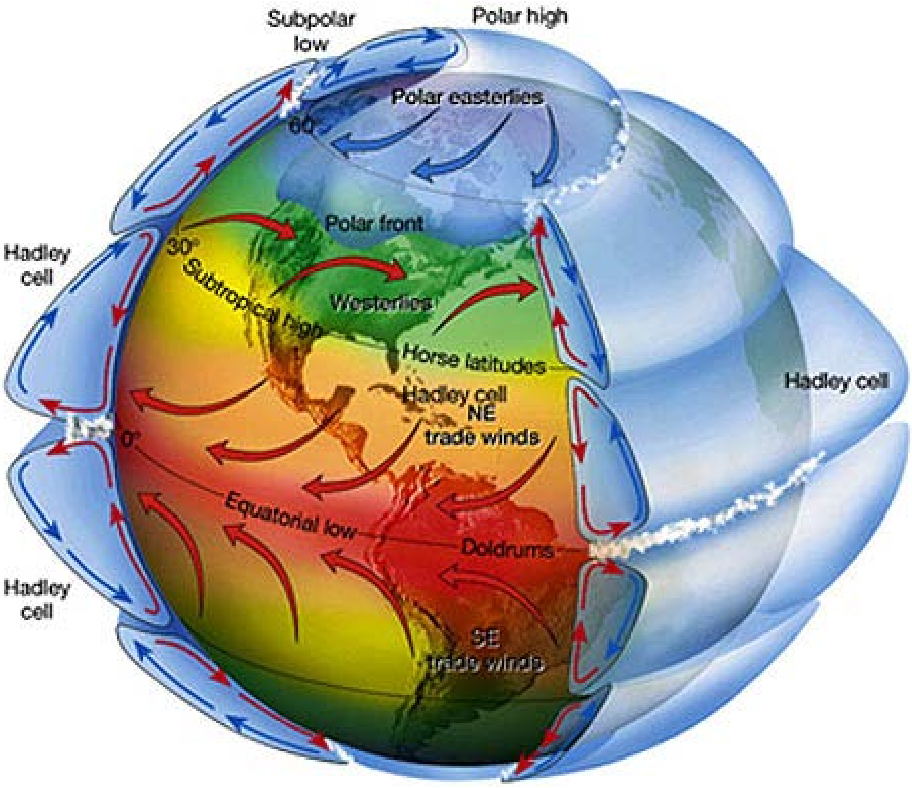
Appendix B
Literature Review of Observed and Projected Climate Changes



U.S. Department of the Interior
Bureau of Reclamation
Upper Colorado Region
Albuquerque Area Office

December 2013

Mission Statements

The U.S. Department of the Interior protects America's natural resources and heritage, honors our cultures and tribal communities, and supplies the energy to power our future.

The mission of the Bureau of Reclamation is to manage, develop, and protect water and related resources in an environmentally and economically sound manner in the interest of the American public.

The U.S. Army Corps of Engineers Mission is to deliver vital public and military engineering services; partnering in peace and war to strengthen our Nation's security, energize the economy and reduce risks from disasters.

Sandia Laboratory Climate Security program works to understand and prepare the nation for the national security implications of climate change.

Cover graphic: Atmospheric circulation in the climate system.
Source: http://serc.carleton.edu/earthlabs/climate/5.html.

Acronyms and Abbreviations

°C	degrees Celsius
°F	degrees Fahrenheit
AMO	Altantic Multidecadal Oscillation
AOGCM	atmosphere-ocean general circulation model
cm	centimeters
CMIP3	Coupled Model Intercomparison Project Phase 3
COOP	Cooperative Observer Program
ENSO	El Niño-Southern Oscillation
HCN	Historical Climatology Network
HDe	Hybrid Delta-ensemble
IPCC	Intergovernmental Panel on Climate Change
NCAR	National Center for Atmospheric Research
NCDC	National Climate Data Center
NOAA	National Oceanic and Atmospheric Administration
NARCCAP	North American Regional Climate Change Assessment Program
PDO	Pacific Decadal Oscillation
PDSI	Palmer Drought Severity Index
PNA	Pacific North American
SRES	Special Report on Emissions Scenarios
SWE	snow water equivalent
T_{max}	maximum high temperature
T_{min}	minimum temperature
USGCRP	U.S. Global Change Research Program
VIC	Variable Infiltration Capacity
WWCRA	West-Wide Climate Risk Assessment

Table of Contents

Page

Tables

Figures

Figures (continued)

Literature Review of Observed and Projected Climatic Changes

This appendix provides a brief overview of the climate of the Upper Rio Grande basin, including its major features, drivers and sources of variation (this discussion is based on Sheppard et al. 2002). The impacts of increasing atmospheric greenhouse gases over the coming decades are anticipated to vary spatially as the warming is mediated by existing and evolving large-scale patterns of atmospheric circulation. The effects of regional and local factors, such as continentality (distance from a large water source), relief, and sea surface temperature patterns, will be superimposed on global scale changes to atmospheric circulation. This is particularly important for the Upper Rio Grande basin because it is located on the boundary between the subtropical dry and temperate mid-latitude climate zones. This boundary is anticipated to shift northward, and with this change, alter the seasonal precipitation patterns in the region.

I. Overview of the Climate of the Upper Rio Grande

The Upper Rio Grande basin is classified as an arid climate, with average annual precipitation in most areas < 15 inches (<38 centimeters [cm]) except in mountain regions. Precipitation is bi-seasonal, with the major peak in summer (July to September), a secondary peak in winter (November to March), and arid spells in spring (April to June) and fall (late September through early November).

Temperature and precipitation vary by latitude and elevation within the Upper Rio Grande (Kunkel et al. 2013b):

- Southern, lower-elevation areas south of Elephant Butte Dam have average annual temperatures of 61 to 65 degrees Fahrenheit (°F) (16 to 18 degrees Celsius [°C]), and receive less than 15 inches (38 cm) of precipitation annually.
- The Albuquerque portion of the Upper Rio Grande has an annual temperature of approximately 51 to 55°F (11 to 13°C) (Figure 1) and receives 11 to 15 inches (28 to 38 cm) of precipitation per year (Figure 2).
- In the San Luis Valley of Southern Colorado, the average annual temperature is 41 to 45°F (5 to 7°C) and precipitation averages <10 inches (25 cm) per year.

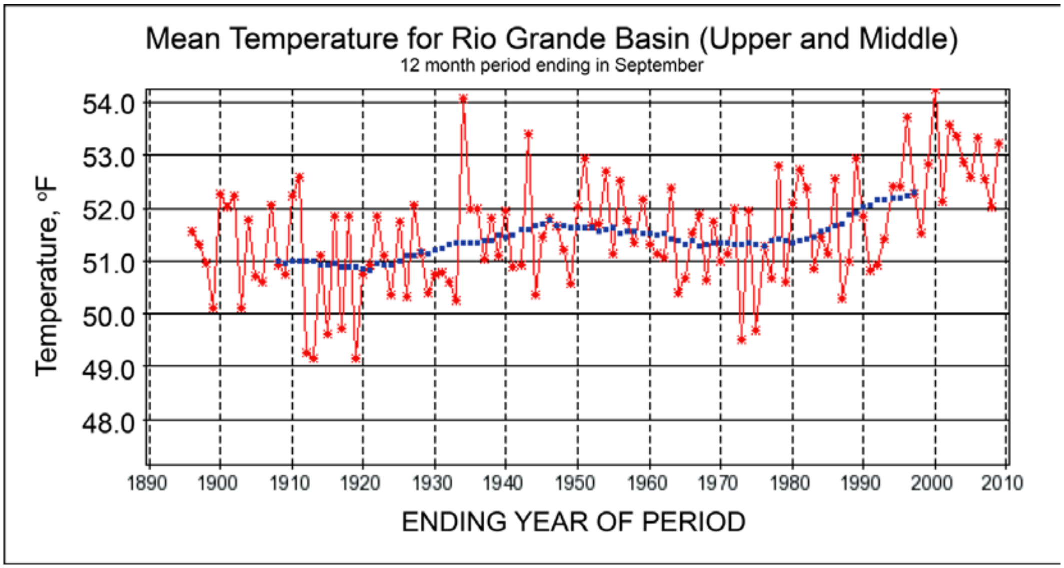


Figure 1.—Observed annual temperature, averaged over the Rio Grande Basin above Elephant Butte. Red line indicates annual time series for the given geographic region. Blue line is 25-year moving annual mean (University of Arizona et al., 2007).

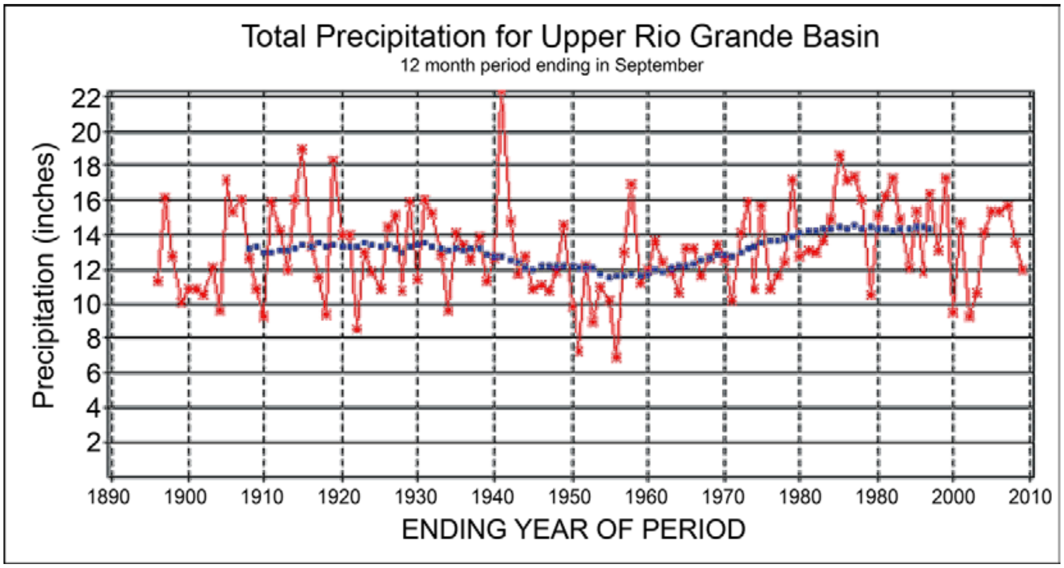


Figure 2.—Observed annual precipitation, averaged over the Rio Grande Basin above Elephant Butte. Red line indicates annual time series for the given geographic region. Blue line is 25-year moving annual mean (University of Arizona et al., 2007).

- In the adjacent San Juan Mountains of southern Colorado, average annual temperatures are as cool as 21 to 30°F (-6 to -1°C), with precipitation in the wettest areas exceeding 40 inches (100 cm) per year.

The basic pattern of New Mexico's climate is driven by its latitude and its position in the continental interior. Solar heating of Earth's surface along the equator causes humid air in this region to rise and to drop its moisture as rain in a band along the equator. A portion of this risen air moves poleward at high altitude, where it cools and eventually descends over the subtropics. As this air descends, it warms and its capacity to retain moisture increases, pulling moisture out of the environment as the air mass descends.[1] The descending dry air returns towards the equator. This convection system moving air between the equator and the subtropics is known as a "Hadley Cell." Most of the world's deserts are located at the descending arm of the Hadley Cell, including the Mohave, Sonoran, Chihuahuan, Sahara, Thar Deserts, and the deserts of Saudi Arabia in the Northern Hemisphere, the Atacama, Kalahari, and central Australian Deserts in the Southern Hemisphere.

The location of the Hadley Cell in the Northern Hemisphere shifts north in the summer and south in the winter due to the tilt of the Earth's axis. During summer months, the northern portion of the descending arm of the Hadley Cell encompasses northern New Mexico and southern Colorado, allowing hot, dry air to settle over the region from March through September. The aridity and heat are reduced in late summer/early fall due to the North American Monsoon, in which diurnal heating of the land surface pulls humid air in from the Gulf of Mexico (sometimes the southeastern Pacific). Heating of this air leads to daily convective storms producing intense, localized cloud-bursts. The location of these storms is strongly mediated by topography, with higher elevations tending to have more reliable monsoonal precipitation than lower, and latitude, with southeastern Arizona falling inside the core monsoon region and the Upper Rio Grande falling outside. Precipitation during the summer monsoon is characteristically more than 50 percent of the annual total in most portions of the Upper Rio Grande Basin. The North American Monsoon tapers off in fall as diurnal heating is reduced, although remnant Tropical Pacific cyclones can bring sustained precipitation to the region, especially in September.

With the onset of winter, the area of maximum heating shifts south of the equator, which causes the northern limit of Hadley Cell circulation to shift south of the study area and enables the jet stream to push mid-latitude cyclonic storms into the region. These storms precipitate rain and snow over wide areas and alternate with

---

[1] As a general rule of thumb, rising air cools and as it cools, the water it contains condenses and eventually precipitates out – so areas underneath rising air get rain. Descending air warms, and as it warms it can hold more moisture, so it becomes relatively drier. Areas underneath descending air do not get rain.

high pressure systems that bring dry, sunny weather to the region. However, the amount of precipitation from these systems is limited because the Upper Rio Grande is located in the interior of North America: it is surrounded by dry land and is distant from warm oceans. This limit is exacerbated by the region's location in the rainshadow of the Sierra Nevada mountains: much of the moisture coming off of the Pacific is wrung out of storm systems as they cross the Sierras, and is only added back in when these storms reach the Plains states and tap into humid air masses originating over the Gulf of Mexico. As a result, winter precipitation across most of the region is less than summer.

