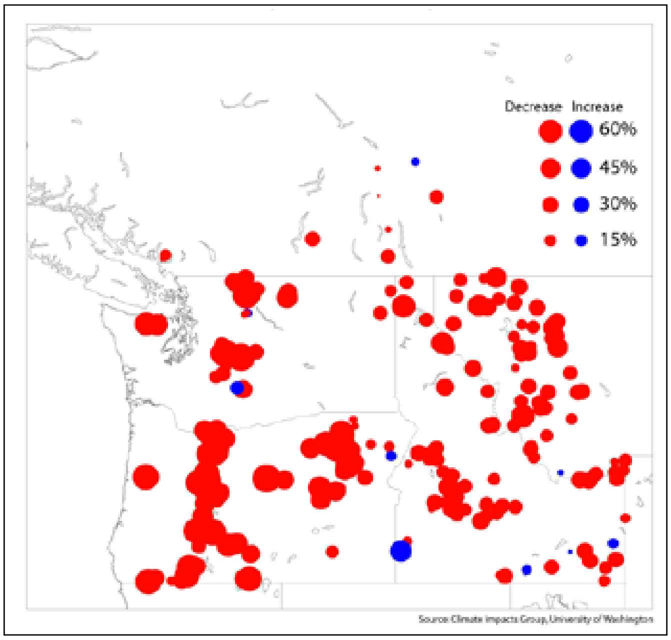


Figure 29. Trends in average annual precipitation, 1920–2000. Increases (decreases) are indicated with blue (red) dots, and the size of the dot corresponds to the magnitude of change. Source: Climate Impacts Group, University of Washington.

7.5.3 Surface hydrology

Spring snowpack and SWE declined throughout the Upper Columbia Basin in the second half of the 20th century. The decline was most pronounced at low and mid-elevations, and declines of more than 40% were recorded for some parts of the region (fig. 30; Hamlet et al. 2005; Mote 2006, 2003). Declines in snowpack and SWE are associated with increased temperatures and declines in precipitation during the same period (Mote et al. 2005; Mote 2003). The timing of peak runoff shifted 2–3 weeks earlier for much of the region during the second half of the 20th century (Stewart et al. 2004), and the greatest shifts occurred in the mountain plateaus of Washington, Oregon, and western Idaho (Hamlet et al. 2007). Because Upper Columbia Basin ecosystems rely on the release of moisture from snowpack, these shifts are significantly impacting plant species, causing some to bloom and leaf out earlier in the spring (Mote et al. 2005; Cayan et al. 2001; Schwartz and Reiter 2000).



Source: Climate Impacts Group, University of Washington

Figure 30. Trends in April 1 SWE, 1950–2000. Increases (decreases) are indicated with blue (red) dots, and the size of the dot corresponds to the magnitude of the change. Pluses and minuses indicate increases or decreases, respectively, that are less than the given scale. Source: Climate Impacts Group, University of Washington.

7.5.4 Ocean-atmosphere interactions

Variations in Upper Columbia Basin climatic conditions are related to ocean-atmosphere and land-surface interactions, namely the ENSO and PDO phenomena. In their warm phases, both the ENSO and PDO increase the chance for a warmer winter and spring in the Upper Columbia Basin and decrease the chance that winter precipitation will reach historical averages. The opposite tendencies are true during a cool-phase ENSO and PDO: they increase the odds that Upper Columbia Basin winters will be cooler and wetter than average (Clark et al. 2001). While strong El Niño years are typically warmer than average, SWE anomalies are often less pronounced then, and winter precipitations are commonly close to historical averages (Clark et al. 2001). Clark et al. (2001) suggested that El Niño circulation anomalies are centered more in the interior West than are La Niña circulation anomalies and are most evident in mid-winter.

8.1.1 Small changes can have large impacts

Changes in the distribution of minimum temperatures and frost-free days illustrate how small changes in temperature (1–2°C, 2–4°F) may result in large changes to surface hydrology (Barnett et al. 2004, 2005) as they contribute to earlier melt-off and diminished spring snowpack (Pederson et al. 2010, 2009; Mote 2006; Barnett et al. 2008; Stewart et al. 2004; McCabe and Clark 2005), increases in the proportion of winter precipitation as rain rather than snow (Knowles et al. 2006; Bales et al. 2006), decreased snow season length at most elevations (Bales et al. 2006), and lower summer flows (Barnett et al. 2008). Evidence from a number of studies suggests that even small temperature increases can have dramatic impacts on water availability for much of the western United States. Along with changes in snowpack and earlier spring runoff, the predicted temperature increases will likely contribute to increased drought severity, duration, and frequency (fig. 31; Hoerling and Eischeid 2007; Barnett and Pierce 2009). Increased winter precipitation predicted for central and northern regions of the study area will likely be inadequate to offset the increased evaporation and plant water use driven by rising temperatures (Gray and McCabe 2010; Hoerling and Eischeid 2007; Seager et al. 2007).

8.1.2 Shifting distributions and new norms

Many parts of the study region are vulnerable to small changes in temperature because the overall climate is arid to semi-arid to begin with, and the water available in these areas depends heavily on the mountain snowpack dynamics (Gray and Anderson 2010). While ecosystems are adapted to natural variations in water availability, a shift in drought frequency and magnitude, or even the occurrence of an especially severe and prolonged dry event, could result in regional ecosystems reaching a tipping point

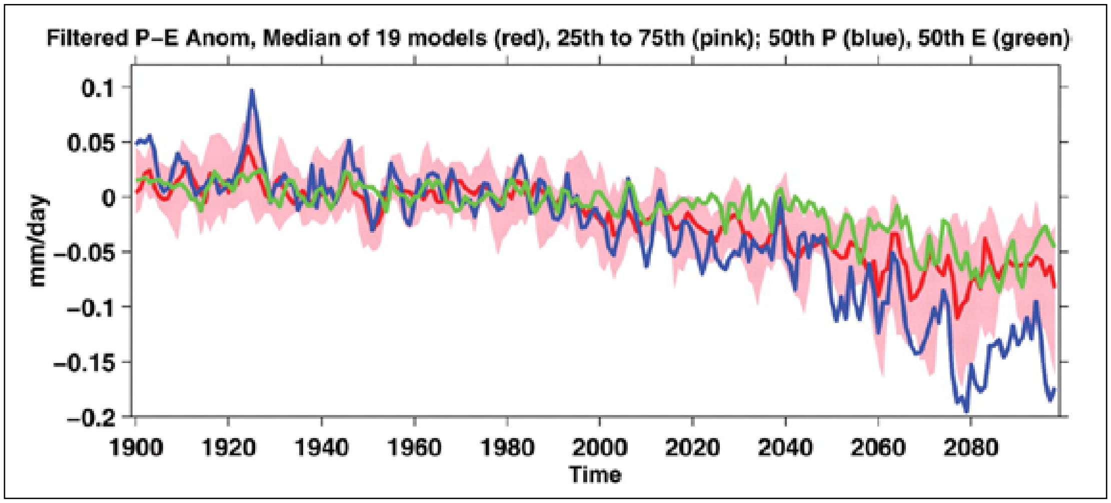


Figure 31. Modeled changes in annual mean precipitation minus evaporation (P–E) over the Southwest (125°W to 95°W and 25°N to 40°N, land areas only), averaged over ensemble members for each of the 19 climate models participating in the Fourth Assessment Report (AR4) of the IPCC (Solomon et al. 2007). The historical period used known and estimated climate forcings; the projections used the SResA1B emissions scenario. The median (red line) and 25th and 75th percentiles (pink shading) of the P–E distribution among the 19 models are shown, as are the ensemble medians of P (blue line) and E (green line) for the period common to all models (1900–2098). Anomalies (Anom) for each model are relative to that model's climatology from 1950 to 2000. Results have been 6-year, low-pass, Butterworth-filtered to emphasize low-frequency variability that is of most consequence for water resources. The model ensemble mean P–E in this region is ca. 0.3 mm/day. Source: Seager et al. 2007; reprinted with permission.

