Monitoring Climate Change in the Rio Grande Basin of New Mexico and Colorado above Elephant Butte Reservoir, New Mexico:
Baseline Report

November 2013

Ariane O. Pinson

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USACE Flood Risk Management Program and the
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Executive Summary

The response of species, habitats, and ecosystems, as well as of surface water and ground water resources to regional climate change is dependent in part on how rapidly that change occurs. Thus, it is essential to understand how fast temperatures and precipitation are changing so that these factors can be accurately represented in project planning. This report addresses the critical question:

How fast is climate change occurring in the Upper Rio Grande relative to model projections?

The report provides an over view of the current regional climate and its major drivers, and then goes on to review observed and projected climate change for the west as whole and the Southwestern United States (SWUS) in particular.

This is followed by an analysis of regional temperature and precipitation trends derived from temperature and precipitation data from the National Weather Service Cooperative Observer Program (COOP), National Oceanic and Atmospheric Administration National Climate Data Center Historical Climatology Network (HCN) and the National Resources Conservation Service Snow Telemetry (SNOTEL) stations within the URG.

For the entire Upper Rio Grande study area, temperatures increased substantially over the four decade period 1971-2012. Average annual temperatures (Tavg) increased at a rate of 0.35°C (0.63°F) per decade, with a faster increase in nighttime minimum temperature (Tmin) of 0.37°C (0.67°F) per decade offset by a slower increase in daytime high temperature (Tmax) of 0.25°C (0.45°F) per decade. Precipitation was unchanged at the regional scale.

Mountain and valley regions responded differently to warming. Mountain areas, particularly the Tusas Mountains, saw large increases in nighttime minimum temperatures (Tmin) of 0.67°C (1.21°F) per decade that were significant in every month but February; daytime high temperatures (Tmax) rose at the slower rate of 0.14°C (0.25°F) decade. Tmin increases were significant throughout the spring and summer (April through September). By contrast, the rates of increase in valley Tmin (0.28°C (0.50°F) per decade) and Tmax (0.34°C (0.61°F) per decade) were similar to each other, with the greatest increases occurring in the months of May through September. March and November were months of rapid change across most of the region, notable for a significant increasing trend in overnight temperatures and a significant decreasing trend in precipitation. The increasing temperatures and decreasing precipitation in March and November are important because these contribute to a longer growing season and decreased period of snowpack accumulation in winter months.

The rate of temperature change (°C/decade) was not constant over the period 1971-2012. In the first 30 years of this period, 1971-2000, positive rates of change in Tmax, Tmin, and, therefore, Tavg occurred across mountain sites, valley sites, and the region as a whole. The rate of increase in Tmin was larger than the gains Tmax for both mountains and valleys.

In the 11 years beginning in 2001, the trend in Tmax (-0.13°C (-.23°F) per decade) and Tmin (-0.38°C (-0.68°F) per decade) has been negative in valley areas. By contrast, mountain regions have been characterized by accelerated increase in rates of warming: Tmax rose from 0.17°C (0.31°F) per decade to 0.39°C (0.70°F) per decade while the rate of increase in Tmin went from 0.62°C (1.12°F) per decade over 1971-2000 to 1.75°C (3.15°F) per decade over the period 2001-2012. It is not immediately clear what is driving these changes in landscape response with topographic position.

Comparison of observed trends with model projections provides a means of assessing the significance of current rates of change, should they continue, with respect to responses of the natural environment. The rates of future change in stream flow and vegetation models are dependent on the rates of change in the climate model(s) driving them.

The observed trends in annual temperature are compared to trends projected by models for areas encompassing the Upper Rio Grande. If temperatures in the Upper Rio Grande basin continue to rise at the rate of the forty year period 1971-2012, average net warming for the period 2010-2039 would be 0.86°C (1.55°F) above the last decades of the 20th century; net warming by 2050 would be 1.94°C (3.49°F) above the last decades of the 20th century. This is the second highest observed rate of change among published studies. Observed rates of change, when multiplied out, are approximately in the middle of the range of model estimates of future warming, reaching approximately 1.75°C (3.15°F) by 2050 and 3.5°C (6.3°F) by 2100.

The observed regional trend is in line with the most recent NARCCAP model projections used in the 2013 National Climate Assessment (USGCRP 2013). These models project that the Upper Rio Grande area will warm 4.1-4.9°C (7.5-8.5°F) by 2070-2099 under the A2 (high emissions) scenario and 2.5-3.1°C (4.5-5.5°F) by 2070-2099 under the B1 (low emissions) scenario.

The rate of change in temperature is likely to result in rapid changes to such dependent conditions as wildfire frequency, vegetation communities, soil moisture, and both surface and ground water quantity and quality. The data suggest that taking climate change projections into account in the design and operation of U.S. Army Corps of Engineers, Albuquerque District projects is likely to be both valuable and cost-effective in the long run.

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I. Introduction

Most climate models currently project temperature increases at the Earth's surface by the year 2100 will average 3-4°C above the average temperature for the period 1950-1999, and that these temperatures will continue to increase in the following decades. The amount of warming depends on the rate of greenhouse gas (GHG) release into the atmosphere by human (anthropogenic) and natural sources above what can be removed by natural processes and, in the future, but novel technologies. Radical reductions in GHG emissions could restrain the temperature increase to about 2°C by 2100; continued accelerating emissions, potentially accompanied by changes in Earth surface processes could result temperature increases closer to 6°C by 2100, with more beyond (IPCC 2007b).

Although alarming, the warming itself is much less of a problem than the major consequences of this warming for the Southwestern United States (SWUS), as outlined in the recent *Draft Third National Climate Assessment* (USGCRP 2013):

- Continental interior regions such as the SWUS are anticipated to warm at a rate that is faster than the global average. Therefore, it is not surprising that many models anticipate SWUS temperatures will increase 1-2°C before mid-century.
- The SWUS is anticipated to experience declines in precipitation across most of the year, although this is not seen in all models (NOAA 2013a).
- Changes are expected to occur in precipitation intensity: precipitation is expected to fall in larger storms interspersed with longer dry spells. Winter and summer storms may become more severe.
- Increases in evaporation and plant transpiration (evapotranspiration) are expected to scale with temperature. Longer growing seasons will translate into higher overall water demand among all users. Some plants may reduce transpiration in response to increases in atmospheric carbon, but this phenomenon is poorly understood. Precipitation increases may not be sufficient in many areas to offset increases in evapotranspiration, leading to a net soil moisture loss.
- Increased evaporation is expected to exceed any increases in temperature, resulting in a shift to a more arid climate characterized by more frequent and longer drought.
- The increases in temperature, and the resulting rise in evapotranspiration rates over a longer growing season, are anticipated to lead to net declines in soil moisture, favoring increasing wildfire frequency and changes in vegetation.
- Shorter, warmer winters are expected to result in a smaller share of winter precipitation falling as snow, coupled with earlier spring runoff (therefore earlier peak river flows, in advance of peak irrigation needs), and the decline and loss of snowfields (reduced late summer runoff and river flows).
- Changes are expected to phenology (the seasonal timing of reproduction among species in response to environmental cues, affecting offspring survival) for both plant and animal species. These changes will occur on a species-by-species basis, and have the potential to disrupt food webs.
- Changes are expected in the distribution of plant and animal species, both terrestrial and aquatic. Changes to migration patterns and to the availability of food and habitat along migration routes are also expected.

The Arctic and subarctic regions are warming at a rate that is at least double that of the world as a whole. The loss of Arctic Sea ice in particular is starting to produce observable changes in midlatitude atmospheric circulation that is likely to affect fall and winter climates in the SWUS. In addition, warming in the tropics, leading to expansion of the subtropical dry zone, may impact summer precipitation. It is unclear if warmer sea surface temperatures will result in a stronger late summer monsoon over the SWUS, or how these changes will affect the El Niño-Southern Oscillation (ENSO), the changes in Pacific Ocean sea surface temperatures that can bring anomalously wet or dry conditions to the SWUS.

Climate change in the SWUS is likely to have profound effects on runoff and stream flow in general and in the Rio Grande basin specifically. Reductions in stream flow will pose special challenges for water resource managers, who must balance competing human uses against the demands of ecosystems and endangered species. Changes to vegetation, soils, and wildfire frequency in the catchment are likely to alter established rainfall-runoff relationships in ways that researchers are only beginning to appreciate.

A. Goals

Given projections of an imminent transition to drier climates in the SWUS (Seager et al. 2007), resource management agencies face the great challenge of planning projects under highly divergent model projections of future climate. Observational data on regional climate trends provides a guide for how fast climate is currently changing, but doesn't give a clear picture of what the region's climate is likely to be in the future.

This study addresses the gap between models and observations by comparing observed trends in temperature and precipitation in the Upper Rio Grande (URG), that portion of the Rio Grande Basin above Elephant Butte Dam, with model projections to answer the question:

How fast is climate change occurring in the Upper Rio Grande relative to model projections?

The answer to this question provides guidance for prioritizing model projections by their likelihood that projections accurately describe the magnitude of future climate change relative to the present. This will enable a better understanding of the likely impacts climate change will have over the lifetime of existing and future ecosystem restoration, water operations, flood risk management and other U.S. Army Corps of Engineers (USACE), Albuquerque District (SPA) projects.

B. About the Study Area

The study area is the Rio Grande Basin above Elephant Butte Reservoir. The Rio Grande rises in southern Colorado, in the San Juan Mountains near Creede. The higher reaches of the watershed extend above 11,000 feet above sea level (asl); conifer forests and grassy alpine meadows dominate the landscape above about 8,000 ft asl. From the headwaters, the river drains southeast to Alamosa, Colorado, before adopting a generally south-trending course through New Mexico to El Paso, Texas, then turning southeast to the Gulf of Mexico. Along the way, the river and its tributaries descend through mixed conifer forest and conifer woodland to valley floors characterized by grassland and desert scrublands. South of Socorro, New Mexico, vegetation gives way to Semidesert Grassland and, further south, grades into Chihuahuan Desertscrub (Brown et al. 1998). Approximately 75% of the river flow above El Paso originates in the San Juan Mountains of Colorado, and in the Sangre de Cristo and Jemez Mountains of New Mexico, all of which are upstream of Elephant Butte Dam. Elephant Butte Dam is located approximately 40 miles below the confluence with the Rio Salado, last major tributary of the Rio Grande in New Mexico, and is north of Las Cruces, New Mexico, and the Mesilla Valley agricultural region.

C. Authority and Support

Authority to undertake write this report derives from the determination by Jo-Ellen Darcy, Assistant Secretary of the Army (Civil Works), that a proactive approach to assessing the risks posed by climate change is essential for USACE to find ways to mitigate or adapt to a changed climate (Darcy 2010):

> *The US Army Corps of Engineers (Corps) recognizes that the entire portfolio of our structural and nonstructural water resources projects will be affected by climate change, necessitating not only mitigation to climate change, but adaptation as well...[The] Corps is responding to water-related risks posed by climate change to water resources infrastructure, including risk and vulnerability assessments, identification of potential adaptation strategies, and collaborative efforts supporting climate change adaptation.*

This project received funding under the USACE Middle Rio Grande Collaborative Program Authority (Section 106, Omnibus Appropriations Act of 2009 (P.L. 111–8)) because the project contributes to fulfilling Reasonable and Prudent Alternatives (RPA) S, T and EE outlined in the U.S. Fish and Wildlife Service (USFWS) 2003 Biological Opinion covering USACE and U.S. Bureau of Reclamation (Reclamation) water operations on the Rio Grande (USFWS, 2003, Consultation #2-22-03-F-0219):

- RPA Element S directs USACE and Reclamation to construct habitat/ecosystem restoration projects in the Middle Rio Grande [the Rio Grande between Cochiti Dam and Elephant Butte Dam] that are depletion neutral. Such projects should create backwaters, oxbows, shallow water habitats, overbank flooding, and stands of native vegetation. Information developed by this project will contribute to RPA Element S by enabling more accurate estimate of trends in climate factors that influence water quality and quantity in the Middle Rio Grande, and therefore affect the abundance and quality of habitat for endangered species in the region.
- RPA Element T directs USACE and Reclamation to construct habitat restoration features to offset adverse environmental impacts resulting from river operation by USACE and Reclamation. Understanding trends in river flows on an annual and seasonal basis is a critical element of restoration projects sensitive to stream discharge (e.g., high and low flow channels, overbanking, native vegetation establishment on destabilized surfaces). River flow is strongly tied to climate factors, such as temperature and precipitation in the river catchment, that determine the timing and volume of runoff. Information collected by this project would shed light on trends on climate factors that influence stream flow in the Middle Rio Grande in terms of the absolute and seasonal abundance of water available for habitat restoration.
- RPA Element EE directs USACE and Reclamation to monitor water quality impacts on the silvery minnow. The major part of this is identifying and monitoring pollution. However, a key component of water quality is water abundance: how much water is available to dilute the pollutant load. By tracking trends in regional climate, this project contributes to RPA Element EE through identifying trends in climate that influence river water abundance on an annual and seasonal basis.

This project contributes to the recovery of the Southwestern Willow Flycatcher (flycatcher) by providing additional information necessary for planning and implementing projects that increase and improve occupied, suitable and potential breeding habitat (USFWS, 2002: Action 1). Temperature, precipitation, and storminess, among other climate variables, directly influence river levels and indirectly influence the hydrologic conditions essential to establishing and maintaining suitable flycatcher habitat. Data on current climate changes trends will help biologists understand how the spatial distribution of prime flycatcher habitat should change in response, and should impact decisions about where to place future restoration features.

Likewise, this project contributes to the recovery of the Rio Grande silvery minnow (minnow) by providing essential information for planning and implementing projects that support 3 of the 5 recovery actions identified by the USFWS (1999, 2007):

- Action 2: Restore, protect, and alter habitats as necessary to alleviate threats to the Rio Grande silvery minnow.
- Action 3. Ensure the survival of the Rio Grande silvery minnow in its current habitat and reestablish the species in suitable habitats within its historical range.
- Action 4. Implement and maintain an adaptive management program so that appropriate research and management activities are implemented in a timely manner to achieve recovery of the Rio Grande silvery minnow.

Flows in the Middle Rio Grande are particularly sensitive to changes in the winter precipitation and the snowpack, which has already started to exhibit long-term declines in volume and advances in the timing of melt. If these changes persist, changes to the geomorphology of the river and the volume of its flows will be significantly affected, leading to changes in the distribution of habitat and suitable flows for minnow life stages. Thus, data on current climate trends (spatial and temporal) are essential to developing plans to alleviate habitat threats, selecting places on the river at which to restore minnow populations, and formulating and executing adaptive management plans.

Funding for this project has also been provided by the USACE Flood Risk Management Program. Projected shifts to a warmer and more arid regional climate are anticipated to raise evapotranspiration rates and decrease soil moisture (see Section IV). Warming is also anticipated to lead to lead to larger, more frequent wildfires and larger, more frequent outbreaks of bark beetles and other pathogens. These changes typically lead to declines in soil infiltration that produce greater runoff and erosion from storms of a given size. Furthermore, under a warming climate, precipitation is anticipated to concentrate in fewer, larger storms even though average annual precipitation may not change. Thus, under a warmer climate, declining base flows are likely to be accompanied by larger flash flood flows resulting from fewer, larger precipitation events falling on slopes with less organic matter in the soil and less canopy cover. Fewer and larger floods are the likely outcome, making climate change an important factor in planning for flood risk management.

