

Chapter 4: Effects of Climate Change on Hydrology, Water Resources, and Soil

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Introduction

Water is critical to life, and many of the effects of climate change on ecosystems are mediated through altered hydrology. Snow accumulation and melt are consistently cited as the most important changes to water in the western United States (Barnett et al. 2005; Service 2004), affecting when water will be available for forests, fish, and people. Changes in summer atmospheric circulation patterns may alter the ability of summer precipitation to provide a midsummer respite from seasonal drought and dampening of wildfire spread (IPCC 2013) (Chapter 8). Declining summer water contributions will challenge municipal and agricultural water supplies (Barnett and Pierce 2009; Dawadi and Ahmad 2012). Aquatic and terrestrial ecosystems—including riparian areas, wetlands, and groundwater-dependent ecosystems—will be affected by lower base flows (Kormos et al. 2016; Rood et al. 2008), earlier snowmelt (Luce et al. 2014), increased periods of drought (Cayan et al. 2009), increased sediment delivery (Goode et al. 2012), and higher midwinter floods (Goode et al. 2013). Soils will likewise be affected by increased temperatures and shifts in precipitation and hydrological processes, with effects on physical and biological processes and attributes of soils.

This chapter describes potential changes to hydrological processes, groundwater resources, and soil attributes and processes in the Intermountain Adaptation Partnership (IAP) region with a changing climate. We specifically discuss potential changes in snowpack and glaciers, streamflow, drought, sediment yield, and groundwater recharge, and in soil temperature, moisture, carbon, nitrogen, biological activity, and chemical properties. The *Soil Resources* section concludes with an example vulnerability assessment method and application.

Hydrological Processes

Climate and Hydrological Processes

Warming temperatures are the most certain consequence of increased carbon dioxide in the atmosphere. The hydrological consequences of warmer temperatures include less

snowpack and greater evaporative demand from the atmosphere. In general, snowpack depth, extent, and duration are expected to decrease, particularly at lower and middle elevations, because of a combination of less precipitation falling as snow (Pierce et al. 2008) and slightly earlier melt (Luce et al. 2014). The degree of change expected as a result of warming varies considerably over the landscape as a function of temperature (Luce et al. 2014). Places that are warm (near the melting point of snow) are expected to be more sensitive than places where temperatures remain subfreezing throughout much of the winter despite warming (Woods 2009).

The relationship of evapotranspiration (ET) to a warming climate is more complicated (Roderick et al. 2014). Warmer air can hold more water, which means that even if relative humidity stays constant, vapor pressure deficit—the difference between actual water content of the air and water content at saturation—increases. That difference between actual and saturation drives a water vapor gradient between leaves and the atmosphere that can draw more moisture out of the leaves. This is likely to cause more evaporation in a warmer climate (Cook et al. 2014; Dai 2013).

However, evaporation is an energy-intensive process, and there is only so much additional energy that will be available for evaporation. In addition, one needs to consider both the water balance and the energy balance when considering future warming (Roderick et al. 2015). The observation that temperatures are warmer during drought is related more generally to the lack of water to evaporate, leading to warmer temperatures, than to warmer temperatures causing faster evaporation (Yin et al. 2014). Unfortunately, when potential ET models based on air temperature (including Penman-Monteith) are applied as postprocessing to global climate model (GCM) calculations, an overestimate of increased ET is likely, because the energy balance is no longer tracked (Milly 1992; Milly and Dunne 2011). The reality is that most of the increased energy from increased longwave radiation will result in warming rather than increased evaporation (Roderick et al. 2015).

Precipitation has a direct effect on hydrological processes, although precipitation is less commonly discussed because climate change projections are uncertain (Blöschl and Montanari 2010; IPCC 2013). Figure 4.1 helps to illustrate how the IAP region is located in an area of high

uncertainty in regard to precipitation projections, slightly overlapping with projected increases to the north and drying to the south (Walsh et al. 2014). The bounds of uncertainty (-20 to +30 percent) are large, making it difficult to accurately project the effects of precipitation on many hydrological processes (e.g., floods, hydrological drought, snow accumulation, groundwater recharge). As a consequence, we use an ensemble average precipitation for streamflow projections here. In this assessment, we also discuss uncertainty surrounding the mean estimate to illustrate which processes or hydrological outcomes are most uncertain and where. Not all processes are sensitive to precipitation, and uncertainty in outcomes caused by uncertainty in precipitation is not the same everywhere for a given process.

Background information can help to clarify where and when some precipitation estimates may be more reliable than others. Two primary concepts are applied for precipitation change: dynamic (referring to changes in wind and atmospheric circulation) and thermodynamic (referring to how much water the air can hold) (Seager et al. 2010). Dynamic drivers of precipitation change include changes in global circulation patterns (e.g., the Hadley cell extent) and changes in mid-latitude storm tracks. Changes in teleconnection patterns (e.g., the North American Monsoon System [NAMS]) fall into this category and are very important for this region. Thermodynamic changes reflect the fact that the atmosphere can hold more water (Held and Soden 2006), leading to an expectation on the order of a 3.9 percent increase in precipitation per 1 °F of temperature change. There are, however, other physical limits on the disposition of energy driving the cycling of water in the atmosphere. These lead to estimates on the order of less than 1 percent per 1 °F of temperature change at the global scale, with individual grid cells being less or potentially negative, particularly over land (Roderick et al. 2014). Different approaches to scaling the thermodynamic contribution are a reason for differences among models, although the dynamic process modeling differences can be great as well.

One outcome of thermodynamically driven changes is that when precipitation occurs, the same total volume is expected to fall with greater intensity, leading to shorter events and longer dry periods between events. The number of consecutive dry days is projected to increase across the western United States, which will affect portions of the IAP region (fig. 4.1). In the Pacific Northwest, this projection is connected to an expected decrease in summer precipitation, but for the Southwest, it is more likely connected to a decrease in monsoonal moisture during the late spring (fig. 4.2). In general, the NAMS is expected to weaken (IPCC 2013), particularly in the early portion of the monsoon season (Cook and Seager 2013), which could have substantial effects across the southern portion of the IAP region. Late spring and early summer precipitation contributions can be an important determinant of the severity of summer drought and length of fire seasons (Abatzoglou and Kolden 2013). Pairing longer periods of precipitation deficit in summer

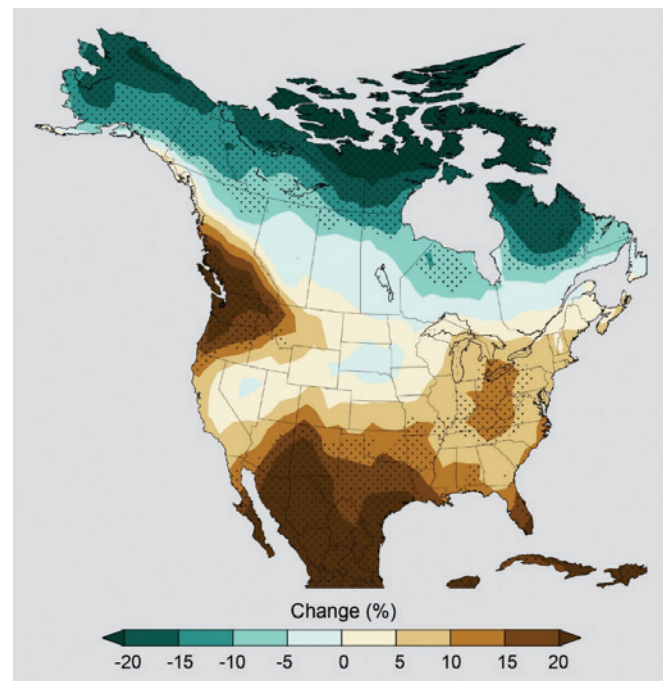


Figure 4.1—Projected change in the number of consecutive dry days for 2071–2099 (compared to 1970–1999) for the RCP8.5 emissions scenario (from Walsh et al. [2014]).

with decreasing snowpack may be particularly challenging for vegetation and aquatic ecosystems.

Changes in orographic enhancement of precipitation over mountainous areas also have dynamic effects. Historical changes in westerly windflows have led to a decrease in the enhancement of winter precipitation by orographic lifting over mountain ranges (Luce et al. 2013), raising the question of whether such a pattern may continue into the future. There is general agreement among GCMs projecting further decreases in windspeed into the 21st century, but the correlation is applicable only to the northern portion of the IAP region. Westerly winds are strongly correlated with precipitation in mountainous areas (fig. 4.3), but valley precipitation is not, nor is precipitation in much of southern Idaho. The historical trend in westerlies was driven by pressure and temperature changes spatially consistent with those expected under a changing climate; however, the rapidity of the changes in the last 60 years may have been partly enhanced by normal climatic variability.

Dynamic downscaling using a regional climate model (RCM) with small (~8 mile) cells provides a means to estimate orographically induced precipitation, which cannot be simulated with the large cell size of GCMs. Although the GCM shows general moistening over most of the area, the RCM shows a pattern of drying or no change on the upwind side of major mountain ranges, with moistening limited to valleys in the lee. Because mountainous areas are where most of the precipitation falls (and streamflow originates), this is a potentially important aspect of future changes. The Variable Infiltration Capacity (VIC) model simulations

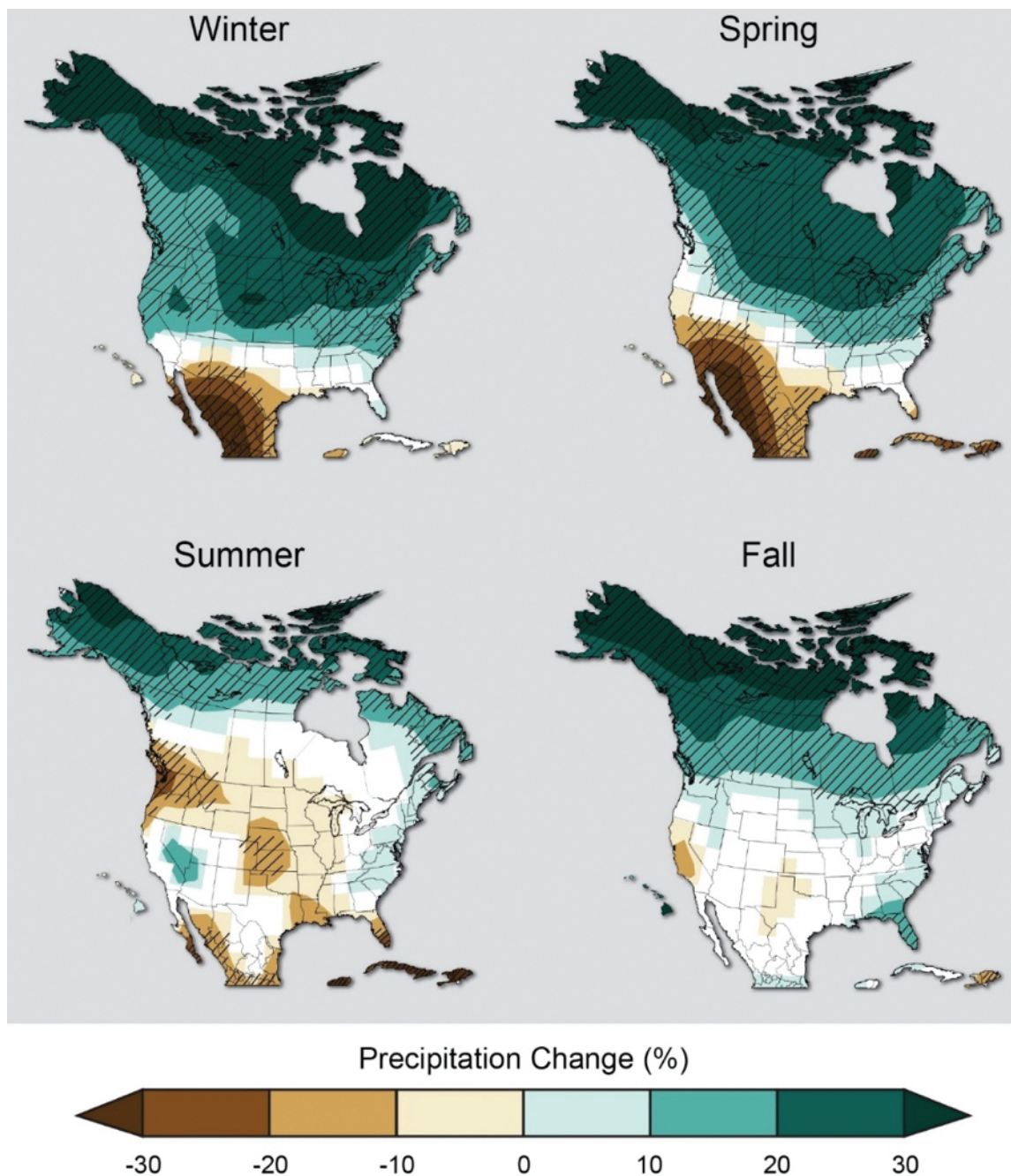


Figure 4.2—Projected change in seasonal precipitation for 2071–2099 (compared to 1970–1999) under the RCP 8.5 emissions scenario (Walsh et al. 2014). Hatched lines indicate model agreement and nonhatched areas have the highest uncertainty. The Intermountain Adaptation Partnership region sits in an area of high uncertainty, between projected moistening to the north and drying to the south.

discussed later in this chapter do not include this effect, so for purposes of general discussion, it can be considered an additional source of uncertainty for precipitation.

The range of potential changes to climate is complex, particularly for such a varied landscape as the IAP region (fig. 4.2), and current climatological settings vary over the landscape at both large and small spatial scales. Precipitation seasonality and amount differ between mountain and valley locations. Trends and drivers for climatic

variation will also differ from the northern to southern portions of the region. Fundamentally, topography is an important factor affecting seasonality, precipitation amount, and potential trends. Because most forests and generation of water supply are generally in mountainous areas, it is important to recognize how topography affects climate. Specific hydrological outcomes of interest are discussed in the following sections.