How do we know if observed changes are related to human-caused climate change?

The Intergovernmental Panel on Climate Change (IPCC, Solomon et al. 2007) included studies to determine whether a detected climate change is significantly different from natural variations of the climate system. Attribution studies seek to establish the principal causes for observed climate phenomena, including trends in temperature and extreme climate events, and whether they are related to human activities. In order to attribute a detected change, scientists must demonstrate that the change is consistent with an identified anthropogenic cause and inconsistent with any alternative, physically plausible explanation that excludes anthropogenic causes (Houghton et al. 2001). If attribution is established, the IPCC may assign a likelihood for the probability that the identified cause resulted in the observed conditions.

Attribution studies use empirical analyses of past climate relationships and evaluate cause-and-effect relationships with climate models. Model simulations of past climate are compared with the observed record using statistical analysis, including estimates of natural variability and trends from climate models, historical observations, or paleoclimate reconstructions. "Fingerprint" methods seek the unique signature of climate change by simultaneously looking at changes in many variables. In studies conducted to determine the cause of the observed warming of temperatures in western and northern North America over the last half-century, annually averaged North American surface temperatures from 1950 to 2007 were computed using the IPCC (CMIP3) models forced with the observed record of greenhouse gases, volcanic aerosols, and solar forcing during 1950 to 1999 and subsequently (2000–2007) with the A1B scenario of greenhouse gas emissions. (The A1B scenario = rapid economic growth, global population that peaks in mid-century and declines thereafter, and the rapid introduction of new and more efficient technologies). Comparison of these modeled temperatures (fig. 1, top panel) with observations (fig. 19) suggests that anthropogenic greenhouse gas emissions have contributed about 1°C (1.8°F) of the observed warming in the last 30 years. Similarities between the modeled climate and the observed trends provide the best available evidence for external (rather than natural variability) forcing of surface temperature change by anthropogenic greenhouse gases because the bulk of the warming occurs after about 1970 in both time series and the externally forced warming of about 1°C (1.8°F) since 1950 is close to the observed rate.

A series of recent studies sought to detect and attribute climate change in the western United States (Bonfils et al. 2008; Pierce et al. 2008; Hidalgo et al. 2009; Das et al. 2009) using the same downscaled projections and PRISM data as the westwide projections shown below (figs. 35, 36). Bonfils and colleagues conducted a detailed analysis of

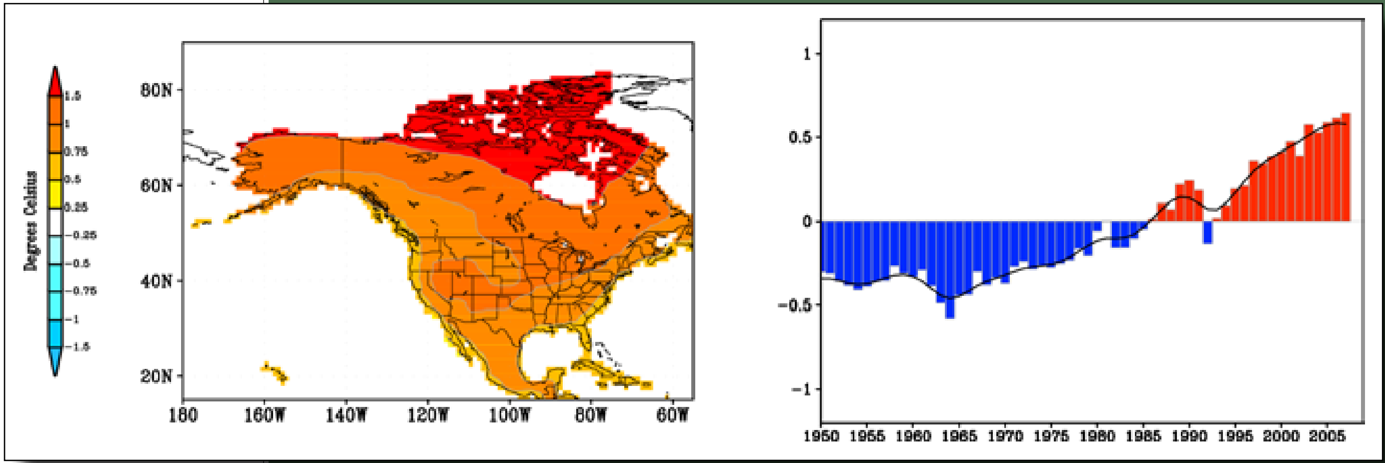


Figure 1. Annual average North American surface temperature (1950–2007) from 22 IPCC model simulations with greenhouse gas, aerosol, solar, and volcanic forcing from 1950 to 1999, and the A1B emissions scenario from 2000 to 2007 (top panel). Annual anomaly values of surface temperature averaged over the whole of North America (compared to 1971–2000 average, lower panel). The smoothed curve highlights multi-year variations. Source: Ray et al. 2008, adapted from CCSP 1.3, figure 3.3; reprinted with permission.

models thought to best simulate the climate of the western United States, and using these models, found that natural variability is insufficient to explain the increase in daily minimum and maximum temperatures, a sharp decline in frost days, a rise in degree days above 0°C (32°F), and a decline in snowpack at low and mid-elevations. They ruled out solar variability and volcanic forcing as a cause. They found that the anthropogenic signal is detectable by the mid-1980s in a signal-noise ratio of minimum temperature. Other attribution papers focus on streamflow (Hidalgo et al. 2009), snowpack (Pierce et al. 2008), and the structure and detectability of hydrological variables (Das et al. 2009). These studies have estimated that up to about half of the trends in temperature and associated hydrologic variables can be attributed to anthropogenic causes (Barnett et al. 2008; Pierce et al. 2008).

whereby a major redistribution of vegetative communities ensues (Gray et al. 2006; Jackson et al. 2009b). Rapid changes in surface hydrology related to altered snowpack dynamics could bring similar impacts to aquatic ecosystems that have developed under distinctly different hydroclimatic regimes (fig. 32).

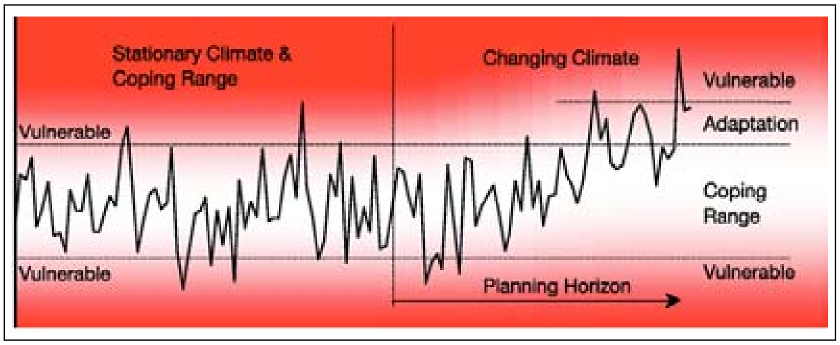


Figure 32. Relationships between climate change, coping range, vulnerability thresholds, and adaptation. Idealized version of a coping range, showing the relationship between climate change and threshold exceedance and how adaptation can establish a new critical threshold, reducing vulnerability to climate change (modified from Jones and Mearns 2005). Source: IPCC AR4 WGI 2007; reprinted with permission.