II. Variation in Winter Climate

Winter precipitation varies from year to year, depending primarily on the Pacific Ocean sea surface temperature. Areas of the ocean with warm sea surface temperatures add a great deal of heat (energy) and moisture to overlying air masses, creating larger storms with greater precipitation potential.[2] Areas with cool sea surface temperatures fail to heat the air much and produce small, weak storms with low or no precipitation potential. Ocean temperatures in areas that matter for Southwestern climate—eastern Pacific, Gulf of Mexico—vary in temperature from year to year, with direct consequences for climate in the Upper Rio Grande.

The most familiar variation in ocean temperature (and in the overlying atmosphere) is the El Niño-Southern Oscillation (ENSO) cycle. In a normal (ENSO-neutral) year, surface winds push warm equatorial Pacific surface waters to the west, creating a pool of warm water near Indonesia and allowing very cold, deep ocean water to rise to the surface in the eastern Pacific from northern Peru to Mexico. Over the warm pool, heat and moisture are contributed to the air, the warm air rises, and heavy precipitation occurs in the western Pacific. At the same time, the air over the eastern Pacific is comparatively cool and dry, and therefore the eastern Pacific and adjacent regions (such as the Southwest) are relatively cool and dry.

In an El Niño year, the warm pool "migrates" to the east, leaving Indonesia cooler and drier, and shutting off the upwelling of cold ocean water in the eastern Pacific. Although most precipitation occurs out to sea, there is a significant increase in atmospheric moisture in the eastern Pacific, which brings more winter precipitation to the Southwest. Winter 2009-2010 was an El Niño year, which brought an increased level of moisture to the Upper Rio Grande Basin.

---

[2] Warmer air can hold more moisture and warmer seas evaporate more moisture, so air masses over warm seas become warm and humid (e.g., over the Gulf of Mexico). Cooler seas have less heat energy to drive evaporation, so evaporation is less. In addition, cooler sea surfaces also contribute less heat to the overlying air. The result is that air masses over cool parts of the ocean tend to be cooler and drier.

ENSO has a third state known as La Niña. In a La Niña phase, the warm pool migrates to the west of its normal position, bringing additional rain to Indonesia and Australia while at the same time bringing hyper-dry conditions to the eastern Pacific. Winter 2010-2011, which was a La Niña winter, was exceptionally warm and arid in the Southwest.

The frequency of El Niño and La Niña events has increased since the 1970s. Before 1970, El Niño and La Niña events occurred in roughly equal frequencies, and were separated by several normal (ENSO-neutral) years. Since the late 1970s, the frequency of El Niño and La Niña events has increased, El Niño events have outnumbered La Niña events by 2:1, the number of "normal" years separating the two have decreased, and El Niño events have increased in strength. The reasons for these changes are poorly understood. They may relate to other large-scale climate phenomena, including long-cycle changes in sea surface temperatures in the north Pacific[3] which operate on multi-decadal (50 to 80 year) cycles, and which can serve to amplify or dampen the different phases of the ENSO cycle. Since the 1970s, Central Pacific El Niño events have become more common, in which the warm pool occurs in the central rather than eastern Pacific. During Central Pacific El Niño events, precipitation in the U.S. is reduced relative to Eastern Pacific El Niño events, leading to winter precipitation in the Upper Rio Grande that is at or only slightly above normal (Jin-Yi and Yuhao 2013). Since 1990, five of the last seven El Niño events have been Central Pacific El Niño events.

ENSO effects on precipitation in the Southwest are primarily a winter phenomenon, and summers are usually characterized by ENSO-neutral or transition states. National Oceanic and Atmospheric Administration (NOAA) maintains a regularly updated discussion of current ENSO status, near-term (about 6 months) ENSO projections, and implications for how changes in ENSO will affect temperature and precipitation across North America (National Weather Service (NWS) 2011).[4]

The strength of El Niño and La Niña are also affected by the interplay of long- and short-term climate cycles. Long-term wet and dry cycles in the Southwest are controlled primarily by Pacific sea surface temperatures (SST), particularly the multi-decadal Pacific Decadal Oscillation (PDO), and Atlantic sea surface temperatures via the Altantic Multidecadal Oscillation (AMO). The phase of the PDO in particular acts to amplify and dampen portions of the ENSO cycle. The negative (cool) phase of the PDO enhances La Niña effects and dampens the

---

[3] These cycles are known as the Pacific Decadal Oscillation (PDO). ENSO cycles are also affected by multi-decadal, cyclical sea surface changes in the Atlantic (Atlantic Decadal Oscillation and others). All of these long-term climate cycles, and the effects of their interactions, are still poorly understood.

[4] This discussion can be found at NOAA's National Weather Service Climate Prediction Center, online at <http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/enso_advisory/>.

increase in precipitation during El Niño events, while the reverse is true under during positive PDO cycles. The PDO has been in a negative phase since May 2010 (Mantua 2013). Historically, the driest periods in the Southwest were associated with cool Pacific sea surface temperatures (negative PDO) and warm Atlantic sea surface temperatures (positive AMO) (McCabe et al. 2004).

III. Variation in Summer Climate

The North American Monsoon is driven by daytime heating of the land surface that, in turn, warms the lower atmosphere leading to atmospheric convection. The rising air cools and, if moisture is present, can lead to precipitation. The monsoon is initiated in mid-summer when surface heating is strong enough over a large enough area to draw in moisture from the Gulf of Mexico and, secondarily, the eastern Pacific/ Gulf of California. The monsoon onset is time-transgressive, beginning mid-June in areas in the southern part of the Southwest and in mid-July in areas in the north.

Monsoon strength increases with elevation, in direct proportion to the amount of increase in daytime air mass rise. All things being equal, higher elevation areas will receive greater—and more consistent—monsoonal precipitation, with many high mountain areas experiencing daily downpours. Lower elevation areas will tend to see less midday precipitation but more evening precipitation, and there will be greater day-to-day and place-to-place variation in precipitation.

The strength of the monsoon varies greatly from year to year for reasons that are not well understood. The strength of the monsoon appears to depend on:

- How hot the Southwest gets (i.e., how much heat is available to drive air convection)
- How warm the sea surface temperatures are the eastern Pacific and Gulf of Mexico, which serve as the principal sources for moist air and therefore determine the amount of moisture in air masses being pulled into the Southwest
- How active the cyclone/hurricane season is in the eastern Pacific and Gulf of Mexico, which can push tremendous amounts of moisture into the Southwest during the late summer and early fall.

Monsoon strength is also affected by sea surface temperatures at the hemispheric scale that govern large-scale movements of air masses at different latitudes. The specific controls on interannual variations in monsoon strength are not well understood.

Monsoon precipitation is typically intense but localized, and rarely has a uniform effect across a large drainage basin area, such as the Upper Rio Grande Basin. However, precipitation can be more widespread if the monsoon is able to tap moisture from a tropical cyclone in the moisture source regions.

IV. Literature Review: Observed and Projected Temperature Change

Recent overviews of climate change in the Southwestern United States have been provided in Garfin (2013), U.S. Global Change Research Program (USGCRP) (2013), and NOAA (2013a). Important syntheses of climate change impacts to New Mexico and Colorado include New Mexico Office of the State Engineer (2006) and Ray et al. (2008).

IV.A. Global, National, and Western U.S. Temperature Trends

Temperatures in the Intermountain West have shown a relatively steady rise beginning in the early 20th century. The rise stalled during the middle part of the century during the post-war economic boom as increasing atmospheric pollution reduced the amount of sunlight entering the lower atmosphere, and then continued to rise following implementation of laws regulating environmental and atmospheric pollution.

Globally, 2010 was tied with 2005 as the warmest year on record, continuing a trend of 34 consecutive years during which average global surface temperatures remained above the 20th century average. The 2010 global land surface temperature average was 1.7°F (0.96°C) above the 20th century mean (NOAA 2011a). The first 10 years of this century constitute 10 of the 11 warmest years in the historical record, and may be warmer than it has been for millennia. In this decade, there were four wet El Niño cycles and three dry La Niña cycles (NOAA 2011b).

Warming has continued, with 2012 constituting the warmest year on record for the contiguous United States (National Climate Data Center [NCDC] 2013). The average temperature was 55.3°F (12.9°C), which was 3.2°F (1.8°C) above the 20th century average (NCDC 2013).

The consensus view is that recent increases in temperature in the Western U.S. exceed observations in the historic record beginning in the late 19th century (USGCRP 2009). In the mountainous West, average annual temperatures for 2001 through 2009 were 1.4°F (0.8°C) higher relative to the average for 1895 through 2000 (MacDonald 2010). Temperature increases were greater in areas to the south and at lower elevation.

Particularly troubling have been increases in winter (January, February, and March) temperatures throughout the mountainous West. The observational record of 1950 through 1999 shows an increase in maximum average winter temperatures of 2.8°F (1.53°C) and an increase in minimum average winter temperatures of 3°F (1.72°C) (Bonfils et al. 2008). Rising winter temperatures have contributed to a contraction of 8 days in the number of days below freezing, and a corresponding lengthening of the frost-free period. Detection and attribution modeling studies indicate that these patterns cannot be replicated in models of natural climate forcing (i.e., models that exclude human greenhouse gas emissions but include the effects of ENSO, Pacific Decadal Oscillation, solar variation, and changes in volcanic aerosol concentrations), but these patterns are robustly replicated in models that also include human greenhouse gas emissions (Bonfils et al. 2008).

IV.B. Southwestern U.S. and Upper Rio Grande Temperature Trends

In the Southwestern U.S. as a whole, encompassing New Mexico, Colorado, Arizona, Utah, Nevada, and California, the decade 2001 through 2010 was the warmest of all decades from 1901 through 2010, with temperatures increasing by approximately 1.6 ± 0.5°F (0.9 ± 0.3°C) over the period 1901 through 2010 (Hoerling et al. 2013). Rising temperatures increased the frequency of heat waves, reduced the frequency of cold waves, and contributed to the expansion of the growing season by 17 days (7%) during 2001 through 2010 compared to the average season length for the 20th century. The period since 1950 in the Southwest has been warmer than any comparable period in at least 600 years, according to paleoclimate records (Hoerling et al. 2013).

At the regional level, several recent studies have examined trends in temperature. Tebaldi and colleagues (2012) use low elevation National Weather Service Cooperative Observer Program (COOP) station data and corrected climate data from the NOAA Historical Climatology Network (HCN) to estimate that average annual temperatures in Colorado rose at a rate of 0.225°F (0.13°C) per decade over the period 1912 to 2011, but rose at the faster rate of 0.483°F (0.27°C) per decade since 1970. The same study shows New Mexico warmed at an average rate of 0.219°F (0.10°C) per decade from 1912 to 2011 but at the faster rate of 0.678°F (0.34°C) per decade since 1970. The same pattern of faster recent warming was also observed in annual average daytime maximum high temperature (T_{max}) and annual average nighttime minimum temperature (T_{min}). In Upper Rio Grande, the increase in average annual temperatures from 2001 through 2009 was 1.5 to

2 standard deviations above the 20th century average in the Upper Rio Grande valley in New Mexico, and between 1 and 1.5 standard deviations above in the Colorado portion of the Upper Rio Grande (see Figure 1, MacDonald 2010).

Enquist and Gori (2008) examine temperature trends as part of a study of changes in habitat and species vulnerability in wilderness areas under a warming climate. They find that over the period 1970 through 2006, the average rate of temperature increase in wilderness areas in Northern New Mexico was 0.684°F/decade (0.36°C /decade), with T_{min} increasing on average at a rate of 0.684 °F /decade (0.38°C /decade), approximately 0.072°F /decade faster (0.04°C/decade) than T_{max}.

In the Upper Rio Grande Basin, a comparison of average monthly temperatures over the 1995 through 2004 period with average monthly temperatures for the period 1961through 2000 (Figure 3) showed increases of 3 to 4°F (1.5 to 2.5°C) in winter, with increases in the April through November period less than approximately 2.0 F (1.1°C) in all months but May (Saunders and Maxwell 2005).

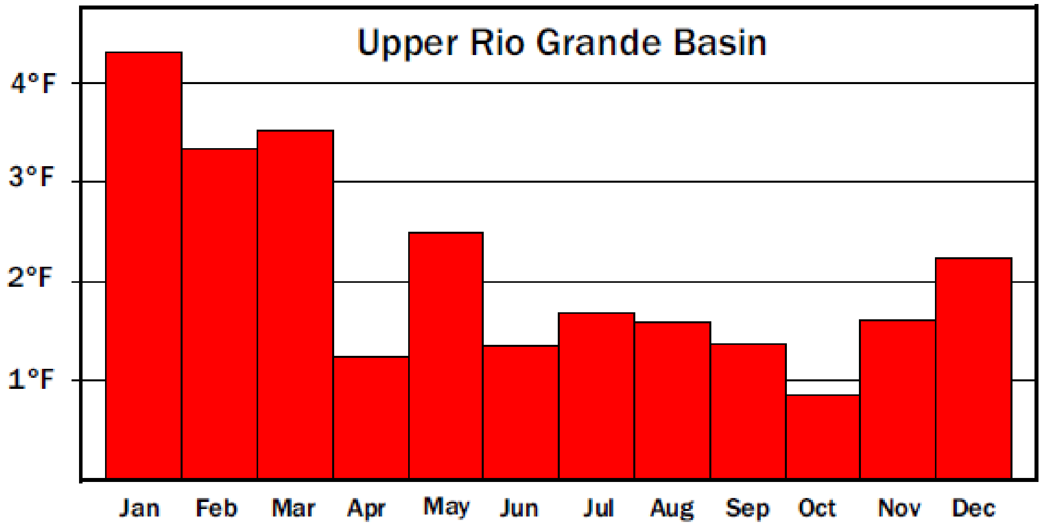


Figure 3.—Average monthly temperature change in the Upper Rio Grande Basin, showing that warming is greatest in the winter months.
(Source: Saunders and Maxwell 2005).