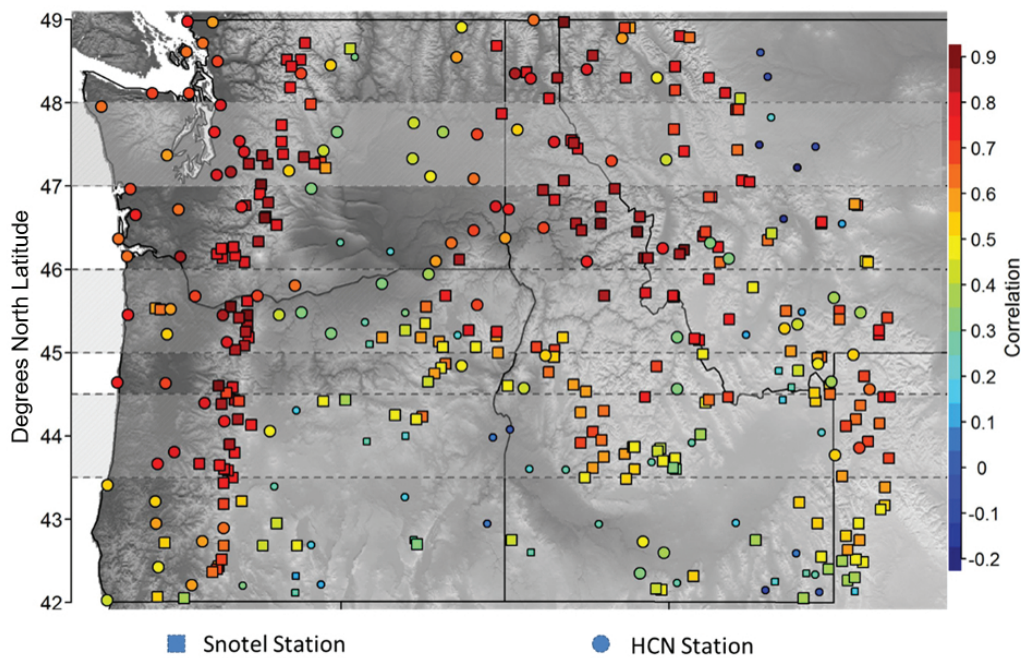


Figure 4.3—Correlation of winter precipitation to winter westerly windspeed across the Pacific Northwest, showing snowpack telemetry (SNOTEL) and Historical Climate Network (HCN) stations (from Luce et al. [2013]).

Snowpack and Glaciers

Snowpack

Snowpack declines are among the most widely cited changes occurring with climate change, through the effect of warmer temperatures on the fraction of precipitation falling as snow (Barnett et al. 2008). About 70 percent of the water supply in the western United States is tied to mountain snowpacks (Service 2004), so changes in snowpack are highly relevant to municipal and agricultural water supplies and timing (Stewart et al. 2005).

Historical trends in snowpack accumulation have been negative across most of the western United States (Mote et al. 2005; Regonda et al. 2005). However, care must be taken when looking at individual sites, which can be influenced by site-specific effects such as vegetation changes, physical site changes, sensor changes, and measurement technique (Clayton and Julander 2015). Temperature sensitivity of the snowpack is highest in places that are already relatively warm (warm snowpacks), and warm snowpacks with high precipitation are likely to undergo some of the largest changes in snow storage as the climate warms (Luce et al. 2014; Nolin and Daly 2006).

The most sensitive locations within the IAP region include the eastern Sierra Nevada and mid- to lower-elevation sites across Idaho, Utah, and Nevada (figs. 4.4–4.7). In contrast, many interior portions of the IAP region are cold enough to be relatively insensitive to warming and strongly sensitive to precipitation variation (Luce et al. 2014; Mote 2006). At the coldest and highest elevations, in the Uinta, Teton, Wind River, and some central Idaho ranges, for instance, there could be increases in snow water equivalent (SWE) if precipitation increases (Rice et al. 2017). Despite warming temperatures, a large proportion of precipitation would still fall as snow in these areas. This means that

the future of snow, and consequently hydrology in these regions, depends on one of the more uncertain parts of GCM projections: the precipitation.

Precipitation uncertainty can be substantial, but it does not translate into equal uncertainty in snowpack changes everywhere (fig. 4.8). We estimated sensitivity of April 1 SWE using data from 524 snowpack telemetry (SNOTEL) stations across the western United States in a space-for-time model (Luce et al. 2014). This allowed us to determine where in the western United States snowpack was more sensitive to variability in precipitation or variability in temperature. We computed an index of uncertainty as the ratio (R_u) of the effects on snow (ΔS) from the likely range of precipitation values (about ± 7.5 percent for 1 standard deviation across models) in the numerator, to ΔS from the relatively certain temperature change in the denominator:

$$R_u = \frac{\Delta S \text{ across precipitation uncertainty } (+7.5\%)}{\Delta S \text{ expected from warming}}$$

We found strong certainty of large changes in April 1 SWE for the Cascade Range, Sierra Nevada, and the Southwest ($R_u < 0.2$). But we found substantial uncertainty ($R_u > 0.6$) in outcomes for interior locations such as the Greater Yellowstone Area and higher elevations in Idaho and Utah, where cold temperatures leave the snowpack more sensitive to precipitation than to temperature changes (fig. 4.8). The uncertainty ratio in these colder areas suggests that relatively large increases in precipitation could help counter the effects of warming on snowpack loss. These results are similar to those seen using a physically based model across the western United States (Gergel et al. 2017).

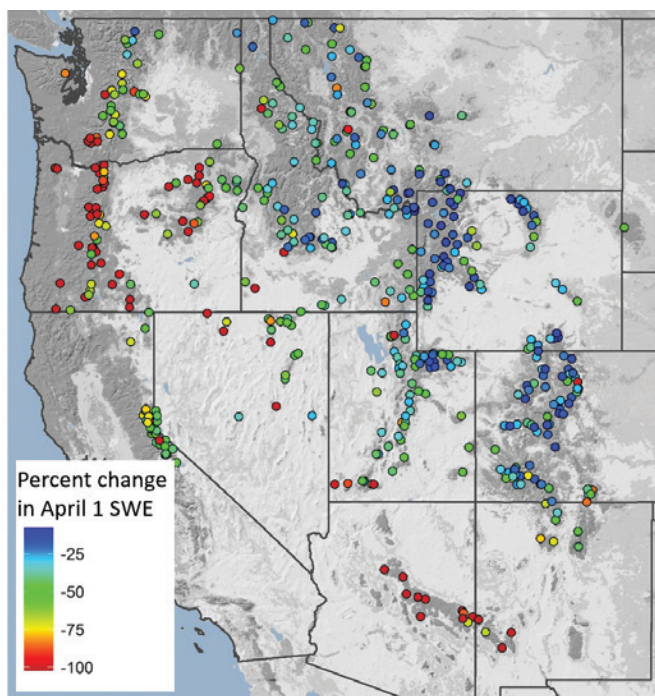


Figure 4.4—Estimated April 1 snow water equivalent (SWE) sensitivity (percentage change) for a 5.5-°F increase in winter average temperature at each snowpack telemetry station (modified from Luce et al. [2014]).

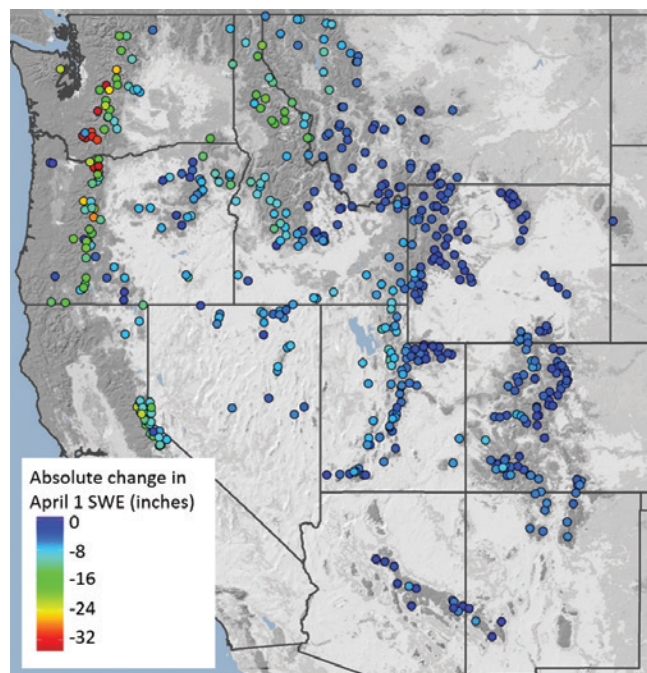


Figure 4.5—Estimated April 1 snow water equivalent (SWE) sensitivity (absolute change in inches) for a 5.5-°F increase in winter average temperature at each snowpack telemetry station (modified from Luce et al. [2014]).

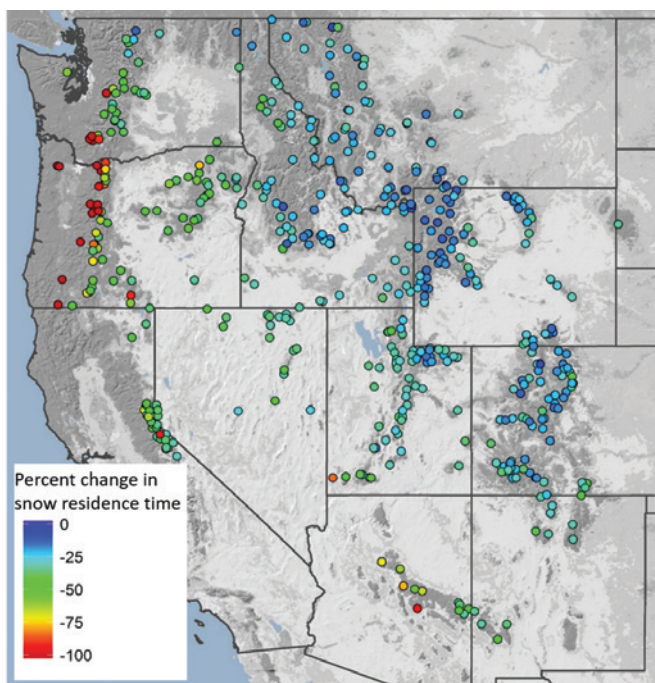


Figure 4.6—Estimated mean snow residence time sensitivity (percentage change) for a 5.5-°F increase in winter average temperature at each snowpack telemetry station (modified from Luce et al. [2014]).

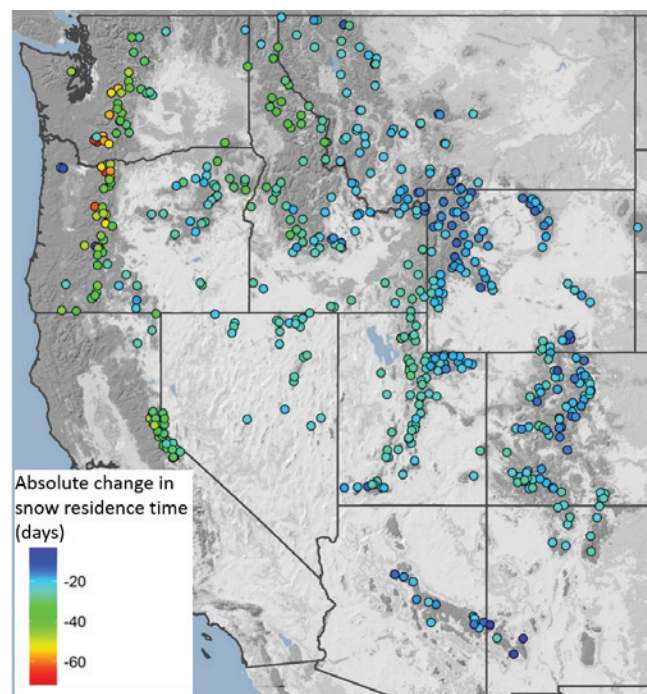


Figure 4.7—Estimated mean snow residence time sensitivity (absolute change in days), for a 5.5-°F increase in winter average temperature at each snowpack telemetry station (modified from Luce et al. [2014]).

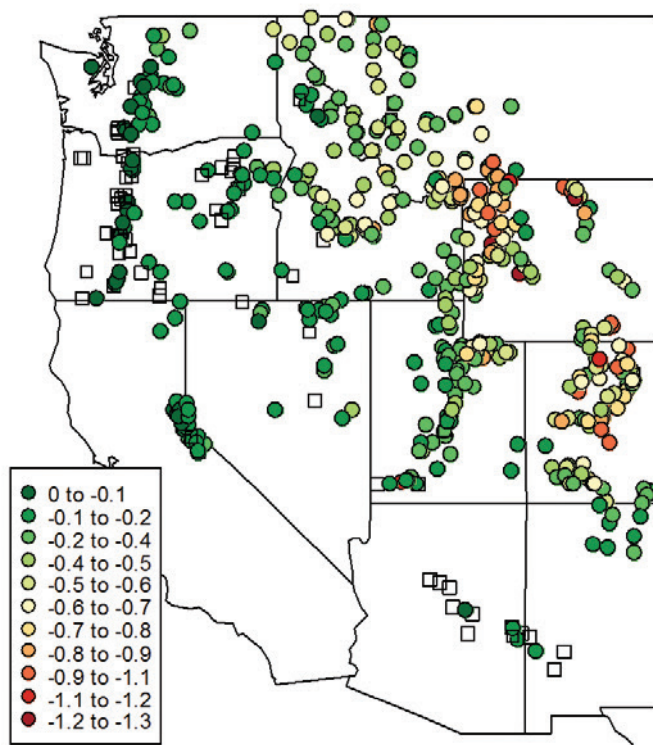


Figure 4.8—Uncertainty ratio for April 1 snow water equivalent. Orange to dark red sites are strongly influenced by precipitation in contrast to temperature. Thus, temperature-based projections for snow water equivalent in those sites may be inaccurate if precipitation changes are large. At dark green (and white) sites, temperature effects will predominate, and precipitation changes in either direction are inconsequential.