Lastly, for all the reasons listed above, stream hydrographs, flood frequency, and historical rainfall-runoff relationships are all anticipated to need alteration as the regional climate changes. Since this project is the first step towards developing an understanding of these changes, support for this project was also provided by the USACE, Albuquerque District, Operations Division.

D. About this Report

This report was prepared using primary technical literature, public documents, and data available from online sources.

- Chapter 1 introduces the report and the study area.
- Chapter 2 provides an overview of USACE climate change authorities, policy and guidance.

- Chapter 3 summarizes the current climate of New Mexico, paying close attention to the global and regional factors that determine spatial and seasonal patterns of temperature and precipitation. The chapter also covers the key sources of cyclical variation in these climate factors.
- Chapter 4 discusses observed and projected climate parameters in the western United States in general. It covers evidence for recent climate change, and emphasizes the impact of warming on precipitation, evaporation, snowpack, snowmelt and stream runoff.
- Chapter 5 provides an overview of observed and projected climate changes in northern New Mexico and southern Colorado.
- Chapter 6 presents the analysis of observed trends in temperature and precipitation based on climate data from in the URG study area. The chapter also presents the comparison between observed trends with model projections, and discusses the implications of the results.

II. USACE Climate Change Authorities, Policies and Guidance

A. Sea Level Rise

The first USACE climate change-related guidance focused on the impact of sea level rise on coastal areas, beginning with the Water Resources Development Act (WRDA) of 1986 (P.L. 99-662) which states:

> *SEC. 731. STUDY OF RISING OCEANS.*
>
> *(a) The Congress finds that increasing scientific evidence indicates the level of the oceans will rise significantly over the next seventy-five years.*
>
> *(b) The Secretary, in cooperation with the National Oceanic and Atmospheric Administration, the Federal Emergency Management Agency, and other appropriate Federal, State, and local agencies and the private sector, is authorized to conduct a study of shoreline protection and beach erosion control policy and related projects of the Secretary, in view of the prospect for long-term increases in the levels of the ocean. Such study shall include, but is not limited to--*
>
> *(1) an assessment of the probability and the extent of coastal flooding and erosion;*
>
> *(2) an appraisal of various strategies for managing relocation, disinvestment, and reinvestment in coastal communities exposed to coastal flooding and erosion;*
>
> *(3) a summary of the legal and institutional impact of rising sea level on riparian lands; and*
>
> *(4) recommendations for new or additional criteria for Federal participation in shoreline protection projects.*

USACE responded by developing guidance for incorporating sea level change in coastal planning by issuing EC 1165-2-211, Water Resource Policies and Authorities Incorporating Sea-Level Change Considerations in Civil Works Programs (1 July 2009), and in EC 1165-2-212, Sea-Level Change Considerations for Civil Works Programs (1 October 2011). A key point of the sea level guidance is that planning alternatives and engineering designs must be evaluated for the entire range of possible future rates of sea level change, without preference for one, where the "low" level is the historic rate and includes a moderate rate, and then a high rate that is at the very upper range of scientific predictions for sea level change. Estimates of sea level change are adjusted for local conditions (such as isostatic rebound) that affect local rates of sea level change.

B. Climate Change Impacts to Freshwater Resources: Initial Concerns Raised

Although concern for climate change impacts to coastal areas preceded that for interior regions, even before EO 13514, Federal Leadership in Environmental, Energy, and Economic Performance (October 5, 2009), USACE had joined with the U.S. Geological Survey (USGS), US Bureau of Reclamation (Reclamation) and the National Oceanic and Atmospheric Administration (NOAA) to assess the impact of climate change on water resources management (Brekke et al. 2009). Among other things, this document recommended adoption of alternatives that perform well over a wide range of future scenarios and advocated for the use of an adaptive management approach to deal with uncertainty introduced by potential climate change.

C. Executive Order 13514 and Implementing Guidelines

As the evidence for, and impacts of, climate change became more evident in interior regions, the Federal Government moved towards a more-comprehensive approach to addressing climate change impacts on planning. Executive Order (EO) 13514, Section 8 states:

> *Agency Strategic Sustainability Performance Plan. Each agency shall develop, implement, and annually update an integrated Strategic Sustainability Performance Plan that will prioritize agency actions based on lifecycle return on investment. Each agency Plan and update shall be subject to approval by the OMB Director under section 4 of this order. With respect to the period beginning in fiscal year 2011 and continuing through the end of fiscal year 2021, each agency Plan shall: …*
>
> > *(i) evaluate agency climate-change risks and vulnerabilities to manage the effects of climate change on the agency's operations and mission in both the short and long term…*

Section 16 of EO 13514 ordered all Federal agencies (including USACE) to participate actively in the Interagency Climate Change Adaptation Task Force (ICCATF) and "shall develop approaches through which the policies and practices of the agencies can be made compatible with and reinforce [the U.S. national strategy for adaptation to climate change]". The ICCATF is required to issue annual progress reports (CEQ, 2010, CEQ, 2011a). The ICCATF has identified five key areas of Federal adaptation progress:

- Integrating adaptation into Federal government planning and activities.
- Building resilience to climate change in communities.
- Improving accessibility and coordination of science for decision making.
- Developing strategies to safeguard natural resources in a changing climate.
- Enhancing efforts to lead and support international adaptation.

The White House Council on Environmental Quality (CEQ) and the Office of Management and Budget (OMB) jointly oversee Federal agency implementation of EO 13514. They have established implementing instructions for EO 13514 (CEQ and OMB, 2011) for use by Federal agencies in climate change planning. These instructions require the head of each agency to:

- Establish an agency climate change adaptation policy.
- Increase agency understanding of how the climate is changing.
- Apply understanding of climate change to agency mission operations.
- Develop, prioritize and implement actions.
- Evaluate and learn through participation in interagency workshops.

CEQ also published a supporting document to assist Federal agencies with meeting CEQ implementing instructions requirements (CEQ, 2011b). This document lays out the guiding principles for climate change adaptation:

- Adopt integrated approaches.

- Prioritize the most vulnerable.
- Use best-available science; don't delay adaptation until better information is available.
- Build strong partnerships across multiple sectors, geographic scales, and levels of government.
- Apply risk-management methods and tools.
- Apply ecosystem-based approaches because healthy ecosystems build resilience and reduce vulnerability of people and livelihoods to climate change impacts.
- Maximize mutual benefits.
- Continuously evaluate performance through measurable goals and performance metrics. Flexible planning is essential.

D. 2011 National Action Plan: Priorities for Managing Freshwater Resources

The 2010 progress report of ICCATF (CEQ, 2010) identified climate change impacts to freshwater ecosystems as a key issue affecting many different Federal actions and areas of responsibility, and recommended the development of a national action plan to address this issue. The ICCATF issued the *National Action Plan: Priorities for Managing Freshwater Resources in a Changing Climate (National Action Plan)* in 2011 (ICCATF, 2011). The goal for managing freshwater resources in a changing climate is stated as:

> *Government agencies and citizens collaboratively manage freshwater resources in response to a changing climate in order to ensure adequate water supplies, to protect human life, health and property, and to protect water quality and aquatic ecosystems.*

Specific recommendations include:

- Establish a planning process to adapt water resources management to a changing climate, including continuing interagency coordination and expanding outreach to, and collaboration with, State, Tribal, and local governments and other stakeholders.
- Improve water resources and climate change information in decision making, including observational data, predictive models, measures of progress, and regular updates of analyses.
- Strengthen assessment of vulnerability of water resources to climate change, including effective communication of data and development of risk assessment tools at multiple scales.
- Expand water use efficiency, including opportunities to develop water efficient technologies and to promote greater efficiency of water use and reuse.
- Support training and outreach to build response capability, including practices to mainstream climate change adaptation into existing programs.

Under the National Action Plan, USACE bears primary responsibility for three actions under Recommendation 5, Integrated Water Resources Management:

- Supporting Action 17, work with partners to incorporate integrated water resources management into planning and programs with attention to climate-change adaptation issues.
- Supporting Action 19, work closely with states to identify flood risk and drought management "best practices" to prepare for hydrologic extremes.

- Supporting Action 20, develop benchmarks for incorporating adaptive management into water project designs, operational procedures, and planning strategies.

In addition, USACE has subsidiary responsibilities for supporting actions under other Recommendations. USACE is co-lead with NOAA for Supporting Action 6, provide coastal states and communities with essential information to identify areas likely to be inundated by sea level rise. In addition, both agencies are responsible for leading efforts under Supporting Action 9, develop a federal internet portal to provide current, relevant, and high quality information on water resources and climate change. USACE is also one of several agencies charged with leading Supporting Action 16, enhance coordination among Federal water efficiency programs and improve program effectiveness, including creating a “toolbox” of key practices.

E. Current USACE Climate Change Adaptation Guidance

The CEQ Support Document (CEQ, 2011b) mandates that each Federal agency issue an agency-wide climate change adaptation policy statement committing the agency to adaptation planning to address challenges posed by climate change risks to the agency’s mission, programs, and operations. The USACE climate change adaptation policy (Darcy 2010) requires consideration of the effects of climate change at every step in the lifecycle for all USACE projects, both existing and planned. The goal is to reduce vulnerabilities and enhance the resilience of the Nation’s water resources infrastructure using the best currently available, actionable science, and be updating the plans as better information becomes available. Actions that integrate adaptation (managing the unavoidable impacts) and mitigation (avoiding the unmanageable impacts) are desirable (USACE, 2012).

The most recent USACE climate change and adaptation report (USACE, 2012) identifies six climate change adaptation priority areas for USACE:

- Integrated Water Resources Management.
- Risk-Informed Decision-Making for Climate Change: A risk-informed decision-making framework is anticipated to be used to address the entire project life cycle so that there is flexibility to respond as new information becomes available. The assessment would include both consequence and likelihood assessments, and emphasizes the formulation of risk management alternatives under changing conditions. Stakeholder involvement throughout the decision process is deemed critical. “The risk management framework will be a foundation for developing strategies to incorporate climate change into the decision making processes of USACE, with FY 12 and FY13 priorities being ecosystem restoration, flood risk management, and water management” (USACE, 2012:8).
- Nonstationarity: Nonstationarity refers to the idea that future climate conditions will not be like the past and therefore the historical record of event frequency and magnitude alone are insufficient for characterizing future conditions in a changing system. Developing methods and procedures to address nonstationarity throughout the project life cycle is a priority action for USACE. In part to address the issue of nonstationarity, the guidance for determining flood flow frequency, Bulletin 17B, Guidelines for Determining Flood Flow Frequency (U.S. Interagency Advisory Committee on Water Data) is currently being revised. USACE is also collaborating with other agencies to develop statistical methods for addressing nonstationarity.
- Portfolio of Approaches: USACE is investigating a portfolio of approaches to adapting to and mitigating the consequences of climate change. Some of these are in the testing phase.
- Continued Vulnerability Assessments: Formal vulnerability assessments complement top-down approaches to climate adaptation/mitigation planning by determining the level of event at which

project damage might occur, then determining if this level of event is likely to occur within the project lifetime given a changing climate. USACE is currently involved in a “screening-level” vulnerability assessment at the HUC-4 watershed scale, with plans to refine method and scale going forward.

- Metrics and Endpoints: USACE is also working on developing appropriate metrics and endpoints for assessing the efficiency and effectiveness of climate change adaptation activities, to ensure solutions are practical, nationally consistent, legally justifiable and cost effective, as well as considering both structural and nonstructural alternatives. Methods for measuring both costs and benefits are in their infancy.

Finally, USACE has established a new guidance series, beginning 31 December 2011: *Series 1100, Global Changes* (see OM 25-1-51 (31 December 2011)) that addresses new challenges for USACE projects, including demographic shifts, changing land use, climate change, sea level variability, increasing State capabilities, aging infrastructure, disappearing wetlands, water availability and changing social values and economic considerations.

In addition, there are three recent documents of relevance to climate change planning at the District level:

- A recent study (Brekke 2011) has outlined a prioritized list of long-term water resources planning and management needs, identifying user needs and 39 information gaps grouped into eight categories reflecting typical steps in the planning process. Most of the needs were prioritized as high or medium priority, reflecting the general complexity of the problem and the overall lack of certainty in how to address it.
- A draft USACE report titled *Risk Informed Decision Making for Climate Change* (Harper n.d.) outlines risk assessment approaches, including consequence or opportunity assessment and probability assessment, and describes qualitative and quantitative techniques for risk analysis and how to choose and evaluate an approach. It also describes risk evaluation, climate change drivers, and other factors in choosing alternatives given climate uncertainty.
- This report responds to the requirements of Section 9506 of the Omnibus Public Lands Act (P.L. 111-11; Appendix A) for a report to Congress describing the impact of global climate change on freshwater resources, identifies actions that could improve the Nation’s ability to detect and predict changes to freshwater resources that may result from climate change (Federal Interagency Panel on Climate Change and Water Data and Information 2011).

F. Executive Order "Preparing the United States for the Impacts of Climate Change"

President Barack Obama issued the Executive Order "Preparing the United States for the Impacts of Climate Change" on November 1, 2013. The EO states: “The Federal Government must build on recent progress and pursue new strategies to improve the Nation's preparedness and resilience. In doing so, agencies should promote: (1) engaged and strong partnerships and information sharing at all levels of government; (2) risk-informed decision making and the tools to facilitate it; (3) adaptive learning, in which experiences serve as opportunities to inform and adjust future actions; and (4) preparedness planning.”

Further, the heads of Federal agencies, including USACE, are tasked within 9 months of this EO, with completing an inventory and assessment of proposed and completed changes to their land- and water-related policies, programs, and regulations necessary to make the Nation's watersheds, natural resources, and ecosystems, and the communities and economies that depend on them, more resilient in the face of a changing climate. As appropriate, agencies should focus on program and policy adjustments that promote the dual goals of greater climate resilience and carbon sequestration, or other reductions to the sources of climate change. USACE is named specifically as a member of the new interagency Council on Climate Preparedness and Resilience that will replace the existing Interagency Climate Change Adaptation Task Force.