Elevation plays an important role in determining the season of greatest warming in the mountains:

- Higher elevation sites experienced their greatest warming during the summer months, with temperatures increasing at a rate of 1.5°C (2.7°F) per decade during this season. Rates of warming in high elevation areas may be considerably greater than the regional average. In a recent analysis of National Weather Service and SNOTEL site data in the San Juan Mountains, Rangwala and Miller (2010) detected a rate of warming of1.8°F (1°C) per decade from 1990 to 2005. Summer maximum temperatures rose faster than summer minimum temperatures at higher elevations.

- Lower elevation sites experienced greatest warming during the winter months, warming in winter at an average rate of 2.7°F (1.5°C) per decade. The differences in the season of greatest warming are due to the cooling effects on air temperatures of snow on the ground. Increases in winter minimum temperatures increased faster than winter maximum temperatures at lower elevations, while summer maximum temperatures rose faster than summer minimum temperatures at higher elevations.

In a longitudinal analysis of annual temperatures in the San Luis Valley, Colorado, average annual temperatures were found to have increased by 1.9°F (1.1°C) over the period 1957 through 2006 (Ray et al. 2008). A “breakpoint” in the year 1994 was identified in the COOP data for sites in the San Luis Valley, and the increase in the mean growing season temperature for the period 1994 through 2008 was 1.7 to 3.53°F (0.4 to 1.96°C) greater than the mean for the period 1958 through 1993 (Mix et al. 2012).

IV.C. Climate Model Temperature Projections

Climate model projections of temperature and precipitation consist of three components:

- A coarse-resolution global model of atmospheric and ocean circulation (atmosphere-ocean general circulation model [AOGCM]).

- Estimates of future concentrations of greenhouse gases in the atmosphere, typically provided by Intergovernmental Panel on Climate Change (IPCC) Special Report on Emissions Scenarios (SRES) models of different combinations of economic, demographic, and technological development, as well as estimates of future globalization, primarily consisting of the A2 (high emissions), A1B (moderate emissions) and B1 (low emissions) scenarios (IPCC 2000). Since the atmosphere is well-mixed, these represent global values and not regional values. These estimates are key to determining the rate and magnitude of climate change modeled using the AOGCMs.

- Statistical or dynamical downscaling of the AOGCM model outputs to produce estimates of climate change at the regional scale and to serve as inputs into regional hydrologic models to estimate changes in streamflow and other parameters.

IV.C.1. Climate Model Projections for the Southwestern U.S.

Model projections indicate that surface temperatures in the Southwest will warm substantially over the 21st century (highly likely), and warming is likely to be higher in summer and fall than in winter and spring (Cayan et al. 2013). This contrasts with warming to date, which has been greatest in winter months (e.g., Saunders and Maxwell 2005). For the Southwest as a whole, compared to the period 1971 through 2000, models used in the most recent national climate assessment (USGCRP 2013) project (Cayan et al. 2013):

- For the 2021 through 2050 period, warming under the low future emissions model scenario (known as B1) will be between 1 to 3°F (0.6 to 1.7°C) while under the higher future emissions model scenario (known as A2) warming is likely to be between 2 to 4°F (1.1 to 2.2°C).
- For the period 2041 through 2070, warming under the B1 scenario is likely to range from 1 to 4°F (0.6 to 3.3°C) and under the A2 scenario from 2 to 6°F (2.2 to 3.3°C).
- For the period 2071 through 2099, warming under the B1 scenario is likely to range from 2 to 6°F (2.2 to 3.3°C) while under the A2 scenario, the projections are 5 to 9°F (2.8 to 5°C).
- Warming is likely to be higher inland and to increase from south to north.

Seasonal differences in warming are likely, although the high variation among models reduces confidence in specific results (Cayan et al. 2013):

- Increases in summer temperatures are likely to be greater than for other seasons, with mean increases across modeled scenarios around:
    - 3.5°F (1.9°C) in 2021 through 2050
    - 5.5°F (3.1°C) in 2041 through 2070
    - 9°F (5°C) 2071 through 2099
- The least amount of warming is anticipated for the winter months, with average increase of 2.5°F (1.4°C) in 2021 through 2050, increasing to almost 7°F (3.9°C) in 2071 through 2099.

IV.C.2. Climate Model Temperature Projections for the Upper Rio Grande Basin

NOAA recently conducted a relatively fine-grained analysis in support of the National Climate Assessment (NOAA 2013a), using downscaled Coupled Model

Intercomparison Project 3 (CMIP3) models and the more recent North American Regional Climate Change Assessment Program (NARCCAP) models. In maps of average annual temperature change using the CMIP3 multi-model mean simulations, the Upper Rio Grande region warms by 7.5 to 8.5°F (4.1 to 4.9°C) by 2070 through 2099 under the higher emissions (A2) scenario, and by 4.5 to 5.5°F (2.5 to 3.1°C) by 2071 through 2099 under the lower emissions (B1) scenario (Figure 14, NOAA 2013a). These changes are considered significant.

For the NARCCAP simulations using the A2 (high emissions) scenario for the period 2041 through 2070, compared to a baseline period of 1971 through 2000, temperature increases by season show that the largest increases are likely to occur in summer, with increases of 5.5 to 6.0°F (3.1 to 3.3°C) in average temperature in the Upper Rio Grande Basin, followed by fall, with increases in average temperature in the range of 5.0 to 5.5°F (2.8 to 3.1°C). In winter and spring, the Upper Rio Grande Basin is likely to see increases in average temperature of 4.0 to 4.5°F (2.2 to 2.5°C) (Figure 15, NOAA 2013a). There is model agreement on the direction and magnitude of these changes.

In addition to changes in average annual and seasonal temperatures, models project changes in other temperature-related variables. The number of days with maximum daytime temperatures greater than 95°F is expected to increase by about 5 days in the northern part of the Upper Rio Grande Basin grading to about 15 to 20 days in the southern portions. There is strong model agreement for changes in the southern portion of the region but not in the northern. Conversely, the number of days with temperatures below freezing is expected to decline by approximately 25 to 30 days throughout most of the Upper Rio Grande Basin, and as high as 30 to 35 days in the Colorado portions of the Upper Rio Grande Basin. The freeze free season will increase by 25 to 30 days throughout the Upper Rio Grande Basin (Figures 18, 20, and 22; NOAA 2013a).

Additional projections of temperature change come from studies focusing specifically on New Mexico (New Mexico Office of the State Engineer 2006) and Colorado (Ray et al. 2008 and Nydick et al. 2012):

- Projected changes to New Mexico temperatures based on the SRES A1B scenario were modeled using an ensemble of 18 global climate models downscaled to finer resolution. The models suggest significant increases in temperature by 2100. Statewide, average annual temperatures are projected to rise more than 5°F (3°C) over the average from 1971 through 2000. This is a change greater than that observed in the instrumental record. Increases in summer temperature are projected to be greater (Gutzler et al. 2006).

- For Colorado as a whole, an increase in annual temperature of 1.5 to 3.5°F (0.8 to 2.0°C) by 2025 relative to 1950 through 1999 average temperatures is expected, with increases of 2.5 to 5.5°F (1.4 to 3.1°C) expected by 2050 (Ray et al. 2008). Summer temperatures are anticipated to increase faster than winter temperatures.

For the San Juan Mountains, modeling has been undertaken by Rangwala and colleagues (Cozzetto et al. 2011) using a series of downscaled models driven by the A2 (high emissions) scenario. They compared the average temperatures and precipitation for the baseline period of 1971 to 2000 against the model reference period of 2041 to 2070. In summer, fall and winter, daytime high temperatures were expected to increase faster than nighttime low temperatures, but the pattern is reversed in the spring.

Table 1.—Model Projections for San Juan Mountain Climate Change, Average for 2041 to 2070 Compared to 1971 to 2000, Median Values of Model Runs (Cozzetto et al. 2011)

	Change in T_{max} (°C)	Change in T_{min} (°C)	Change in Precipitation (%)	Change in Precipitation (cm)
Winter	2.5	3.2	4.0	0.5
Spring	2.8	2.5	-5.0	-1.0
Summer	3.7	3.1	-17.0	-2.3
Fall	3.2	2.7	-9.0	-1.3

In a 2007 study, Hurd and Coonrod (2007) use three global climate models driven by the A1B “business as usual” SRES scenario to model hydrology and stream flow changes for the periods 2020 through 2039 and 2071 2070 through 2089. The three models are chosen because one represents a slightly “wetter” projection, one a slightly “drier” projection and one a “middle of the road precipitation” projection.

In their models, average annual temperatures increased by 1.7 to 3.2°F (0.95 to 1.76°C) by 2030 (Figure 4) and 5.5 to 7.9°F (3.06 to 4.40°C) by 2080 (Figure 5). Temperature increases are projected to be greatest in summer under the dry scenario, presumably reflecting changes in summer cloudiness resulting from a reduced monsoon (under the dry scenario, precipitation declines steeply in the summer months).

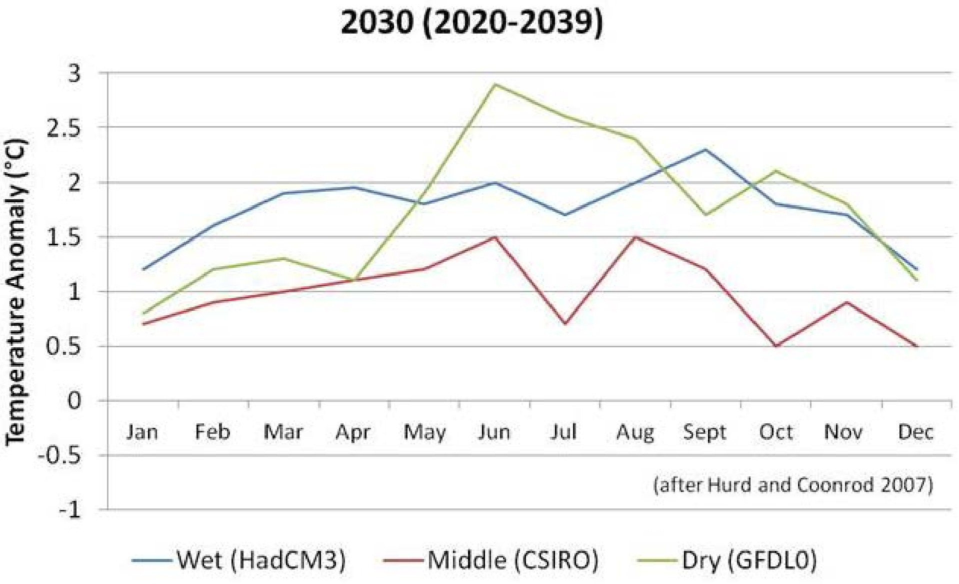


Figure 4.—Three scenarios for temperature change projected for the Rio Grande basin in 2020 through 2039 (source: Hurd and Coonrod 2007).

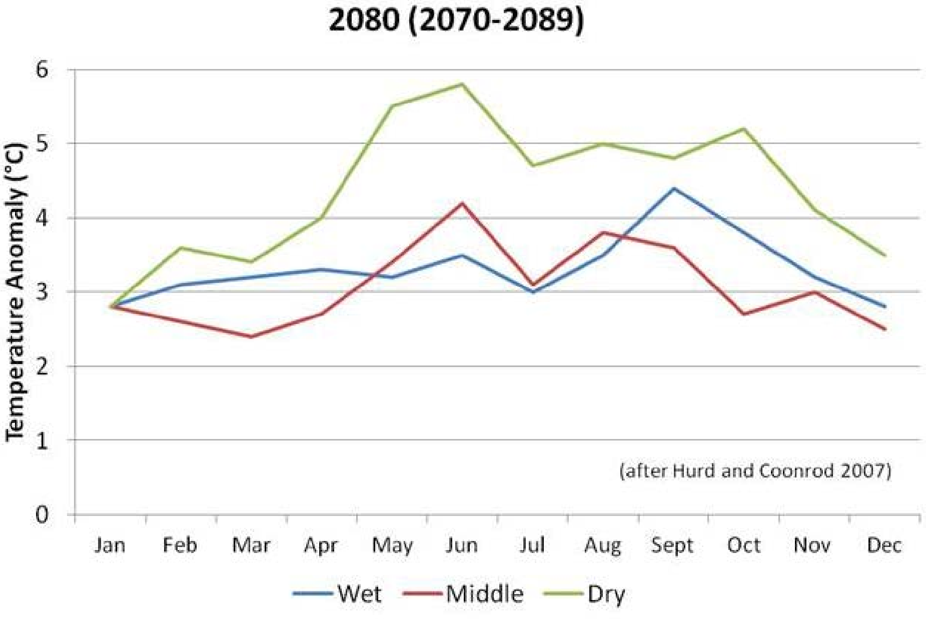


Figure 5.—Three scenarios for temperature change projected for the Rio Grande basin in 2070 through 2089 (source: Hurd and Coonrod 2007).