Glaciers

Glaciers are limited throughout the IAP region but do occur in central Idaho, in western Wyoming, and in an isolated location at Wheeler Peak in Great Basin National Park (see maps at Portland State University 2009). Declines in the extent of glaciers in the Wind River Range have been observed over the 20th century (Marston et al. 1991).

Estimating future changes in glaciers is complex (Hall and Fagre 2003), but empirical relationships derived for glaciers indicate a brief future for them, with many glaciers becoming fragmented or disappearing by the 2030s. Increasing temperatures yield a rising equilibrium line altitude (ELA), decreasing the effective contributing area for each glacier as warming progresses. Warming of 5.5 °F can translate to an elevation rise of 1,000 to 1,600 feet in snow-rain partitioning and summer temperatures. Those changes do not directly equate to a shift in ELA, which depends on the geometry and topography of the contributing cirque.

Temperate alpine glaciers are well known for being as, or more, sensitive to precipitation variations as they are to temperature variations (McCabe and Fountain 1995), which has very likely contributed to changes in glacial dynamics across the Pacific Northwest. Westerlies and their

contribution to winter precipitation have changed over the northern part of the region since the 1940s (Luce et al. 2013), and April 1 SWE at these elevations and latitudes is relatively insensitive to temperature. However, summer temperature is a strong predictor of glacial behavior, and changes in summer temperatures could affect the melt rate and additional snow contributions in glaciers because this area receives significant spring and summer precipitation (Hall and Fagre 2003).

Streamflow

Streamflow changes of significance for aquatic species, water supply, and infrastructure include annual yield, summer low flows (average, extreme), peakflows (scouring floods), peakflow seasonality, and center of runoff timing. Irrigation water for crops and urban landscapes is typically needed in summer months. Annual yield, summer low flows, and center of runoff timing are important metrics with respect to water supply, but they are most relevant to surface water supplies rather than groundwater supplies, although changes in long-term annual means could be informative for the latter. The mean summer yield (June through September) is used for summer low flows. Center of runoff timing is the date on which 50 percent of the annual runoff has flowed out of a basin and is an effective index for the timing of water availability in snowmelt-driven basins. Shifts to earlier runoff in the winter or spring disconnect streamflow timing from water supply needs such as agricultural irrigation. Center of timing can be redundant with other metrics that measure impact more directly, but with care in interpretation, it can help clarify different potential causal mechanisms, such as changing precipitation versus changing temperature.

Peakflows are important to fish and infrastructure. Scouring flows can damage eggs in fish redds if they occur while the eggs are in the gravel or alevins are emerging (Goode et al. 2013; Tonina et al. 2008). Winter peakflows can affect fall-spawning fish (chinook [*Oncorhynchus tshawytscha*], bull trout [*Salvelinus confluentus*], and brook trout [*S. fontinalis*]), whereas spring peakflows affect spring-spawning cutthroat trout (*O. clarkii*), resident rainbow trout (*O. mykiss*), and steelhead (*O. m. gairdneri*) (Wenger et al. 2011a,b). Spring peakflows associated with the annual snowmelt pulse are typically muted in magnitude compared to winter rain-on-snow events for two reasons. The rain-on-snow events can generate larger water input rates (rainfall precipitation plus high melt rate), and they tend to affect much larger fractions of a basin at a time, so scouring is less of a risk to spring-spawning fishes. Consequently, a shift to more midwinter events can yield higher peakflow magnitudes, which can also threaten infrastructure such as roads, recreation sites, and water management facilities (e.g., diversions, dams).

Historical changes in some of these streamflow metrics have been examined in northern portions of the IAP region, specifically earlier runoff timing (Cayan et al. 2001; Stewart

et al. 2005) and declining annual streamflows (Clark 2010; Luce and Holden 2009). Declining low flows (7Q10) have also been observed in the western half of the Northern Rockies (Kormos et al. 2016), associated more with declining precipitation than warming temperature effects for the historical period. Low-flow changes and timing changes in projections are generally associated with expected changes in snowpack related to temperature (e.g., more melt or precipitation as rain in winter, yielding a longer summer dry period). Low-flow changes driven by these timing changes are strongly dependent on groundwater conditions in the basin (Tague and Grant 2009), which vary considerably across the IAP region as discussed later in the Groundwater Resources section.

Streamflow Projections

Streamflow projections for an ensemble of Coupled Model Intercomparison Project Phase 3 (CMIP3) models under the A1B scenario (Littell et al. 2011) were produced from the VIC model (Liang et al. 1994) for the western United States (University of Washington, Climate Impacts Group 2017)) (figs. 4.9–4.13). Differences between the climate described by CMIP3 projections and the more recently developed CMIP5 projections are minimal with respect to temperature (Chapter 3). The gridded VIC data were used to estimate streamflow by using area-weighted averages of runoff from each VIC grid cell within a given basin, following the methods of Wenger et al. (2010), to accumulate flow and validate. Streamflow metrics were calculated for stream segments in the National Hydrography Dataset Plus (version 2) stream segments (USDA FS n.d.).

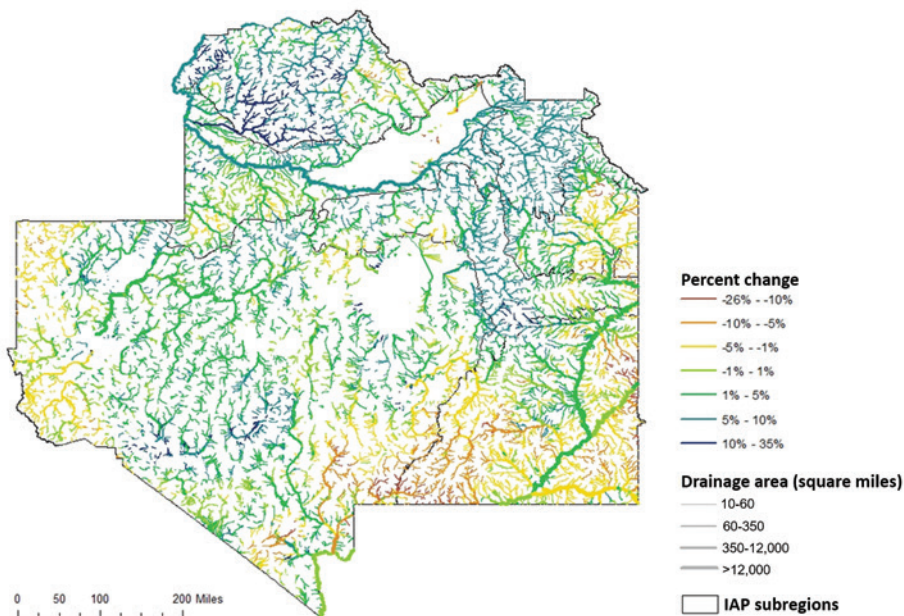


Figure 4.9—Percentage change in mean annual flow projections in the Intermountain Adaptation Partnership (IAP) region between a historical period (1970–1999) and the 2040s. Projections are from the Variable Infiltration Capacity model, following the methods of Wenger et al. (2010).

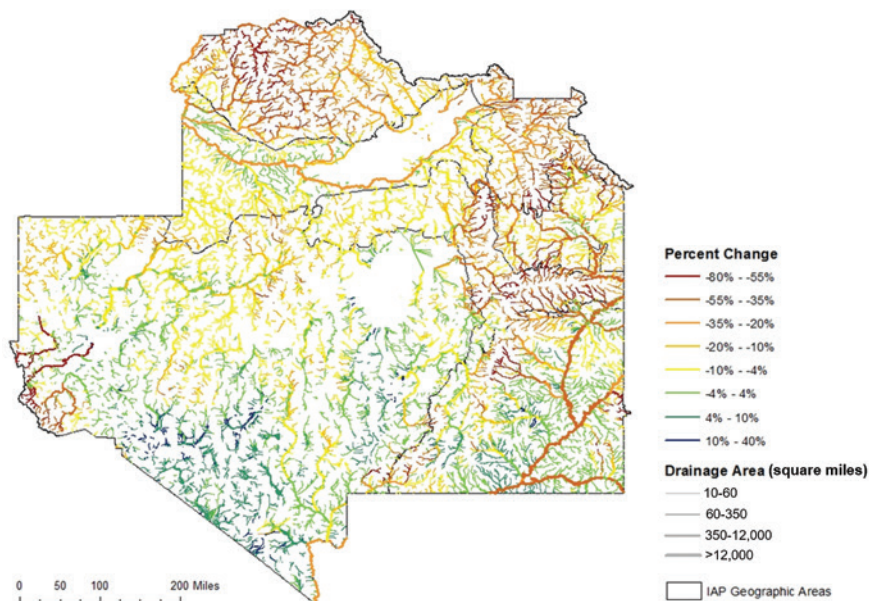


Figure 4.10—Percentage change in mean summer flow projections in the Intermountain Adaptation Partnership region between a historical period (1970–1999) and the 2040s. Projections are from the Variable Infiltration Capacity model, following the methods of Wenger et al. (2010).

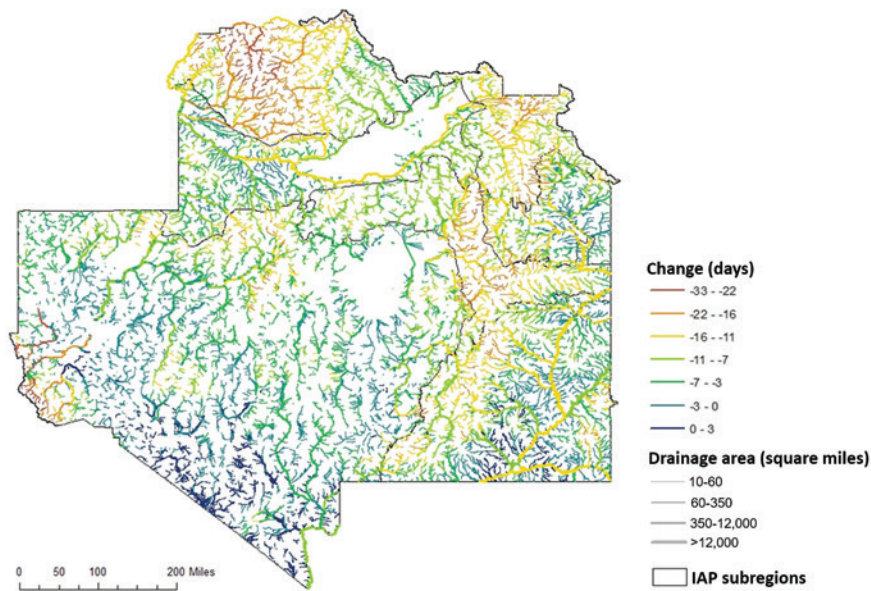


Figure 4.11—Change (days) in the center of flow mass projections in the Intermountain Adaptation Partnership region between a historical period (1970–1999) and the 2040s. Projections are from the Variable Infiltration Capacity model, following the methods of Wenger et al. (2010).

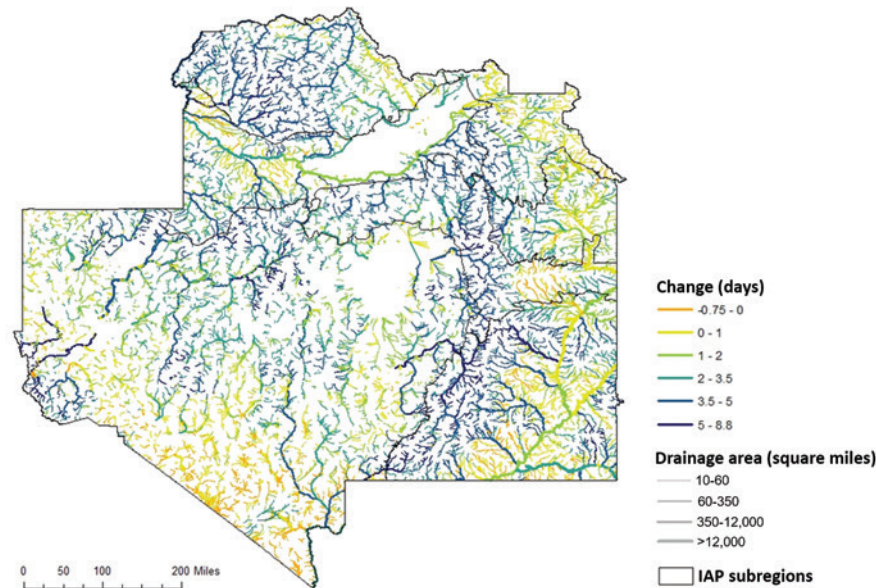


Figure 4.12—Projections of change (days) in the number of mid-winter floods (95th-percentile flow) in the Intermountain Adaptation Partnership region between a historical period (1970–1999) and the 2040s. Projections are from the Variable Infiltration Capacity model, following the methods of Wenger et al. (2010).