Small increases in temperature (e.g., 1–2°C, 2–4°F) will result in greater evaporative losses from lakes, streams, wetlands and terrestrial ecosystems, and it is likely that this enhanced evaporation will lead to significant ecosystem and water management impacts (Arnell 1999; Gray and Andersen 2010). In the central and southern portions of the study area, the increases in winter precipitation predicted by some models are not expected to offset this increased evaporation and transpiration. For example, models from Gray and McCabe (2010) estimate a 15–25% decrease in average Yellowstone River flows from a 1.5–3°C (2.7–5.4°F) temperature increase, and it would require the equivalent of the wettest years in the last millennium to offset the impacts of increased evapotranspiration on this system (Gray and McCabe 2010). As seen in many recent observational records, seemingly small changes in mean conditions can lead to an increased frequency of hot weather relative to historical conditions and extreme precipitation events (Karl et al. 2009; Solomon et al. 2007; Groisman et al. 2005; Kunkel et al. 2003; Madsen and Figdor 2007; fig. 33). Multiple assessments point to a potential shift in precipitation such that storms will become more intense but less frequent (Groisman et al. 2005; Kunkel et al. 2003; Madsen and Figdor 2007). This, in turn, would increase the number of dry days between precipitation events, while also altering runoff, infiltration, and erosion rates.

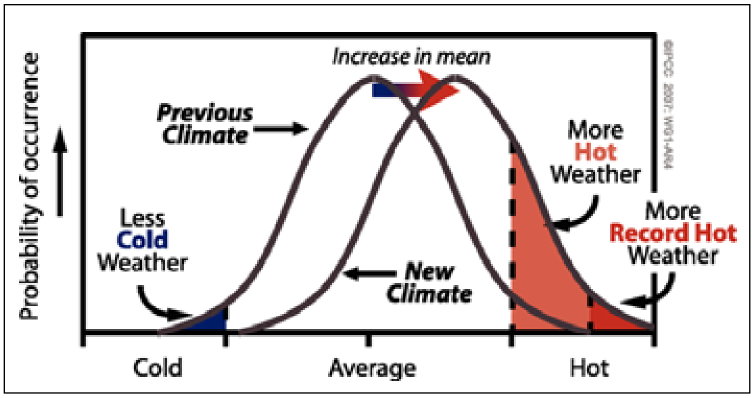


Figure 33. Schematic for a normal temperature distribution showing the effect on extreme temperatures when the mean temperature increases. Source: IPCC AR4 WGI 2007; reprinted with permission.

Many of the trends in climate evident in the last century are expected to continue in the future. Projections shown in this report are based on the global climate model (GCM) projections done for the IPCC (Solomon et al. 2007), a coordinated large set of climate model runs known as the Coupled Model Intercomparison Project, Phase 3 (CMIP3), performed at modeling centers worldwide using 22 global climate models. Output of most of these models is at large resolution, often a 200-kilometer grid. Although it was common in past climate impact studies to present the results of only one or two global climate models, research now suggests that the average of multiple models provides a better approximation, and use of ensembles is made possible by increasing computing capacity and technical abilities for analyzing multiple model simulations (Salathé et al. 2010). The CMIP3 models and knowledge from these comparisons are the current state of the art in climate modeling and assessments.

9.1.1 GCM projections for North America

These global models project broad-scale increases in temperature in North America through the mid-21st century. Projected changes compared to a recent baseline (1950–1999 average) through mid-century (2040–2060 average) are shown in figure 34. For much of the interior western United States, the multi-model average projects an annual mean warming of about 2°C (4°F, in orange) by 2050. Individual global models also show a broad-scale pattern of warming, though of different magnitudes across models. The range of individual GCM projections (10th and 90th percentiles of the model projections) is from about +2.5F to +5.5°F (1.4–3.1°C). GCM projections show summers warming by about +5°F (range: 3–7°F, 2–4°C) and winters by about +3°F (range: 2–5°F, 1–3°C) (fig. 34, top row). The multi-model average and many individual global models show less warming within several hundred kilometers of the Pacific coast. This

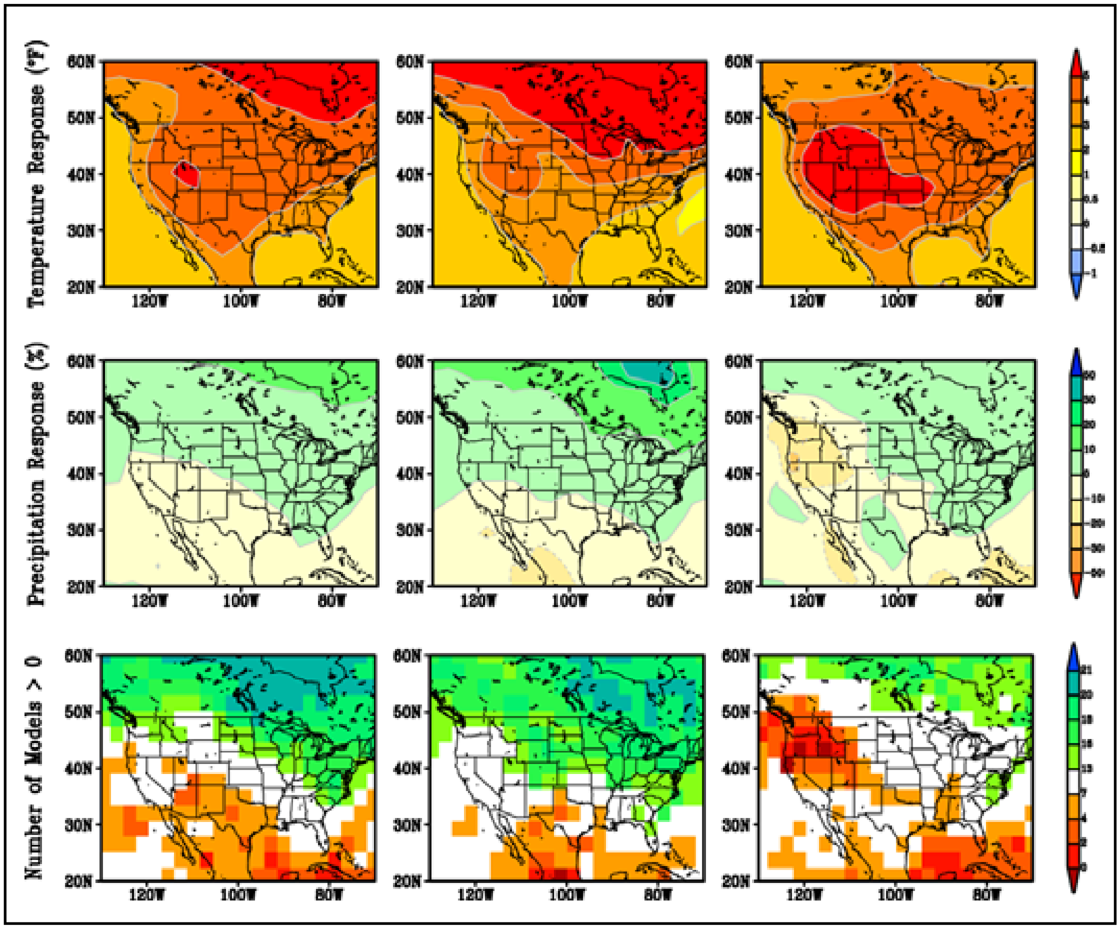


Figure 34. Temperature and precipitation changes in North America projected for 2050 (2040–60 average) by an ensemble of 22 climate models used in the IPCC AR4. Changes are shown relative to the 1950–1999 average. The top row is the multi-model average temperature change for the annual mean (left), winter (center), and summer (right). The second row shows the percentage change in total precipitation (data source: CMIP3 multi-model dataset, PCMDI). The bottom row shows multi-model agreement. Source: Ray et al. 2008; reprinted with permission.

feature may be a result of the inability of the global models to simulate the effects of the coastal mountain ranges and hence the moderating coastal influence penetrates too far inland. Regional climate modeling studies corroborate this (Salathé et al. 2010), showing large values of summertime warming much closer to the coast than for the global models.