Climate change in the Upper Rio Grande Basin was modeled by Reclamation (2011a and c) using the Hybrid Delta-ensemble (HDe) approach (Brekke et al. 2010) employing output from 16 models from the CMIP3 multi-model dataset. The outputs are average monthly precipitation and surface air temperature generated from a suite of 16 CMIP3 models forced by three IPCC SRES scenarios for future greenhouse gas emissions. The scenarios chosen are the A2 (high emissions), A1B (business-as-usual emissions) and B1 (low emissions) scenarios. The baseline period is the 1990s. The spatial resolution of the model is 1/8° (about 12 x 12 kilometers).

The basin-average mean-annual temperature is projected to increase by approximately 5 to 6°F (1.8 to 3.3°C) during the 21st century (Reclamation 2011a) relative to the 1990s. Temperature changes are anticipated to be uniform over the basin and to increase steadily through time.

IV.D. Summary of Projected Temperature Changes

By the end of the century, temperatures in the Upper Rio Grande are anticipated to increase by about 9°F (5°C) over 20th century values under high emissions scenarios, and by close to 5.4°F (3°C) under the B1 (low emissions) scenario. There is consensus that temperature increases will be greater in summer and fall. In mountain areas, overnight temperatures (T_{min}) are likely to rise faster than daytime high temperatures (T_{max}). Changes in precipitation are likely to affect net warming across the year because evaporation and condensation processes consume energy that would otherwise go to land surface heating, and also indirectly affect warming through the density and composition of vegetation cover and the persistence of snow cover. By the 20th century's end, temperature increases are anticipated to expand the freeze-free (growing) season by 25 to 30 days; to cause more frequent, longer heat waves (>95°F); and to cause less frequent, shorter cold spells (<0°F).

V. Literature Review: Observed and Projected Changes to Precipitation

V.A. Recent Precipitation Trends

Warming-driven changes to global atmospheric circulation will affect when, where, and by how much precipitation will change. These changes will be superimposed on already highly-variable precipitation patterns resulting from the interplay of long- and short-term climate cycles. Long-term wet and dry cycles in the Southwest are controlled primarily by Pacific sea surface temperatures particularly the multi-decadal PDO. Atlantic Ocean sea surface temperatures are also important. The driest phases in the Southwest are associated with cool Pacific sea surface temperatures (negative PDO) and warm Atlantic sea surface temperatures (positive AMO) (McCabe et al. 2004). Interannual (i.e., time scales of 1 to less than 10 years) variation in winter precipitation is controlled by the ENSO cycle, with either El Niño or La Niña amplified, depending on the state of the PDO. Because of the high variability in precipitation in the Southwest at multiple scales, detecting changes in precipitation has been more challenging than detecting changes in temperature.

V.A.1. National Precipitation Trends

At the national scale, precipitation has increased by 5 percent over the past 50 years, driven by increased evaporation from warmer ocean surfaces putting more moisture into warmer air that, in turn, enables bigger storms with more precipitation to form. Most of the precipitation gain has been in the Northeastern U.S. from the eastern Dakotas to the Atlantic Ocean, with decreases in the Southeast. New Mexico overall had a slight increase in November to March precipitation over the period 1950 through 1999 (Mote et al. 2005). Attribution studies have so far concluded that precipitation trends in the region currently cannot be attributed solely (or directly) to anthropogenic causes, because the magnitude of the trend so far is swamped by the magnitude of variation due to long-term and short-term shifts in Pacific and Atlantic sea surface temperatures (Dominguez et al. 2010).

V.A.2. Southwestern U.S. and Upper Rio Grande Precipitation Trends

In the Southwest during the 20th century:

- 1905 through 1930 had wetter winters than average.
- 1931 through 1941 was approximately average.
- 1942 through 1964 was drier than average, with peak dryness occurring during the drought from 1950 through1956 when average annual precipitation remained below the long term average (Swetnam and Betancourt 1998, Sheppard et al. 2002, and Gutzler 2003).
- Average years from 1965 through 1975 were followed by the period from 1976 through 1997/1998 when warm, wet winters and erratic summer precipitation were the norm (Swetnam and Betancourt 1998, Sheppard et al. 2002, and Gutzler 2003).

- These conditions gave way by 1999/2000 to conditions that were warmer and drier than at any period in the 20th century or the preceding 1,200 and more years (MacDonald et al. 2008 and Woodhouse et al. 2010).

Since 2001, large portions of the Southwest have experienced drought, with particularly widespread and severe drying in 2002, 2003, 2007, 2009, 2011 and 2012. During these extremes, precipitation across the region averaged 22 to 25% below the average for the 20th century (MacDonald 2010), leading to a significant reduction in soil moisture and stream flow. For instance, at Lee's Ferry on the Colorado River, annual flow in the early 20th century was approximately 17.0 million acre feet, but averaged only 11.2 million acre feet for 2001 through 2006, and for 2002 alone, flow declined to approximately 6.2 million acre feet (Reclamation 2011b).

Changes in PDO and AMO correspond to the major dry and wet periods (McCabe et al. 2004):

- From 1944 through 1963, combination of a negative PDO and positive AMO were major contributors to Southwestern drought.

- From 1964 through 1976, negative PDO and negative AMO contributed to average precipitation conditions

- From 1977 through 1994, the combination of positive PDO and negative AMO contributed to wetter-than-average precipitation.

- Since 2000, PDO has been primarily negative (Mantua 2013) and AMO has been strongly positive (National Center for Atmospheric Research [NCAR] 2012), contributing to the reemergence of drought across the Southwest.

- The decade 2001 through 2010 has had the second-largest area affected by drought (after the period 1951 through 1960) and the most severe average drought conditions of any decade since 1901 (Hoerling et al. 2013). This drought is ongoing (National Drought Mitigation Center 2013). No trends have been observed in annual water year precipitation from 1895/96 through 2010/11 for the six-state Southwest (NOAA 2013a). Seasonal time series show no trends for winter or spring and summer, and fall shows a slight upward, but not statistically-significant, trend.

For wilderness areas in northern New Mexico, Enquist and Gori (2008) found precipitation changes were highly variable with respect to direction: a 4.5 percent change in mean annual precipitation for 1991 through 2005 compared to the mean for 1961 through 1990 was observed across sites in the Upper Rio Grande Basin in northern New Mexico. However, for the same sites, comparing the mean for 2000 to 2005 against the mean for the period 1961 through 1990 showed a 7.56 percent decrease.

In all parts of Colorado, including the northern portion of the Upper Rio Grande Basin, no consistent long-term trend in annual precipitation have been detected (Ray et al. 2008). High variability in precipitation makes detection of trends difficult.

In addition, there has been no overall trend in the frequency of extreme precipitation events across the Southwest (NOAA 2013a). Throughout the 20th century and into the early 21st century, the number of 1-day-duration and 5-year return interval precipitation events fluctuated but remained within the range of early 20th century values.

V.B. Model Projections of Late 21st Century Precipitation

V.B.1. Projected Changes in Precipitation for the Southwestern U.S.

Climate models are highly confident that the Southwest will become drier. “Highly confident” means that most models agree that drying will occur, even if there is disagreement about the magnitude of drying and the amount of change in precipitation. Drying will be driven by increased evaporation due to warmer temperatures, and by changes in the factors discussed in detail below.

Predictions of precipitation levels have much greater uncertainty than for temperature because there are great uncertainties with respect to how warming might impact ENSO and multi-decadal ocean oscillations in the Pacific, Atlantic, and Arctic Oceans. Small changes in one place can be amplified by changes elsewhere in ways that are poorly understood for current systems. The North American Monsoon and cloud cover in general are also poorly handled in most models.

The general rule of thumb is that warming will intensify precipitation patterns: wet areas, such as the northeastern U.S., will get wetter and dry areas, such as northern Mexico and southern Arizona, will get drier (USGCRP 2009 and 2013). But what will happen in areas lying on the current boundary between subtropical and mid-latitude climates, such as New Mexico, west Texas, Oklahoma and Kansas, is harder to project because these changes depend on estimates of how far north the storm tracks may be displaced by the poleward expansion of the subtropical dry zone (Lu et al. 2007), and how far north the monsoon may penetrate. Model projections range from essentially no change in precipitation to reductions of about 10 percent (Barnett and Pierce 2009).

Researchers at the U.S. Global Change Research Program (2009 and 2013) project a 10 to 20 percent decline in precipitation by 2080 through 2090, primarily in the winter and spring. This decline results from the northward (poleward) shift of midlatitude winter storm tracks bringing the Southwest into the subtropics year-round. Land and ocean warming should bring more moisture into New Mexico during the summer months, providing stronger monsoons, but this is only projected by some models. Modeling by Dominguez and colleagues (2010) suggests that the distribution of drying will be uneven across the Southwest: the southern part of the Southwest will become drier, and the northern part slightly wetter, but the modeled trends were not significant.

Model projections show that precipitation would continue to be highly variable in time and place and that the region would still be vulnerable to unusually wet and dry spells (Cayan et al. 2013). Overall, model simulations used in the most recent National Climate Assessment show changes in precipitation that range from -13% to +10% across all model runs (Cayan et al. 2013). Confidence in model projections is medium-low, reflecting the variation in the magnitude and direction of projected changes.

A key change projected by models is that precipitation will become concentrated in a smaller number of larger-magnitude precipitation events. This is borne up by data that show that the frequency and intensity of heavy downpours in the U.S. has increased, with the share of total precipitation falling in major storm events increasing by nearly 20 percent. This pattern has also been observed in the Southwest. From 1958 to 2007, there was a 9 percent increase in the amount of rainfall falling in very heavy precipitation events across the Southwest, the lowest rate of increase in the country (the Northeast has seen a 67 percent increase and the Midwest a 31 percent increase over this same timeframe). Climate models project that the share of precipitation falling in heavy rainfall events will continue to increase, while a decreasing share will fall during low-intensity events (USGCRP 2009).

V.B.2. Projected Precipitation Changes in the Upper Rio Grande

In the Upper Rio Grande Basin, projected changes to precipitation have no greater certainty than the projections for the Southwest as a whole:

- Global climate models driven by the A2 (high emissions) scenario project an annual precipitation decrease in New Mexico by 2100 of 4.8 percent (29.3 mm), driven mainly by decreases in winter precipitation, but offset slightly by gains in summer precipitation (Gutzler et al. 2006). In the

San Juan Mountains, small gains in winter precipitation are more than offset by declines in precipitation over the remainder of the year (Cozzetto et al. 2011).

- As elsewhere in the West, winter precipitation is expected to increasingly fall as rain rather than snow as warming delays the onset of freezing and advances the start of the growing season (Gutzler et al. 2006). This is expected to be particularly pronounced in the Southwestern states because winter temperatures are already not far below freezing in many areas (Gutzler et al. 2006). Models are split between those showing declines in winter precipitation and those showing small increases. However, temperature-driven increases in evaporation are expected to exceed any increases in precipitation, driving a negative shift in the overall water balance (Nash and Gleick 1993).

- Models showing reductions in winter precipitation show that the mechanism for this is likely to be the northward migration of the winter storm track, particularly in the late winter/early spring. This shift may already be underway, as the data show that the late winter/early spring storm track in the Western states has moved north of the long-term average between 1978 and 1998, contributing to declines in late winter precipitation in New Mexico (McAfee and Russell 2008). Some models suggest changes in ENSO cycles may also drive declines in winter precipitation. However, there is no model agreement on projected changes to ENSO cycles (Vecchi and Wittenberg 2010).

Reclamation's modeling for the Upper Rio Grande Basin suggests a gradual decline in precipitation over the basin over the 21st century (Reclamation 2011a). Rainfall events are anticipated to become more frequent over the course of the year while snowfall events are projected to become less frequent, reflecting expansion of the freeze-free season and warmer overall winter temperatures.

A recent study projects an increase in the size of the probable maximum precipitation event for most of the world using AOGCMs driven by the largest and smallest future greenhouse gas emissions scenarios (Kunkel et al. 2013a). Maximum daily precipitation in the mapped area corresponding to the Upper Rio Grande, under the maximum emissions scenario, sees an increase of 10 to 30 percent in the maximum daily precipitation value in 2071 through 2100, compared to the period 1971 through 2000 (remembering that the 1980s and 1990s were historically the wettest on record in the Southwest). Some of this increase may be mitigated by topographic effects. The increase was halved under the moderate emissions scenario used). The increased storm intensity is anticipated to occur mainly in July/August in the Southwestern U.S.

Climatologically, this would seem to indicate more intense, localized monsoon storm events (e.g., bigger flash floods) and not increased spring runoff flood events. The driving force in this increase in storm intensity is increased global atmospheric moisture content.

V.B.3. Projected Changes in Winter Precipitation Due to Expansion of the Subtropical Dry Zone and Changes to the Jet Stream

Changes in the location of the jet stream, driven by expansion of the tropics and warming of the Arctic, are expected to contribute to increasingly arid conditions in the Southwestern U.S., primarily through reductions in winter precipitation. The Southwestern U.S. is located on the boundary between the arid subtropics (the poleward portion of the tropics), and the more temperate mid-latitudes whose weather systems are dominated by large-scale cyclonic systems. Seasonal changes in atmospheric circulation bring the Southwestern U.S. more deeply into the subtropics in summer, when precipitation is mainly due to local convection (monsoon). In winter, the poleward boundary of the subtropics in the Northern Hemisphere shifts towards the equator, allowing the jet stream to move over the northern Southwestern U.S. and permitting mid-latitude storm systems to cross the region. Thus, winter precipitation in the region is very sensitive to the location of the boundary between the subtropics and mid-latitudes.