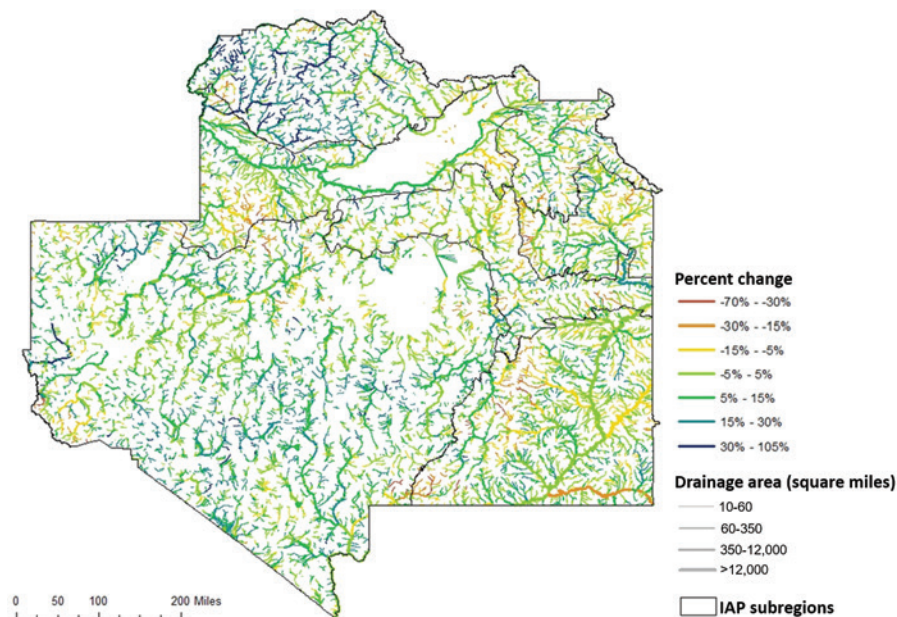


Figure 4.13—Percentage change in 1.5-year flood magnitude (approximate "bankfull" flow) in the Intermountain Adaptation Partnership region between a historical period (1970–1999) and the 2040s. Projections are from the Variable Infiltration Capacity model, following the methods of Wenger et al. (2010).

Uncertainty in climate model inputs can be a significant factor in uncertainty for outcomes related to natural resources (Wenger et al. 2013). Downscaling for these runs was done statistically, not dynamically, using an RCM to account for orographic enhancement changes, so GCM expectations for precipitation are implicit in the streamflow estimates. No effects of change in orographic enhancement are inherent in these images; thus, uncertainty may be higher (in a drier direction) on the windward side of mountain ranges.

Mean annual flow projections (fig. 4.9) suggest a slight increase across the northern portion of the IAP region, which ties back to the general moistening predicted by CMIP3 GCM runs (also illustrated in fig. 4.2). Minor changes are displayed through the central part of the region. The decreases shown in the southern part of the region are associated with changes in the Hadley cell circulation, which has also been described as an expansion of the mid-latitude deserts.

Despite projections of slightly increased annual flow across much of the region, summer low flows are expected to decline (fig. 4.10), with relatively uniform changes in mountainous areas, particularly in wetter ranges. The primary mechanism expected to drive lower summer base flows is reduced snowpack in winter, leading to less stored water. The VIC model simulations do not include the effects of large groundwater reserves; thus, this effect could be moderated in systems where groundwater flow contributes a substantial volume of water to late summer flows (see more discussion in the *Groundwater Resources* section on where this may be important). Although such groundwater support could moderate the percentage declines as shown in figure 4.10, actual low-flow runoff rates could have greater declines in such places because the fractional decline is applied to a larger pre-change low-flow rate (e.g., Tague and Grant 2009). This is an important consideration when dealing with water rights, in which actual volumes or flows, rather than percentages, are allocated to individual rights.

Places where summer precipitation plays an important role, particularly the southern portion of the region, are more likely to see low flows affected by summer precipitation patterns. Shifts in circulation that affect how moisture flows from the Gulf of Mexico in summer are expected to negatively affect precipitation. Increased spacing between precipitation events (IPCC 2013; Luce et al. 2016) and decreased moisture in the early portion of the monsoon season (Cook and Seager 2013) are other likely occurrences. These summer wet areas are also more likely to have greater losses of precipitation with increased evaporation, but it is important to recognize energy balance constraints when estimating the degree of loss (Roderick et al. 2014). This is not done in the VIC modeling, which uses only the temperature outputs from GCMs without reevaluating the change in energy balance from a different hydrological formulation; loss by evapotranspiration may thus be overestimated (Milly and Dunne 2011).

Generally, areas showing a change in summer low flows also show a shift to earlier center of flow mass timing (2–4

weeks) (fig. 4.11) and a shift to stronger changes in mountains dominated by snowmelt runoff. Changes in timing are related to snow residence time and earlier snowmelt runoff (figs. 4.6, 4.7).

Projected changes in the number of winter floods (95th percentile flow) show more of an effect in mid- and lower-elevation mountain ranges. Higher elevation and colder ranges, which will preserve more snowpack, show less change (fig. 4.12). The shift to more midwinter rain and more rain-on-snow flooding depends strongly on the elevational range of each basin. At middle elevations, temperatures are projected to increase enough that rain is likely on snowpacks, even in midwinter. Consequently, projected peakflow increases are generally stronger in these mid-elevation mountainous areas (fig. 4.13). Greater midwinter flooding could increase both the occurrence and magnitude of peakflows (fig. 4.14), as well as the potential for scour in gravel riverbeds (Goode et al. 2013).

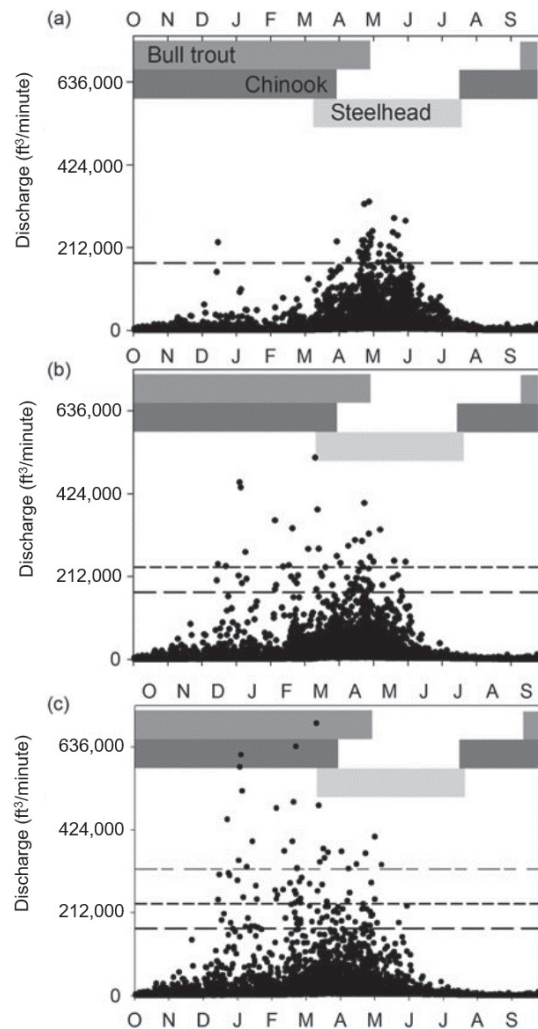


Figure 4.14—Illustration of increased mid-winter flooding potential. Projected streamflows are from Variable Infiltration Capacity modeling for (a) current conditions, (b) 2040s, and (c) 2080s. The long, short, and gray dashed lines indicate the 2-year flood for each period (current, 2040s, and 2080s, respectively) (from Goode et al. [2013]).

Drought

Several studies help to provide a paleoclimatic context for evaluating drought in the IAP region. For example, both an early and an updated reconstruction of streamflow in the Colorado River Basin indicate that water allocation agreements were developed during one of the wettest periods in the last 500 years (Stockton and Jacoby 1976; Woodhouse et al. 2006), and that droughts were more severe before the 20th century (Woodhouse et al. 2006). Similarly, DeRose et al. (2015) found that in the Bear River of the Great Basin, the latter half of the 20th century was the second wettest period in the last 1,200 years. Other studies have also demonstrated high variability and severe droughts in the Uinta Mountains (MacDonald and Tingstad 2007), Weber River (Bekker et al. 2014), Logan River (Allen et al. 2013), and Great Salt Lake (Wang et al. 2012). Figure 4.15 illustrates a general correlation in wet and dry cycles between these basins over time, but also some unique differences based on onsite-specific factors (DeRose et al. 2015).

Understanding long-term climate dynamics is critical for sustainable management of environmental resources. In combination with projections for climate change, knowledge of past climatic conditions can help inform water and land management decisions. For a more extensive discussion of drought, paleoclimatic history, and effects on forests and streams, see Luce et al. 2016.

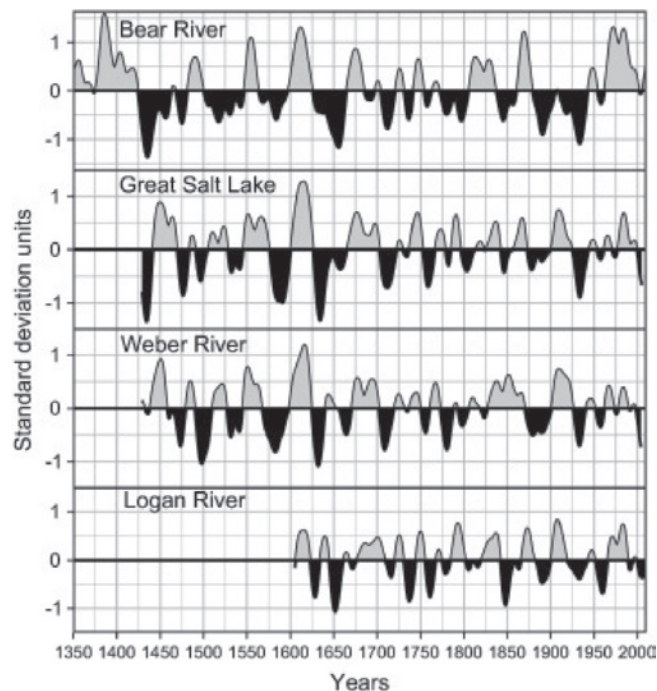


Figure 4.15—Comparison between the Bear River and other Wasatch Front hydroclimate reconstructions, illustrating the cyclical nature of wet and dry periods in the Intermountain Adaptation Partnership region (from DeRose et al. [2015]).

Sediment Yield

The delivery and transport of sediment through mountain rivers affect aquatic habitat and water resource infrastructure. Although climate change is expected to produce significant changes in hydrology and stream temperature, the effects of climate change on sediment yield have received less attention. Climate change is expected to increase sediment yield primarily through the effects of temperature and hydrology on vegetation disturbances (wildfire, insects, drought-related mortality) (Goode et al. 2012).

A conceptual model (fig. 4.16) of sediment yield (solid black line) relative to climate can help to illustrate the regulating role of vegetation. The dashed lines indicate the relative increase in resistance to erosion that vegetation provides as the driving force of precipitation increases. The biggest divergence in the lines occurs in semiarid climates where sufficient precipitation is available to drive erosion, but there is a limited amount of vegetation to stabilize hillslopes from erosion. The result is higher sediment yield in semiarid climates. The red arrow and circle on the plot depict the potential shift of current temperate forest climates to more semiarid climates, increasing overall erosion potential and sediment yield (Goode et al. 2012) (fig. 4.16).

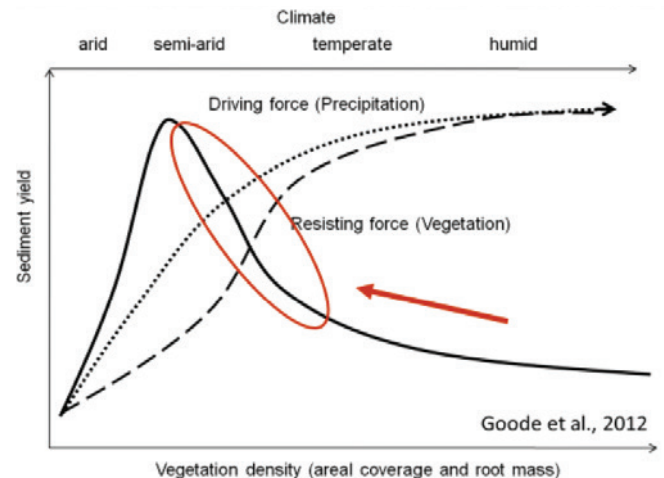


Figure 4.16—Conceptual plot of sediment yield (solid black line) relative to hydroclimate and the regulating role of vegetation. The red arrow and circle illustrate the potential shift of current temperate forest climates, which under a warming climate could become more like semiarid climates, increasing erosion potential and sediment yield (from Goode et al. [2012]).

Groundwater Resources

Climate change is likely to have significant, long-term implications for groundwater resources in the IAP region. Climate change is expected to cause a transition from snow to rain, resulting in diminished snowpack, shifts in streamflow to earlier in the season (Leibowitz et al. 2014; Luce et

al. 2012), and changes in groundwater recharge to aquifers and groundwater discharge to groundwater-dependent ecosystems (GDEs). In this section, we synthesize existing information about occurrence of groundwater resources in five of the six subregions, describe potential effects of climate change, and describe how climate change can affect GDEs, including aquifers, streams, wetlands, and springs.

Groundwater is broadly defined as “all water below the ground surface, including water in the saturated and unsaturated zones” (USDA FS 2012). Groundwater resources include water residing in the subsurface, as well as ecosystems that depend on the presence or discharge of groundwater.

Groundwater-dependent ecosystems are “communities of plants, animals and other organisms whose extent and life processes are dependent on access to or discharge of groundwater” (USDA FS 2012). In the IAP region, GDEs include springs, springbrooks, groundwater-supported lakes, fens, streams, and rivers with base flow and riparian wetlands or phreatophytic vegetation along segments known as “gaining river reaches.” **Fens** are wetlands supported primarily by groundwater with a minimum depth (usually 12–16 inches) of accumulated peat (Chadde et al. 1998; USDA FS 2012a). **Springs** are entirely supported by groundwater. These GDEs contribute significantly to local and regional biodiversity (Murray et al. 2006).