Although temperatures are expected to increase across the landscape, natural short-term variation (i.e., years to decades) is still expected. One example would be the unusually cool Northern Hemisphere winter of 2009–10, which is likely linked to a natural climate variation called the North Atlantic Oscillation (NAO; Hurrell 1995), a large-scale circulation feature that can alternately block or allow cold air from the Arctic to enter the mid-latitudes of North America, Europe, and Asia. In 2009–10, the NAO was positioned to spawn cold winters as well as strong storms and heavy snowfalls. However, when considering the global average, the December 2009–February 2010 period still ranks as the 13th warmest in the last 131 years (http://www.ncdc.noaa.gov/sotc/). Rates of warming in the Northern Hemisphere have slowed somewhat in the last decade, even though the 9 of the 10 warmest years on record occurred during this period. This phenomenon has been largely attributed to a temporary decrease in the water vapor in the lower stratosphere which acts as a greenhouse gas (Solomon et al. 2010). Because of this natural short-term variability, some scientists prefer to report the averages of projections for 20 to 30 years, and this is done for the projections presented here.

Changes in the amount and spatial distribution of precipitation are still poorly understood and thus difficult to project (Solomon et al. 2007), especially against a background of substantial year-to-year variability. This is largely because precipitation is controlled by complex interactions between global and hemispheric-scale circulation features and ocean-and-land-surface atmospheric interactions that occur across a range of spatial and temporal scales. Moreover, the complex terrain of the study region will likely alter the impact of any broad-scale shifts in precipitation pattern. Predicting future precipitation is further complicated by the fact that human activities may also be altering natural controls (e.g., ENSO and PDO) on atmospheric circulation patterns and storm tracks (Barnett et al. 2008; Bonfils et al. 2008).

For total annual precipitation, the dominant pattern in North America projects a wetter climate in the northern tier and a drier climate in the southwestern United States (fig. 34, middle row), with small (≤10%) but important changes in annual precipitation in much of the four climate regions, although individual models (not shown) exhibit a range of projected changes. The importance of even modest changes relates to the timing of increased or decreased precipitation. Summer precipitation is expected to decrease for much of the western United States, causing increasingly dry warm-season conditions, whereas increased precipitation in the winter could result in dramatically higher volume streamflows in parts of the Upper Columbia Basin and northern U.S. Rocky Mountains. While models are in better agreement for projected increased winter precipitation for parts of the northern Rockies and the Upper Columbia Basin and decreases in the Southwest, overall uncertainty remains high (Solomon et al. 2007). There is only weak agreement among the models as to whether *annual* precipitation will increase or decrease (fig. 34, bottom row), but there is an indication of a *seasonal* decrease in summer precipitation for parts of the four climate regions, and an increase in winter precipitation (and more agreement among the models for the latter). In the central Rockies and GYA, model results vary widely, and it is unclear how conditions might change in coming decades. In addition, all of these areas feature pronounced natural variability at multiyear to multidecadal scales, and this natural variation may mask or enhance the regional expression of any broad-scale precipitation trends.

The CMIP3 projections, which document a broad spatial scale of warming, are at large resolution (e.g., 200-kilometer [124-mi] grid) that is of limited use for assessing impacts more locally. Hence, we present results from downscaled modeling efforts more relevant for regional or local planning: westwide at

4 kilometers, temperatures at specific sites driven by three emissions scenarios (B2, A1B, and A2) for three future periods, and projections of mid-century conditions for three climate and hydroclimate variables based on the A1B scenario (rapid economic growth, global population that peaks in mid-century and declines thereafter, and the rapid introduction of new and more efficient technologies).

9.1.2 Westwide climate: Statistically downscaled projections

For much of the West, GCMs project about a 2°C (4°F) rise in temperatures for 2050 (the orange shading in fig. 34, top row), with somewhat less warming near the Pacific coast. As part of a project for the U.S. Fish and Wildlife Service, NOAA used statistically downscaled projections to illustrate what the projected rise in temperatures would mean for the western regional climate compared to the existing north-south and elevational gradients of climate in the West (Ray et al. 2010). Downscaled temperature data from the CMIP3 22 model average projection for the A1B emissions scenario (from IPCC AR4) were added to the PRISM climatology (~4 km, 2 mi) for the June–August season. This downscaling method makes minimal, physically based corrections to the global simulation while preserving much of the statistics of interannual variability in the climate model (described by Salathé 2005). This method is similar to the so-called "delta method," in which the temperature changes (the "deltas") from GCMs are spatially interpolated and added to a high-resolution climatology.

The maps depict average daily temperature for the northern and central U.S. Rocky Mountains and Greater Yellowstone Area (fig. 35) and the southern Rockies (fig. 36) for the 1950–1999 climatology and projections for 20-year averages around 2025, 2050, 2090. These graphics illustrate that at large spatial scales, by 2050 the projected changes in summer climate can be visualized as a shift of temperature zones northward and upward in elevation (3rd panel in each figure). This shift of temperature zones continues through the end of the 21st century

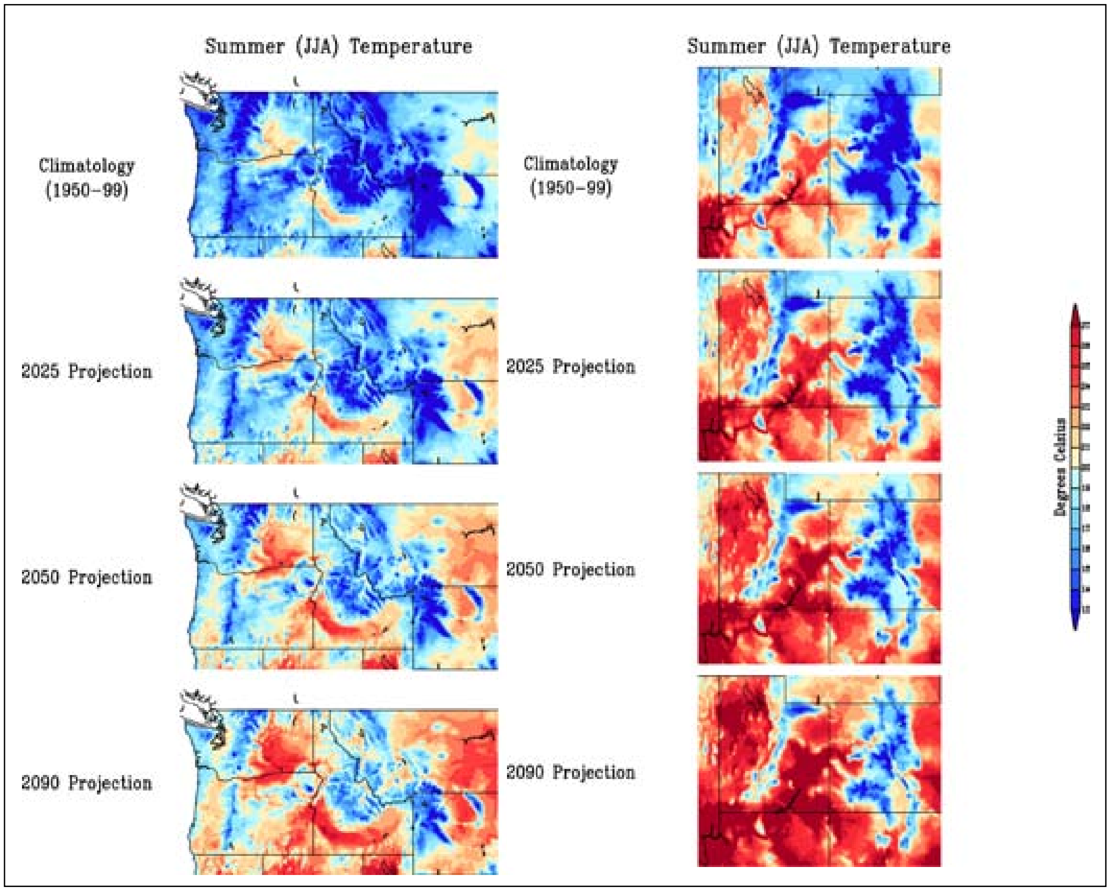


Figure 35. Summer observed average temperatures and statistically downscaled projections for the northern and central U.S. Rockies and Greater Yellowstone Area (left) and Southern Rockies (right). Observed average June–August temperature for 1950–1999 (top panel). Projections were calculated by adding the multi-model average temperature changes to the 4-km PRISM climatology. Observed climatological averages are from PRISM (DiLuzio et al. 2008), projected changes from the IPCC (CMIP3) 22-model average for the A1B emissions scenario. Source: Ray et al. 2010; used with permission.