Climate models project the expansion of the subtropical belt leading to predictable decreases in winter snowpack in the region: under climate warming scenarios, the jet stream and associated wind and precipitation patterns moves poleward under global warming by a variety of mechanisms. Models have projected an expansion of the tropics by as much as 2 degrees of latitude over the 21st century (about 1° degree poleward in each hemisphere) (Lu et al. 2007).

However, a series of trends studies examining changes in atmospheric composition, wind speed, and other parameters suggest that in the period from 1979 through 2005, the subtropical dry zone already expanded poleward between 2 and 8 degrees of latitude (about 0.8 to 4 degrees poleward in the Northern hemisphere) depending on the measure used (Seidel et al. 2008 and Fu and Lin 2011). The reason for the accelerated expansion of the subtropical dry zone relative to model projections is not clear. In the Southern Hemisphere, stratospheric ozone depletion in addition to greenhouse gas forcing has been suggested as a cause, while in the Northern Hemisphere increases in both black carbon (soot) and tropospheric ozone as a result of human activities may be important, contributing causes in addition to greenhouse gas forcing (Allen et al. 2012).

Some researchers see evidence in the current climate data that warming-driven expansion of the subtropical dry zone is already under way (Seager et al. 2007). One study has shown that a northward shift in the jet stream began in 1978, allowing more rain to fall to the north and east, and leaving the Southwest drier in the early spring (McAfee and Russell 2008). As a result, the northern Great Plains states have seen a small increase in spring precipitation. Another study notes changes in precipitation and evaporation in the tropical atmosphere since 1979 that are consistent with warming-forced expansion of the subtropical dry zone (Seager and Naik 2012).

Changes in the speed of the jet stream and the amplitude of the Rossby waves in the jet stream are also occurring (Francis and Vavrus 2012). Rossby waves are the north-south meanders in the jet stream, which have a characteristic amplitude. The speed of the jet stream and amplitude of the Rossby waves are affected by changes in the radiation balance in the Arctic (caused by changes in Arctic sea ice thickness and extent) and changes in Northern Hemisphere snow cover extent and season duration. In particular, recent reductions in summer sea ice extent have allowed the Arctic Ocean to warm more in summer and to gradually release this additional heat to the atmosphere during the autumn. The effect of this warming has been to reduce the temperature gradient between the polar and mid-latitudes during much of the year, but particularly from October through December. Since the speed of the jet stream is directly related to the temperature gradient, the result has been a decrease in the speed of the jet stream. This decrease in the jet stream speed has resulted in a slowing of the eastward progression of Rossby waves in the jet stream that influence the formation and movement of mid-latitude storms, noticeably in autumn (Francis and Vavrus 2012).

Warming in the Arctic is also likely to increase the amplitude of the Rossby waves in the jet stream by causing the northern peaks of the Rossby waves to extend further poleward. A storm following the jet stream will thus have a higher amplitude wave to track and as a result of this extra north-south movement, it will take storms longer to make net easterly progress across the country. This phenomenon appears to be occurring during both the summer and fall.

Recent effects of changes to the jet stream as a consequence of Arctic sea ice loss and earlier Arctic snowmelt have been to cause midlatitude storms and anticyclones (high pressure systems that bring clear, dry weather) to “linger” over regions producing longer wet periods and longer dry periods between, along with protracted heat waves and cold spells (Francis and Vavrus 2012). Continued decreases in Arctic sea ice and Northern Hemisphere snow cover are likely to amplify these effects.

V.B.4. Projected Changes in the North American Monsoon

Over the period 1948 through 2004, a significant delay in the beginning, peak, and closing stages of the monsoon was observed, corresponding to a decrease in rainfall during July and a corresponding increase in rainfall during August and September. Dry preceding winters led to decreased soil moisture. Grantz et al.

(2007) proposed that, since early season monsoonal precipitation depends on moisture derived from evaporation, low soil moisture precluded sufficient evaporation to initiate early monsoon precipitation. Consequently, the onset of the monsoon was delayed until sufficient moisture could be drawn into the region by convection. Modeling studies suggest that an enhanced convective barrier may form due to low soil moisture in the early summer (leading to declines in early summer precipitation) and is followed by higher late summer/early fall monsoonal precipitation (Seth et al. 2011).

The delayed monsoon model of Grantz et al. (2007) is directly contradicted by studies showing that monsoons are strengthened following dry winters (e.g., Gutzler 2000). Such discrepancies arise because the fundamental drivers of variations in the North American Monsoon are poorly understood:

- Global circulation models cannot resolve the North American Monsoon as a distinct process because they cannot key resolve regional processes, or if dynamically downscaled, the models can only do so at a very coarse resolution (Cayan et al. 2013).
- ENSO and PDO exert effects on the North American Monsoon, and how these may change in the future is unclear.
- It is not clear how strengthening of Hadley cell circulation leading to enhanced subsidence in the subtropical dry zone (including New Mexico) will affect North American Monsoon formation and strength.

Recent modeling reported by NOAA (2013a) using the NARCCAP models under the A2 (high emissions) scenario indicate declines in spring and summer precipitation of about 5 to 10 percent in the Upper Rio Grande Basin in 2041 through 2070 compared to 1971 through 2000, but model agreement was poor and the changes were not significant in most models. These losses are offset by small gains in fall and winter precipitation.

V.B.5. Summary of Projected Precipitation Changes

Overall, models project that precipitation in the Upper Rio Grande Basin—and the Southwest as a whole—will remain unchanged, or will decline slightly (with a maximum of approximately 13 percent reduction) over the 21st century. More precipitation likely will fall as rain; less will fall as snow. Slight gains in fall and winter precipitation may be offset by losses in summer precipitation. The frequency of extreme precipitation events is likely to be unchanged. Precipitation may become more concentrated in larger precipitation events, but this change in distribution is likely to affect only a small fraction of storms. Projections for precipitation are limited by uncertainties in factors driving variability in the North American Monsoon, ENSO, PDO, and AMO. Additional uncertainties arise with respect to the impacts of the loss of Arctic sea ice, the reductions in Northern Hemisphere snow cover, and the poleward expansion of the subtropical dry zone, all three of which appear to be occurring at a rate faster than predicted by current global circulation models.

VI. Literature Review: Projected Changes to Drought Frequency and Intensity

Regardless of whether precipitation increases or decreases, the consensus is that temperature-driven increases in evaporation will lead to greater evapotranspiration, a net decrease in soil moisture, and a persistently negative water balance for the region. Increases in precipitation would act as a negative feedback, slowing down these impacts; decreases in precipitation would act as a positive feedback, accelerating these changes.

Three classes of drought are generally recognized (Dai 2011):

- “Meteorological drought” refers to a period of months or years in which precipitation is below normal, whether or not this condition is accompanied by increased temperatures. Direct precipitation measurements are used to assess meteorological drought.

- “Agricultural drought” refers to a period when soils are dry, which can be a result of a decrease in precipitation (meteorological drought), an increase in evaporation (e.g., due to increased temperatures), or changes in land use, vegetation cover, or other factors in the watershed. Agricultural drought is usually measured using an index, such as the Palmer Drought Severity Index (PDSI). The PDSI includes both a precipitation term and a temperature term (as a proxy for evaporation). Therefore, this index reflects the balance of moisture inputs or loss to the soil in an area.

- “Hydrological drought” refers to declines in streamflow and water storage in lakes and reservoirs. Hydrological drought is measured in terms such as discharge (cfs), stream flow (feet per second), or storage (acre-feet) of a water body. Hydrological drought is sensitive to a variety of factors, including precipitation, temperature, evapotranspiration, surface and ground water management, erosion, grazing, and other changes in land use and vegetation cover in the watershed. Hydrologic drought develops more slowly and may be partially masked by natural and artificial storage.

Because both agricultural and hydrological droughts are measures of water balance and not of absolute precipitation, it is possible to have both kinds of drought in the absence of a meteorological drought: warming atmospheric temperatures that drive up atmospheric moisture demand (evaporation) can tip the balance towards agricultural and hydrologic drought even if precipitation stays the same or even increases slightly (the "Global Change Type Drought" of Breshears and colleagues [2005]). Modeling studies have shown that this process may have been happening during latter half of the 20th century, when increasing temperature in the Southwest led to declines in both soil moisture and runoff in spite of precipitation increases (Andreadis and Lettenmaier 2006).

Changes in precipitation intensity can also affect soil moisture and stream flow even if total precipitation is unchanged: some climate models predict an increase in frequency of heavy precipitation and a reduction in light to moderate precipitation events. This would lead to longer and more intense dry spells between larger precipitation events, causing vegetation stress and die-off. As precipitation falls in more intense showers, this heavy rain falls on increasingly bare ground, leading to higher runoff to precipitation ratios, lower infiltration rates, and greater erosion than previously. Decreased infiltration reduces the amount of surface moisture that can be subsequently evaporated and precipitated in a region, a positive feedback further enhancing the length of the period between storm events. Decreased infiltration also contributes to reductions in groundwater recharge, contributing to regional near-surface water table declines and decreases in soil moisture, with follow-on impacts to springs, streams, and woody vegetation.

VI.A. Recent and Past Droughts

Historically, droughts were common in the Southwest. Between 1916 and 2008, there were 11 extreme drought years covering all or part of the region. An extreme drought year is defined as a water year in which the area-averaged soil moisture falls below the 10th percentile of the 1951 through 1999 historical period (Cayan et al. 2010). Extreme drought years in the 20th century have usually been embedded in longer dry periods, with the droughts building up and subsiding over several years. These dry periods historically ranged from 47 to 123 months (about 4 to 10 years). Three of the 11 extreme drought years occurred in the 1st decade of the 21st century (in 2002, 2007, and 2008), nestled within a period of elevated temperatures beginning in 2000 and continuing through 2012.

Although most years in the first decade of the 21st century have been exceptionally dry, overall the drought through 2010 did “not have an unusual precipitation deficit,” but the “warmth of the … drought [was] exceptionally strong and consistent” (Cayan et al. 2010). The results have been persistent soil moisture deficits and runoff levels that are below average extreme dry levels: for example, in the first decade of the 21st century, the Colorado River has experienced its lowest 5-year mean flows on record. The start of the current drought is variably placed by researchers, with some arguing for the onset of drought by late 1999 (Cook et al. 2004). Modeling by Seager and Vecchi (Seager and Vecchi 2010) suggests that the early 21st century drought is within the range of natural climate variation and cannot be attributed to anthropogenic warming. Since 2010, precipitation has declined strongly, with drought currently due to both elevated temperatures and reduced precipitation.

The duration the current drought is not remarkable considering the tree ring records of climate change covering the last 1,200 years (back 2,000 years in some areas). In a widely cited work, Cook and colleagues (2004) used annually-resolved tree-ring records from throughout North America to reconstruct annual summer-season PDSI for the last 1,200 years, which includes the Medieval Warm Period, a warm climate interval between approximately AD 800 and 1300 when Northern Hemisphere temperatures increased due to natural forcing.[5] In the warmest part of the Medieval Warm Period (from AD 950 to 1150), average Northern Hemisphere temperatures was higher by 0.36 to 0.72°F (0.2 to 0.4°C) than the mean annual temperature for 1850 through 2006; by comparison, late 20th / early 21st century global average temperatures are 1.44°F (0.8°C) above the same mean (Mann et al. 2008).

In the Southwest, average annual temperatures during the Medieval Warm Period may have been 0.72 to 1.44°F (0.4 to 0.8°C) above the mean annual temperatures recorded for 1850 through 2006. During the warmest intervals in the Southwest, temperatures may have approached 1.8°F (1°C) above this mean, a value equal to the 1961 through 1990 mean and well below the average temperature for the first decade of the 21st century (Woodhouse et al. 2010).

The strongest of the multi-decadal droughts during the Medieval Warm Period megadrought occurred between 1140 and 1159. Based on tree-ring records (Meko et al. 2007), the warmest, driest period of the 12th century was AD 1146 through 1150. During this period, 65.5 percent of the Southwest was under drought conditions, and average annual maximum temperatures for the region were 60.2°F (15.65°C). By comparison, over the 20th century, the average annual maximum temperature (1909 through 2008) has been 60.3°F (15.72°C) and average temperature from 1999 through 2008 was 63.6°F (17.54°C). Moreover, during the 1146 through 1150 period, 32.6 percent of the region was under drought conditions, while during the period 1999 through 2008, 48.4 percent of the region was in drought (see Woodhouse et al. 2010). The drought has persisted through fall 2013.

Medieval Warm Period warming is thought to have been the result of increased solar irradiance and reduced volcanic activity, which forced the Pacific into a persistent "La Niña"-like state. El Niño and La Niña climate swings occurred

---

[5] Detection and attribution studies have assessed whether the same factors responsible for the Medieval Warm Period (high solar irradiance and reduced volcanism [Cook et al. 2004]) might account for today's warming, and have consistently found that they do not.

relative to this drier base state. If recent warming has a similar effect on tropical Pacific sea surface temperature patterns, a similarly more arid base state will emerge that could potentially last for centuries (the duration of projected warming). While some researchers propose changes to ENSO as the primary driver of future droughts, both the intensification of tropical-subtropical circulation and the expansion of the resulting subtropical dry zone are important features projected to contribute to future aridity in climate models of the Southwest (Seager et al. 2007).