The fundamental hydrological processes that influence GDEs are: (1) amount, timing, and type of precipitation (rain or snow); (2) groundwater recharge; (3) groundwater quality; (4) groundwater discharge; and (5) evapotranspiration (Lins 1997). Along stream segments referred to as “gaining reaches,” groundwater enters the stream from the banks or the channel bed, and the volume of downstream streamflow is subsequently increased (Winter 2007; Winter et al. 1996). Groundwater can contribute substantially to late summer streamflow (Gannett 1984) and is the source for cool-water upwellings that serve as refugia for cold-water aquatic species (Lawrence et al. 2014; Torgersen et al. 1999, 2012).

Hydrogeologic Setting

Hydrogeologic setting provides a context for assessing potential climate-induced changes to groundwater resources. Geologic units respond differently to changes in precipitation because of differences in hydraulic conductivity, transmissivity, primary versus secondary porosity, and fracture patterns. In a study that combined aerial photography (over 50–80 years) and climate analysis, Drexler et al. (2013) showed that five fens in the Sierra Nevada (California) decreased 10 to 16 percent in area. This decrease in area occurred over decades with high mean minimum air temperature and low SWE and snowpack longevity. However, two fens in the southern Cascade Range, underlain by different geology than the Sierra Nevada, did not change in area, suggesting that the hydrogeologic setting plays an important role in mediating GDE functionality.

Several different hydrogeologic settings have been delineated in the IAP region, including igneous/metamorphic, sedimentary, karst, and unconsolidated sediments. Igneous and metamorphic rocks with low permeability and porosity, with low-volume groundwater discharges, and that are recharged only during large infrequent precipitation or snowmelt events may not be vulnerable to changes in temperature and precipitation. However, aquifers in sedimentary formations, karst formations (fig. 4.17), and unconsolidated sediments may be more sensitive to climate change because they have high permeability, high porosity, and larger volume discharges to GDEs.

Groundwater Systems in the Intermountain Adaptation Partnership Region

Middle Rockies Subregion

Located in central Idaho, the Middle Rockies subregion is underlain with predominantly igneous and volcanic rocks with carbonate (fig. 4.17) and other sedimentary rocks in the southeast. Groundwater occurs in fractured and weathered crystalline rocks and sedimentary rocks (USGS 2000). Sand and gravel aquifers are found in floodplains and terraces in the valleys. Bedrock aquifers are the only source of groundwater across much of the subregion. Igneous, metamorphic, and sedimentary rocks that underlie the mountains generally yield little water to wells. Recharge to the basin aquifer system is by precipitation that falls directly on basin floors and by snowmelt that runs off the surrounding mountains and is transported into the basins by tributary streams.

Southern Greater Yellowstone Subregion

The Southern Greater Yellowstone subregion in western Wyoming and eastern Idaho consists primarily of sedimentary rocks but also contains igneous and metamorphic rocks (USGS 2000). Aquifers include sedimentary rocks, sand, and gravel along streams and basin-fill aquifers adjacent to the mountain blocks. Groundwater occurs in pore spaces, joints, fractures, faults, and solution openings in carbonate rocks. Many basins are bounded by mountain front faults. The most important aquifers are basin-fill aquifers, but they are recharged mainly from the mountain blocks. Deposits that fill the basins are mostly alluvium derived from the weathering and erosion of consolidated rocks that underlie the mountains bordering the basins. Primary recharge areas are generally located along the mountain fronts and extend into some mountain valleys. Groundwater is obtained principally from the basin-fill deposits. Basin margin faults are likely to influence flow from the bedrock aquifers to the basin-fill aquifers by forming barriers and highly permeable pathways, depending on the fault-zone geometry.

Uintas and Wasatch Front Subregion

Consolidated rocks of Precambrian to Tertiary age, which form the Wasatch Range and other mountain ranges in the Uintas and Wasatch Front subregion, yield water chiefly through complex systems of fractures, joints, solution

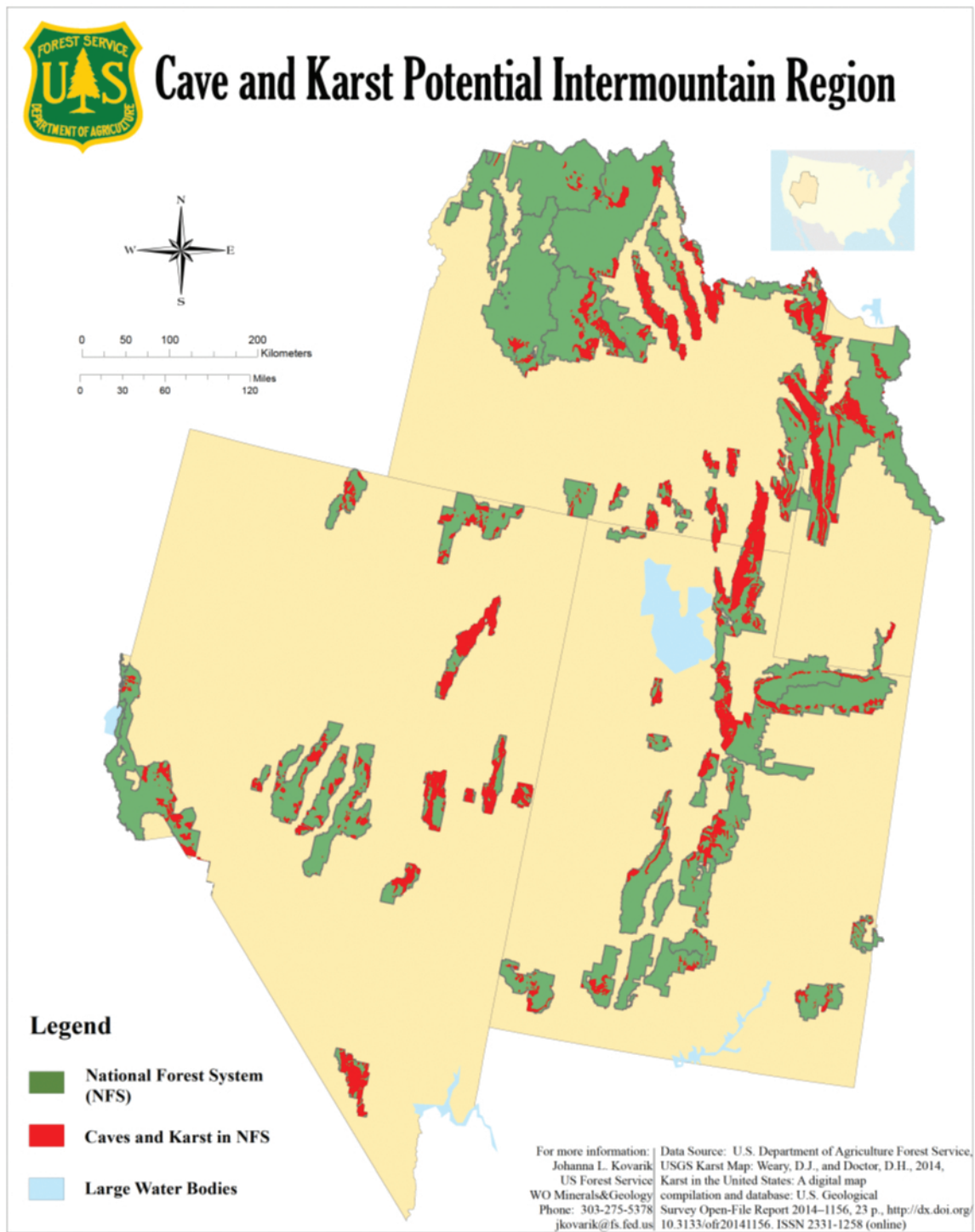


Figure 4.17—Areas with cave and karst potential composed of carbonate and minor volcanic rocks in the U.S. Forest Service Intermountain Region (from Weary and Doctor [2014]).

cavities, fault zones, and vesicles (Price 1985). These water-bearing zones, which are not present at all locations, are difficult to find and delineate. Wells in consolidated rocks also have small yields, and the depth to the saturated zone can be great. Consequently, the consolidated rocks in the Wasatch Front area are not considered to be favorable sources of water for withdrawal from wells. As a unit, however,

they do absorb, store, and transmit large volumes of water to downstream aquifers. This is particularly true for carbonate units (fig. 4.17). In Utah, the aquifers in Cache Valley, the lower Bear River area, and along the Wasatch Front provide water to 84 percent of the population of Utah.

Geologic conditions vary considerably throughout the Wasatch Front area, and thus, groundwater occurrence,

movement, quality, and availability also vary. Most of the wells that obtain water from the consolidated rocks are used for domestic supply and produce only a few gallons of water per minute. However, some of the springs that discharge from these rocks (especially carbonate rocks) produce several hundred to more than 1,000 gallons per minute. Alluvial fans make up much of the valley fill near mountain fronts.

Plateaus Subregion

Colorado Plateau aquifers underlie the Plateaus subregion of eastern Utah. Principal aquifers in the Colorado Plateau include the Uinta-Animas, Mesaverde, Dakota-Glen Canyon, and Coconino-De Chelly (USGS 2000). Distribution of aquifers on the Colorado Plateau is controlled in part by the structural deformation and erosion that have occurred since deposition of the sediments composing the aquifers. Much of the land is underlain by rocks that contain aquifers capable of yielding usable quantities of water of quality suitable for most agricultural or domestic use. In general, the aquifers in the Colorado Plateau area are composed of permeable, moderately to well-consolidated sedimentary rocks. These rocks range in age from Permian to Tertiary and vary greatly in thickness, lithology, and hydraulic characteristics.

Relatively impermeable confining units separate each of the four principal aquifers in the Colorado Plateau. The two thickest units are the Mancos (underlying the Mesaverde aquifer) and Chinle-Moenkopi (underlying the Dakota-Glen Canyon aquifer). Groundwater recharge to the Uinta-Animas aquifer generally occurs at higher elevations along the margins of each basin. The Mesaverde aquifer is at or near the land surface in extensive areas of the Colorado Plateau.

Great Basin and Semi Desert Subregion

The Great Basin and Semi Desert Subregion lies within the Basin and Range Physiographic Province, which contains three principal aquifer types: (1) volcanic-rock aquifers, which are primarily tuff, rhyolite, or basalt of Tertiary age; (2) carbonate-rock aquifers (fig. 4.17), which are primarily limestones and dolomites of Mesozoic and Paleozoic age; and (3) basin-fill aquifers, which are primarily unconsolidated sand and gravel of Quaternary and Tertiary age (USGS 2000). These aquifers underlie most of Nevada, western Utah, and southern Idaho. All the precipitation that falls in the area is returned to the atmosphere by ET, and streams do not carry water to the oceans.

Fracturing in carbonate rocks (limestone and dolomite) may enable groundwater to circulate through the fractures where the water can dissolve the slightly soluble rock and enlarge and increase the size and number of pathways for water movement through the rock. Such dissolution eventually can change a relatively impermeable carbonate rock into a permeable water-yielding unit. Carbonate rocks predominate in a 20,000- to 30,000-foot thick sequence of Paleozoic and Lower Mesozoic rocks in an extensive area of southern and eastern Nevada and are present on all National

Forests in the region. The location of solution-altered zones of enhanced permeability within these carbonate rocks is poorly known. Although extrusive igneous rocks (primarily basalt) can be permeable in local areas, most other types of consolidated rock are not sufficiently permeable to transmit large volumes of water, and bedrock generally forms a relatively impermeable boundary to the Basin and Range aquifers.

The groundwater flow systems of the Basin and Range area are in individual isolated basins or in two or more hydraulically connected basins. The impermeable rocks are boundaries to the flow system, and most of the groundwater flows through basin-fill deposits. If carbonate rocks underlie the basins, substantial quantities of water can flow between basins through the carbonate rocks and into the basin-fill deposits. Most recharge to the basin-fill deposits originates in the mountains as snowmelt. Major faults that cut the alluvial deposits can act as partial barriers to the movement of groundwater.

Dependence of Special Habitats on Different Water Sources

Groundwater-dependent ecosystems occur in locations with abundant growing-season water. Because precipitation is the ultimate source of water and directly influences streamflow characteristics and groundwater dynamics, it is expected that climate-induced changes in precipitation will affect riparian areas, wetlands, and GDEs. Availability of water is also influenced by physical watershed characteristics that affect infiltration and surface and hillslope runoff, including lithology, soil depth, and topography (Jencso et al. 2009).

Groundwater Recharge in Mountain Aquifers in the Western United States

Most aquifers in Western mountains are small compared to the major aquifer systems in the basins. Despite being small, these aquifers are essential in storing and transmitting groundwater that becomes recharge to the adjacent major aquifer systems. Altered recharge caused by climate change will translate into altered mountain aquifer storage and discharge, which will, in turn, directly influence recharge to downgradient aquifers and stream base flows. Between 61 and 93 percent of diffuse mountain catchment recharge becomes streamflow available for downstream aquifer recharge by stream loss (Meixner et al. 2016).

Snowmelt is likely to contribute the majority of recharge in most mountainous regions of the western United States, either because most of the precipitation falls as snow, or snowmelt infiltrates below the root zone more effectively than rainwater (Earman et al. 2006). Snowmelt can compose a large fraction of recharge because much of the water released from the snowpack occurs over a prolonged period in early spring when ET is low (Ajami et al. 2012; Earman et al. 2006). Consequently, mountain recharge is sensitive

to the climatic shifts that result in changes in SWE noted earlier in the chapter.