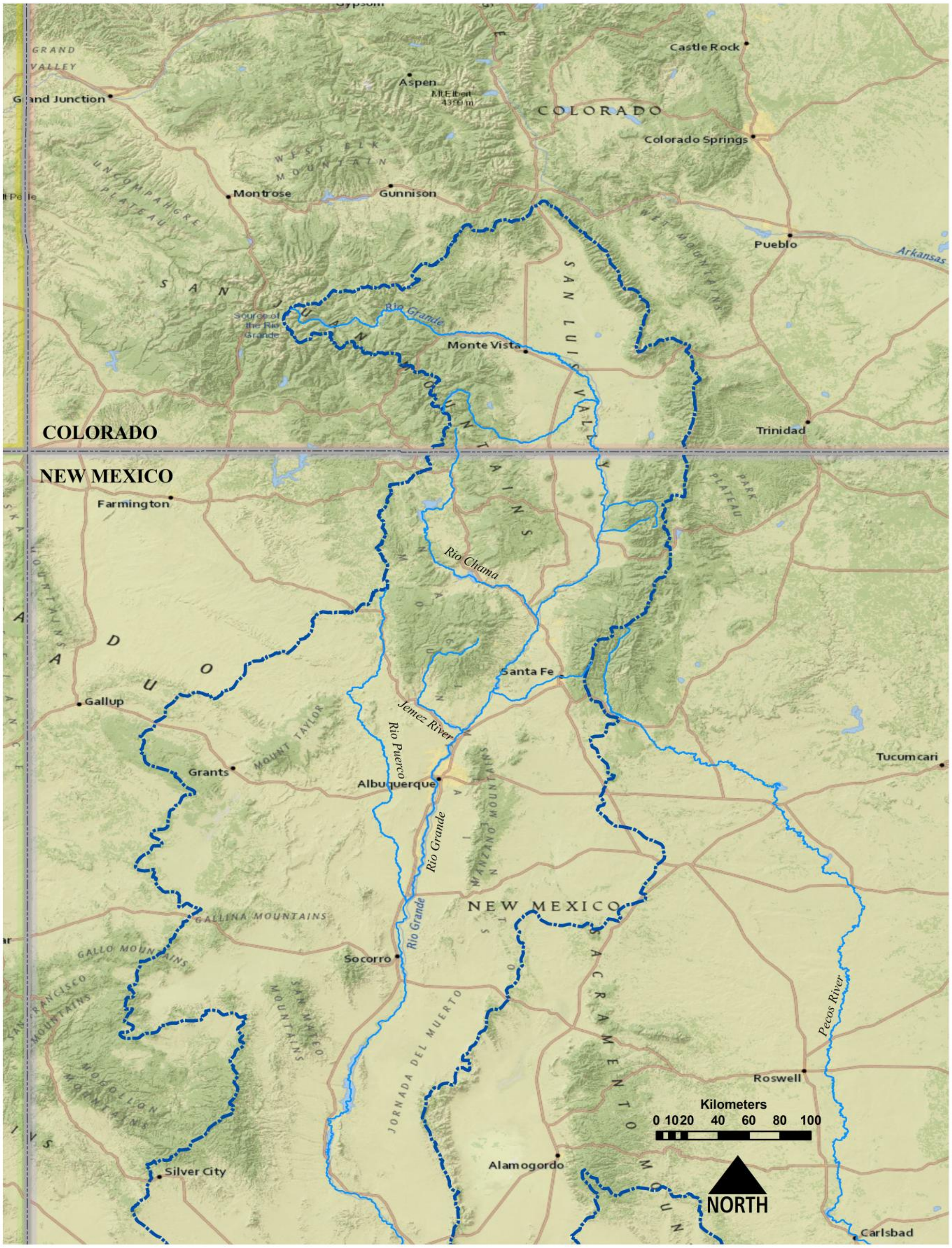

Figure 1:Map showing the location of the Rio Grande Basin in New Mexico and Colorado (blue line).

III. Current Climate of New Mexico

Atmospheric warming over the next century will have significant impacts on the Southwestern landscape. However, these impacts will vary by place and season in ways governed by the large scale features of the modern climate system. Therefore, an understanding of how climate change will manifest in the State of New Mexico (Figure 1) must rest on an understanding of how the modern climate system works. This section of the report provides a brief overview of New Mexico's climate, its major features and drivers (this dicussion is based on Sheppard et al. 2002).

New Mexico is classified as an arid climate, with an average annual precipitation approximately 13.4". That amount is greater (approaching 20") in northern areas and at higher elevations, and significantly less in southern areas and at lower elevations. 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 can be classified as warm. Average daily low temperatures in winter are 20°F (-7°C) while daily summer temperatures average 80 to 95°F (27 to 35°C). Again, areas to the north and at higher elevation are cooler year-round and areas to the south and at lower elevation are warmer year-round.

The basic pattern of New Mexico's climate is driven by its subtropical latitude and its position in the continental interior (Figure 2). In summer, New Mexico sits at the northern edge of the subtropics, a latitudinal zone located between approximately 20° and 40° north of the equator. Solar heating of Earth's surface along the equator causes humid air in this region to rise and to drop its moisture as rain. A portion of this air migrates northward, and eventually descends over the subtropics. As it descends, it warms and its capacity to retain moisture increases (pulling moisture out of the environment as the air mass descends)[1]. New Mexico's summer is therefore hot and largely dry.

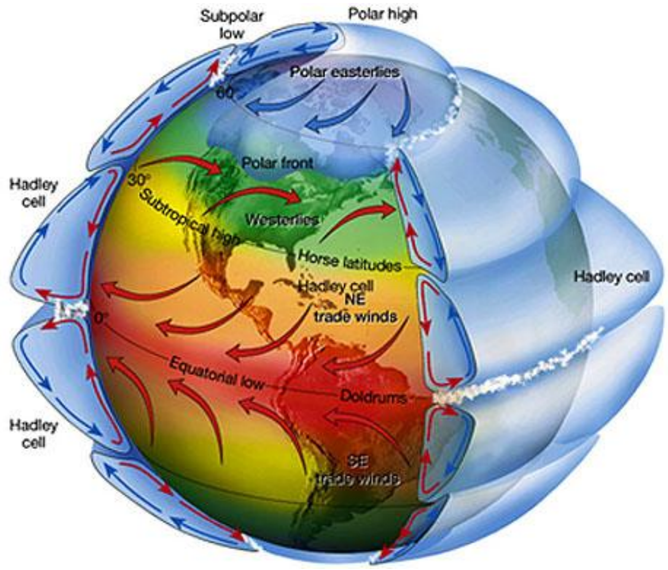

Figure 2: Map of atmospheric circulation (source: SERC/Carleton College).

In winter, the area of maximum heating shifts south of the equator, which causes the zone of subtropical descending air to shift south as well. In New Mexico, this shift allows the southern movement of the temperate climate zone, bringing New Mexico into the reach of the jet stream and the alternating pattern of high pressure (clear) and low pressure (storm) systems that the jet stream drives west to east across mid-latitude North America. The weather in New Mexico in winter resembles that of Wyoming and Montana, although it is warmer and drier.

In addition to its latitudinal position, New Mexico also sits in the interior of North America: it is surrounded by dry land and is

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[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.

distant from warm oceans. Being in the continental interior limits how much moisture is available for precipitation. 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.

A. Winter Climate

Although winter brings mid-latitude, temperate climate to New Mexico, the state is along the southern margin of this climate zone. Consequently, most storms bring cold air, wind, and clouds, but little precipitation. Some storms penetrate farther south, bringing widespread soaking rain or snow to lower elevations and, as demonstrated in February 2011, occasional subfreezing temperatures. Precipitation from these storms may continue on and off over a 24 hour or longer period and may impact a large region.

Winter precipitation varies from year to year depending on the temperature of the ocean surface to the southwest and west. 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 Albuquerque District.

The most familiar variation in ocean temperature (and in the overlying atmosphere) is the El Niño-Southern Oscillation (ENSO) cycle (Figure 3). In an 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.

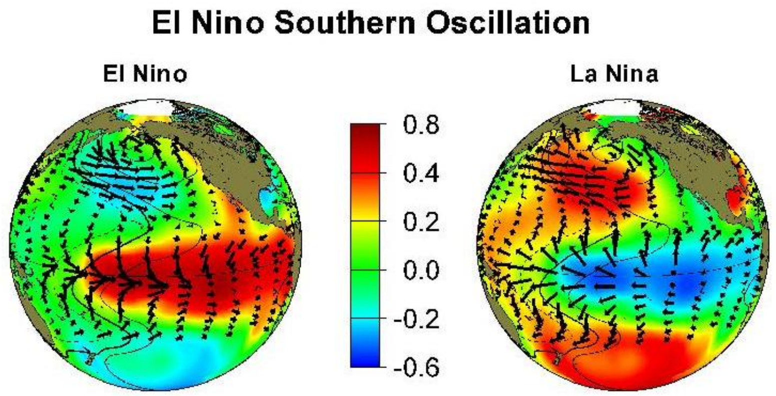

Figure 3: Changes in sea surface temperature in the Pacific as the result of ENSO cycles (source: JISAO/University of Washington).

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.

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[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 is 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 (50-80 year) cycles, and which can serve to amplify or dampen the different phases of the ENSO cycle.

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

B. Summer Climate

Summer precipitation in New Mexico has a very different origin than winter precipitation. In summer, the mid-latitude storm track migrates northward and New Mexico falls under the influence of subtropical climates. This is most evident in the hot, dry period April through June. Temperatures remain elevated through September, but in many years are moderated by the development of the North American Monsoon. A monsoon is defined as a seasonal reversal of wind and air movement of at least 120°. Whereas in winter, most weather descends on the region from the northwest, in the summer the flow is predominantly from the south.

Figure 4: Moisture sources for the North American Monsoon (source: NOAA).

Summer precipitation in the Albuquerque District is driven entirely by sunlight falling on Earth's surface, which heats the overlying air by conduction, and then the heated air rises. As this convection cycle gets repeated day after day, this eventually draws in moist air from over the eastern Pacific and Gulf of Mexico (Figure 4). As this air is drawn into this daily cycle of heating and rising, the rising air cools, water condenses from it and falls as rain over the Southwest. The monsoon onset is time-transgressive, beginning mid-June in southwestern New Mexico, and in mid-July in the

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[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/ (Accessed March 16, 2011).

Upper Rio Grande.

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 high mountain areas experiencing daily downpours. Lower elevation areas will tend to see less mid-day precipitation but more evening precipitation, and there will be greater day-to-day and place-to-place variation in precipitation.

The strength of the monsoonal rains depends very much on: 1) how hot the Southwest gets (how much heat is available to drive air convection); 2) 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; and, 3) related to point 2, 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 altitudes.

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

IV. Literature Review: Observed and Projected Climate Change in the Western United States

Any discussion of projected impacts of climate change in the Rio Grande basin must take into account the range of future climate projections. Because future greenhouse gas emissions are dependent on future economic, demographic, technological, political and other considerations, any projection of future runoff must account for these factors. This chapter begins with a discussion of future climate scenarios typically used in climate change models. It then goes on to discuss impacts to the West in general and New Mexico in particular of projected climate changes given different future atmospheric greenhouse gas concentrations.

A. Projecting Climate Warming under Different Economic Scenarios

Projecting climate change is a complex, multi-step process involving:

- Construction of a coarse-resolution, global-scale model of Earth's climate system (atmospheric general circulation model or atmospheric GCM). Because of the tremendous influence of sea surface temperatures on climate, to more "realistically" mimic the climate system, the atmospheric GCM has to be coupled to a model of Earth's ocean circulation (oceanic GCM), producing what is known as a "coupled" model (coupled GCM). The completed model is refined by comparison with modern climate systems (e.g., "forcing[5]" the model with El Niño conditions to see if it reproduces known El Niño climate patterns), the historic instrumental record, and other records of past climate change (e.g., the tree ring record of past southwestern climates, which extends back at least 1,200 years (Cook et al. 2004, Woodhouse et al. 2010)).
- Construction of models of future global economic, social, demographic, and technological change to project future carbon emissions. These are referred to as scenarios. These scenarios were developed by the IPCC in its *Special Report on Emissions Scenarios* (IPCC 2000) to use as common climate model inputs for both comparing how robust model predictions are, and assessing how sensitive climate will be to different carbon emissions levels. The SRES scenarios used for modeling are described in Figure 5.
- The coupled GCM is then forced with the carbon dioxide (and other) emissions of one of the IPCC scenarios and run until the model stabilizes. This yields a projection for future climate at a particular point in time (e.g., the mean value for the period 2060-2079) under a particular scenario. Because there are stochastic (random) elements to climate, the model is run a number of times, and the reported value is the average of all the model runs. Some aspects of projected climate change recur across model runs, and are considered "likely" or "very likely"; some aspects of projected climate change differ among model runs and are considered "less likely" (less certain).
- For larger scale assessments, model ensembles are used and the results from multiple runs of climate models for a given scenario are grouped together and treated as a single data set.
- No matter what the baseline period is, the baseline data are the data simulated by the model and evaluated against the historical record. The model is optimized so that its output best represents

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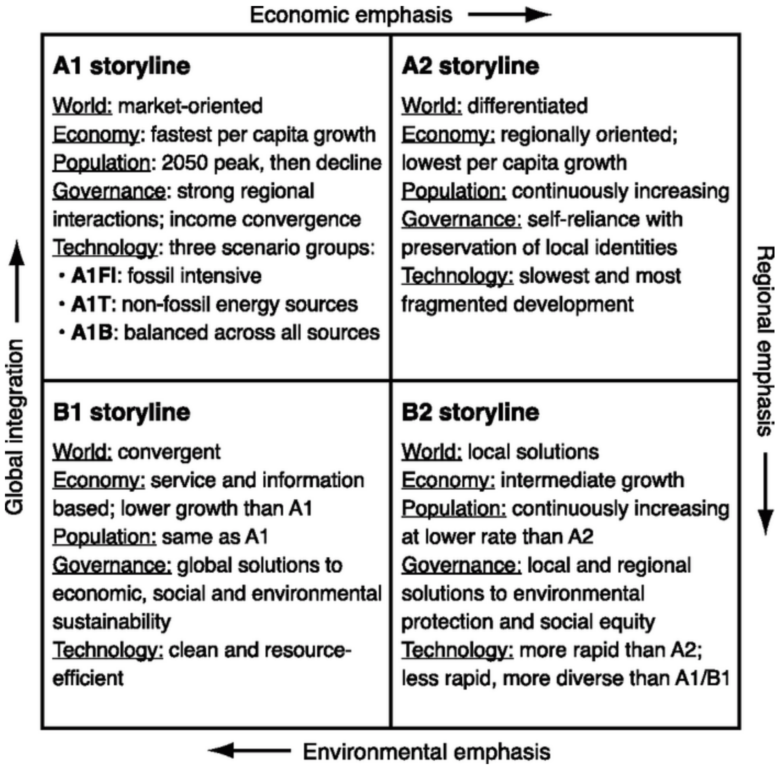
[5] In this situation, forcing a model refers to altering the input parameters to see how the model responds. In this instance, altering the pattern of sea surface temperatures in the Pacific enables researchers to determine whether there are appropriate corresponding changes to atmospheric circulation, surface temperature, precipitation, and so forth.

the climate record for a given historical period forms the baseline data against which model projections are compared.

- Computing power limits the spatial resolution of coupled GCMs. To get a more detailed representation (for instance, in areas of significant differences of relief, such as the Southwest), a finer-resolution, local model has to be tucked into the coupled model so that the larger-scale model outputs for that area of the world become inputs to the local scale model. This process is known as “downscaling”. Downscaling is essential to modeling local and regional landscape responses to climate change [such as the soil moisture changes captured using the Variable Infiltration Capacity (VIC) model (Gao et al. 2009)].



A1 storyline
World: market-oriented
Economy: fastest per capita growth
Population: 2050 peak, then decline
Governance: strong regional interactions; income convergence
Technology: three scenario groups:
- **A1FI**: fossil intensive
- **A1T**: non-fossil energy sources
- **A1B**: balanced across all sources

A2 storyline
World: differentiated
Economy: regionally oriented; lowest per capita growth
Population: continuously increasing
Governance: self-reliance with preservation of local identities
Technology: slowest and most fragmented development

B1 storyline
World: convergent
Economy: service and information based; lower growth than A1
Population: same as A1
Governance: global solutions to economic, social and environmental sustainability
Technology: clean and resource-efficient

B2 storyline
World: local solutions
Economy: intermediate growth
Population: continuously increasing at lower rate than A2
Governance: local and regional solutions to environmental protection and social equity
Technology: more rapid than A2; less rapid, more diverse than A1/B1

Figure 5: IPCC SRES model scenarios (IPCC, 2007a: Figure TS.2).