A retrospective analysis has shown that, since 1980, there has been a highly statistically significant trend toward increased drought in the American Southwest, particularly over the Colorado River Basin, dependent on teleconnections with the Pacific North American (PNA) pattern and the AMO, which primarily influence winter precipitation (Balling and Goodrich 2010).

VI.B. Model Projections of Late 21st Century Drought

In a review of 19 models used by the IPCC in its most recent assessment report, Seager and colleagues (2007) examined trends in precipitation minus evaporation over the period 1900 through 2098 (modeling included both the historic record and projected 21st century climate) in the Southwest. They found that under the A1B (moderate emissions) scenario, models project a sustained transition to drier climate beginning in the 1990s or early in the 21st century. This change is driven by declines in precipitation and increases in evaporation. Most of the projected drying occurs in winter. This modeling effort suggests that the average climate of the Southwest by mid-21st century will resemble that of climate during a multi-year drought today. “The most severe future droughts will still occur during persistent La Niña events, but they will be worse than any since the Medieval period, because the La Niña conditions will be perturbing a base state that is drier than any state experienced recently” (Seager et al. 2007).

Seager and Vecchi (2010) also reviewed 24 IPCC models with robust representations of precipitation and evaporation in the Southwest through 2099. They found that the models project a steady decline in both winter (October through March) and summer (April through September) precipitation in the 21st century relative to the 20th century. In winter, warming causes evaporation to increase steadily, resulting in projections of an increasingly negative value for precipitation-evaporation over the 21st century. Decreases in the value of winter precipitation-evaporation occur in all models—regardless of precipitation trends, showing the projected dominance of temperature-forced increases in evaporation over any increase in precipitation. In the models, the primary causes of changes in precipitation-evaporation are expansion of the subtropical dry zone and the poleward retreat of the temperate wet zone, driven by global-scale warming. Worst-case drying scenarios occur in models predicting a shift to a persistent

La Niña state in the Pacific, while the wettest scenarios occur in models predicting a persistent El Niño state. However, because of the overprinting of a gradual drying in the Southwest, not even the wettest future models predict a return to the two wet decades preceding the 1997-98 El Niño. Finally, recent trends in carbon emissions exceed the levels used in this study (based on the A1B SRES scenario), so the drying may be greater than projected in this study.

More recently, a series of 19 models were used to assess projections of future drought over the U.S. under the SRES A1B (moderate emissions) scenario (Wehner et al. 2011):

> *All models, regardless of their ability to simulate the base-period drought statistics, project significant future increases in drought frequency, severity, and extent over the course of the twenty-first century under the SRES A1B emissions scenario. Using all 19 models, the average state in the last decade of the twenty-first century is projected under the SRES A1B forcing scenario to be conditions currently considered severe drought (PDSI < -3) over much of continental United States and extreme drought (PDSI < -4) over much of Mexico. . . . Periods of drought intensity comparable to the massive droughts of the 1930s or 1950s are replicated in the simulated twentieth century by the corrected models, albeit less frequently than observed. By the end of the twenty-first century, this condition becomes the normal one (Wehner et al. 2011:1374).*

Part of differences in model projections of drought at any point in time is affected by differences in the rate of change inherent in the models: models with faster rates of change predict higher temperatures (and therefore more drought) than models with slower rates of change for a given point in time. To adjust for this, the models were used to project drought for a given temperature, without regard to when this temperature is reached by the models:

> *At a 2.5 K [2.5°C, 3.6°F] global increase in surface air temperature relative to the 1900-09 average, an all-model projection exhibits moderate drought conditions over most of the western United States and severe drought over southern Mexico as the mean climatological state (Wehner et al. 2011:1375).*

The dates at which these models reach 3.6°F (2.5°C) above the 1900 through 1909 average ranges from 2029 to 2110, with 11 of 19 models falling between 2045 and 2060 (Wehner et al. 2011:Table 5). In the models, drought intensity is greatest in the Intermountain West and Plains.

VII. Hydrologic Changes

In the West, most of the water flowing year-round in streams originates as mountain precipitation (via winter snow pack) or from localized upstream precipitation during the summer monsoon, primarily in headwaters areas. Snowmelt is 50 to 80 percent of flow volume in this region (Stewart et al. 2005). Snowmelt is the dominant source of flow in the Rio Grande above its confluence with the Rio Chama, while below this confluence both snowmelt and summer precipitation are important. The river is fully allocated and flows in the river are tightly regulated.

VII.A. Observed Hydrologic Changes

VII.A.1. Changes to Snowpack

Two important variables with regard to snowpack are the quantity of precipitation falling as snow, and the amount of water contained in a given volume of snow, snow water equivalent (typically 5 to 20 percent in freshly fallen snow).

Throughout much of the West, warming winter temperatures have contributed to declines in snowpack (Mote et al. 2005). Warmer late fall and early spring temperatures mean that precipitation that formerly fell as snow during these periods now often falls as rain, particularly at lower elevations and in more southerly mountainous regions. Thus the percent of annual mountain precipitation that falls as snow has declined, reducing the amount of water available for runoff in the spring and summer months.

There has also been a long-term decline in the ratio of winter-total snow water equivalent (SWE) to winter total precipitation. The most significant reductions have occurred where winter wet-day minimum temperatures averaged for the period 1949 through 2004 were warmer than -5°C, with the greatest loss between -3°C and 0°C. The changes were most pronounced in spring (Knowles et al. 2006).

In a major review of the data from 1950 to 1997, the Southwestern mountains showed a 60 percent gain in precipitation (Mote et al. 2005), but this is an artifact of a trend line that begins in the last major Southwestern drought (1950 through 1956) and ends in the wettest period of the historical record (1976 through 1997/1998). A study combining observational data and modeled historic snowpack has shown a post-1980 decline in snowpack conditions in the West that has no precedent in 20th century temporal variability in precipitation, temperature, and estimated snow water equivalence. Winter temperatures since 1980 are, on average, higher than any other decade of the 20th century, while the average April 1 snow water equivalence and the ratio of snow water equivalence to precipitation are lower (McCabe and Wolock 2009).

In a recent study, tree ring records were used to estimate annual snow water equivalence since AD 1200 in the Rocky Mountains. Prior to the 1980s, there was a pronounced dipolar character to snow water equivalence: dry years in the

Northern Rocky Mountains (Wyoming and north) corresponded to wetter years to the Southern Rocky Mountains (Colorado and New Mexico), and vice-versa. Since 1980, this pattern has broken down, and declines in snow water equivalence are evident across the entire cordillera (mountain range) (Pederson et al. 2011). The authors conclude that their data suggest “a fundamental shift from precipitation to temperature as the dominant influence on snowpack in the North American Cordillera.”

The importance of snowmelt to runoff has been changing in northern New Mexico. A study of runoff trends over the period 1948 through 2008 shows that streams draining the Sangre de Cristo Range and Jemez Mountains have shifted from clearly snowmelt dominated to increasingly rain dominated over this time period, a trend that has not emerged in the San Juan Mountains (Fritze et al. 2011).

Snowpack accumulation is also related to regional vegetation cover, with maximum accumulation occurring in forests with canopy densities between 25 and 40 percent, and along north-facing canopy edges (Veatch et al. 2009). Canopies of this density effectively intercept snowfall and shade it from direct solar radiation. Anticipated changes to mountain vegetation due to drought and wildfire (Williams et al. 2010) have the potential to change the way snowpack accumulates by replacing forests with bare ground, grassy meadows, shrublands, and woodlands in large portions of mountain catchments.

VII.A.2. Advances in Snowmelt

The observational record of 1948 through 2000 reveals a steady advance in the initiation of snowmelt across the West, with greater advances occurring in the northern tier of Western states (Stewart et al. 2005). The data show earlier beginning of snowmelt, and advances in the center of mass of the annual hydrograph (peak spring runoff) by one to four weeks (see also Fritze et al. 2011). The earlier onset of snowmelt is accompanied by decreased spring and early summer (April, May, June, and July) fractional flows (i.e., flows as a portion of the annual total) as a greater portion of the runoff occurs earlier in the water year (due to earlier snowmelt and warmer late winter temperatures permitting snow to fall as rain and earlier mountain snowmelt). Importantly, the advance in timing correlates strongly with an increase in temperature over this time period but correlates poorly with long-term changes in Pacific Ocean sea surface temperatures. Model projections suggest continued advances in snowmelt timing, with advances of as much as a month or more projected for 2080 through 2099 relative to baseline data from 1951 through 1980 (Stewart et al. 2004).

Other processes associated with aridity can affect the rate of snowmelt. Increased aridity is likely to reduce vegetation cover, leaving soil exposed to erosion by wind and water. On the Colorado Plateau, researchers measured dust emissions from different vegetation communities. The communities were selected as analogs for vegetation changes expected with increasing aridity. The researchers found that increased temperatures due to climate change will increase wind erosion across the Colorado Plateau, leading to much higher dust emissions in areas with low vegetation cover and low rates of biological soil crust (Munson et al. 2011). The dust can move large distances, and can readily be blown onto areas of mountain snow, changing snowfield albedo (reflectivity) and thereby helping to accelerate spring snowmelt (Seager and Vecchi 2010).

VII.A.3. Declines in Runoff

Southwestern flood magnitudes over the last 85 year have declined strongly, with the strongest decreases along the Rio Grande, Colorado, and Salt-Gila rivers (Hirsch and Ryberg 2011). These declines cannot be wholly explained by reference to ENSO, PDO, AMO, or other natural forcing, or to changes in water allocation or land use practices.

The Colorado River has received greater research attention than the Rio Grande, and serves as a proxy for regional stream flows in many analyses. The rivers are similar in that both streams receive most of their flow from Rocky Mountain runoff rather than from precipitation in downstream reaches. But they differ in a crucial way that suggests projections of future flow based on Colorado River data will underestimate reductions in flows on the Rio Grande. The Colorado River receives runoff from northern Utah, and northern and western Wyoming, areas that are likely to see increases in precipitation that partially offsets reduced precipitation in southern Utah and western Colorado (USGCRP 2009). By contrast, the Rio Grande headwaters lie in the San Juan Mountains of southern Colorado, a place that is likely to see overall reductions in precipitation and increases in evaporation due to the northward expansion of the subtropical dry zone. In the 2011 La Niña winter, heavy precipitation in the Northern Rockies coincided with much-reduced precipitation in the Southern Rockies. As a result, flows in the Colorado River increased from the prior year while flows in the Rio Grande remained low.

During the first decade of the current drought (2001 through 2010), flows declined on both rivers (Hoerling et al. 2013). At Lee's Ferry on the Colorado, average naturalized flows were 12.6 million acre-feet/year, compared to the 1901 through 2000 average of 15.0 million acre-feet/year, representing a 16 percent decadal deficit. On the Rio Grande at El Paso, observed flows for 2001 through 2010 were about 23 percent lower than the period from 1941 through 2000.

VII.B. Projected Hydrologic Changes

VII.B.1. Projected Changes for the Southwestern U.S.

Reductions in snowpack, declines in snow water equivalence, and advances in snowmelt are all projected to contribute to substantial declines in flows in the Southwest's rivers (Cayan et al. 2013). Studies of the Colorado River show that flow on the Colorado River is likely to be reduced by 10 to 30 percent (see discussion in Barnett and Pierce 2009). Since the headwaters of the Rio Grande are located in a region that will likely see no increases in winter precipitation as well as significant declines in precipitation for the rest of the year (USGCRP 2009), it is probable that projected declines in flow in the Rio Grande will equal or exceed those for the Colorado River (Cayan et al. 2013).

Models of future Colorado River flows consistently show reductions in average flow across the 21st century. Coupled ocean-atmosphere global climate models downscaled to the Western U.S. were used to drive a Variable Infiltration Capacity (VIC) model to study changes in streamflow as a result of climate change (Christensen et al. 2004 and Leung et al. 2004). Modelers drove the model using a moderate emissions scenario (close to the mean of models used in the 2009 IPCC reports). For the Colorado River basin, annual predicted runoff was 14, 18, and 17 percent below the historical average for the periods 2010 through 2039, 2040 through 2069 and 2070 through 2098, respectively. However, due to earlier spring snowmelt and higher evaporation rates, it is predicted that the total basin storage in regional reservoirs could decline by as much as 36, 32, and 40 percent for these periods, respectively.

A more recent effort used a simple water budget model that calculated the net effects of inflows and outflows on a monthly time step (Barnett and Pierce 2009). The model incorporates reductions in evaporation from reservoirs as surface area shrinks, as well as changes in river management in response to altered flows. The model shows that, by 2050, if runoff is reduced by 10 percent and consumption is unchanged, water managers will be unable to deliver all of the promised water 58 percent of the time. A reduction in runoff of 20 percent leads to a failure in water delivery approximately 88 percent of the time if consumption patterns are unchanged. The shortfall ranges from at least 970,000 to 1.5 million acre-feet per year (1.2 to 1.9 billion cubic meters per year) to approximately 1.8 to 2.8 million acre-feet per year (2.2 to 3.4 billion cubic meters per year) by 2050 out of a total request of 14 million acre-feet per year (17.3 billion cubic meters per year) (Barnett and Pierce 2009). The magnitude of the shortfall is small enough that it could be compensated for by reductions in demand. Although average flows may decline only a small amount, flow deficits in multi-year drought years have the potential to exceed flow deficits in the observational record by as much as 60 to 70 percent (Cayan et al. 2010).