Recharge in many mountainous areas is permeability limited rather than recharge limited where thin soils overlie low-permeability crystalline bedrock (Flint et al. 2008). Lower maximum annual SWE in these areas may decrease overland flow of snowmelt to streams, but has little influence on recharge because spring snowmelt substantially exceeds the unsaturated zone storage capacity (Blankinship et al. 2014). Conversely, bedrock permeability in karst areas is so high that most snowmelt and rainfall infiltrates into the porous rock, flows in conduits, and is discharged to the surface as springs or discharges directly to fill aquifers.

Higher minimum temperatures will reduce the longevity of snowpack, and decrease the length of time aquifer recharge can occur, potentially leading to less groundwater recharge. Some watersheds will be shifting from snow-dominated to rain-dominated, which may result in declines in groundwater recharge (Earman and Dettinger 2011; Safeeq et al. 2013, 2014). Recharge could also increase in these areas as a result of a more gradual release of water from snowpack from enhanced winter melting (Byrne et al. 2014; Musselman et al. 2017). Projecting future mountain recharge requires knowledge of groundwater flow systems that is generally unavailable.

In summary, recharge is likely to decrease in the southern IAP region, but changes in other parts of the region are uncertain because of limited understanding of mountain recharge processes and groundwater flow in mountains (Meixner et al. 2016). However, there are existing approaches (e.g., Safeeq et al. 2014) that can be used to develop sensitivity maps from available information about geology, stream recession behavior, and other factors. These approaches could be used to evaluate sensitivities for future mountain recharge in the IAP.

Current Resource Conditions

Steep elevation gradients, varied bedrock, and glacial landforms influence the distribution, characteristics, and water chemistry of groundwater-dependent features. Existing information on the condition and distribution of GDEs in National Forests of the IAP region is limited. Here, we rely on data compiled by the Spring Stewardship Institute, the National Hydrography Dataset (USGS 2017), and the National Wetlands Inventory (NWI) (USFWS n.d.) to assess the current condition of springs, wetlands, and GDEs in the region.

Springs are usually small, averaging less than 0.5 acre, with few spring habitat patches larger than 2 acres (Kreamer et al. 2015). Thus, springs fall below the scale of most landscape mapping efforts and are therefore neglected in remote sensing, soil, and floristic mapping. Recently, National Forests in the U.S. Forest Service Intermountain Region and the Spring Stewardship Institute have begun to map springs and other GDEs, but the known occurrence of these are limited, and many more springs certainly exist (table 4.1).

Table 4.1—Area of palustrine emergent wetlands with saturated water regime and number of mapped springs in the U.S. Forest Service Intermountain Region. Differences in wetland area reflect different mapping accuracy and limitation among national forests; wetlands less than 1 to 3 acres are generally not included (USFWS n.d.).

National Forest	Wetland area	Springs
	<i>Acres</i>	<i>Number</i>
Ashley	18,388	426
Boise	92	442
Bridger-Teton	3,397	140
Caribou-Targhee	262	1,467
Dixie	1,048	652
Fishlake	1,259	622
Humboldt-Toiyabe	1,275	4,286
Manti-La Sal	1,894	481
Payette	218	155
Salmon-Challis	205	692
Sawtooth	224	1,211
Uinta-Wasatch-Cache	19,269	917
Total	47,530	11,491

Springs play a key role as groundwater discharge zones that deliver cool water to warming streams, support late-season streamflows in summer, and deliver relatively warm water during winter months (Lawrence et al. 2014; Winter 2007). Most streams and rivers in the region are at least partially groundwater dependent (Santhi et al. 2008). Locations of groundwater discharge to streams have been identified by using remote sensing (Torgersen et al. 1999) and field techniques (Torgersen et al. 2012), but have not been systematically mapped.

Wetlands can be identified by targeting palustrine/emergent wetlands with a saturated water regime (Cowardin et al. 1979) in the NWI database (table 4.1). To ascertain whether these wetlands are indeed fens, each wetland would require a field visit to determine if it is supported (at least in part) by groundwater and is peat-forming (Chadde et al. 1998; USDA FS 2012a, b). Fens occupy a small portion of the landscape, but contribute substantially to biodiversity of plants and animals (Blevins and Aldous 2011). In an otherwise arid region, perennially saturated fens are critical habitat for invertebrate and amphibian species. Although not explicitly differentiated as fen vegetation, several herbaceous-dominated plant associations frequently occur in fens.

Since 2008, GDEs, mostly springs and fens, have been inventoried and documented in National Forests by using draft and final versions of the Groundwater-Dependent Ecosystems Level I and Level II inventory methods

(USDA FS 2012a, b). The Level I guide (USDA FS 2012a) describes basic methods for assessment of GDEs within a given area (e.g., National Forest, ranger district, specific project area) and is intended to qualitatively document the location, size, and basic characteristics of each GDE site. The Level II guide presents more detailed inventory (USDA FS 2012b) in addition to protocols for more comprehensive characterization of the vegetation, hydrology, geology, and soils at a given site.

Inventories have been conducted in Ashley, Caribou-Targhee, Dixie, Humboldt-Toiyabe, Manti-La Sal, and Sawtooth National Forests. Inventories in Humboldt-Toiyabe National Forest targeted springs with terrestrial and aquatic flora and fauna values. In the GDE Level I protocol (USDA FS 2012a), a series of 25 management indicator statements assist in identifying potential concerns and needs for management action based on observations recorded during field inventories. Three of the most important indicators are aquifer functionality, soil integrity, and vegetation composition. Most inventories targeted sites where proposals for water development could be damaging, or portions of grazing allotments and watersheds with specific management concerns. Findings of inventories commonly show notable resource impacts through water diversion, soil disturbance, and effects of livestock on vegetation.

Potential Climate Change Effects on Groundwater and Groundwater-Dependent Ecosystems

Groundwater recharge has been examined in only a few locations (Tague and Grant 2009), and little is known about groundwater recharge processes in many watersheds, including those that may be shifting from snow-dominated to more rain-dominated hydrological regimes (Safeeq et al. 2013, 2014). Depending on elevation and hydrogeologic setting, slowly infiltrating precipitation that includes both rain and snow may recharge some groundwater aquifers as effectively as rapid, seasonal snowmelt runoff. Although rain-on-snow zones are expected to shift upwards in elevation, the influence of these shifts on groundwater recharge is unknown.

Small, unconfined aquifers (especially surficial and shallow aquifers) are more likely to have renewable groundwater on shorter time scales and may respond rapidly to changes in climate (Healy and Cook 2002; Lee et al. 2006; Sophocleous 2002; Winter 1999). Larger, deeper, and confined aquifers are more likely to have nonrenewable groundwater, may be less sensitive to the direct effects of climate change, and are projected to have a slower response (Wada et al. 2012).

Groundwater storage can moderate surface water response to precipitation changes (Maxwell and Kollet 2008), and changes to groundwater levels can alter the interaction between groundwater and surface water systems (Hanson et al. 2012). Climate-induced changes in connectivity between groundwater and surface water could directly affect

stream base flows and associated wetlands and other GDEs (Earman and Dettinger 2011; Kløve et al. 2012; Tujchneider et al. 2012). Short flow-path groundwater systems, including many that provide base flow to headwater streams, could change substantially in timing of discharge in response to changes in seasonality of recharge (Waibel et al. 2013). In contrast, regional-scale aquifer systems, with flow paths on the order of tens of miles, are much less affected by shifts in seasonality of recharge. These effects may be highly variable, depending on local hydrogeology.

Altered groundwater levels in wetlands can reduce groundwater inflow, leading to lower water table levels and altered wetland water balances. For local and intermediate-scale systems, the spatial extent of some GDEs is likely to contract in response to decreasing surface water and groundwater and increasing temperatures. Changes in groundwater and surface water will also vary depending on location within the watershed, as well as future land use.

Effects of changing climate on the ecology of GDEs will depend on changes in groundwater levels and recharge rates, as influenced by the size and position of groundwater aquifers (Aldous et al. 2015). GDEs supported by small, local groundwater systems tend to exhibit more variation in temperature and nutrient concentrations than regional systems (Bertrand et al. 2012). It is likely that larger systems will be more resilient to climate change.

Freshwater springs depend on continuous discharge of groundwater, forming ecotones between subsurface-surface water and aquatic-terrestrial environments (Ward and Tockner 2001). Springs and springbrooks support locally unique biological communities (Barquin and Death 2006). However, climate-induced changes in recharge may cause decreased summer flows with possible drying, as well as increased winter flow and inundation of biological communities (Green et al. 2011).

Many biogeochemical processes are temperature dependent, so climate-induced changes in groundwater temperature may negatively affect the quality of groundwater, and, in turn, influence aquatic communities (Figura et al. 2011). However, because the thermal regime of groundwater systems is less dependent on air temperature patterns than surface waters, the effects of rising air temperatures are likely to be less pronounced in springs and other GDEs.

Peat-accumulating processes in fens will be influenced by increasing temperatures and local and regional changes in hydrological regime. Reduced groundwater levels tend to promote soil aeration and organic matter oxidation. Generation and maintenance of peat soils over time depend on stable hydrological conditions. In recent studies of peatlands exposed to groundwater lowering, responses such as soil cracking, peat subsidence, and secondary changes in water flow and storage patterns have been observed (Kværner and Snilsberg 2011). Wetland plant species can respond to even slight changes in water table elevation (Magee and Kentula 2005; Shipley et al. 1991; Vitt and Slack 1984), and shifts in composition of vascular and bryophyte species could occur with lowered water tables.

Some riparian ecosystems depend on the presence of flowing water, although streamflow may not be perennial along all stream segments and can vary considerably with season, physical features of the watershed, and water source. Depending on physical characteristics of a given stream segment, the volume of streamflow can also drive seasonal changes in water table elevation of the adjacent riparian area (Jencso et al. 2011). These hydrological and fluvial processes and resulting geomorphic surfaces are essential for the persistence of riparian vegetation (Naiman et al. 2005). According to long-term daily flow data, different streams in the region are supported by perennial runoff, snow plus rain, and stable groundwater levels (Poff 1996).

Changes in water table elevations and streamflow volumes may affect riparian areas and their plant communities (Jencso et al. 2009; Naiman et al. 2005). Examples of changes in flow systems are decreased summer base flows (see earlier *Hydrological Processes* section), lower riparian water table elevations, and reduced hydrological connectivity between uplands and riparian areas (Jencso et al. 2009, 2011). Streamflow volume along gaining reaches increases with the inflow of groundwater to the channel. Stream water can also drain from the channel bed and banks to the groundwater system (losing reaches), resulting in a loss of downstream surface flow volume (Winter et al. 1996). Gaining and losing stream reaches result in different aquatic communities in the channels and different riparian plant communities on the floodplains. The extent to which gaining or losing characteristics of specific reaches may change in response to climate-induced changes in precipitation, streamflow characteristics, and groundwater discharge is unknown.

In wetlands and riparian ecosystems, hydrological variables are consistently the strongest predictors of plant species distributions (e.g., Cooper and Merritt 2012). Current understanding of the water sources used by riparian and wetland plants is limited to a few highly valued or highly invasive species (mostly woody). However, riparian and wetland plant species use water from multiple sources (surface water, soil water, groundwater), depending on life stage and season (Busch and Smith 1995; Cooper et al. 1999; Goslee et al. 1997). In assessing the vulnerability of riparian and wetland species to climate-induced changes in streamflow or groundwater, the availability of water at all life stages must be considered, from plant recruitment and establishment, to reproducing adults, to persistence at later life stages (Cooper and Merritt 2012).

Climate-induced changes in precipitation, drought, and streamflow will influence the distribution of riparian vegetation via changes in local hydrological regimes, especially if summer base flows decrease. If water table elevation can be assumed to be in equilibrium with water levels in the stream, reduced base flows could result in lower riparian water table elevations and subsequent drying of streamside areas, particularly in wide valley bottoms. Wetland and riparian plant communities will respond to climate-induced changes in hydrological variables differently as a function

of species composition (Merritt et al. 2010; Weltzin et al. 2000).

Although ET is not expected to increase substantially from the landscape generally as outlined earlier in the chapter, water supplies around riparian areas and GDEs are consistently high. Riparian areas and GDEs compose a very small fraction of the landscape, so they affect the land-surface energy balance only very slightly. Consequently, the increased net radiation and atmospheric demand will induce higher ET rates in riparian areas and GDEs, and this higher ET rate will not substantially feed back into the regional energy balance. Higher ET will result in drying in these ecosystems, potentially stressing characteristic plant species and resulting in compositional shifts in vegetation. If plant cover is reduced in riparian areas, erosion may increase.

Soil Resources

The potential effects of climate change on soils are multifaceted; changes in soil physical, biological, and chemical processes can occur with changing climate, which may in turn affect other processes such as carbon cycling and vegetation growth. Soil responses to climate change will vary by geographic location and are determined by the interactions of soil, vegetation, and the degree of management intervention. The following sections provide a summary of potential effects of climate change on soils in the IAP region.

Soil Temperature and Moisture

Soil temperature and moisture are the primary drivers of change for all soil processes. Potential changes to these soil properties with climate change have been well studied, but where and when the changes may occur is difficult to predict. The magnitude of projected change is variable depending on existing soil resources and existing climate. The properties and processes of soils are not independent, and a change to one soil property will affect other soil properties and processes. For example, changes to soil temperature and moisture will affect carbon and nitrogen cycles, which can in turn affect soil properties such as water holding capacity, cation exchange capacity, soil nutrient content, and aggregate stability (Brevik 2013).

An increase in soil temperature will generally produce an increase in soil biological activity and soil respiration. However, the rate and magnitude of change are dependent on soil moisture (Kardol et al. 2010). In the current semiarid soils of the IAP region, an increase in soil temperature without an increase in soil moisture is likely to create soils that have reduced biological activity and less potential to store carbon. If soil moisture is limiting with increased soil temperatures, the soils may become a net source of carbon until equilibrium is reached. However, an increase in soil temperature could be offset by an increase in water available for biological activity and vegetation production, resulting in little change or a possible increase in carbon storage. In

the colder and wetter areas of the IAP region, an increase in soil temperature may lead to longer growing seasons if soil moisture is not limiting (Kurylyk et al. 2014). Thus, the timing and type of moisture will determine soil biological activity, respiration, and ultimately vegetation supported by the soil.