Some points to consider:

- Greenhouse gas emissions due to human actions are small compared to the natural greenhouse flux. However, the natural flux is balanced – as much is taken up by natural processes as is given off. Human-caused emissions are unbalanced – they accumulate in the atmosphere because they are not removed by natural processes (e.g., photosynthesis, marine shell formation, burial, etc.).
- There is a lag between the emission of carbon and its effect on climate. Therefore, even if we went to zero carbon emissions today, the climate would continue to warm (Armour and Roe 2011).
- Changes can be amplified through positive feedbacks. For instance, snow and ice reflects sunlight almost like a mirror, so little heat is absorbed and little heat given off to the overlying air. Patches of bare ground and ocean water, however, are very good at absorbing sunlight, heating up, and then transmitting the heat to the overlying air so that it warms. In the Arctic, the decline in summer sea ice means that summer air over the Arctic Ocean is much warmer and more humid

than previously, which is accelerating the warming of the Arctic (Screen and Simmonds 2010). It is therefore also contributing to a shortening of the Arctic winter; changing ocean circulation between the Arctic Sea and surrounding oceans; and changing summer climates throughout the subarctic.

- Changes can be dampened by negative feedbacks. For instance, aerosol pollutants in the atmosphere reflect sunlight back into space, and have a net cooling effect on the planet. Ironically, cleaning the air of pollutants will actually slightly increase warming.
- Climate change is not necessarily linear (gradual), but can be stepwise (rapidly shift between two very different states).
- The rate of warming is a critical issue because faster warming reduces the time humans, plants and animals have to adapt to climate change physiologically, or through migration, or, for humans, technologically.

B. Overview of the Current Climate of the Upper Rio Grande

This section provides a brief overview of the climate of the Upper Rio Grande basin, including its major features, drivers and sources of variation (this dicussion 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 midlatitude climate zones. This boundary is anticipated to shift northward, and with this change, alter the seasonal precipitation patterns in the region.

The Upper Rio Grande basin is classified as an arid climate, with average annual precipitation in most areas < 15” (<38 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. 2013). Southern, lower-elevation areas south of Elephant Butte Dam have average annual temperatures of 61-65°F (16-18°C), and receive less than 15” (38 cm) of precipitation annually. The Albuquerque portion of the Upper Rio Grande has an annual temperature of approximately 51-55°F (11-13°C) (Figure 6) and receives 11-15” (28-38 cm) of precipitation per year (Figure 7). In the San Luis Valley of Southern Colorado, the average annual temperature is 41-45°F (5-7°C) and precipitation averages <10” (25 cm) per year. 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” (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 it descends, it warms and its capacity to retain moisture increases, pulling moisture out of the environment as the air mass descends[6]. 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.

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[6] 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.

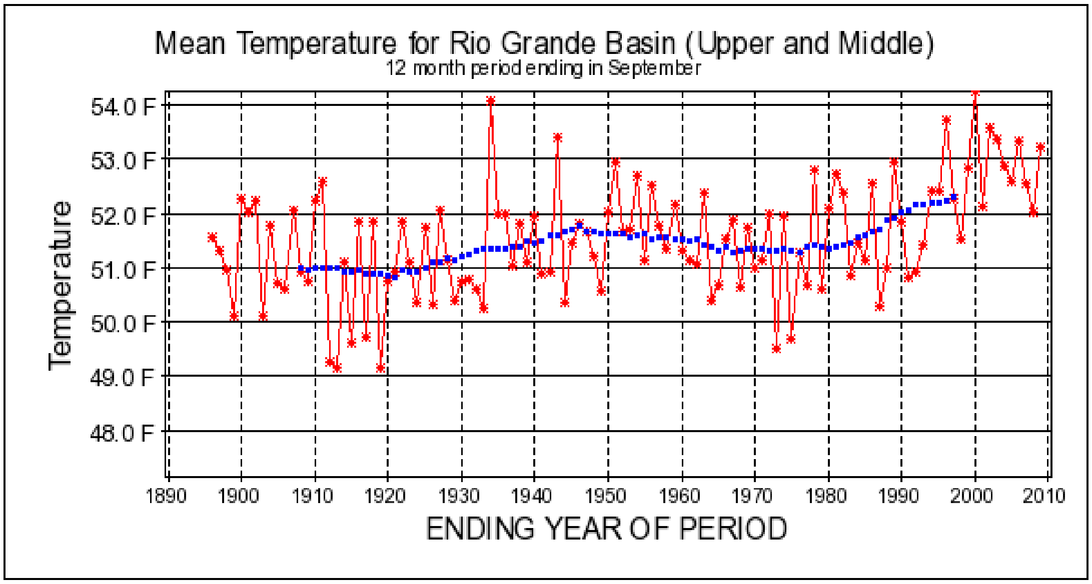
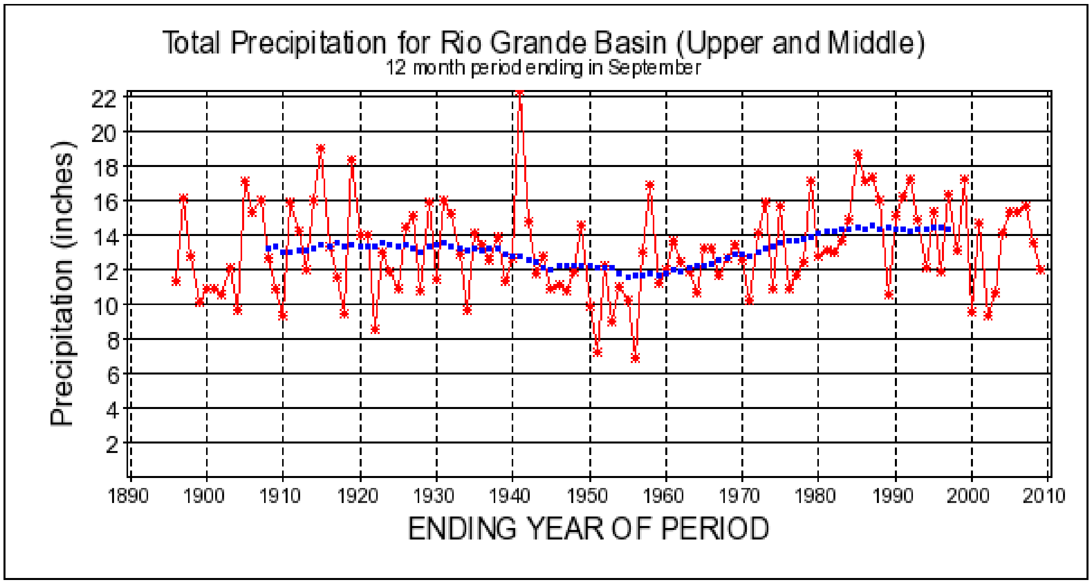


Figure 6: Observed annual temperature, averaged over the Rio Grande Basin above Elephant Butte. Blue line is 25-year moving annual mean.



Source: Western Climate Mapping Initiative (WestMap) available at: http://www.cefa.dri.edu/Westmap/. Red line indicates annual time series for the given geographic region. Blue line indicates 25-year moving annual mean.

Figure 7: Observed annual precipitation, averaged over the Rio Grande Basin above Elephant Butte. Blue line is 25-year moving annual mean.

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 in the core monsoon region and the Upper Rio Grande falling outside. Precipitation during the summer monsoon is characteristically more than 50% of the annual total in most portions of the Upper Rio Grande. 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 midlatitude cyclonic storms into the region. These storms precipitate rain and snow over wide areas, and alternate with 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.

C. 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[7]. 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.

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

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[7] 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.

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[8] which operate on multi-decadal (50-80 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 7 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. NOAA maintains a regularly updated discussion of current ENSO status, near-term (about 6 month) ENSO projections, and implications for how changes in ENSO will affect temperature and precipitation across North America[9].

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 (SSTs), particularly the multi-decadal Pacific Decadal Oscillation (PDO), and Atlantic SSTs via the Atlantic 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 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 SSTs (negative PDO) and warm Atlantic SSTs (positive AMO) (McCabe et al. 2004).

D. 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

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[8] 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.

[9] 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/ (Accessed March 16, 2011).

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 1) how hot the Southwest gets (how much heat is available to drive air convection); 2) 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; and, 3) 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. However, precipitation can be more widespread if the monsoon is able to tap moisture from a tropical cyclone in the moisture source regions.

E. Literature Review: Observed and Projected Temperature Change

Recent overviews of climate change in the Southwestern United States (SWUS) have been provided in (Garfin et al. 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).

i. Recent Temperature Trends

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 0.96°C (1.7°F) 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 (NCDC 2013). The average temperature was 12.9°C (55.3°F), which was 1.8°C (3.2°F) above the 20th century average (National Climate Data Center 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 2001-2009 were 0.8°C (1.4°F) higher relative to the average for 1895-2000 (MacDonald 2010). Temperature increases were greater in areas to the south and at lower elevation.

Particularly troubling have been increases in winter (January, February, March, or JFM) temperatures throughout the mountainous West. The observational record of 1950-1999 shows an increase in maximum average JFM temperatures of 1.53°C (2.8°F) and an increase in minimum average JFM temperatures of 1.72°C (3°F) (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 (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 are robustly replicated in models that also include human greenhouse gas emissions (Bonfils et al. 2008).

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 2001-2010 was the warmest of all decades from 1901-2010, with temperatures increasing approximately 0.9°C±0.3°C over the period 1901-2010 (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 2001-2010 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 et al. (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 0.13°C (0.225°F) per decade over the period 1912 to 2011, but rose at the faster rate of 0.27°C (0.483°F) per decade since 1970. The same study shows New Mexico warmed at an average rate of 0.10°C (0.219°F) per decade from 1912 to 2011 but at the faster rate of 0.34°C (0.0.678°F) per decade since 1970. The same pattern of faster recent warming was also observed in annual average daytime maximum high temperature (Tmax) and annual average nighttime minimum temperature (Tmin). In Upper Rio Grande, the increase in average annual temperatures over 2001-2009 was 1.5 to 2 standard deviations above the 20th Century average in the 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).

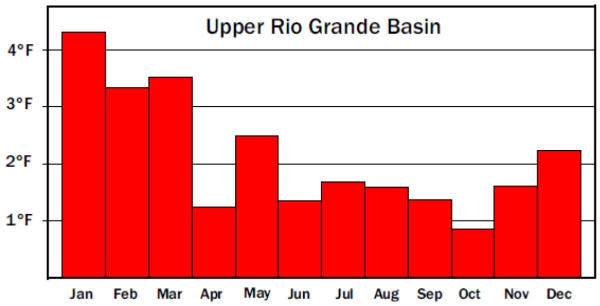

Figure 8: Average monthly temperature change in the Upper Rio Grande Basin, showing that warming is greatest in the winter months (source: Saunders and Maxwell 2005).

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 1970-2006, the average rate of temperature increase in wilderness areas in Northern New Mexico was 0.36°C/decade (0.684°F/decade), with Tmin increasing on average at a rate of 0.38°C/decade (0.684 °F/decade), approximately 0.04°C/decade faster (0.072°F/decade) than Tmax.

In the Upper Rio Grande Basin, a comparison of average monthly temperatures over the 1995-2004 period with average monthly temperatures for the period 1961-2000 (Figure 8) showed increases of 1.5-

2.5°C (3-4°F) in winter, with increases in the April through November period less than approximately 1.1°C (2.0°F) in all months but May (Saunders and Maxwell 2005).

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) detect a rate of warming of 1°C (1.8°F) per decade from 1990 to 2005. Elevation plays an important role in determining the season of greatest warming in the mountains. Lower elevation sites experienced greatest warming during the winter months, warming in winter at an average rate of 1.5°C (2.7°F) per decade. 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. 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 adjacent San Luis Valley, average annual temperatures were found to have increased 1.1°C (1.9°F) over the period 1957-2006 (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 1994-2008 was 0.4 to 1.96°C greater than the mean for the period 1958-1993 (Mix et al. 2012).

ii. 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 or 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 stream flow and other parameters.

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 1971-2000, models used in the most recent national climate assessment project (USGCRP Cayan et al. 2013, 2013):

- For the 2021-2050 period, warming under the low future emissions model scenario (known as B1) will be between 0.6-1.7°C (1-3°F) while under the higher future emissions model scenario (known as A2) warming is likely to be between 1.1-2.2°C (2-4°F).

- For the period 2041-2070, warming under the B1 scenario is likely to range from 0.6-3.3°C (1-4°F) and under the A2 scenario from 2.2-3.3°C (2-6°F).

- For the period 2070-2099, warming under the B1 scenario is likely to range from 2.2-3.3°C (2-6°F) while under the A2 scenario, the projections are 2.8-5°C (5-9°F).

- 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 1.9°C (3.5°F) in 2021-2050, 3.1°C (5.5°F) in 2041-2070, and 5°C (9°F) 2070-2099. The least amount of warming is anticipated for the winter months, with average increase of 1.4°C (2.5°F) in 2021-2050 increasing to almost 3.9°C (7°F) in 2070-2099.

Upper Rio Grande Temperature Projections

A relatively fine-grained analysis was recently conducted by NOAA, 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 URG region warms 4.1-4.9°C (7.5-8.5°F) by 2070-2099 under the higher emissions (A2) scenario, and 2.5-3.1°C (4.5-5.5°F) by 2070-2099 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 2041-2070, compared to a baseline period of 1971-2000, temperature increases by season show that (Figure 15, NOAA 2013a) the largest increases are likely to occur in summer, with increases of 3.1-3.3°C (5.5-6.0°F) in average temperature in the URG, followed by fall, with increases in average temperature in the range of 2.8-3.1°C (5.0-5.5°F). In winter and spring, the URG is likely to see increases in average temperature of 2.2-2.5°C (4.0-4.5°F). 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 URG grading to about 15-20 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 25-30 days throughout most of the URG, and as high as 30 to 35 days in the Colorado portions of the URG. The freeze free season will increase 25-30 days throughout the URG (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, 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 3°C (5°F) over the average from 1971 to 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 0.8 to 2.0°C (1.5 to 3.5°F) by 2025 relative to 1950 to 1999 average temperatures is expected, with increases of 1.4 to 3.1°C (2.5 to 5.5°F) 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 Tmax (°C)	**Change in Tmin (°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 2020-2039 and 2070-2089. 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.

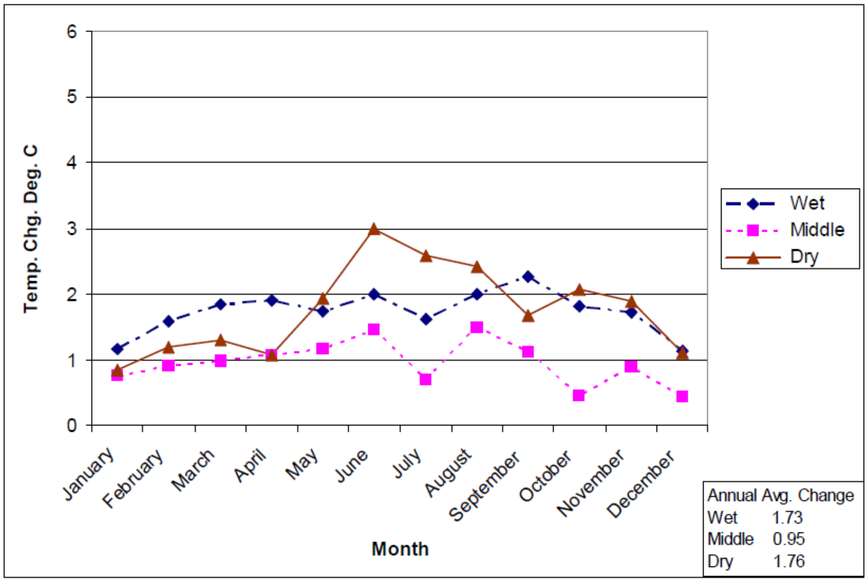

Figure 9: Three scenarios for temperature change projected for the Rio Grande basin in 2020-2039 (source: Hurd and Coonrod 2007).