Reduced runoff and changes in snowpack have a secondary effect on groundwater systems by reducing the amount of water available for recharge. As surface water sources become scarce, groundwater sources may be increasingly relied upon to satisfy water needs. As aquifers are drawn down, the relationship between surface water and ground water may change, reducing surface flow in rivers where groundwater is a significant contributor to surface flow.

Reduced total runoff will likely be accompanied in the future by increases in peak discharge. Precipitation is expected to become more concentrated in time, with fewer but larger storms separated by periods of increased aridity. Aridity will significantly alter vegetation structure, with more xeric vegetation and larger patches of exposed earth. During high-precipitation events, the exposed surfaces may funnel greater share of runoff to streams, contributing higher peak flows than at present.

Studies that detect change and attribute it to causes (detection and attribution studies) have had less success with precipitation, snowmelt, runoff, and other hydroclimate variables than with temperature. In the northern Intermountain West, modelers engaged in detection and attribution studies discovered a clear anthropogenic signal to earlier peak runoff during the period 1950 through 1999 (Hidalgo et al. 2009). However, the observed changes in the southern Intermountain West could not be clearly distinguished by cause: anthropogenic changes appear to be one of several causes contributing to earlier peak spring runoff, declines in snow water equivalent, and other hydroclimate changes in the region.

VII.B.2. Projected Changes in the Upper Rio Grande

There are fewer projections of hydrologic change in the Upper Rio Grande than for the Colorado River, reflecting different definitions of the Southwest used by researchers, and the smaller population dependent on the Rio Grande than on the Colorado River. Changes in temperature and precipitation patterns are expected to drive changes in snowpack:

- Overall, the freezing altitude is projected to rise and snowpack volume to decrease as temperatures rise. Higher temperatures will delay the date at which precipitation falls as snow in the fall and cause a 4 to 6 week earlier shift in the date at which precipitation reverts to rain in the spring. The altitude at which a winter snowpack will develop is anticipated to rise. In the 2005, the Rocky Mountain Climate Organization (Saunders and Maxwell 2005) noted that 10 of the previous 16 years in the Rio Grande Basin had snowpack below the long-term average, a trend that has continued since.

- The snow water content of the snowpack has also declined (Mote et al. 2005), and this trend is anticipated to continue. Compared to the water content of the April snowpack for the period 1950 through 1999, modeling studies of the Colorado River watershed project determined that that the content of water contained in April snowpack will decline by approximately 38 percent by the end of the 21st century in models driven by the A2 (high emissions) scenario (Christensen and Lettenmaier 2007). Similar reductions in snow water equivalence are predicted for all watersheds in the West.

- Regional climate models driven by the A2 (high emissions) scenario indicate that the snowpack may be non-existent south of 36°N (approximately the latitude of the City of Española, New Mexico) by 2100 (Gutzler et al. 2006). The same study showed reductions in snow water equivalence of approximately one-third to one-half (approximately 50 to 200 mm of water) compared to the 1961 through 1985 average in the San Juan Mountains.

Increases in temperature and increases in evaporation will lead to increasing soil moisture deficit:

- In many modeling studies, the increase in summer evaporation appears to plateau—but only because there is no more surface soil moisture to evaporate (Diffenbaugh et al. 2005). Evaporation over reservoirs and other open water is expected to increase directly with temperature. Prolonged droughts relative to those of the 20th century are expected (Gutzler et al. 2006).

- Regional models driven by the A2 (high emissions) scenario show a pronounced soil moisture deficit in the spring (March through May) season, particularly in northwest New Mexico, where soil moisture is projected to decrease by 5 mm water (20 percent relative to 1961 through 1985 simulated baseline). In the models, this deficit is driven by earlier spring snow melt accompanied by higher temperatures and greater evaporation (Gutzler et al. 2006).

The future flows in the Rio Grande are expected to decline, as discussed in recent studies:

- For the Rio Grande basin above Elephant Butte, declines in snow water equivalence, annual runoff, December through March runoff, and April through July runoff are all anticipated. Reclamation conducted the most detailed hydroclimate modeling specific to the Rio Grande has been under its West-Wide Climate Risk Assessment (WWCRA) program as required under the SECURE Water Act ((Reclamation 2011c, summarized in Table 2). Reclamation used data from 112 CMIP3 models that were bias corrected and spatially downscaled to 1/8° cells and then input into a VIC model, with the flows subsequently routed down the Rio Grande. The median changes from their modeling effort, at specific gages, are provided in the table, below (Reclamation 2011c).

Table 2.—Modeling Results from Reclamation (2011c) Showing Hydrologic Changes to the Rio Grande Basin

Location	Precip. (%)	Mean Temp (°F)	April 1 SWE (%)	Annual Runoff (%)	Dec.-Mar. Runoff (%)	Apr.-July Runoff (%)
2020-2029						
Rio Grande near Lobatos	-0.47	1.84	-25.63	-4.98	-7.12	-2.87
Rio Chama near Abiquiu	0.91	1.79	-87.13	-0.24	4.76	-1.27
Rio Grande near Otowi	-0.54	1.82	-42.20	-4.45	-3.07	-2.48
Rio Grande at Elephant Butte Dam	-0.53	1.79	-93.16	-4.05	-3.59	-1.64
Pecos R. at Damsite #3	-1.48	1.79	-100.00	-2.45	-0.63	-1.39
2050-2059						
Rio Grande near Lobatos	-2.29	2.98	-49.46	-18.89	-20.55	-15.37
Rio Chama near Abiquiu	-1.07	3.83	-96.37	-7.28	5.53	-13.85
Rio Grande near Otowi	-2.42	3.82	-63.92	-14.40	-10.41	-15.91
Rio Grande at Elephant Butte Dam	-2.31	3.82	-98.37	-13.48	-8.95	-15.42
Pecos R. at Damsite #3	-0.72	3.76	-100.00	-2.75	-3.76	-3.63
2070-2079						
Rio Grande near Lobatos	-2.23	5.18	-68.97	-22.41	-23.69	-20.13
Rio Chama near Abiquiu	-1.12	5.19	-98.50	-10.96	8.61	-21.68
Rio Grande near Otowi	-2.40	5.19	-84.56	-19.90	-12.00	-21.83
Rio Grande at Elephant Butte Dam	-2.25	5.17	-99.72	-16.41	-10.86	-20.01
Pecos R. at Damsite #3	-1.91	4.97	-100.00	-4.36	-9.42	-5.06

Although these numbers are very precise, they provide only general guidance for future change because the range of variation around each of these numbers is very large; the range for temperature by 2070 through 2079 is approximately 7 to 8°F while models report both gains and losses in precipitation over the basin. Proportionately similar variation exists around all of the figures presented in Table 2 (see Reclamation 2011c: Figure 46).

- A sensitivity study was conducted to assess how snowmelt runoff in the Rio Grande might be affected by a 7.2°F (4°C) increase in temperature in wet, normal and dry years, as well as for a "normalized year" based on the average condition for the period 1957 through 1994 (Rango and Martinec 2008). For the Upper Rio Grande, a greater share of runoff is projected to occur in the winter (October through March) than in the summer (April through September) and the runoff peak was shifted from May to April. Overall runoff also decreased (Table 3).

Table 3.—Redistribution of runoff in warmer climates (adapted from Rango and Martinec 2008, Tables 2 and 3)

	October-March		**April-September**		**Hydrological Year**	
Base Year	**Runoff million m^3**	**Runoff % of Total**	**Runoff million m^3**	**Runoff % of Total**	**Runoff million m^3**	**Runoff % of Total**
1979 (wet)						
Computed T	91.87	7.6	1120.15	92.4	1212.02	100
Computed T+4°C	146.76	12.3	1046.16	87.7	1192.92	100
1976 (average)						
Computed T	93.22	13.1	616.52	86.9	709.74	100
Computed T+4°C	192.95	28.1	494.80	71.9	687.75	100
1977 (dry)						
Computed T	63.56	24.3	198.17	75.7	261.71	100
Computed T+4°C	77.34	29.2	187.42	71.8	264.76	100
"Normalized Year"						
Computed T	74.66	11.7	561.66	88.3	636.32	100
Computed T+4°C	153.06	24.2	479.58	75.8	632.64	100

- In addition to advancing the date of peak spring flood, increases in summer surface temperatures are expected to strengthen convection over the region, producing a more vigorous hydrologic cycle in which storms are more intense (Carnell and Senior 1998). Whether storm frequency declines as well is not clear. Larger magnitude summer storms may drive bigger magnitude flood events, while concentrating spring runoff earlier in the season may increase the magnitude of spring floods. However, lower overall snowpack volume and snow water equivalence, and earlier snowpack melting, are expected to drive down low summer flows (Gleick 2000). In other words, the stream's base flows decline but are punctuated by larger magnitude summer flood events.

VIII. References

Allen, R. J., S. C. Sherwood, J. R. Norris, and C. S. Zender. 2012. Recent Northern Hemisphere Tropical Expansion Primarily Driven by Black Carbon and Tropospheric Ozone. Nature 485:350-354.

Andreadis, K. M. and D. P. Lettenmaier. 2006. Trends in 20th Century Drought over the Continental United States. Geophys. Res. Lett. 33:L10403.

Balling, R. C. and G. B. Goodrich. 2010. Increasing Drought in the American Southwest? A Continental Perspective using a Spatial Analytical Evaluation of Recent Trends. Physical Geography 31:293-306.

Barnett, T. P. and D. W. Pierce. 2009. Sustainable Water Deliveries from the Colorado River in a Changing Climate. Proceedings of the National Academy of Sciences of the United States of America 106:7334-7338.

Bonfils, C., B. D. Santer, D. W. Pierce, H. G. Hidalgo, G. Bala, T. Das, T. P. Barnett, D. R. Cayan, C. Doutriaux, A. W. Wood, A. Mirin, and T. Nozawa. 2008. Detection and Attribution of Temperature Changes in the Mountainous Western United States. Journal of Climate 21:6404-6424.

Brekke, L., T. Pruitt, and D. Smith. 2010. Climate Change and Hydrology Scenarios for Oklahoma Yield Studies. U.S. Department of the Interior, Bureau of Reclamation, Technical Memorandum 86-68210-2010-01.

Breshears, D. D., N. S. Cobb, P. M. Rich, K. P. Price, C. D. Allen, R. G. Balice, W. H. Romme, J. H. Kastens, M. L. Floyd, J. Belnap, J. J. Anderson, O. B. Myers, and C. W. Meyer. 2005. Regional Vegetation Die-Off in Response to Global-Change-Type Drought. Proceedings of the National Academy of Sciences of the United States of America 102:15144-15148.

Carnell, R. E. and C. A. Senior. 1998. Changes in Mid Latitude Variability Due to Increasing Greenhouse Gases and Sulphate Aerosols. Climate Dynamics 14:369-383.

Cayan, D. R., T. Das, D. W. Pierce, T. P. Barnett, M. Tyree, and A. Gershunov. 2010. Future Dryness in the Southwest US and the Hydrology of the Early 21st Century Drought. Proceedings of the National Academy of Sciences of the United States of America 107:21271-21276.

Cayan, D. R., M. Tyree, K. E. Kunkel, C. Castro, A. Gershunov, J. Barsugli, A. J. Ray, J. T. Overpeck, M. Anderson, J. Russell, B. Rajagopalan, I. Rangwala, and P. Duffy. 2013. The Southwest Climate of the Future - Projections of Mean Climate. Page 509 in G. Garfin, editor. Assessment of Climate Change in the Southwest United States: A Report Prepared for the National Climate Assessment. NCA Regional Input Reports. Island Press, Washington, D. C.

Christensen, N. S. and D. P. Lettenmaier. 2007. A Multimodel Ensemble Approach to Assessment of Climate Change Impacts on the Hydrology and Water Resources of the Colorado River Basin. Hydrology and Earth System Sciences 11:1417-1434.

Christensen, N. S., A. W. Wood, N. Voisin, D. P. Lettenmaier, and R. N. Palmer. 2004. The Effects of Climate Change on the Hydrology and Water Resources of the Colorado River Basin. Climatic Change 62:337-363.

Cook, E. R., C. A. Woodhouse, C. M. Eakin, D. M. Meko, and D. W. Stahle. 2004. Long-Term Aridity Changes in the Western United States. Science 306:1015-1018.

Cozzetto, K., I. Rangwala, and J. Neff. 2011. Downscaled Air Temperature and Precipitation Projections for the San Juan Mountain Region. Narrative on Regional Climate Model Projections Submitted to the San Juan Public Land Center, Durango, Colorado.

Dai, A. 2011. Drought Under Global Warming: A Review. Wiley Interdisciplinary Reviews: Climate Change 2:45-65.

Diffenbaugh, N. S., J. S. Pal, R. J. Trapp, and F. Giorgi. 2005. Fine-Scale Processes Regulate the Response of Extreme Events to Global Climate Change. Proceedings of the National Academy of Sciences of the United States of America 102:15774-15778.

Dominguez, F., J. Canon, and J. Valdes. 2010. IPCC-AR4 Climate Simulations for the Southwestern US: the Importance of Future ENSO Projections. Climatic Change 99:499-514.

Enquist, C. A. F. and D. F. Gori. 2008. Implications of Recent Climate Change on Conservation Priorities in New Mexico. The Nature Conservancy, New Mexico.

Francis, J. A. and S. J. Vavrus. 2012. Evidence Linking Arctic Amplification to Extreme Weather in Mid-Latitudes. Geophysical Research Letters 39:L06801.