Changes to soils with climate change will not be uniform across the IAP region. Soil responses to temperature and moisture are highly dependent on the soil parent material. Soils derived from coarse-textured granitic soils will transfer heat more efficiently downward into the soil profile than fine-textured limestone soils. The heat transfer downward can affect soil processes and even groundwater temperatures, and it could ultimately affect surface water temperatures where groundwater is the source for surface water. Fine-textured soils, which are capable of storing water longer in the soil profile, will generally have a higher buffering capability to changes in soil temperature and moisture.

The timing of soil moisture can also affect soil erosional processes. A shift away from winter precipitation as snow to greater amounts of rain and more intense rain storms can generate a higher frequency of runoff and erosional process from disturbance events such as fire (Litschert et al. 2014). Runoff from extreme rain events could increase for shallow soils with little capacity to store water.

Soil Carbon and Nitrogen

Soils are a major component of carbon and nitrogen cycles. Changes in soil temperature and moisture will affect carbon and nitrogen cycles. Management of soil organic matter can affect both of these cycles at local and global scales. The greenhouse gases carbon dioxide, methane, and nitrous oxide are regulated to some extent by the soil organic matter. Soils provide both a source and sink for carbon and nitrogen and the greenhouse gases associated with carbon and nitrogen. Changes to the carbon and nitrogen cycles may include an increase or reduction in cycling rates or storage of carbon and nitrogen. Soil temperature and moisture drive the type of change that will occur as they affect microbial activity and plant composition.

Soil Carbon Pool

Soil organic carbon (SOC) is derived from soil organic matter (SOM). The SOM is composed of plant and animal residues, cells and tissues of soil organisms, and substances produced by decomposing organisms. The SOC is the carbon component of SOM. Generally, about 58 percent of SOM is SOC by weight. Most soil orders within the IAP region have a near-surface SOC content (by mass) of 0.5 percent (for the hotter and drier areas) to 8 percent (for the cooler and wetter areas; Histosols excluded) (Brady 1999). Hereafter, we use SOC to represent both SOC and SOM, as these properties are likely to have similar response to climate change.

Soil organic carbon may be the best indicator and contributor to soil health because SOC supports many soil processes and functions. These include providing nutrients for plants, binding soil particles together and thereby maintaining structure, providing an energy source for microbes, increasing water infiltration and retention, and providing cation-anion exchange for retention of ions and nutrients. Climate change will affect SOC and ultimately the functions and processes supported by SOC.

Globally, SOC may contain more than three times as much carbon as is found in the atmosphere or terrestrial vegetation. In forest ecosystems, SOC may be as much as 80 percent of the total terrestrial carbon pool, and in nonforest ecosystems, SOC may be as much as 95 percent of the total terrestrial carbon pool (Meyer 2012). Soils can store and release carbon at the same time. If soils store more carbon than they release back to the atmosphere, the soil is a carbon sink. If soils are releasing more carbon than what is being stored, they become a carbon source. Therefore, the management of SOC is critical to the management of atmospheric carbon concentrations (Woodall et al. 2015).

Carbon is stored in soils in organic or inorganic forms. Soil organic carbon originates from carbon fixation during photosynthesis and microbial decomposition (Thomey et al. 2014). Inorganic carbon (IC) is in rocks and minerals. An example of IC is limestone, or calcium carbonate. The stable IC is released slowly through weathering or anthropogenic manipulation, such as mining and conversion to other chemical compounds. Although IC is slow to change, it represents a large portion of stored carbon in many ecosystems, such as drier shrublands and grasslands. Many of the drier rangeland soils include carbonate horizons within the soil profile. Climatic changes may affect the release of IC. Higher atmospheric carbon dioxide concentrations and a warmer and wetter climate will increase weathering of rocks and acidification in the carbonate layers in the soil, which will release greater amounts of carbon into the active carbon cycle (USDOE 2014).

Different soils have different capacity to store carbon. The differences are related to parent material, existing climate, and terrestrial ecological community types. Shrublands have a higher percentage of SOC stored lower in the soil profile (below 3 feet). Forests have most of their SOC in the first 3 feet of the soil. Changes to vegetative composition can affect long-term carbon storage. A shift from shrublands to annual grasslands will eventually move the bulk of carbon from deep in the soil profile to the upper 8 inches (USDOE 2014). This may be happening with conversions of shrublands to cheatgrass (*Bromus tectorum*) in the IAP region. This process is very slow, however, and respiration of carbon deep in the soil profile is much slower than near the soil surface. A shift from shrublands to forest will increase near-surface carbon pools as a result of litter addition and deeper reserves being tapped by roots for the production of biomass (Nave et al. 2013).

Soils have SOC storage limits set by soil physical and chemical composition as well as microbial and plant

community types, all of which are determined by soil moisture and temperature. Most soils in the IAP region are at SOC saturation for the existing climate (Woodall et al. 2015). Soils that are degraded or furthest from potential SOC saturation have the greatest ability to sequester additional carbon. These are generally areas with vegetation that has been altered for long periods of time, such as agricultural fields. Most of the soils in the IAP region could sequester additional carbon if soil temperatures decrease and soil moisture increases. This is particularly true with the lower-elevation soils. However, soil moisture may not increase in a warming climate.

Soil Physical Properties Related to Carbon and Climate Change

Changes to SOC can alter several soil properties, including soil structure, bulk density, and soil porosity (Pal Singh et al. 2011). These soil properties affect water infiltration, rooting depth, soil erosional losses, and water holding capacity. A reduction in SOC will change soil structure through reducing the bonds between soil particles and the microbial “glue” that helps hold soil particles together. This can lead to less resistance of the soil to erosive forces of wind and water. A change in soil structure can also lead to changes in soil porosity and bulk density. Soil porosity and pore size distribution are important for soil water management and maintaining release rates of greenhouse gases (carbon dioxide, methane, and nitrous oxide) within the soil. A loss of macropores with reduction in SOC could lead to slower water infiltration rates, increased runoff, decreased nutrient cycling, reduced plant growth (above and below ground), and poor aeration of the soil, resulting in a decreased capacity to oxidize greenhouse gases, specifically methane. A reduction in SOC also leads to increased risk of surface compaction with management activities through an increase in surface bulk density of soils. Surface compaction restricts water infiltration and increases surface runoff.

Although the changes to soil physical properties with loss of SOC are highly variable across the landscape, they do provide potential indicators that can be used to prioritize management in a changing climate; the soil types where current management is already having negative effects on soil physical properties could be the soil types that are prioritized for climate change adaptation actions. For example, areas where excessive runoff and soil loss have occurred because of grazing management may be a priority. These areas would be expected to have a higher risk of soil quality loss under a warmer and drier climate with reduction in plant growth and SOC development.

Soil Nitrogen

SOM typically contains about 5 percent nitrogen; therefore, the distribution of soil nitrogen closely parallels that of SOM (Brady 1999). Nitrogen cycles are closely tied to carbon cycles, although they may respond differently to changes in climate. On average, nitrogen fixation occurs at a rate of about 9 pounds per acre for forested sites and 13

pounds per acre for grasslands (Brady 1999). Forest soils may contain 15 times as much nitrogen as the standing vegetation, including roots (Brady 1999). About 29 to 56 percent of the soil nitrogen pool is found in the upper 4 inches (Page-Dumroese and Jurgensen 2006), making the soil nitrogen pool highly susceptible to loss from erosion.

Although most of the nitrogen in terrestrial systems is found within the soil, the mineralization of nitrogen is required to provide a form of nitrogen that plants can utilize. Nitrogen mineralization occurs through decomposition of organic material by soil micro-organisms. Warmer soil temperatures increase decomposition by microbial activity, thus increasing nitrogen mineralization. However, soil moisture may have a greater effect on net nitrogen mineralization through changes in the form of nitrogen (Emmett et al. 2004).

Nitrogen could be limiting to plants in some soils of the IAP region even if conditions for plant growth improve with changes in climate. If plant growth increases with increased atmospheric carbon dioxide, then organisms that decompose residual plant material will need more nitrogen to survive. If nitrogen is tied up by soil microflora and microfauna, the nitrogen would be unavailable to plants. Thus, any positive effects of increased atmospheric carbon dioxide may be offset by the reduction in available soil nitrogen (Brevik 2013), particularly on nitrogen-limited soils. In the IAP region, those areas most susceptible to reduction in nitrogen are forested areas on soils with coarse-textured parent material.

Soil Biological Activity

Soil organisms perform many functions in the soil, including decomposition and nutrient cycling. As with other soil processes, the soil biology is affected by other soil processes and the inherent soil composition and climate. Thus, the effects of climate change on soil biology will be variable. Some models project that a warming of the soil will create greater microbial activity, resulting in more carbon being released to the atmosphere because of increased decomposition (Kardol et al. 2010). Other models suggest that a warming of the soil will result in lower microbial growth and less carbon released through respiration (Wieder et al. 2013). In the warm and dry desert ecosystems, such as those in the Great Basin and Semi Desert and Intermountain Semi Desert subregions, the effects of soil warming are expected to increase microbial activity and carbon dioxide released to the atmosphere (Thomey et al. 2014). In cooler and wetter ecosystems, projections are mixed (Steinweg et al. 2013).

Vegetation composition can affect soil biology, soil processes, and soil response to climate change. Some soil organisms prefer specific plant types, and plant diversity increases soil biological diversity. Greater biodiversity in soils is likely to increase soil resilience to climate change (Kardol et al. 2010).

Soil Chemical Properties

Potential effects of climate change on soil chemical properties are linked to other biological and physical changes in the soil, all of which are driven by soil temperature and moisture. Soil pH is closely tied to organic matter, parent material, and soil moisture. A warming climate with additional or similar precipitation will lower soil pH (Pal Singh et al. 2011). A warming climate with less precipitation may increase pH on some higher-elevation soils and have little effect on existing lower-elevation high-pH soils (Pal Singh et al. 2011).

In areas that are expected to experience increased drought, such as drier shrubland and grassland systems, an increase in the accumulation of carbonates and salts in the soil profile is expected. This would result in a salinization of the soil and have significant effects on vegetation composition. In wetter areas of the IAP region, an increase in soil temperatures may cause an increase in acidification from the decomposition of organic matter. This could change species composition and diversity to favor species more adapted to acidic soils.

Cation exchange capacity (CEC) is the ability of the soil to retain nutrients such as calcium, potassium, and magnesium and make them available to the soil solution and plants. It also provides the capacity to retain and immobilize some cations that may be toxic to soil microbes and plants in high amounts, such as aluminum and manganese. The CEC is generally low in coarse-textured soils or soils with low amounts of SOC. Soils with a subsurface argillic horizon (higher CEC) are likely to be able to moderate a shift in nutrient exchange, specifically a loss of SOC and other major soil nutrients requiring cation exchange sites. An increase in soil temperatures could lead to a reduction of SOC and the CEC of soils. A reduction in CEC would result in loss of base cations in the soil that are released to groundwater and surface water (Pal Singh et al. 2011).

Managing soil resources for optimum SOC will limit the effects of climate change to the CEC. Areas with low SOC are the most susceptible to CEC changes in soils due to climate change, as they have poor buffering capabilities. In the IAP region, these are primarily drier rangeland soils, particularly those that have been vegetated for many decades with annual shallow-rooted grasses.

Weathering of rock and erosion of soil is a continuous process. Changes to rainfall and wind as well as changes to chemical, physical, and biological properties of soils can affect the type, amount, and rates of runoff and erosion of soil. More precipitation may not have any long-term effects on erosion and runoff, because vegetation will respond positively to increases in available soil moisture. However, an increase in the number of intense rain events may result in an increasing rate of erosion. A reduction in the amount of precipitation generally reduces the rate of erosion. However, lower vegetation cover in response to low soil moisture may result in increased rates of erosion from wind and water. Areas of the IAP region that are most susceptible

to increased erosion are the lower-elevation shrublands and grasslands, where a warmer and drier climate will reduce the potential for vegetative cover.

Regional-Scale Soil Vulnerability Ratings

The potential magnitude of change to soil resources in a changing climate is extremely variable because of the heterogeneity of soil types and their potential response across the landscape. Identifying the degree of potential change and risk to soils spatially across a landscape is difficult, even with detailed soils data. However, general projections can be made across large landscapes by using different soil and landscape attributes, particularly vegetation composition and productivity. The following section provides a general rating of soil vulnerability to climate change across the IAP region. The rating is based on general soil characteristics and data from Natural Resources Conservation Service STATSGO datasets (NRCS 2017).

Several assumptions were used to develop a regionwide rating of soil vulnerability to climate change. These include:

- The existing climate will generally be warmer and drier in the future across all subregions in the IAP.
- Soils that are currently capable of holding water longer and deeper within the soil profile have a greater buffering capacity to change.
- Soils that are currently cooler and wetter, or have higher SOC, are less susceptible to climate change.
- Many soil properties will change in a changing climate.
- The ratings of soil vulnerability to climate change are based on general surrogates of landscape and soil conditions. Data on detailed soil properties, such as available water holding capacity, were not available across the region to make predictions.
- Soil properties used to derive a vulnerability map were soil temperature and moisture regimes (combined into subclasses), and SOC from depths of 0 to 12 inches.
- Soil polygons contain many soil components. The components were combined by dominant condition for soil temperature and moisture subclasses and by weighted averages for SOC.