(lower panel in each figure). These maps do not illustrate the year-to-year or day-to-day variability that will also occur. Furthermore, there are a number of unknowns about how climate effects may reduce or amplify the large-scale pattern of widespread warming that is projected over the western United States. It is unclear how the details will play out at any given location.

9.1.3 *Climate projections downscaled to specific alpine sites*

As part of the USFWS project, NOAA also generated temperature projections statistically downscaled to 22 mountain ranges in the western United States (Ray et al. 2010). Graphics for four sites not illustrated in that report are presented below (Glacier National Park and the Gallatin Mountains, Montana; Niwot Ridge, Colorado; and Clearwater Mountains, Idaho). This analysis illustrates implications of model-projected changes for the seasonal cycle, the relationship of projected climate change to historical climate variability, the spread of the individual model projections, and the evolving nature of the ensemble of projections throughout the century. This analysis used a modified version of the statistically

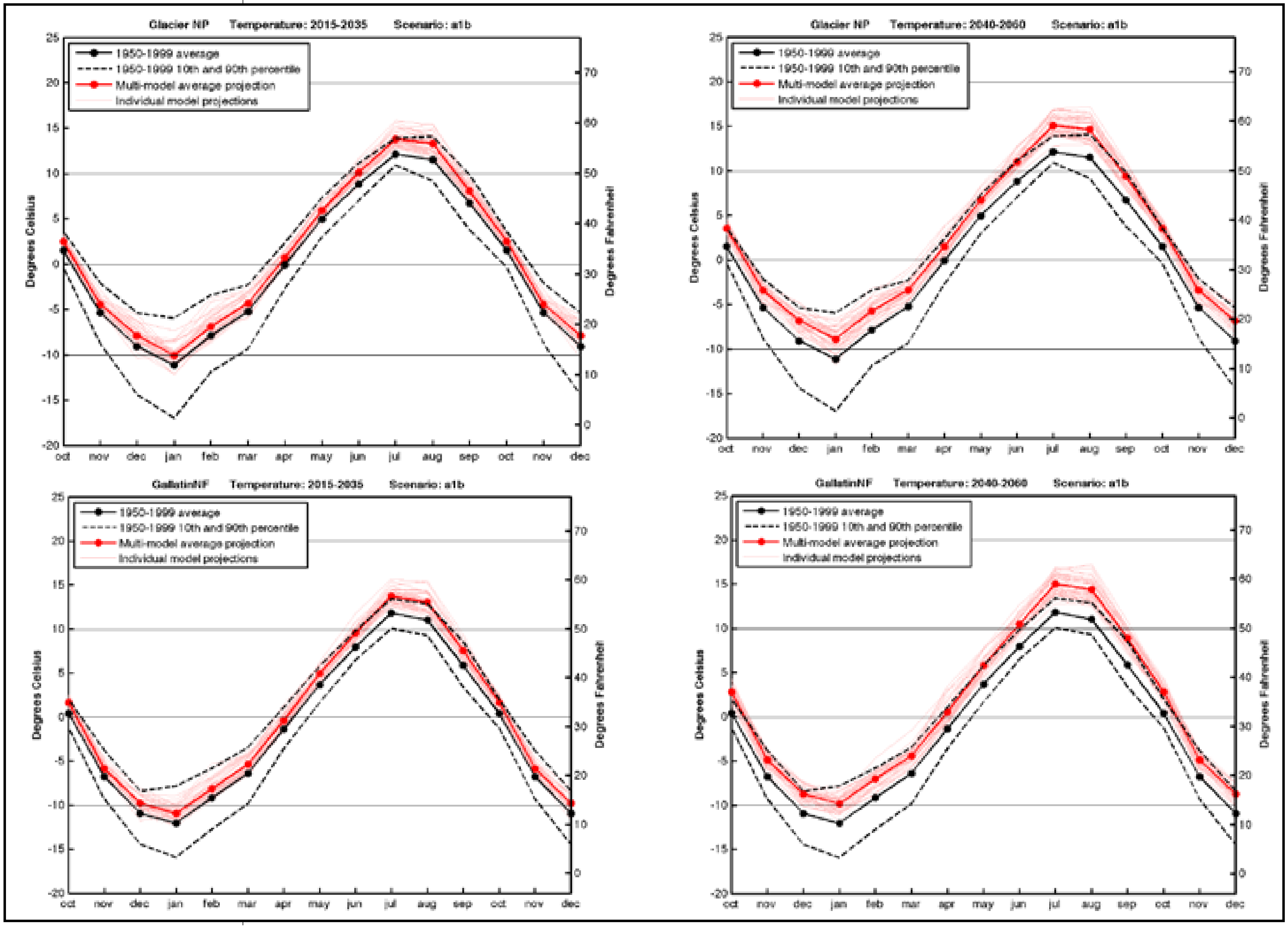


Figure 36. June–August 20-year temperature projectionscentered on 2025 (left panels), 2050 (right panels) for Glacier National Park (top panels, elevation 1866m) and the Gallatin Mountains, Montana (elevation 2778m, bottom panels) for a 4-km grid cell (approximately 30 x 40 mile). Each graphic compares observed monthly average temperatures to projections for the period. The observed monthly averages (solid black) and 10th and 90th percentiles values (dashed black lines) are based on observations during 1950–1999. Projected monthly climatologies (thin red lines) are from the multi-model ensemble for the 20-year period centered on 2050. The average of the projections is shown as a heavy red line. Data are derived from Maurer 2007. Note that the magnitude of projected temperature change is comparable to or greater than variations in the historical record. Source: Ray et al. 2010, used with permission.

downscaled CMIP3 Climate Projections created by the Department of the Interior Bureau of Reclamation and the University of Santa Clara. The statistical downscaling technique is known as “bias corrected spatial disaggregation” (BCSD) and was originally developed for hydrologic impact studies (Wood et al. 2004; Maurer 2002). This dataset downscales the projections to a 1/8° (12-km, 7-mi) grid (see details in Ray et al. 2008), which we adapted to the 4-km (2-mi) PRISM climatology (DiLuzio et al. 2008) to be smaller in scale for ecological applications in mountainous regions.

The resulting estimates adjusted to PRISM are among the best inferences available for temperature at this scale, albeit representative of a 4-km (2-mi) average and based on interpolation from station observations that may be distant from the grid box (See Ray et al. 2008 p 30). The results are shown in Figures 36–37. Projected temperatures from the BCSD/PRISM downscaling for the A1B emissions scenario are shown in red (thin lines) with the average projection (heavy line). (A1B scenario = rapid economic growth, global population that peaks in mid-century and declines thereafter, and