Fritze, H., I. T. Stewart, and E. Pebesma. 2011. Shifts in Western North American Snowmelt Runoff Regimes for the Recent Warm Decades. Journal of Hydrometeorology 12:989-1006.

Fu, Q. and P. Lin. 2011. Poleward Shift of Subtropical Jets Inferred from Satellite-Observed Lower-Stratospheric Temperatures. Journal of Climate 24:5597-5603.

Garfin, G., editor. 2013. Assessment of Climate Change in the Southwest United States: A Report Prepared for the National Climate Assessment. NCA Regional Input Reports. Island Press, Washington, D.C.

Gleick, P. H. 2000. Water: The Potential Consequences of Climate Variability and Change for the Water Resources of the United States. The Report of the Water Sector Assessment Team for the National Assessment on the Potential Consequences of Climate Variability and Change for the Water Resources of the United States for the U.S. Global Change Research Team.

Grantz, K., B. Rajagopalan, M. Clark, and E. Zagona. 2007. Seasonal Shifts in the North American Monsoon. Journal of Climate 20:1923-1935.

Gutzler, D. S. 2000. Covariability of Spring Snowpack and Summer Rainfall Across the Southwest United States. Journal of Climate 13:4018-4027.

______. 2003. Drought in New Mexico: History, Causes and Future Prospects. Pages 101-105 in P. S. Johnson, L. A. Land, L. G. Price, and F. Titus, editors. Water Resources of the Lower Pecos Region, New Mexico—Science, Policy, and a Look to the Future Decision-Makers Field Guide New Mexico Bureau of Geology and Mineral Resources, Socorro, New Mexico.

Gutzler, D. S., G. Garfin, and B. Zak. 2006. Observed and Predicted Impacts of Climate Change on New Mexico's Water Supplies. Pages 4-32 in A. Watkins, editor. The Impact of Climate Change on New Mexico's Water Supply and Ability to Manage Water Resources. New Mexico Office of the State Engineer/Interstate Stream Commission, Santa Fe, New Mexico.

Hoerling, M. P., M. D. Dettinger, K. Wolter, J. Lukas, J. Eischeid, R. Nemani, B. Liebmann, and K. E. Kunkel. 2013. Evolving Weather and Climate Conditions of the Southwest United States. Page 509 in G. Garfin, editor. Assessment of Climate Change in the Southwest United States: A Report Prepared for the National Climate Assessment. NCA Regional Input Reports. Island Press, Washington, D.C.

Hurd, B. H. and J. Coonrod. 2007. Climate Change and its Implications for New Mexico's Water Resources and Economic Opportunities. New Mexico State University, Agricultural Experiment Station Technical Report 45, Las Cruces, New Mexico.

Hydrologic Preocesses 20:723-739.

Intergovernmental Panel on Climate Change (IPCC). 2000. Special Report on Emissions Scenarios. Page 599 in N. Nakicenovic and R. Swart, editors. Cambridge University Press. Cambridge, United Kingdom. Online: http://www.grida.no/climate/ipcc/emission/.

Jin-Yi, Y. and Z. Yuhao. 2013. The Enhanced Drying Effect of Central-Pacific El Niño on US Winter. Environmental Research Letters 8:014019.

Kunkel, K. E., T. R. Karl, D. R. Easterling, K. Redmond, J. Young, X. Yin, and P. Hennon. 2013a. Probable Maximum Precipitation and Climate Change. Geophysical Research Letters 40:1402-1408.

Kunkel, K. E., L. E. Stevens, S. E. Stevens, L. Sun, E. Janssen, D. Wuebbles, K. T. Redmond, and J. G. Dobson. 2013b. Regional Climate Trends and Scenarios for the U.S. National Climate Assessment. NOAA Technical Report NESDIS 142-5.

Leung, L. R., Y. Qian, X. Bian, W. M. Washington, J. Han, and J. O. Roads. 2004. Mid-Century Ensemble Regional Climate Change Scenarios for the Western United States. Climatic Change 62:75-113.

Lu, J., G. A. Vecchi, and T. Reichler. 2007. Expansion of the Hadley Cell Under Global Warming. Geophysical Research Letters 34:L06805.

MacDonald, G. M. 2010. Water, Climate Change, and Sustainability in the Southwest. Proceedings of the National Academy of Sciences of the United States of America 107:21256-21262.

MacDonald, G. M., D. W. Stahle, J. V. Diaz, N. Beer, S. J. Busby, J. Cerano-Paredes, J. E. Cole, E. R. Cook, G. Endfield, G. Gutierrez-Garcia, B. Hall, V. Magana, D. M. Meko, M. Ménendez-Pérez, D. J. Sauchyn, E. Watson, and C. A. Woodhouse. 2008. Climate Warming and 21st Century Drought in Southwestern North America. EOS Transactions 89:82-83.

Mann, M. E., Z. H. Zhang, M. K. Hughes, R. S. Bradley, S. K. Miller, S. Rutherford, and F. B. Ni. 2008. Proxy-based Reconstructions of Hemispheric and Global Surface Temperature Variations Over the Past Two Millennia. Proceedings of the National Academy of Sciences of the United States of America 105:13252-13257.

Mantua, N. 2013. PDO Index. Online: http://jisao.washington.edu/pdo/PDO.latest. Accessed March 21, 2013.

McAfee, S. A. and J. L. Russell. 2008. Northern Annular Mode Impact on Spring Climate in the Western United States. Geophysical Research Letters 35.

McCabe, G. J., M. A. Palecki, and J. L. Betancourt. 2004. Pacific and Atlantic Ocean Influence on Multidecadal Drought Frequency in the United States. Proceedings of the National Academy of Sciences of the United States of America 101:4136-4141.

McCabe, G. J. and D. M. Wolock. 2009. Recent Declines in Western U.S. Snowpack in the Context of Twentieth-Century Climate Variability. Earth Interactions 13:1-15.

Meko, D. M., C. A. Woodhouse, C. A. Baisan, T. Knight, J. J. Lukas, M. K. Hughes, and M. W. Salzer. 2007. Medieval Drought in the Upper Colorado River Basin. Geophysical Research Letters 34.

Mix, K., V. Lopes, and W. Rast. 2012. Growing Season Expansion and Related Changes in Monthly Temperature and Growing Degree Days in the Inter-Montane Desert of the San Luis Valley, Colorado. Climatic Change 114:723-744.

Mote, P. W., A. F. Hamlet, M. P. Clark, and D. P. Lettenmaier. 2005. Declining Mountain Snowpack in Western North America. Bulletin of the American Meteorological Society 86:39-+.

Munson, S. M., J. Belnap, and G. S. Okin. 2011. Responses of Wind Erosion to Climate-Induced Vegetation Changes on the Colorado Plateau. Proceedings of the National Academy of Sciences of the United States of America 108:3854-3859.

Nash, L. L. and P. H. Gleick. 1993. The Colorado River Basin and Climatic Change: the Sensitivity of Streamflow and Water Supply to Variations in Temperature and Precipitation. U.S. Environmental Protection Agency EPA230-R-93-009.

National Center for Atmospheric Research (NCAR). 2012. Atlantic Multidecadal Oscillation (AMO). Online: http://www.cgd.ucar.edu/cas/catalog/climind/AMO.html. Accessed 28 March 2013.

National Drought Mitigation Center. 2013. U.S. Drought Monitor: West March 26, 2013. University of Nebraska-Lincoln. Online: http://droughtmonitor.unl.edu/DM_west.htm. Accessed 28 March 2013.

National Weather Service. 2011. El Niño/Southern Oscillation (ENSO) Diagnostic Discussion. http://www.cpc.ncep.noaa.gov/products/analysis_monitoring/enso_advisory/(Accessed March 16, 2011).

New Mexico Office of the State Engineer, editor. 2006. The Impact of Climate Change on New Mexico's Water Supply and Ability to Manage Water Resources. New Mexico Office of the State Engineer/Interstate Stream Commission, Santa Fe, New Mexico.

NOAA National Climate Data Center. 2013. NCDC Announces Warmest Year on Record for Contiguous U.S. Online: http://www.ncdc.noaa.gov/news/ncdc-announces-warmest-year-record-contiguous-us. Accessed 27 March 2013.

Nydick, K., J. Crawford, M. Bidwell, C. Livensperger, I. Rangwala, and K. Cozetto. 2012. Climate Change Assessment for the San Juan Mountain Regions, Southwestern Colorado, USA: A Review of Scientific Research. Prepared by Mountain Studies Institute in Cooperation with USDA San Juan National Forest Service and USDOI Bureau of Land Management Tres Rios Field Office. Durango, CO. Online: <http://www.mountainstudies.org>.

Pederson, G. T., S. T. Gray, C. A. Woodhouse, J. L. Betancourt, D. B. Fagre, J. S. Littell, E. Watson, B. H. Luckman, and L. J. Graumlich. 2011. The Unusual Nature of Recent Snowpack Declines in the North American Cordillera. Science 333:332-335.

Rango, A. and J. Martinec. 2008. Predictions for Snow Cover, Glaciers and Runoff in a Changing Climate.in HydroPredict 2008, Prague, Czech Republic, 15-18 September 2008.

Rangwala, I. and J. R. Miller. 2010. Twentieth Century Temperature Trends in Colorado's San Juan Mountains. Arctic Antarctic and Alpine Research 42:89-97.

Ray, A. J., J. J. Barsugli, K. B. Averyt, K. Wolter, M. Moerling, N. Doesken, B. Udall, and R. S. Webb. 2008. Climate Change in Colorado: A Synthesis to Support Water Resources Management and Adaptation. A Report for the Colorado Water Conservation Board, Boulder, Colorado.

Saunders, S. and M. Maxwell. 2005. Less Snow, Less Water: Climate Disruption in the West. Rocky Mountain Climate Organization, Louisville, CO.

Seager, R. and N. Naik. 2012. A Mechanisms-Based Approach to Detecting Recent Anthropogenic Hydroclimate Change. Journal of Climate 25:236-261.

Seager, R., M. F. Ting, I. Held, Y. Kushnir, J. Lu, G. Vecchi, H. P. Huang, N. Harnik, A. Leetmaa, N. C. Lau, C. H. Li, J. Velez, and N. Naik. 2007. Model Projections of an Imminent Transition to a More Arid Climate in Southwestern North America. Science 316:1181-1184.

Seager, R. and G. A. Vecchi. 2010. Greenhouse Warming and the 21st Century Hydroclimate of Southwestern North America. Proceedings of the National Academy of Sciences of the United States of America 107:21277-21282.

Seidel, D. J., Q. Fu, W. J. Randel, and T. J. Reichler. 2008. Widening of the Tropical Belt in a Changing Climate. Nature Geoscience 1:21-24.

Seth, A., S. A. Rauscher, M. Rojas, A. Giannini, and S. J. Camargo. 2011. Enhanced Spring Convective Barrier for Monsoons in a Warmer World? Climatic Change 104:403-414.

Sheppard, P. R., A. C. Comrie, G. D. Packin, K. Angersbach, and M. K. Hughes. 2002. The Climate of the US Southwest. Climate Research 21:219-238.

Stewart, I. T., D. R. Cayan, and M. D. Dettinger. 2004. Changes in snowmelt runoff timing in western North America under a “business as usual” climate change scenario. Climatic Change 62:217-232.

______. 2005. Changes Toward Earlier Streamflow Timing Across Western North America. Journal of Climate 18:1136-1155.

Swetnam, T. W. and J. L. Betancourt. 1998. Mesoscale Disturbance and Ecological Response to Decadal Climatic Variability in the American Southwest. Journal of Climate 11:3128-3147.

Tebaldi, C., D. Adams-Smith, and N. Heller. 2012. The Heat is On: U.S. Temperature Trends. Palo Alto, California.

U.S. Global Change Research Program (USGCRP). 2009. Global Climate Change Impacts in the United States. Cambridge University Press, Cambridge, United Kingdom.

______. 2013. Draft Third National Climate Assessment Report.

University of Arizona, the Western Regional Climate Center/Desert Research Institute, and the PRISM Climate Group at Oregon State University. 2007. Climate Analysis & Mapping Toolbox. Online: <http://www.cefa.dri.edu/Westmap/Westmap_home.php>

Veatch, W., P. D. Brooks, J. R. Gustafson, and N. P. Molotch. 2009. Quantifying the Effects of Forest Canopy Cover on Net Snow Accumulation at a Continental, Mid-Latitude Site. Ecohydrology 2:115-128.

Vecchi, G. A. and A. T. Wittenberg. 2010. El Nino and Our Future Climate: Where do We Stand? Wiley Interdisciplinary Reviews-Climate Change 1:260-270.

Wehner, M., D. R. Easterling, J. H. Lawrimore, R. R. Heim, R. S. Vose, and B. D. Santer. 2011. Projections of Future Drought in the Continental United States and Mexico. Journal of Hydrometeorology 12:1359-1377.

Williams, A. P., C. D. Allen, C. I. Millar, T. W. Swetnam, J. Michaelsen, C. J. Still, and S. W. Leavitt. 2010. Forest Responses to Increasing Aridity and Warmth in the Southwestern United States. Proceedings of the National Academy of Sciences.

Woodhouse, C. A., D. M. Meko, G. M. MacDonald, D. W. Stahle, and E. R. Cooke. 2010. A 1,200-Year Perspective of 21st Century Drought in Southwestern North America. Proceedings of the National Academy of Sciences of the United States of America 107:21283-21288.