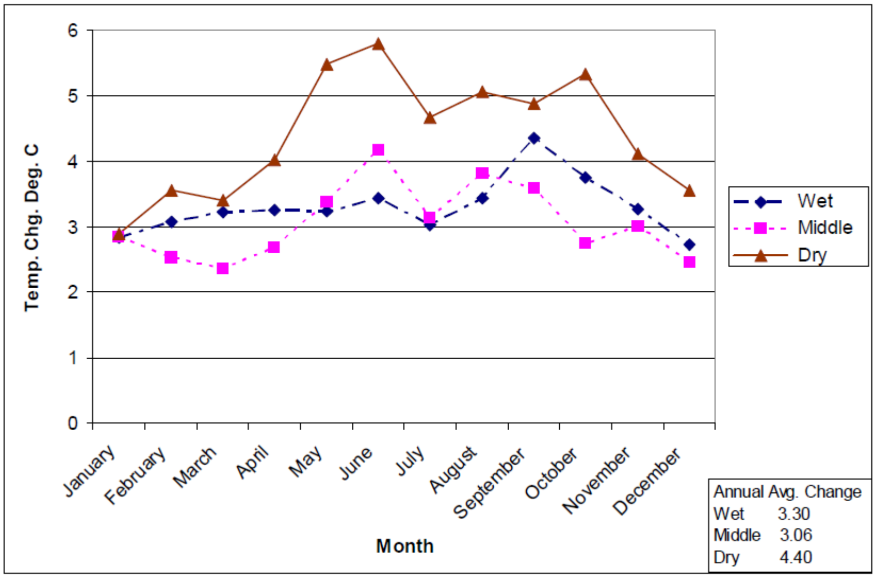

Figure 10: Three scenarios for temperature change projected for the Rio Grande basin in 2070-2089 (source: Hurd and Coonrod 2007).

In their models, average annual temperatures increased by 0.95 to 1.76°C (1.7-3.2°F) by 2030 (Figure 9) and 3.06 to 4.40°C (5.5-7.9°F) by 2080 (**Error! Reference source not found.**). 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).

Climate change in the Upper Rio Grande basin was modeled by Reclamation (2011c, a) 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 3 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 km).

The basin-average mean-annual temperature is projected to increase by approximately 1.8-3.3°C (5-6°F) 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.

Summary of Projected Temperature Changes

By the end of the century, temperatures in the URG are anticipated to increase by about 5°C (9°F) over twentieth century values under high emissions scenarios, and by close to 3°C (5.4°F) under the B1 (low emissions) scenarios. There is consensus that temperature increases will be greater in summer and fall. In mountain areas, overnight temperatures (Tmin) are likely to rise faster than daytime high temperatures (Tmax). 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 century's end, temperature increases are anticipated to expand the freeze-free (growing) season by 25-30 days; to cause more frequent, longer heat waves (>95°F); and to cause less frequent, shorter cold spells (<0°F).

F. Literature Review: Observed and Projected Changes to Precipitation

i. 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 (SSTs), particularly the multi-decadal Pacific Decadal Oscillation (PDO). Atlantic Ocean SSTs are also important. The driest phases in the Southwest are associated with cool Pacific SSTs (negative PDO) and warm Atlantic SSTs (positive Atlantic Multidecadal Oscillation (AMO)) (McCabe et al. 2004). Interannual (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.

National Precipitation Trends

At the national scale, precipitation has increased 5% 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 1950-1999 (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).

Southwestern U.S. and Upper Rio Grande Precipitation Trends

In the Southwest during the 20th century, the period 1905-1930 had wetter winters than average, 1931-1941 was approximately average, and from 1942-1964 it was drier than average, with peak dryness occurring during the drought from 1950-1956 when average annual precipitation remained below the long term average (Swetnam and Betancourt 1998, Sheppard et al. 2002, 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, 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 1200+ years (MacDonald et al. 2008, 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 22-25% 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 2001-2006, 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 1964-1976, negative PDO and negative AMO contributed to average precipitation conditions, and 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 (NCAR 2012), contributing to the reemergence of drought across the Southwest. The decade 2001-2010 has had the second-largest area affected by drought (after the period 1951-1960) and the most severe average drought conditions of any decade since 1901 (Hoerling et al. 2013). This drought is ongoing through March 2013 (National Drought Mitigation Center 2013) and is anticipated to persist into summer 2013 (NOAA 2013b).

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, 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 1991-2005 compared to the mean for 1961-1990 was observed across sites in the 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 1961-1990 showed a 7.56% decrease.

In all parts of Colorado, including the northern portion of the URG, 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 2011a). 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.

ii. Model Projections of Late 21st Century Precipitation

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: poleward expansion of the tropics (including subtropics).

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 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, 2013). But what will happen in areas lying on the current boundary between subtropical and midlatitude 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% (Barnett and Pierce 2009).

Researchers at the U.S. Global Change Research Program project a 10 to 20% decline in precipitation by 2080-2090 primarily in the winter and spring, resulting 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 et al. (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 will continue to be highly variable in time and place, and that the region will 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%. This pattern has also been observed in the Southwest (USGCRP 2009). From 1958 to 2007, there was a 9% 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% increase and the Midwest at 31% 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.

Projected Precipitation Changes in the Upper Rio Grande

In the Upper Rio Grande, 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% (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).

Modeling by the Reclamation for the 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. 2013). Maximum daily precipitation in the mapped area corresponding to the Upper Rio Grande, under the maximum emissions scenario, sees an increase of 10-30% in the maximum daily precipitation value in 2071-2100 compared to the period 1971-2000 (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 United States. Climatologically, this would seem to indicate more intense, localized monsoon storm events (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.

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 SWUS, primarily through reductions in winter precipitation. The SWUS is located on the boundary between the arid subtropics (the poleward portion of the tropics), and the more temperate midlatitudes whose weather systems are dominated by large-scale cyclonic systems. Seasonal changes in atmospheric circulation bring the SWUS 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 SWUS and permitting midlatitude 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 midlatitudes.

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 (~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 1979-2005, the subtropical dry zone already expanded poleward between 2 and 8 degrees of latitude (~0.8 to 4 degrees poleward in the Northern hemisphere) depending on the measure used (Seidel et al. 2008, 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 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 has 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 midlatitudes 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 midlatitude 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 has 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.

Projected Changes in the North American Monsoon

Over the period 1948-2004, 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 variation 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). In addition, ENSO and PDO exert effects on the North American Monsoon, and how these may change in the future is unclear. In addition, it is not clear how strengthening of Hadley cell circulation leading to enhanced subsidence in the subtropical dry zone (including New Mexico) will affect 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 ~5-10% in the URG in 2041-2070 compared to 1971-2000, 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.

Summary of Projected Precipitation Changes

Overall, models project that precipitation in the Upper Rio Grande, and the Southwest as a whole, will remain unchanged, or will decline slightly (with a maximum of approximately 13%) 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.

G. 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 the data 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). PDSI includes both a precipitation term and a temperature term (as a proxy for evaporation), and therefore reflects the balance of moisture inputs or loss to the soil in an area.
- Hydrological drought refers to declines in stream flow and water storage in lakes and reservoirs. Hydrological drought is measured in terms such as discharge (cfs), stream flow (f/s), 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 drought 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 et al. (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 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 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.

i. Recent and Past Drought

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 1951-1999 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 (approx. 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 has experience 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 (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 (as of may 2013) 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 1200 years (back 2000 years in some areas). In a widely cited work, Cook et al. (2004) used annually-resolved tree-ring records from throughout North America to reconstruct annual summer-season PDSI for the last 1200 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[10]. In the warmest part of the Medieval Warm Period (from AD 950 to 1150), average Northern Hemisphere temperatures rose 0.2-0.4°C (0.36-0.72°F) above the mean annual temperature for 1850-2006; by comparison, late 20th / early 21st Century global average temperatures are 0.8°C (1.44°F) above the same mean (Mann et al. 2008).

In the Southwest, average annual temperatures during the Medieval Warm Period may have been 0.4-0.8°C (0.72-1.44°F) above the mean annual temperature for 1850-2006. During the warmest intervals in the Southwest, temperatures may have approached 1°C (1.8°F) above this mean, a value equal to the 1961-1990 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 1146-1150. During this period, 65.5% of the Southwest was under drought conditions, and average annual maximum temperatures for the region were 15.65°C (60.2°F). By comparison, over the 20th Century, the average annual maximum temperature (1909-2008) has been 15.72°C (60.3°F) and 32.6% of the region was under drought conditions, while during the period 1999-2008, 48.4% of the region was in drought and the average temperature was 17.54°C (63.6°F) (Woodhouse et al. 2010). The drought has persisted through March 2013, at which time almost the entire Southwest,

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

except for portions of California, was under moderate to extreme drought (National Drought Mitigation Center 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 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 Atlantic Multidecadal Oscillation (AMO), which primarily influence winter precipitation (Balling and Goodrich 2010).

ii. Model Projections of Late 21st Century Drought

In a review of 19 models used by the IPCC in its most recent assessment report, Seager et al. (2007) examined trends in precipitation minus evaporation (P-E) over the period 1900-2098 (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 (Oct.-March) and summer (Apr.-Sept.) precipitation in the 21st Century relative to the 20th. In winter, warming causes evaporation to increase steadily, resulting in projections of an increasingly negative value for precipitation-evaporation (P-E) over the 21st Century. Declines in summer precipitation are also projected. Decreases in the value of winter P-E 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, expansion of the subtropical dry zone and the poleward retreat of the temperate wet zone, driven by global-scale warming, are the primary causes of changes in P-E. 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 2.5°C (3.6°F) above the 1900-1909 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.

H. 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% 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.

i. Observed Hydrologic Changes

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% for 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 1949-2004 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% gain in precipitation (Mote et al. 2005), but this is an artifact of a trend line that begins in the last major Southwestern drought (1950-1956) and ends in the wettest period of the historical record (1976-1997/1998). 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 twentieth century temporal variability in precipitation, temperature, and estimated SWE. Winter temperatures since 1980 are, on average, higher than any other decade of the 20th Century, while the average April 1 SWE and the ratio of SWE to precipitation are lower (McCabe and Wolock 2009).

In a recent study, tree ring records were used to estimate annual SWE since AD 1200 in the Rocky Mountains. Prior to the 1980s, there was a pronounced dipolar character to SWE: 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 SWE 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 1948-2008 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%, 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.

Advances in Snowmelt

The observational record of 1948-2000 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 (AMJJ) fractional flows (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 2080-2099 relative to baseline data from 1951-1980 (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).

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, the Colorado River experienced an increase in flow relative to the prior year while flows in the Rio Grande remained low.

During the first decade of the current drought (2001-2010), 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 1901-2000 average of 15.0 million acre-feet/year, representing a 16% decadal deficit. On the Rio Grande at El Paso, observed flows for 2001-2010 were about 23% lower than the period from 1941-2000.

ii. Projected Hydrologic Changes

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% (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, 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% below the historical average for the periods 2010-2039, 2040-2069 and 2070-2098, 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% 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% and consumption is unchanged, water managers will be unable to deliver all of the promised water 58% of the time. A reduction in runoff of 20% leads to a failure in water delivery approximately 88% of the time if consumption patterns are unchanged. The shortfall ranges from at least 1.2 to 1.9 billion cubic meters per year to approximately 2.2 to 3.4 billion cubic meters per year by 2050 out of a total request of 17.3 billion cubic meters per year (Barnett and Pierce 2009). The magnitude of the shortfall is not so great that it could not 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% (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 to 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.

Projected Changes in the Upper Rio Grande

There are fewer projections of hydrologic change in the Rio Grande than for the Colorado, 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 4-6 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 RMCO (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 1950-1999, modeling studies of the Colorado River watershed project that the content of water contained in April snowpack will decline by approximately 38% 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 50-200 mm of water) compared to the 1961-1985 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) emissions scenario show a pronounced soil moisture deficit in the spring (March-May) season, particularly in northwest New Mexico where soil moisture is projected to decrease by 5 mm water (20% relative to 1961-1985 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-March runoff and April-July runoff are all anticipated. The most detailed hydroclimate modeling specific to the Rio Grande has been conducted by Reclamation under its WWCRA program as required under the SECURE Water Act (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 2070-2079 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 4°C (7.2°F) increase in temperature in wet, normal and dry years, as well as for a "normalized year" based on the average condition for the period 1957-1994 (Rango and Martinec 2008). For the Rio Grande, a greater share of runoff is projected to occur in the winter (October-March) than in the summer (April-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).

Base Year	October-March		April-September		Hydrological Year	
	Runoff 106m3	Runoff % of total	Runoff 106m3	Runoff % of total	Runoff 106m3	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 SWE, 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.

V. Observed Climate Trends in the Upper Rio Grande Basin

Observed climate trends for the Upper Rio Grande basin above Elephant Butte Dam were analyzed to better understand current rates of climate change in the study area. Topographic diversity is a key factor as this region encompasses the headwaters of the Rio Grande in the San Juan and Sangre de Cristo Mountains of Colorado, both with peaks exceeding 14,000 ft asl; the Tusas and Jemez Mountains of New Mexico, with peaks rising as above 11,000 ft asl; the Rio Grande Rift extending from the San Luis Valley of southern Colorado past the southern boundary of the study area at Elephant Butte Dam at approximately 4200 ft asl; and areas to the west and east of the central valley that are nonetheless part of the drainage basin. The region is home to one of the largest remaining stretches of riparian cottonwood forest in the western U.S. and includes critical habitat for the Federally-endangered Southwestern Willow Flycatcher (*Empidonax traillii extimus*) and Rio Grande silvery minnow (*Hybognathus amarus*).

i. Data and Methodology

Three sources of climate data were used to investigate recent climate trends in the Upper Rio Grande:

- USDA Natural Resource Conservation Service SNOTEL (snowpack telemetry) stations provided temperature and precipitation data beginning in 1989 (slightly earlier for some stations). SNOTEL sites in this region are positioned to provide a representative spatial sample of snowpack conditions (Molotch and Bales 2006) and may not provide a spatially representative sample of climate data. Data from 13 SNOTEL sites were used in this study, providing the majority of data from high elevation settings. Monthly average values for temperature and precipitation were obtained from the National Climate Data Center (NOAA National Climate Data Center 2013) for the period of record ending in December 2012.
- NOAA National Weather Service Cooperative Observer Network (COOP) sites provided the bulk of the data from lower elevation settings. COOP sites are located to collect agriculturally-relevant climate data. Data are collected on a voluntary basis and COOP data at most sites contain recording gaps, notably during World War I, the Great Depression and World War II. Consequently, although data exist prior to 1950, it is mainly discontinuous. The data collected since 1950 are more complete, and therefore the year 1950 is taken as the earliest reliable date for most COOP site data in the study area. Monthly average values for temperature and precipitation were obtained from the National Climate Data Center (NOAA National Climate Data Center 2013). The period of record for COOP sites in this study is January 1971 through December 2012.
- NOAA National Weather Service Historical Climatology Network 2 (HCN) data were used, where possible. Eleven HCN sites occur in the study area, primarily but not exclusively in valley floor settings. HCN data were originally collected as part of the COOP system, but have been extensively corrected for station inhomogeneities and gaps in the data have been rectified. Monthly average values for temperature and precipitation were obtained from the National Climate Data Center (NOAA National Climate Data Center 2013). The period of record for the HCN sites used in this study is January 1971 through December 2012.