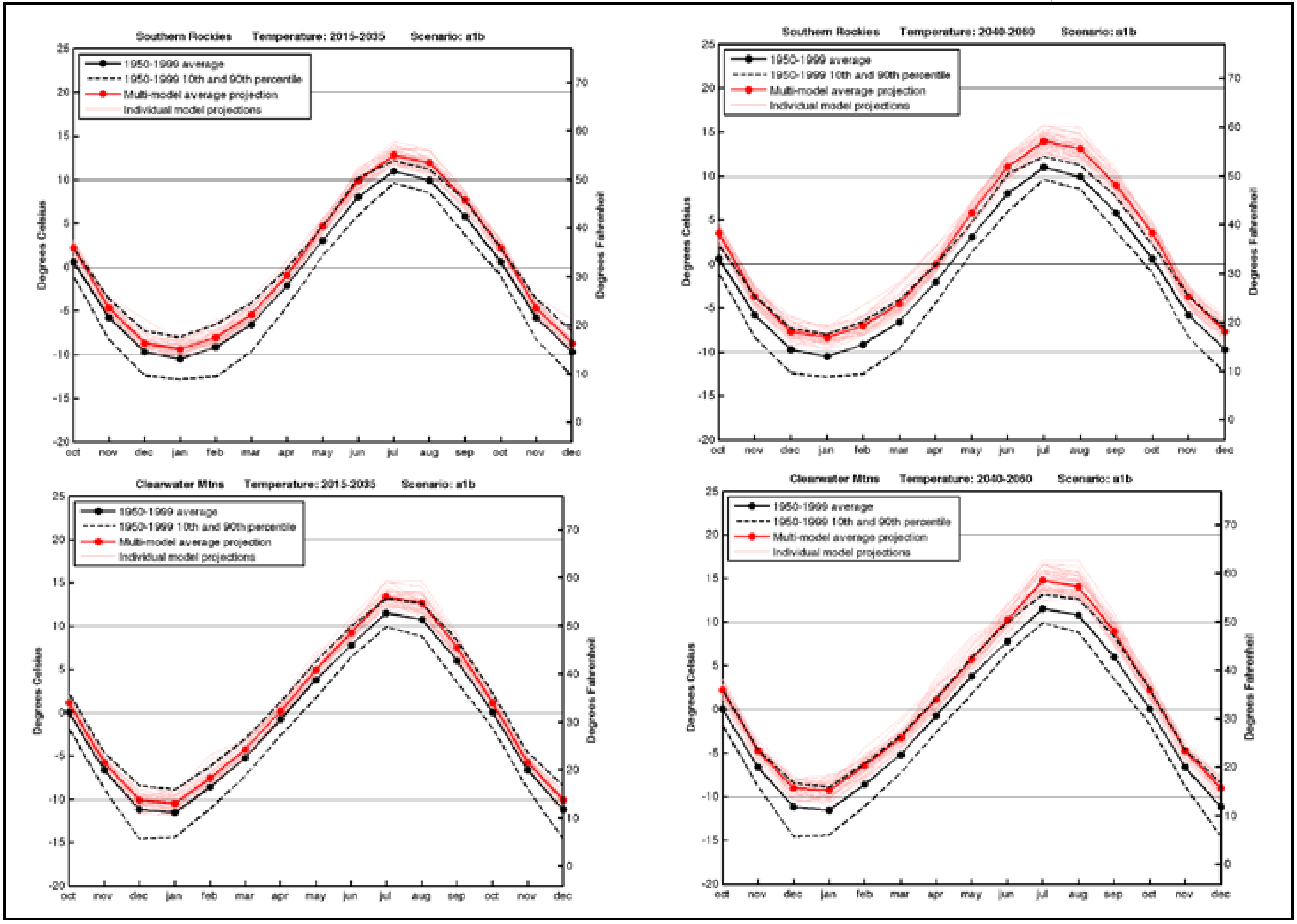


Figure 37. June–August 20-year temperature projections centered on 2025 (left panels), 2050 (right panels) for Niwot Ridge, Colorado (top panels, elevation 3267m) and Clearwater, Idaho (bottom panels, elevation 2467m) for a 4-km grid cell (approximately 30 x 40 mile). Each plot shows observed monthly average temperatures compared to projections for that period. The observed monthly averages (solid black) and 10th and 90th percentiles values (dashed black lines) are based on observations during 1950–1999. Projected monthly climatologies (thin red lines) are from the multi-model ensemble for the 20-year period centered on 2050. The average of the projections is shown as a heavy red line. Data are derived from Maurer 2007. Note that the magnitude of projected temperature change is comparable to or greater than variations in the historical record. Source: Ray et al. 2010, used with permission.

the rapid introduction of new and more efficient technologies.) For comparison purposes, the 1950–1999 PRISM climatology of the monthly average temperature (solid black line) and the 10th and 90th percentiles (dashed black lines) are also shown. These percentiles represent the five warmest and coolest months from 1950 to 1999.

At all four sites (as well as the other 16 not shown here), the temperature increases are largest in summer. The July temperatures from almost all the model projections at these sites lie at or above the 90th percentile of the present climate. Most of the projections suggest that typical summer temperatures will equal or exceed the extreme warm summers of the last half of the 20th century. The projected temperature changes are somewhat smaller in winter and the year-to-year variations are larger. While the proportion of warm winter months is projected to increase, most years, even by 2050, will not be above the 90th percentile of the present climatology. Winter warming will be manifest in the relative absence of months colder than the current average and in the cumulative effects of consecutive warm winters, with an increase in the number of extreme warm winter months.

The spread of the light red lines in figures 36–37 indicate the range of the individual models and the uncertainty in the projection of 20-year average climates, even for a given emissions scenario. High-end projections are approximately 1°C warmer than the multi-model average, and would indicate increased risk at a number of sites. However, the uncertainty implied by this spread may be larger than the true uncertainty due to differences in climate sensitivity among the models studied. The spread about the average projection is a result of two factors: differences in model climate sensitivity (the response of a particular model to climate forcing) and model-simulated multidecadal variability. That is why many scientists prefer to emphasize the multi-model average projection. Because the BCSD/ PRISM downscaling method is based on the CMIP3 projections, the multi-model average projections shown in these figures are consistent with the large-scale patterns of warming in the GCM temperature change maps (see fig. 34, top panel).

The overall pattern that emerges is for hotter summers and somewhat warmer winters. The 2050 summer projection is consistently about 3°C (5°F) higher than in recent climatology, which is the westwide projected increase. The low model projection (the 10th percentile of the distribution) is in most cases higher than the 90th percentile of the recent climatology, suggesting that the coolest summers of the mid-21st century will be warmer than the warmest summers of the recent past. Precipitation projections are not provided, but a recent similar downscaling effort for Colorado found that, unlike temperature projections, potential changes in precipitation are smaller than the year-to-year and decade-to-decade variations observed in the historical record (Ray et al. 2008).

9.1.4 Model projections of future climatic and hydrologic conditions

As part of a larger U.S. Forest Service and U.S. Fish and Wildlife Service project evaluating future conditions, the Climate Impacts Group (CIG, University of Washington) is modeling future climatic and hydrologic conditions (e.g., SWE, soil moisture, and evapotranspiration) for much of the western United States (Littell et al. in press). Preliminary results for a few key variables from this study are shown in figure 38.

9.1.5 Downscaled model methodology

The CIG project applied a range of climate change projections from the WCRP CMIP3 multi-model dataset used for IPCC AR4 (Solomon et al. 2007) to hydrologic model simulations and evaluated the impact of climate change on the hydrology of the region (after Elsner et al., in review). These models were drawn from a common set of simulations of 21st century climate archived from 21 GCMs (Mote and Salathé 2010), using greenhouse gas emissions scenarios as summarized in the IPCC's Special Report on Emissions Scenarios (SRES) (Nakićenović and Swart 2000). CMIP3 simulations were archived predominantly for three SRES emissions scenarios (A1B, B1, and A2) for