Mountain climates are complex and vary over short distances due to aspect and relief, which influence temperature and precipitation via cold air drainage, down and up-canyon winds, variation in the duration of direct vs. indirect insolation, vegetation cover, duration of snow cover, and other factors (Beniston 2006, Barry 2008). Changes at individual stations may differ from regional climate trends (Pepin et al. 2005) in ways that are strongly influenced by landscape position, topography and elevation (Lundquist and Cayan 2007). Valley floors may lag regional warming, particularly in winter months, due to the increasing frequency and severity of temperature inversions under more stable, anticyclonic conditions (Daly et al. 2010), as are anticipated to become more common in the southwestern United States (Seth et al. 2011).

Because of these complexities, additional data processing was not undertaken: some locations in each data set exhibited trends counter to the remainder of the sites, and these data may reflect real, but local climate differences. They may also reflect changes to station equipment, setup and location, and NCDC data are corrected for many of these factors.

Because of the landscape diversity in the 300 km wide by 600 km long study area, the sites were grouped into physiographic units for analysis. Mountain sites include (Figure 11, Table 4):

- San Juan Mountains – Seven sites are located in the eastern San Juan Mountains within the Rio Grande basin or near the drainage divide in adjoining drainages. Six of these sites are SNOTEL stations and the other is an HCN site. Site elevations range from 9048 to 11600 ft. asl, with the HCN site at the lowest elevation in this region.
- Sangre de Cristo Mountains – Two sites are located in the southern Sangre de Cristo Mountains, consisting of one SNOTEL site and one HCN site. Because these mountains mark the boundary between the Southern Rocky Mountains and the Plains, they may be subject to different climate influences in some portions of the year from high elevation sites to the west in the San Juan Mountains. Site elevations range from 8676 ft. asl at the HCN site to 9800 at the SNOTEL site.
- Tusas Mountains – Four sites are located in the Tusas Mountains. The Tusas Mountains have a lower average elevation from mountain ranges to the north. These sites include four SNOTEL sites (between 8400 and 10,040 ft asl).
- Jemez Mountains – Three sites are located in the Jemez Mountains, which are southwest of the Tusas Mountains. These consist of two SNOTEL sites and one COOP site. The two SNOTEL sites are located in high elevation settings at 8600 and 9500 ft. asl while the COOP site is at 8220 ft asl. In addition, this category includes one COOP site located at Los Alamos on the Pajarito Plateau at 7424 ft. asl.

Valley sites used in the URG study were grouped into the following physiographic units (Figure 11, Table 5):

- Northern Valleys – Five sites are located in the San Luis and Rio Grande Valleys in southern Colorado. These consist of two COOP sites and three HCN sites, and they range in elevation from 7533 to 8183 ft. asl.
- Rio Chama and Jemez River Valleys – This category includes three sites located in the Rio Chama Valley. These sites consist of one HCN site and two COOP sites ranging in elevation from 6380 to 7850 ft. asl. This category also includes one COOP site in the Española Basin at Alcalde (5680 ft. asl), in the vicinity of the Rio Chama-Rio Grande confluence, and the HCN Jemez Springs site in the Jemez River Valley at 6262 ft asl.
- Middle Rio Grande – This category includes the COOP site of Albuquerque IAP and the HCN site of Elephant Butte Dam located on the bajada above the floodplain at 4576 and 5310 ft, respectively. It also includes the two HCN sites of Los Lunas and Socorro, which are located directly in the floodplain of the Rio Grande at 4585 and 4840 ft. asl, respectively. The Middle Rio Grande also includes the COOP site of Grants Milan Airport at 6520 ft. asl in the Rio Puerco Valley, and the COOP site of Augustine (7000 ft. asl) in the Plains of San Agustin.

- Plains – Three sites within the Rio Grande basin located east of the Manzano Mountains in a an area potentially subject to different climate conditions from Middle Rio Grande sites. Two of these are COOP sites (6140 and 6150 ft. asl), the other is an HCN site (6520 ft. asl).

Table 4 Mountain sites used for trends analysis.

Map Number	Station	Id	Type	Latitude	Longitude	Elev. (ft.)	Aspect (deg.)	Slope (deg.)
San Juan Mountains								
1	Beartown	07M32S 327	SNOTEL	37.700000	-107.500000	11600	301.80	28.10
2	Hermit	53951	HCN	37.771670	-107.109720	9048	87.40	17.77
3	Middle Creek	07M21S 624	SNOTEL	37.77167	-107.033333	11250	106.54	20.81
4	Slumgullion	07M30S 762	SNOTEL	37.983330	-107.200000	11440	99.58	7.92
5	Upper Rio Grande	07M16S 839	SNOTEL	37.720000	-107.250000	9400	347.79	16.55
6	Upper San Juan	06M03S 840	SNOTEL	37.483330	-106.833330	10200	334.99	10.79
7	Wolf Creek Summit	06M17S 874	SNOTEL	37.466670	-106.800000	11000	60.51	8.72
Sangre de Cristo Mountains								
13	Gallegos Peak	05N18S 491	SNOTEL	36.180000	-105.550000	9800	287.37	22.48
14	Red River	297323	HCN	36.705830	-105.403610	8676	229.14	1.25
Tusas Mountains								
15	Bateman	06N04S 316	SNOTEL	36.500000	-106.316670	9300	268.95	8.09
16	Chamita	06N03S 394	SNOTEL	36.950000	-106.650000	8400	17.56	5.31
17	Cumbres Trestle	06M22S 431	SNOTEL	37.020000	-106.450000	10040	118.89	1.00
18	Hopewell	06N14S 532	SNOTEL	36.700000	-106.250000	10000	50.89	9.02
Jemez Mountains								
24	Los Alamos	295084	COOP	35.864440	-106.321390	7424	36.47	5.32
25	Quemazon	06P01S 708	SNOTEL	35.920000	-106.383330	9500	191.63	25.02
26	Senorita Divide	06P10S 744	SNOTEL	36.000000	-106.833330	8600	85.84	8.98
27	Wolf Canyon	299820	COOP	35.947780	-106.746940	8220	227.03	11.95

Table 5 Valley sites used for trends analysis.

Map Number	Station	Id	Type	Latitude	Longitude	Elev. (ft.)	Aspect (deg.)	Slope (deg.)
Northern Valleys								
8	Alamosa	50130	COOP	37.438890	-105.861390	7533	147.69	0.11
9	Del Norte	52184	HCN	37.674170	-106.324720	7864	13.70	2.24
10	Great Sand Dunes	53541	COOP	37.733330	-105.511940	8183	303.97	3.58
11	Manassa	55322	HCN	37.174170	-105.939170	7690	12.99	0.26
12	Saguache	57337	HCN	38.085800	-106.144400	7701	87.47	0.34
Rio Chama and Jemez River Valleys								
19	Abiquiu Dam	290041	COOP	36.240280	-106.427780	6380	131.33	4.43
20	Alcalde	290245	COOP	36.090830	-106.056670	5680	268.89	4.12
21	Chama	291664	HCN	36.917780	-106.578060	7850	208.58	1.25
22	El Vado Dam	292837	COOP	36.592780	-106.730000	6740	159.58	2.58
23	Jemez Springs	294369	HCN	35.778330	-106.687220	6262	179.38	6.17
Middle Rio Grande								
28	Albuquerque IAP	290234	COOP	35.041670	-106.615280	5310	326.44	0.75
29	Augustine	290640	COOP	34.075000	-107.621110	7000	38.39	0.21
30	Elephant Butte Dam	292848	HCN	33.146110	-107.184440	4576	2.48	6.88
31	Socorro	298387	HCN	34.082780	-106.883060	4585	147.48	0.40
32	Los Lunas	295150	HCN	34.767500	-106.761110	4840	119.59	0.83
33	Grants Milan AP	293682	COOP	35.166390	-107.899170	6520	181.79	1.43
Plains								
34	Estancia	293060	COOP	34.824170	-106.034440	6140	134.96	0.17
35	Mountainair	295965	HCN	34.520830	-106.260560	6520	119.09	5.26
36	Pedernal	296687	COOP	34.615280	-105.473890	6150	128.80	2.62

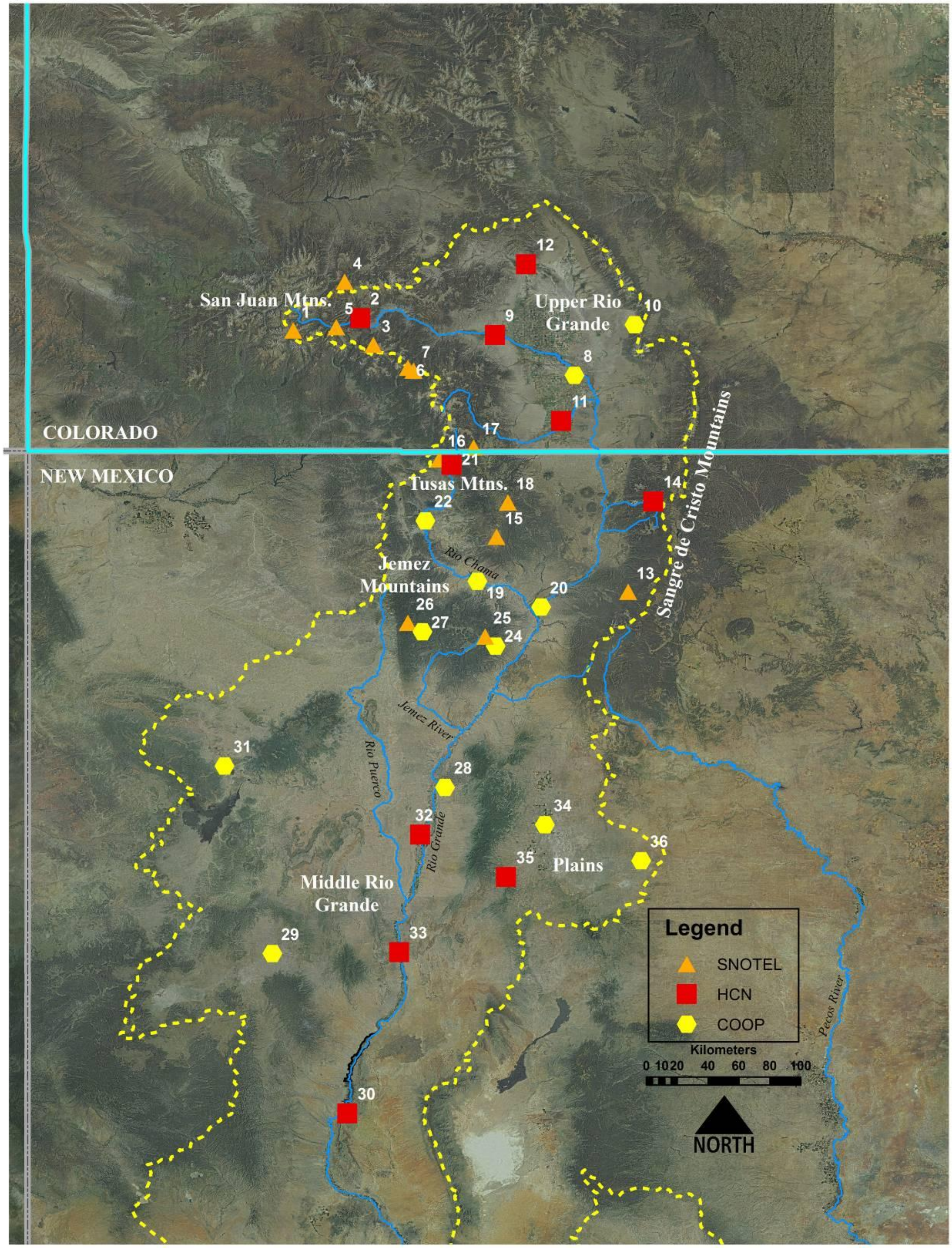


Figure 11: Map showing sites used in the analysis (site numbers keyed to Tables 6 and 7).

Because the distribution of monthly means is skewed, trends are assessed nonparametrically using the Regional Kendal Test (Helsel and Frans 2006). For this analysis, the Regional Kendall Test yields the annual trend (Thiel-Sen's slope) and statistical significance of the trend by physiographic unit. All analyses are conducted using the RKT package in R (Marchetto 2012). Statistical significance was evaluated at the 0.1 (90% confidence) level. Annual trends are computed as the median of the monthly trends.

ii. Observed Trends for the Period 1971-2012

Despite the noise in the data introduced by measurement changes, errors, instrumentation, changes in station microclimate due to movement and wildfire, and other problems, a coherent regional picture of temperature and precipitation emerges when the data are aggregated into mountain and valley sites.

Annual Trends

For the entire Upper Rio Grande study area, temperatures increased substantially over the four decade period 1971-2012. Average annual temperatures (Tavg) increased at a rate of 0.35°C (0.63°F) per decade (Table 6), with a faster increase in nighttime minimum temperature (Tmin) of 0.37°C (0.67°F) per decade (Table 8) offset by a slower increase in daytime high temperature (Tmax) of 0.25°C (0.45°F) per decade (Table 7). Precipitation was unchanged at the regional scale (Table 9).

Mountain and valley regions responded differently to warming. Mountain Tavg increased at a rate of 0.37°C (0.67°F) per decade over the period 1971-2012. This change was driven by increases in nighttime minimum temperatures (Tmin) of 0.67°C (1.21°F) per decade that were significant in every month but February; daytime high temperatures (Tmax) rose at the slow rate of 0.14°C (0.25°F) decade, and this trend was not significant in most areas. By contrast, valley Tavg temperatures increased at a rate of 0.33°C (0.39°F) per decade over the period 1971-2012, driven by both increases in Tmax (0.34°C (0.61°F) per decade) and Tmin (0.28°C (0.50°F) per decade). At valley sites, increases in May-September temperatures were statistically significant, increasing at a rate of 0.3-0.5°C (0.54 – 0.90°F) per decade in these months.

Among the mountain sites, temperature increases were greatest at the four sites in the Tusas Mountains, where Tavg increased at a rate of 0.81°C (1.46°F) per decade, driven by increases in Tmin at a rate of 1.39°C (2.50°F) per decade. The San Juan and Sangre De Cristo Ranges saw temperatures increase at approximately half this rate; further south in the Jemez Mountains, temperatures increased at about a quarter of the rate of the Tusas Mountains.

Table 6 Rate of change in average monthly temperature (Tavg) in °C/year for 1971-2012.