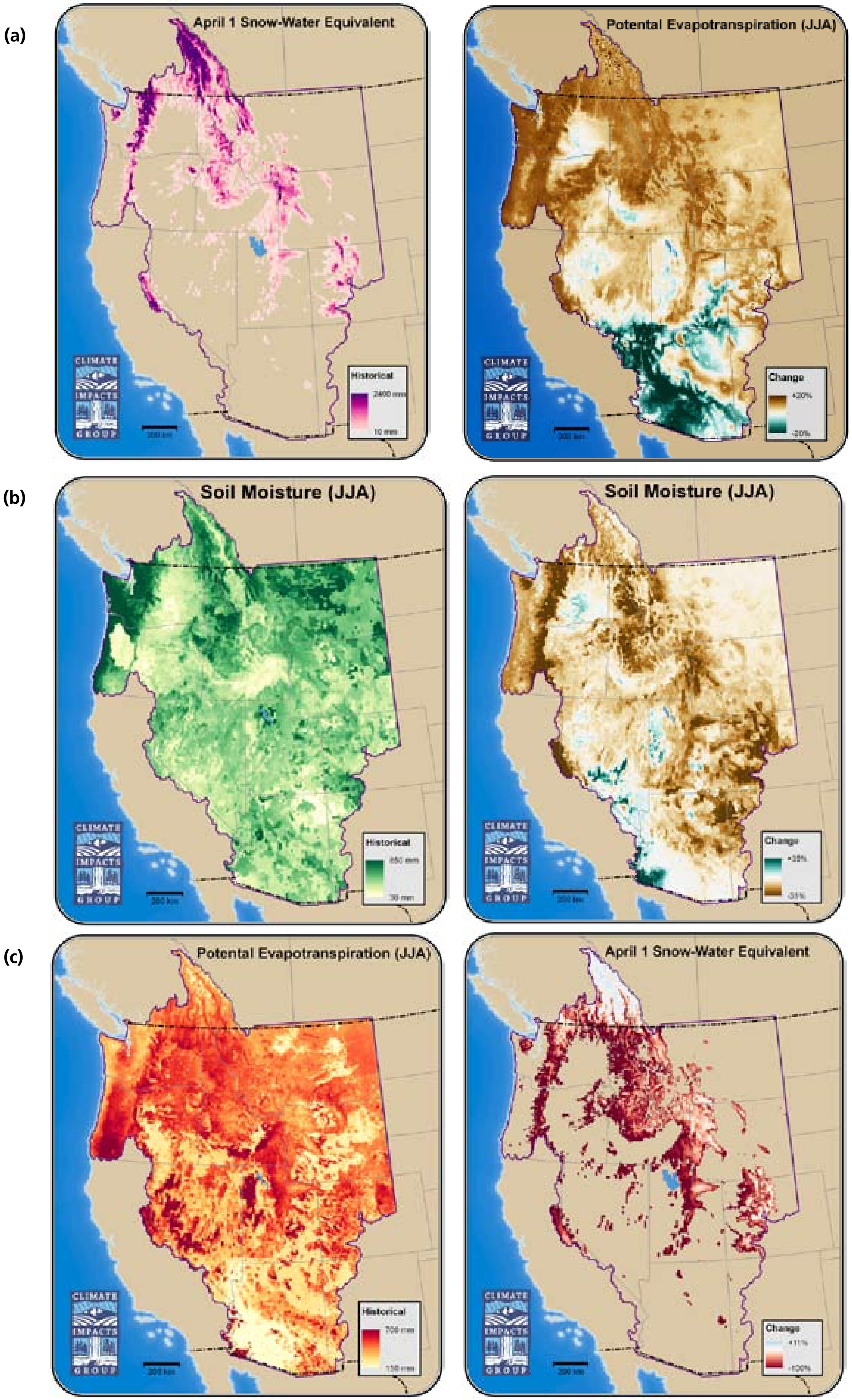

(a)
April 1 Snow-Water Equivalent
Historical
2400 mm
10 mm
200 km
Potental Evapotranspiration (JJA)
Change
+20%
-20%
200 km
(b)
Soil Moisture (JJA)
Historical
850 mm
30 mm
200 km
Soil Moisture (JJA)
Change
+35%
-35%
200 km
(c)
Potental Evapotranspiration (JJA)
Historical
700 mm
150 mm
200 km
April 1 Snow-Water Equivalent
Change
+11%
-100%
200 km

Figure 38. Estimate of (a) mean SWE, (b) mean soil moisture (June–August), and (c) mean potential evapotranspiration (June–August) for 1916–2006 and 2030–2059. Data derived from an ensemble of 8 GCMs that perform the best over the four climate regions. The temperature and precipitation data from the GCMs is used to drive the VIC hydrologic model, and this gives the SWE values. Great Basin and lower Colorado are modeled at 1/8° (~12 km), the rest at 1/16° (~6 km). Source: University of Washington Climate Impacts Group (USFS, and USFWS, in prep.); used with permission.

most of the 21 GCMs, with A2 following the highest trajectory (most warming) for future CO^2 emissions at the end of the 21st century. This work focuses on A1B (moderate warming) because it was simulated by the most GCMs, and our study focuses on mid-21st century change, at which point none of the scenarios is consistently the highest and for which a larger source of uncertainty is the variability in GCM models. We chose to use the eight GCMs for this study that have the best fit to the observed seasonal cycle of climate as well as the lowest bias for the observed precipitation and temperature records in all three modeled basins: Columbia, Missouri, and Colorado.

The spatial resolution of GCM output is generally too coarse to be meaningful for hydrological studies. Therefore, we downscaled the GCM output to 1/16° (~6 km, 4 mi) spatial resolution and applied a delta method approach to develop an ensemble based on the average of the eight scenarios (see e.g., Hamlet and Lettenmaier 1999; Snover et al. 2003). In the delta method, projected changes in precipitation and temperature as determined by GCM simulations are applied to the historical record at the resolution of hydrologic models. We performed hydrologic simulations using the historical record perturbed by these monthly changes in the Variable Infiltration Capacity (VIC) model (Liang et al. 1994; Nijssen et al. 1997) at 1/16° latitude by longitude spatial resolution over the entire region. The VIC model is a macroscale model intended for application to relatively large areas, typically from 10,000 km^2 (3,861 mi) to continental and even global scales. A key underlying model assumption is that sub-grid scale variability (in vegetation, topography, soil properties, etc.) can be modeled rather than represented explicitly.

10.1.1 Climate conditions

- Increasing temperatures (figs. 34–37).
- Increased but highly variable precipitation for parts of the Upper Columbia Basin and the northern and central U.S. Rocky Mountains (figs. 34, 38).
- Highly variable annual precipitation for parts of the central and southern U.S. Rocky Mountains, with possible decreases; however, recent changes in most regions are not statistically significant (figs. 34, 38).
- Increased evapotranspiration for most of the western United States which is unlikely to be offset by increased precipitation (figs. 31, 38; Hoerling and Eischeid 2007; Seager et al. 2007).

10.1.2 Surface hydrology

- Larger proportion of winter precipitation falling as rain rather than snow (Knowles et al. 2006; Bales et al. 2006).
- Decreased snow season length at most elevations (Bales et al. 2006).
- Less spring snowpack (fig. 38; Pederson et al. submitted; Mote 2006, 2003; Mote et al. 2005).
- Earlier snowmelt runoff (Stewart et al. 2005, 2004; Hamlet et al. 2005; Clow 2007).
- Increased frequency of droughts and low summer flows (Gray and Andersen 2010; Meko et al. 2007).

10.1.3 Extreme conditions: droughts, floods, heat waves

- More episodes of extreme temperatures (Parson 2001; Karl et al. 2009).
- Increased frequency of extreme precipitation events, rain-on-snow events, and consequent winter and spring floods in mountains (Madsen and Figdor 2007; Groisman et al. 2005; Kunkel et al. 2003).
- More frequent dry periods as a result of increased temperatures, evapotranspiration, and changes to surface hydrology (Gray and Andersen 2010; Meko et al. 2007).

10.1.4 Productivity and phenology

- Earlier blooming dates for many plant species (by as much as two weeks; Cayan et al. 2001; Schwartz and Reiter 2000).
- Longer growing season and increased productivity where moisture/soil fertility and other factors are not limiting (Bales et al. 2006).

10.1.5 Disturbance

- Higher frequency of large fires, longer fire seasons, and increased area burned by wildfires in the western United States (Westerling et al. 2006; Morgan et al. 2008; Littell et al. 2008, 2009a; Spracklen et al. 2009; Higuera et al. 2010).
- Greater drought stress will likely result in more insect infestations and disease affecting forests (Black et al. 2010; Nordhaus 2009; Romme et al. 2006; Logan et al. 2003; Carroll et al. 2004; Breshears et al. 2005).

Planning for future conditions that are highly uncertain presents a significant challenge for land managers. However, techniques developed for business, finance, and military applications offer a roadmap for planning in the face of large uncertainties. "Scenario planning" is one such approach, and it uses a combination of scientific input, expert opinion, and forecast data to develop alternative scenarios for the future (Schwartz 1991; van der Heijden 1996). This contrasts with more traditional attempts at developing precise, quantitative assessments of future conditions, which are often of limited value for understanding climate change because of compounded uncertainties. In scenario planning, a suite of alternative scenarios can be used as a starting point for exploring species or ecosystem vulnerabilities under a wide range of possible future conditions, and as a means for examining how management strategies might address multiple drivers of change.

Jackson et al. (2009a) developed an example to illustrate this process in which alternative futures are arrayed along two axes: integrators of potential climate change (drought frequency) and potential changes in disturbance regimes (fire size). In concert with monitoring and modeling, studies of past climates can define the range of drought frequency we might reasonably expect, and studies of fire history can place bounds on potential fire size. This exercise yields four quadrants, each comprising a distinct combination of climatic and fire-regime change (fig. 39). Each quadrant provides a contrasting scenario or "storyline" for exploring potential impacts on species or ecosystems and examining the relative costs and benefits of various mitigation and adaptation measures.

At one extreme, major climate change and altered disturbance regimes interact to drive emergence of novel ecosystems. Given limited experience with ecosystem turnover in many of the climate regions, consideration of long-term paleoenvironmental records serves as a primary means for adding context to scenarios. It also helps to determine the likelihood of any of the four quadrants. For example, transition to "novel ecosystems" is analogous to the transition evident 11,000 years ago in Yellowstone when open tundra vegetation was replaced by closed forests (Millspaugh et al. 2000), while the transitions to "inevitable surprises" are analogous to the late Holocene changes in fire regimes (Whitlock et al. 2003; Romme and Despain 1989). The greatest value in scenario planning comes from uncovering vulnerabilities and potential responses, particularly those common to the range of conditions characterizing a set of scenarios. Hence, despite high levels of uncertainty, scenario planning

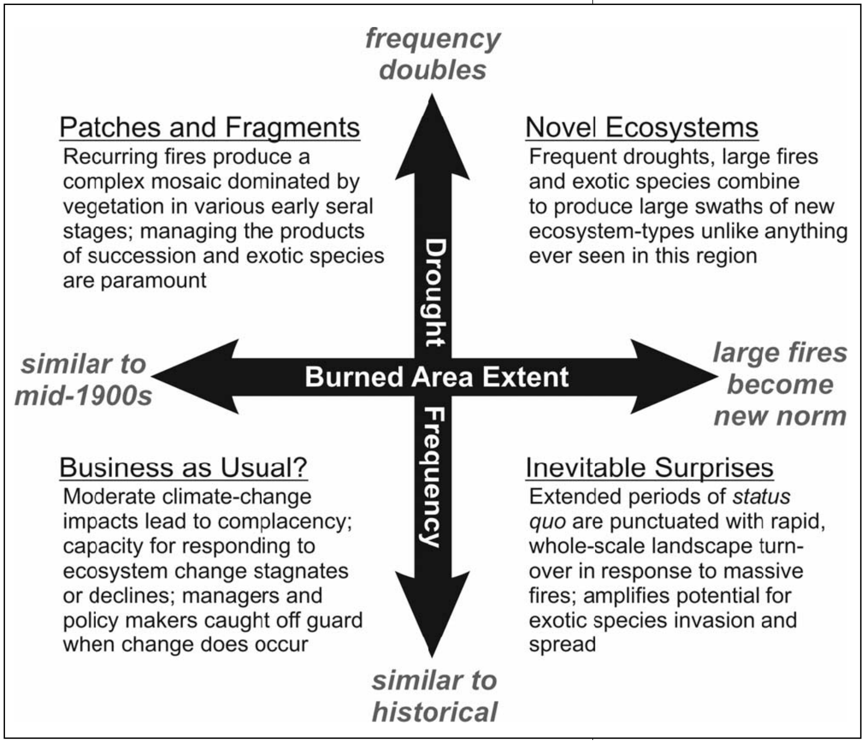

frequency doubles
Patches and Fragments
Recurring fires produce a complex mosaic dominated by vegetation in various early seral stages; managing the products of succession and exotic species are paramount
Novel Ecosystems
Frequent droughts, large fires and exotic species combine to produce large swaths of new ecosystem-types unlike anything ever seen in this region
Drought Frequency
similar to mid-1900s
Burned Area Extent
large fires become new norm
Business as Usual?
Moderate climate-change impacts lead to complacency; capacity for responding to ecosystem change stagnates or declines; managers and policy makers caught off guard when change does occur
Inevitable Surprises
Extended periods of *status quo* are punctuated with rapid, whole-scale landscape turnover in response to massive fires; amplifies potential for exotic species invasion and spread
similar to historical

Figure 39. Example of a scenario planning matrix. Each axis represents a critical driver of system change or a significant trend in the environment. In common practice, the variables chosen for analysis are likely to have the strongest influence on the system or they are associated with a high degree of uncertainty (Shoemaker 1995). In the case presented here, the axes represent a continuum between conditions that are similar to those observed in the historical record and conditions that are significantly altered from those seen today. Combining these two drivers produces four alternative scenarios for future conditions (e.g., frequent drought and large fires in the upper right) that can then be further developed into storylines that provide details about how each scenario might unfold. Depending on the application and available data, the axes and the resulting storylines may be defined quantitatively or based on qualitative assessments alone. Source: Jackson et al. 2009a; reprinted with permission.

may reveal that management response may be similar for a wide range of possible outcomes. Managers can then move beyond the challenges presented by an uncertain future to identify management responses that address ecological responses to a range of climatic conditions.

Chapter 12: Summary conclusions

The influence of human activities on the climate system, primarily through increased greenhouse gases and aerosol concentrations, will be superimposed on natural drivers of climatic change. As records of the past and model projections for the future suggest, our certainty about the patterns of climate change varies widely. Temperatures will most likely increase over large spatiotemporal scales, but the patterns and direction of change in other climatic variables are less clear. Investigating past change also instructs us that, at smaller spatio-temporal scales, changes in future climatic conditions will likely vary greatly, especially in heterogeneous mountain environments characterized by steep biophysical gradients. Temperatures are predicted to increase in the ensuing decades, making much of the western United States vulnerable to increased frequency and duration of drought (Barnett et al. 2008; Seager et al. 2007; Diffenbaugh et al. 2005). Small increases in temperature (e.g., 1–2°C, 2–4°F) will require large increases in precipitation to offset increased evapotranspiration, especially in settings where increased temperatures significantly alter available moisture and surface-energy feedbacks (Hoerling and Eischeid 2007).

High levels of uncertainty about how ecosystems will respond to changing conditions should not prevent managers from planning for the future. Scenario planning can be an effective approach for considering a wide range of possible future conditions that oftentimes indicate similar management actions. In many ways, the outcomes of scenario planning reinforce fundamental principles of adaptive resource management. For example, management practices that focus on maintaining diversity, increasing connectivity, providing buffer habitat around protected areas, and mediating human impacts within and around public lands will also facilitate the dynamic and heterogeneous redistribution of vegetation that will likely occur across the West.

Paleoenvironmental records suggest that ecosystems are responsive to climatic change over millennial to decadal time and spatial scales, but they also suggest that we should anticipate novel vegetation assemblages as species respond individualistically to climate change and communities reorganize (Whitlock et al. 2003; MacDonald et al. 2008; Williams et al. 2007; Jackson et al. 2009b). These changes also suggest that natural disturbances, non-native species, disease, and unforeseen synergistic interactions will be associated during periods of ecological disequilibrium. Management actions that work to accommodate dynamic and difficult to predict redistributions of vegetation by protecting a full suite of biophysical and environmental gradients, increasing connectivity, and minimizing human impacts in landscapes surrounding protected areas may best protect valuable resource attributes in the face of climate change.

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