Region	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual °C/yr	°C/ 10 yr.
San Juan Mountains	0.083	-0.012	0.071	0.033	0.030	0.039	0.038	0.017	0.038	0.005	0.088	0.044	0.04	0.38
Sangre de Cristo Mtns.	0.100	0.033	0.043	0.050	0.050	0.039	0.024	0.037	0.028	0.025	0.071	0.047	0.04	0.41
Tusas Mountains	0.148	0.006	0.065	0.050	0.081	0.100	0.092	0.079	0.080	0.067	0.200	0.100	0.08	0.81
Jemez Mountains	0.032	-0.012	0.040	0.029	0.042	0.028	0.028	0.033	0.029	0.014	0.037	-0.012	0.03	0.29
All Mountain Sites	***0.075***	***0.000***	***0.050***	***0.036***	***0.044***	***0.039***	***0.037***	***0.033***	***0.037***	***0.020***	***0.077***	***0.030***	***0.04***	***0.37***
Upper Rio Grande	0.058	-0.005	0.034	0.027	0.040	0.019	0.026	0.025	0.017	0.003	0.036	-0.005	0.03	0.26
Rio Chama / Jemez Valleys	0.060	0.016	0.036	0.029	0.037	0.024	0.023	0.034	0.025	0.000	0.023	-0.004	0.02	0.25
Middle Rio Grande	0.050	0.024	0.033	0.050	0.078	0.055	0.048	0.056	0.050	0.036	0.045	0.016	0.05	0.49
Plains	0.037	-0.007	0.008	0.033	0.045	0.032	0.025	0.036	0.032	0.010	0.017	0.000	0.03	0.29
All Valley Sites	***0.050***	***0.012***	***0.030***	***0.036***	***0.050***	***0.033***	***0.031***	***0.039***	***0.032***	***0.014***	***0.033***	***0.000***	***0.03***	***0.33***
Region (All Sites)	***0.058***	***0.007***	***0.036***	***0.036***	***0.050***	***0.034***	***0.033***	***0.037***	***0.033***	***0.015***	***0.043***	***0.011***	***0.04***	***0.35***

Tan: Increasing, with correlation significant at 90% (0.1) confidence level; Purple: Decreasing, with correlation significant at 90% (0.1) confidence level.
**Significance not calculated (sample size too small).*
Decadal trend (°C/10 yr.) calculated as Annual Trend x 10.

Table 7 Rate of change in monthly maximum temperature (Tmax) in °C/year for 1971-2012.

Region	**Jan**	**Feb**	**Mar**	**Apr**	**May**	**Jun**	**Jul**	**Aug**	**Sep**	**Oct**	**Nov**	**Dec**	**Annual °C/yr**	**°C/ 10 yr.**
San Juan Mountains	0.054	-0.064	0.006	-0.024	-0.011	0.009	0.018	-0.025	0.008	-0.052	0.080	0.025	0.01	0.07
Sangre de Cristo Mtns.	0.044	-0.012	0.033	0.012	0.037	0.023	0.000	0.012	0.014	0.007	0.060	0.008	0.01	0.13
Tusas Mountains	0.100	-0.037	0.000	-0.012	0.014	0.068	0.029	0.040	0.016	-0.024	0.173	0.036	0.02	0.23
Jemez Mountains	0.003	-0.046	0.047	0.026	0.042	0.020	0.013	0.023	0.021	-0.011	0.037	-0.044	0.02	0.21
All Mountain Sites	***0.042***	***-0.043***	***0.027***	***0.000***	***0.019***	***0.022***	***0.013***	***0.006***	***0.014***	***-0.023***	***0.075***	***0.000***	***0.01***	***0.14***
Upper Rio Grande	0.036	-0.033	0.037	0.013	0.029	0.013	0.020	0.011	0.006	-0.014	0.030	-0.021	0.01	0.13
Rio Chama / Jemez Valleys	0.067	0.008	0.067	0.032	0.052	0.035	0.022	0.033	0.032	0.009	0.045	0.000	0.03	0.33
Middle Rio Grande	0.050	0.000	0.033	0.029	0.064	0.044	0.036	0.058	0.053	0.038	0.054	0.007	0.04	0.41
Plains	0.067	0.007	0.047	0.047	0.060	0.050	0.033	0.043	0.060	0.026	0.072	0.013	0.05	0.47
All Valley Sites	***0.056***	***0.000***	***0.045***	***0.029***	***0.050***	***0.033***	***0.027***	***0.035***	***0.036***	***0.014***	***0.048***	***0.000***	***0.03***	***0.34***
Region (All Sites)	***0.050***	***-0.060***	***0.150***	***-0.040***	***-0.220***	***0.052***	***-0.081***	***0.087***	***0.000***	***0.073***	***0.175***	***-0.020***	***0.03***	***0.25***

Tan: Increasing, with correlation significant at 90% (0.1) confidence level; Purple: Decreasing, with correlation significant at 90% (0.1) confidence level.

**Significance not calculated (sample size too small).*

Decadal trend (°C/10 yr.) calculated as Annual Trend x 10.

Table 8 Rate of change in Tmin (°C/year) by region for 1971-2012.

Region	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual °C/yr	°C/ 10 yr.
San Juan Mountains	0.118	0.040	0.126	0.100	0.067	0.070	0.073	0.055	0.073	0.071	0.114	0.073	0.07	0.73
Sangre de Cristo Mtns.	0.150	0.068	0.059	0.077	0.056	0.061	0.051	0.064	0.044	0.059	0.084	0.106	0.06	0.63
Tusas Mountains	0.193	0.054	0.133	0.122	0.103	0.150	0.160	0.120	0.138	0.140	0.228	0.159	0.14	1.39
Jemez Mountains	0.057	0.015	0.036	0.033	0.040	0.036	0.046	0.038	0.037	0.038	0.044	0.013	0.04	0.38
All Mountain Sites	***0.108***	***0.036***	***0.075***	***0.070***	***0.060***	***0.064***	***0.067***	***0.057***	***0.062***	***0.067***	***0.095***	***0.067***	***0.07***	***0.67***
Upper Rio Grande	0.073	0.019	0.029	0.036	0.050	0.025	0.032	0.040	0.025	0.015	0.033	0.011	0.03	0.31
Rio Chama / Jemez Valleys	0.054	0.024	0.003	0.030	0.023	0.012	0.025	0.036	0.015	0.000	0.000	-0.003	0.02	0.19
Middle Rio Grande	0.037	0.038	0.033	0.075	0.087	0.067	0.056	0.056	0.045	0.038	0.035	0.024	0.04	0.42
Plains	-0.007	-0.032	-0.022	0.020	0.020	0.018	0.023	0.030	0.000	0.000	-0.029	-0.018	0.00	0.00
All Valley Sites	***0.039***	***0.018***	***0.014***	***0.043***	***0.050***	***0.033***	***0.037***	***0.043***	***0.023***	***0.015***	***0.013***	***0.006***	***0.03***	***0.28***
Region (All Sites)	***0.058***	***0.022***	***0.029***	***0.050***	***0.050***	***0.041***	***0.044***	***0.045***	***0.033***	***0.029***	***0.033***	***0.023***	***0.04***	***0.37***

Tan: Increasing, with correlation significant at 90% (0.1) confidence level; Purple: Decreasing, with correlation significant at 90% (0.1) confidence level.
**Significance not calculated (sample size too small).*
Decadal trend (°C/10 yr.) calculated as Annual Trend x 10.

Table 9 Net change in precipitation (cm) by region for 1971-2012.

Region	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Net Ann. Change	Decadal Rate
San Juan Mountains	0.007	-0.015	-0.074	-0.001	-0.053	-0.053	-0.016	-0.026	-0.010	0.000	-0.096	0.060	-0.28	-2.77
Sangre de Cristo Mtns.	0.002	0.027	-0.080	0.008	-0.042	-0.048	0.001	0.013	0.033	-0.001	-0.079	0.004	-0.16	-1.62
Tusas Mountains	-0.060	-0.099	-0.306	0.031	-0.040	-0.078	-0.051	-0.202	0.037	-0.022	-0.225	0.112	-0.90	-9.03
Jemez Mountains	-0.027	0.000	-0.053	0.018	-0.035	-0.007	-0.023	-0.014	-0.023	0.021	-0.050	0.048	-0.15	-1.45
All Mountain Sites	***-0.009***	***-0.007***	***-0.085***	***0.009***	***-0.043***	***-0.041***	***-0.019***	***-0.031***	***-0.001***	***0.001***	***-0.087***	***0.048***	***-0.27***	***-2.65***
Upper Rio Grande	-0.004	0.001	0.000	0.011	-0.014	-0.009	-0.011	0.008	0.011	0.000	-0.020	0.000	-0.03	-0.27
Rio Chama / Jemez Valleys	-0.006	0.004	-0.015	0.029	-0.017	-0.011	0.010	0.001	0.001	0.009	-0.028	0.020	0.00	-0.03
Middle Rio Grande	-0.009	-0.005	0.000	0.000	-0.011	0.001	0.017	-0.039	-0.024	-0.004	-0.011	0.003	-0.08	-0.82
Plains	-0.005	-0.003	0.000	0.000	0.000	0.003	0.000	-0.004	-0.033	0.017	-0.023	0.014	-0.03	-0.34
All Valley Sites	***-0.006***	***0.000***	***-0.001***	***0.005***	***-0.011***	***-0.003***	***0.004***	***-0.007***	***-0.007***	***0.002***	***-0.019***	***0.007***	***-0.04***	***-0.36***
Region (All Sites)	***-0.006***	***-0.001***	***-0.010***	***0.006***	***-0.016***	***-0.009***	***-0.001***	***-0.011***	***-0.006***	***0.002***	***-0.028***	***0.012***	***-0.07***	***-0.68***

Tan: Increasing, with correlation significant at 90% (0.1) confidence level; Purple: Decreasing, with correlation significant at 90% (0.1) confidence level.
**Significance not calculated (sample size too small).*

Among valley sites, the rates of temperature increase were greatest for sites in the Middle Rio Grande than elsewhere, with Middle Rio Grande Tavg increasing at a rage of 0.49°C (0.88°F) per decade from 1971-2012, with comparable increases in both Tmin and Tmax. On the Plains, Tmin was unchanged over this period, but Tmax increased at 0.47°C (0.85°F) per decade, the fastest increase in Tmax among the regions studied.

Monthly and Seasonal Trends

The rates of increase in Tmin, reflecting warming of overnight temperatures, are significant for most months in most mountain regions. February is the only month where change is positive but consistently not significant. The rate of increase in Tmin is significant across all spring (April, May, June) and summer (July, August, September) months. By contrast, changes in mountain Tmax are smaller. February shows a declining trend in Tmax across all four mountain regions; strong, positive increases in Tmax occur in November, which also shows a strong increase in Tmin as well as statistically-significant declining precipitation trends across all mountain regions. Precipitation also declined significantly in March in all mountain areas except the Sangre de Cristo Mountains, which coincides with statistically-significant increases in Tmin but not Tmax. The increasing Tmin and decreasing precipitation in March and November are important because these contribute to a longer growing season and decreased period of snowpack accumulation in winter months.

Valley regions exhibit statistically significant increases in late spring (May and June) and summer temperatures: a rate of about 0.3-0.5°C/decade in both Tmin and Tmax occurs across all valley sites in spring and summer months. Rates of increase in Fall and Winter Tmax are comparable (except for February), but the rate of increase in Tmin is lower (0-0.4°C (0-0.72°F) per decade). As with mountain areas, the trend of decreasing precipitation in November is significant across the region (except in the Middle Rio Grande), and coincides with a rate of increase in Tmax of 0.3-0.7°C (0.54-1.26°F) per decade. The rate of increase in Tmin in November is smaller (0.0-0.35°C (0-0.63°F) per decade) in valley sites, and the rate of change in Tmin is negative on the Plains.

The monthly patterns of change mountain and valley Tmin are similar, but differ in magnitude. Two factors may be at play. Valley Tmin is affected by cold air drainage; under warming, nighttime inversions may be becoming more frequent (Daly et al. 2010) and this may reduce the rate of gain in valley Tmin. By contrast, warming in mountain areas in the presence of soil moisture or snowpack contributes to daytime evaporation of that moisture; condensation under cooler, nighttime temperatures releases heat in the atmosphere and may contribute to faster nighttime warming in higher altitude settings, particularly in winter (Rangwala 2012).

Table 10 Median rates of temperature change (°C per decade) for different time periods.

		Early 1971-2000	Late 2001-2012	1971-2012
Tmax	Mountains	0.17	0.39	0.14
	Valleys	0.25	-0.13	0.34
	Region	***0.22***	***0.25***	***0.25***
Tmin	Mountains	0.62	1.75	0.67
	Valleys	0.36	-0.38	0.28
	Region	***0.42***	***0.75***	***0.37***
Tavg	Mountains	0.42	1.07	0.37
	Valleys	0.39	-0.07	0.33
	Region	***0.36***	***0.07***	***0.35***

The rate of temperature change (°C/decade) was not constant over the period 1971-2012 (Table 10). This was assessed by computing the Regional Mann-Kendall test for two periods: 1971-2000 and 2001-2012 for both mountain and valley sites in aggregate. In the first 30 years of this period, 1971-2000, positive rates of change in Tmax, Tmin, and, therefore, Tavg occurred across mountain sites, valley sites, and the region as a whole. The rate of increase in Tmin was larger than the gains Tmax for both mountains and valleys.

In the 11 years beginning in 2001, the trend in Tmax (-0.13°C (-.23°F) per decade) and Tmin (-0.38°C (-0.68°F) per decade) has been negative in valley areas. By contrast, mountain regions have been characterized by accelerated increase in rates of warming: Tmax rose from 0.17°C (0.31°F) per decade to 0.39°C (0.70°F) per decade while the rate of increase in Tmin went from 0.62°C (1.12°F) per decade over 1971-2000 to 1.75°C (3.15°F) per decade over the period 2001-2012. It is not immediately clear what is driving these changes in landscape response with topographic position.

iii. Comparison of Observed Rates of Temperature Change

Temperature rises observed in this study are comparable to two other regional studies (Figure 12, Figure 13, and

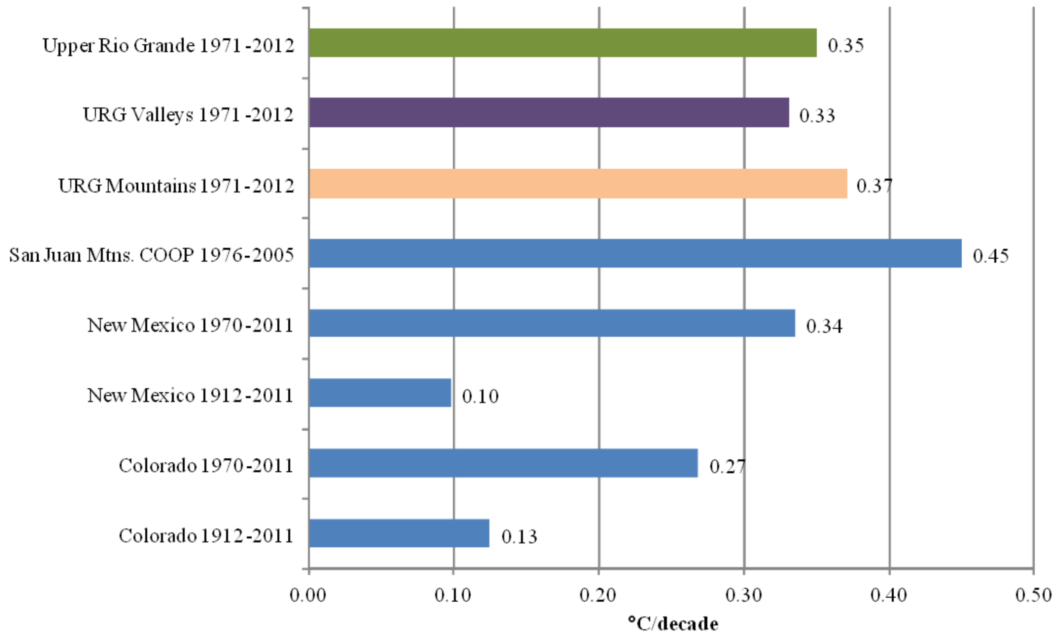
Figure 14). Tebaldi et al. (2012) use linear regression with HCN data to estimate the rate of change in temperature for the period 1912-2011 as compared to the period 1970-2011 for the states of New Mexico and Colorado. For New Mexico, the rate of change in Tavg from 1912-2011 was 0.10°C (0.177°F) per decade and for Colorado, 0.13°C (0.225°F) per decade. For New Mexico, the rate of change in Tavg from 1970-2011 was 0.34°C (0.603°F) per decade, more than three times as fast as the century average. Over this shorter period, the rate of increase in Colorado was 0.27°C (0.483°F) per decade. The same accelerating pattern occurs in the Tmax and Tmin data taken separately (Tebaldi et al. 2012).

In the San Juan Mountains and adjacent valleys for the period 1990-2005, (Rangwala and Miller 2010) find an average warming of nearly 1°C across a combination of COOP and SNOTEL site data. Tmin and Tmax increase at approximately the same rate. Warming at high elevation SNOTEL sites was gradual over the period, but occurred primarily from 1995-2000 at the lower elevation COOP sites, with negligible change in temperature at low elevations after 2000. The authors conclude that the spring and summer warming in the San Juan Mountain region over 1995-2005 is unprecedented, but winter warming is not outside the range of variation. Parsing the data into progressively shorter intervals shows a pattern of accelerated change since 1931 (Table 11).

Table 11 Trends (°C/decade) in climate change in the San Juan Mountains (modified from Table 1, Rangwala and Miller 2010).

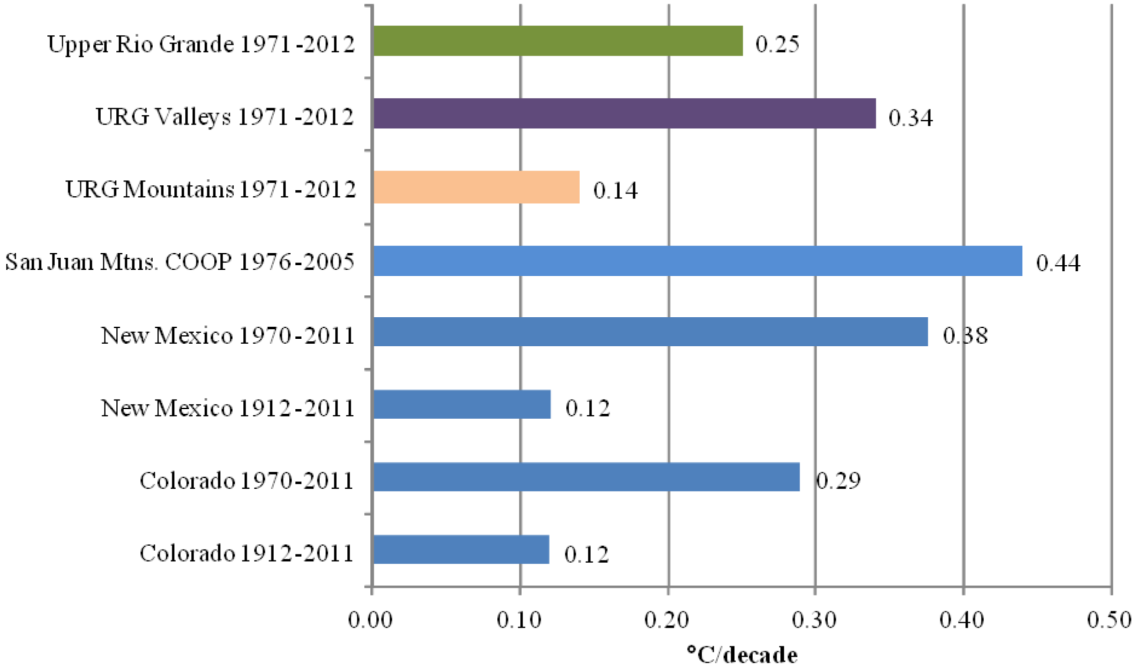
Time Period (Sites)	Tavg (°C/decade)	Tmax (°C/decade)	Tmin (°C/decade)
1931-2005 (NWS COOP)	0.08	-0.02	0.17
1956-2005 (NWS COOP)	0.16	0.11	0.20
1976-2005 (NWS COOP)	0.45	0.44	0.51
1990-2005 (NWS COOP)	1.03	1.15	0.87
1990-2005 (SNOTEL)	1.00	0.94	1.04

Red =increase significant at the 90% (0.1) confidence level (Mann-Kendall test); Purple = decrease significant at the 90% (0.1) confidence level (Mann-Kendall test).



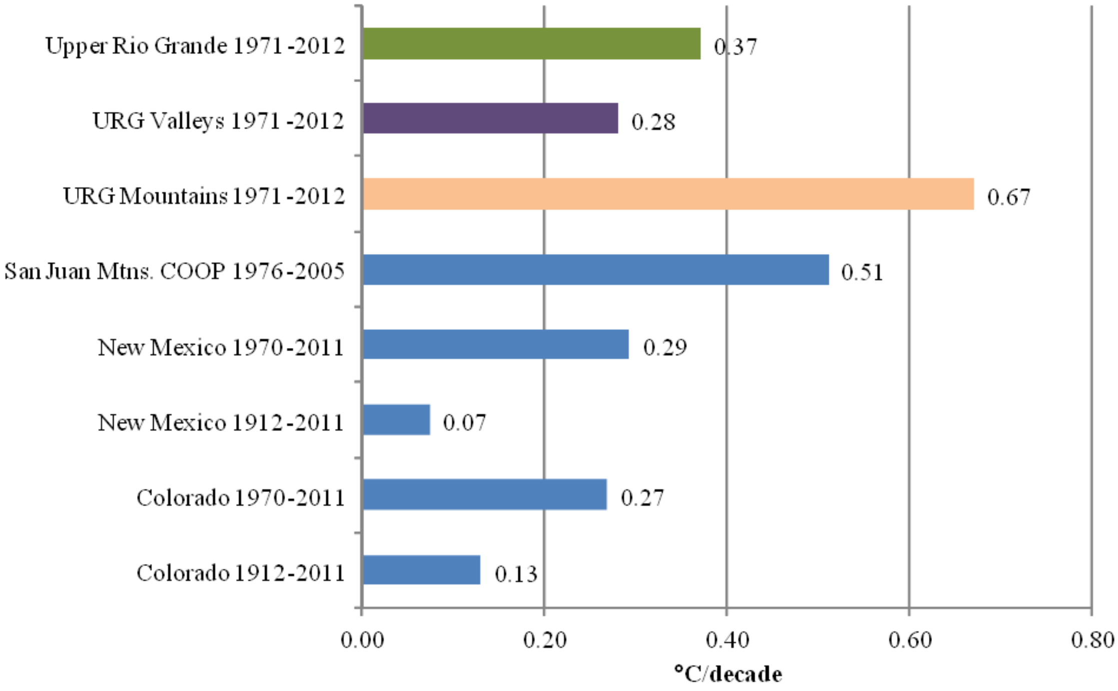
Sources: This study, Rangwala and Miller (2010), Tebaldi et al. (2012).

Figure 12: Comparison of rates of observed change in Tavg with values reported in other studies.



Sources: This study, Rangwala and Miller (2010), Tebaldi et al. (2012).

Figure 13: Comparison of the rates of observed change in Tmax with values reported in other studies.



Sources: This study, Rangwala and Miller (2010), Tebaldi et al. (2012).

Figure 14: Comparison of the rates of observed change in Tmin with values reported in other studies.

The trends in Tavg, Tmin, and Tmax in low elevation settings in the Upper Rio Grande are comparable to those observed by Tebaldi et al. (2012) for the period after 1970, reflecting overlapping datasets. Although the rate of change in Upper Rio Grande mountain Tavg is similar between the two studies, there are large differences in Tmin and Tmax. Mountain Tmax in the Upper Rio Grande is increasing at the relatively slow rate of 0.14°C (0.25°F) per decade, approximately 1/3 the rate of the Tebaldi et al rate of 0.38°C (0.68°F) per decade in New Mexico and 0.29°C (0.52°F) per decade in Colorado. Upper Rio Grande mountain Tmin grew at twice the rate of Tmin increase observed in the Tebaldi et al. (2012) study.

For a broader region encompassing the entire San Juan Mountain Range, Rangwala and Miller (2010) investigated temperature trends using a similar mix of SNOTEL and COOP sites as used in this study, and also computed trends using the Thiel-Sen’s nonparametric slope estimator. The 30-year trend (1976-2005) for NWS COOP data in their study area yielded trends slightly larger than, but comparable to the Upper Rio Grande study. However, the trend estimates for the high elevation SNOTEL sites are much larger than observed in this study. Interestingly, they observe no strong differences in rates of increase in Tmin and Tmax in the data from the San Juan Mountains SNOTEL sites.

The rate of temperature change in the Upper Rio Grande is approximately double that of the world as a whole. A recent study observed a global trend of 0.16°C (0.29°F) per decade for the period 1980-2011, and 0.18°C (0.32°F) per decade for 1990-2011 (Foster and Rahmstorf 2011, Rahmstorf et al. 2012). The observed rate of warming in the Upper Rio Grande basin appears to be in alignment with climate model projections for continental interior regions such as the Southwestern United States under warming scenarios.

iv. Comparison of Observed Trends with Model Projections

Comparison of observed trends with model projections provides a means of assessing the significance of current rates of change, should they continue, with respect to responses of the natural environment. The rates of future change in stream flow and vegetation models are dependent on the rates of change in the climate model(s) driving them. In other words, projections of vegetation and stream flow change for particular decades make critical assumptions about the rate of future change in temperature and precipitation. In short, vegetation and stream flow display a given sensitive to a given amount of temperature and precipitation change, and would change faster if under faster climate change and slower under slower rates of climate change. Thus, it is critical to understand how fast climate is actually changing relative to climate model projections to better understand the likely rates of resulting environmental change.

Table 12 Observed rates of change vs. model projections.

Area	Source	Tavg Change (°C/decade)	Tavg 2010-2039 (°C)	Tavg 2020-2039 (°C)	Tavg 2041-2070 (°C)	Tavg 2050 (°C)	Change in precip. (%)	Notes
Model Projections (SRES scenario)								
Rio Grande Basin (A1B)	Hurd and Coonrod (2007)	--	--	2.35	--	--	-9.07	Dry model, baseline 1971-2000
Rio Grande Basin (A1B)	Hurd and Coonrod (2007)	--	--	1.27	--	--	-0.03	Medium model, baseline 1971-2000
Rio Grande Basin (A1B)	Hurd and Coonrod (2007)	--	--	2.31	--	--	0.97	Wet model, baseline 1971-2000
New Mexico (A1B)	Gutzler et al. (2006)	0.30	0.75	0.90	1.65	--	--	At least 3°C by 2100 ≈0.30C/decade
Colorado (B1, A2B, A2)	Ray et al. (2008)	--	--	--	--	1.4	--	Low estimate, baseline 1950-1999
Colorado (B1, A2B, A2)	Ray et al. (2008)	--	--	--	--	3.1	--	High estimate, baseline 1950-1999
Upper Colorado River Basin (B1)	Ray et al. (2008)	--	--	1.30	--	--	1.00	Difference, baseline period vs. 2020-2039
Upper Colorado River Basin (A2)	Ray et al. (2008)	--	--	1.20	--	--	1.00	Difference, baseline period vs. 2020-2039
San Juan Mountains (A2)	Cozzetto et al. (2011)	--	--	--	2.95	--	-6.75	Median values of model runs
Observed Trends								
New Mexico and Colorado	Tebaldi et al. (2012)	0.12	0.29	0.35	0.67	0.60	--	Average of rates for NM and CO HCN sites, 1912-2011.
New Mexico and Colorado	Tebaldi et al. (2012)	0.31	0.76	0.91	1.72	1.55	--	Average of rates for NM and CO HCN sites, 1970-2011.
San Juan Mountains	Rangwala and Miller 2010	0.45	1.10	1.33	2.50	2.25	-7.56	Average of rates for NWS sites, 1976-2005.
Upper Rio Grande	This report	0.35	0.86	1.03	1.94	1.75	---	Across all HCN, COOP and SNOTEL sites, 1971-2012.

**Estimated from Figure 5-9 (Ray et al. 2008).*

Observed trends in annual temperature are compared to trends projected by models for areas encompassing the Upper Rio Grande (Table 12). If temperatures in the Upper Rio Grande basin continue to rise at the rate of the forty years, average net warming for the period 2010-2039 would be 0.86°C (1.55°F) above the last decades of the 20th century; net warming by 2050 would be 1.94°C (3.49°F) above the last decades of the 20th century. This is the second highest observed rate of change among published studies. Observed rates of change, when multiplied out, are approximately in the middle of the range of model estimates of future warming, reaching approximately 1.75°C (3.15°F) by 2050 and 3.5°C (6.3°F) by 2100.

The observed regional trend is in line with the most recent NARCCAP model projections used in the 2013 National Climate Assessment (USGCRP 2013). These models project that the Upper Rio Grande area will warm 4.1-4.9°C (7.5-8.5°F) by 2070-2099 under the A2 (high emissions) scenario and 2.5-3.1°C (4.5-5.5°F) by 2070-2099 under the B1 (low emissions) scenario.

v. Discussion

The observed trends in temperature indicate warming is occurring at the middle end of model projections. However, whether the true average regional rate of change is 0.35°C/decade, or higher as some models project, warming of 1 to 2.5°C (1.8-4.5°F) by 2040 is likely to exert profound changes on every part of the landscape and is likely to cause significant changes to the availability and quality of surface and ground water in the region. Warming in early spring and late fall contributes to an expansion of the growing season and, therefore, greater transpiration demand and more demand for soil moisture. Declines in soil moisture are likely to contribute to altered fire regimes and changes in vegetation communities, changes that are likely to alter existing rainfall-runoff relationships. Concomitant changes to flood frequency curves and other relationships are likely, with increases in both the frequency of low flow and highest flow years. The current rate of warming is an order of magnitude faster than the rate of warming at the end of the last Ice Age of 0.25 to 0.5°C per century (Porinchu et al. 2005). As during that time, the changes are widely expected to contribute to both species and habitat loss on both global and local scales.

Although mitigation measures may yet reduce net warming by 2100, significant reductions in anticipated warming by 2030 or 2040 are much less likely as much of the warming that will occur in this time frame will be due to greenhouse gases already in the atmosphere. Thus, adaptation will likely be necessary to address climate changes in a region that is likely to be 1 to 2.5°C (1.8-4.5°F) warmer by 2040. Whether the true average regional rate of change is more or less than the 0.35°C/decade observed in this study, such rapid warming is likely to exert profound changes on every part of the landscape and is likely to cause significant changes to the availability and quality of surface and ground water in the region in both the short and long term. Observed warming in early spring and late fall indicates an expansion of the growing season and, therefore, greater transpiration demand and more demand for soil moisture. Declines in soil moisture are likely to contribute to altered fire regimes and changes in vegetation communities, changes that are likely to alter existing rainfall-runoff relationships. Concomitant changes to flood frequency curves and other relationships are likely, with increases in both the frequency of low flow and highest flow years (Reclamation, unpublished data).

Although mitigation measure may yet reduce net warming by 2100, significant reductions in anticipated warming by 2030 or 2040 are much less likely as the majority of the warming that will occur in this time frame will be due to greenhouse gases already in the atmosphere. Thus, climate change impacts to projects in the near term (20-30 year horizon, possibly to 50 years) should be included when making planning decisions.

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