A combined STATSGO soil map was made for all lands within the IAP region. Soil temperature and moisture classes were determined for each polygon based on the dominant condition within the polygon. The soil temperature and moisture classes were further combined according to soil taxonomy to create 44 different subclasses. These subclasses were combined qualitatively based on common soil temperature and moisture breaks to create four class ratings of low, moderate, high, and very high susceptibility to a warming and drying climate. The same STATSGO soil map was used to create a SOC map for the 0 to 4 inches soil depth (the database was poorly populated for depths deeper than 4 inches). Each polygon received a value for SOC. The

Table 4.2—Final rating classes for soil vulnerability to climate change, based on a combination of soil temperature/moisture rating and soil organic carbon rating.

	Organic carbon rating = Low	Organic carbon rating = Moderate	Organic carbon rating = High	Organic carbon rating = Very high
Temperature/moisture rating = Low	Low	Low	Moderate	High
Temperature/moisture rating = Moderate	Low	Moderate	Moderate	High
Temperature/moisture rating = High	Moderate	Moderate	High	Very High
Temperature/moisture rating = Very high	High	High	Very High	Very High

polygon values were assigned to one of four classes (low, moderate, high, very high) of SOC such that all classes contained the same number of polygons in each class. This method of creating general SOC classes was chosen because threshold class values for SOC are not available.

The ratings for soil temperature and moisture were then combined with the ratings for SOC into a four-class rating of soil vulnerability to climate change. A simple matrix was used to determine the final rating (table 4.2). The final soil vulnerability rating was applied to each polygon, and a general soil vulnerability map to climate change was developed (fig. 4.18). The map represents soils that may or may not be capable of sustaining existing ecosystems in a changing climate. The map suggests that soils at higher elevations and deeper soils are not as susceptible to climate change as the soils in warmer and drier areas. But this does not mean soils will not change in wetter or cooler climates or in locations high in SOC.

The regional-scale soil vulnerability map is a coarse estimation of potential change to soils with climate change. Local data and information are needed to estimate vulnerability at a local scale. Other soil properties that should be considered in creating a local map include: available water holding capacity, soil depth, hydrological soil group, erodibility (K) factor, soil texture, parent material, SOC, and calcium carbonate content (inorganic carbon), as well as vegetation type, geology, slope, aspect, and elevation. An example of how to create a soil vulnerability rating at a forest-project scale is given next.

Example Project-Scale Assessment of Soil Vulnerability to Climate Change

At the individual forest level, local soils information could be utilized to create maps of soil vulnerability to climate change at the project, watershed, or landscape scale. The Uinta-Wasatch-Cache National Forest used available soil data and applied a soil vulnerability assessment to a watershed-scale project during the project planning stage. The assessment was used initially to identify potential projects that would help create more resilient ecosystems and then to stratify fieldwork. Vegetation manipulation is

one example of a specific adaptive management strategy to maintain or enhance soil health. Quaking aspen (*Populus tremuloides*) and pinyon-juniper vegetative communities were examined for this example.

A soil map was created for two watersheds within a project boundary. Vegetation and geology layers were added, as well as a digital elevation model to create slope and aspect. Soil available water holding capacity, soil depth to a restrictive layer, hydrological group, parent material, and soil temperature and moisture regimes were included in a matrix and rated for each soil type to derive vulnerability to climate change. Using soil indicators, along with vegetation and geology layers, we can estimate nutrient content, SOM, and how well a soil retains moisture. These estimates were used to assign a rating for soil vulnerability to climate change (table 4.3). The higher the point value rating, the more resilient and resistant the soil resources are to the effects of a warmer and drier climate. This information was added to the soil attribute table in a geographic information system (GIS). Potential focus areas for fieldwork verification and for recommended vegetation projects to meet desired conditions were identified by creating intersects for vegetation attributes of interest. An example of an interpretive map is shown in figure 4.19.

The value ranges or ratings will be used to focus attention on soil resources that best meet vegetation management objectives. Soils that are more suitable or resistant to change (less vulnerable to climate change) are those expected to better maintain soil moisture and nutrient conditions favorable for the vegetative communities present. These soils will be areas of opportunity for vegetation management designed to sustain existing vegetation community types.

Conclusions from the soil vulnerability analysis include:

1. Sustaining aspen vegetative communities within the project area will be one of the objectives of the project. Vegetative treatments will be implemented to promote aspen retention and increase diversity of aspen age classes. The climate change soil vulnerability rating was determined for each of the aspen polygons in the treatment area. An examination of the rating criteria showed some aspen stands with

Region 4 Forest Service Soil Vulnerability To Climate Change Based On Soil Temperature/Moisture Classes and Soil Organic Carbon Content

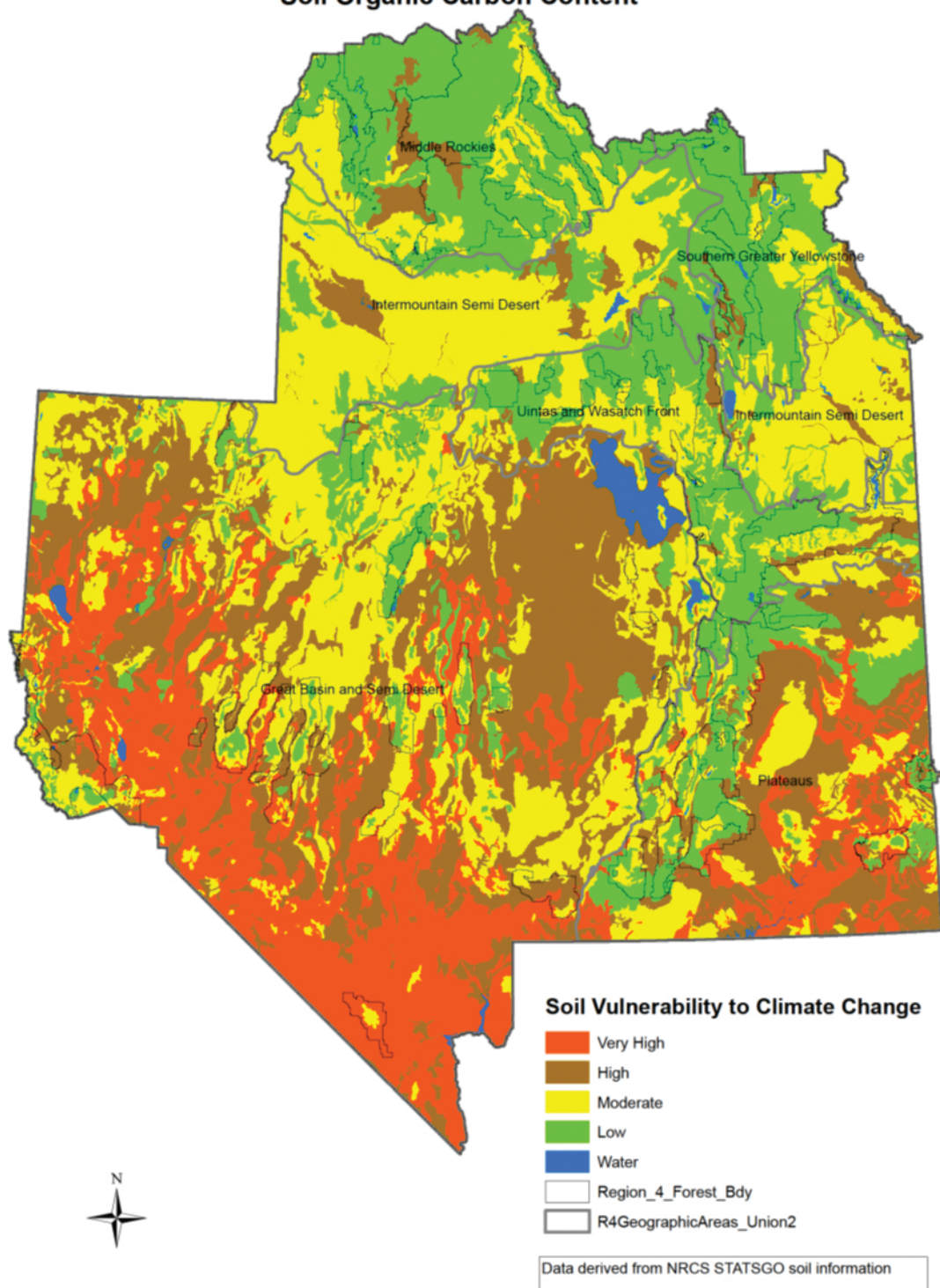


Figure 4.18—Soil vulnerability to climate change in the Intermountain Adaptation Partnership region, based on ratings in table 4.2 (combination of soil temperature/moisture and soil organic carbon).

Table 4.3—Forest-level soil vulnerability indicators and ratings.

Water holding capacity		Soil depth		Hydrological group		Aspect		Parent material		Soil temperature/ moisture regime		Point value range		Suitability rating		Vulnerability rating	
<i>Value^a</i>	<i>Pts^b</i>	<i>Inches</i>	<i>Pts</i>	<i>Group</i>	<i>Pts</i>	<i>Group</i>	<i>Pts</i>	<i>Texture</i>	<i>Pts</i>	<i>Value^a</i>	<i>Pts</i>						
VH	12	60+	12	A	9	N	9	Fine	9	L	9	46–60	VH			VL	
H	9	40–60	9	B	6	E	6	Medium	6	M	6	33–45	H			L	
M	6	20–40	6	C	3	W	3	Coarse	3	H	3	20–32	M			M	
L	3	11–20	3	D	0	S	0			VH	0	7–19	L			H	
VL	0	0–10	0									0–6	VL			VH	

^a VH = very high, H = high, M = moderate, L = low, VL = very low.^b Pts = points.

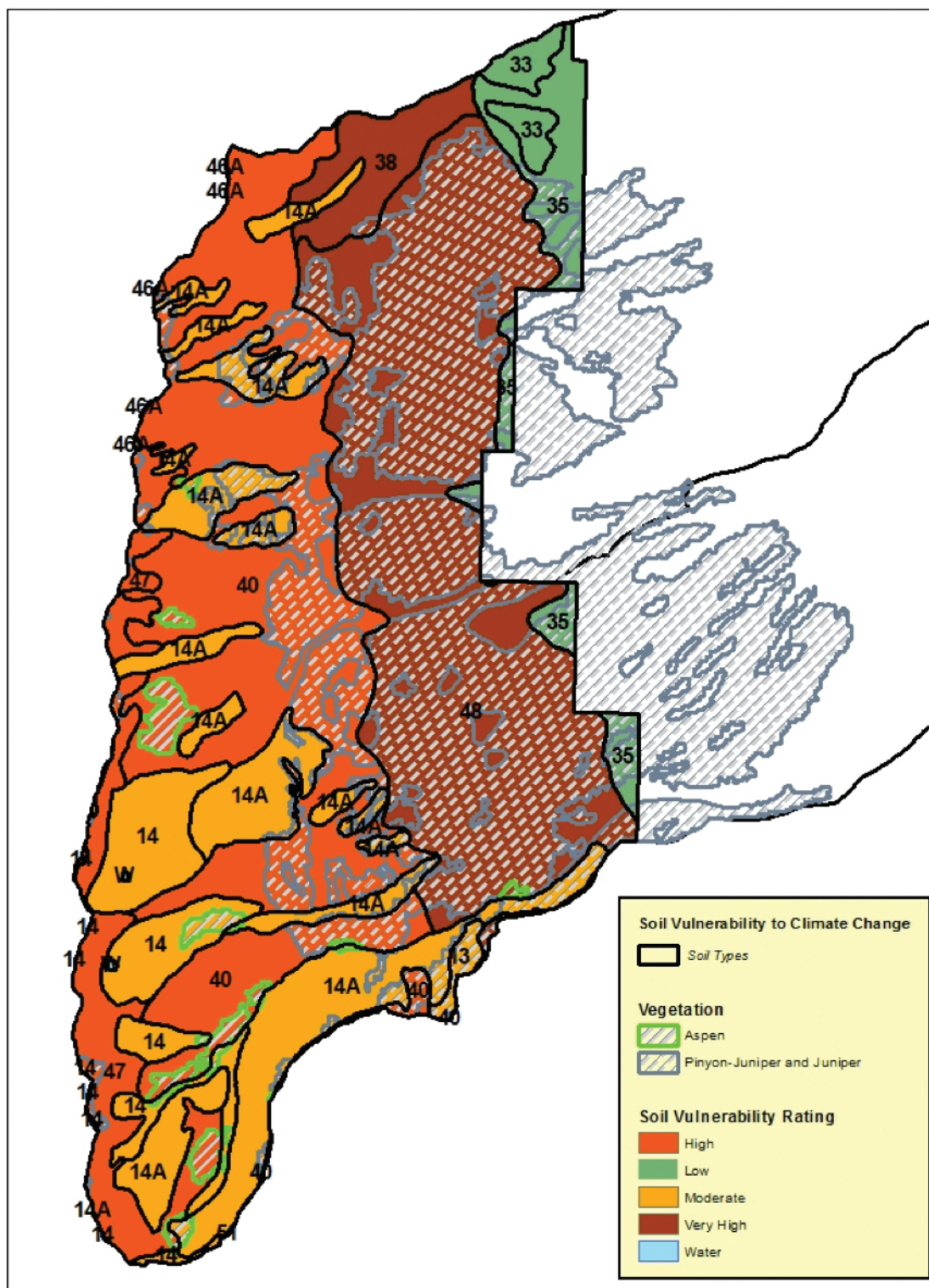


Figure 4.19—Example of soil vulnerability to climate change at a small spatial scale, based on ratings in table 4.2 (combination of soil temperature/moisture and soil organic carbon). Numbers indicate soil series mapping units (not discussed here).

a very high to high vulnerability rating and some with a moderate vulnerability rating. Aspen stands with a moderate vulnerability rating will be areas for recommended project work to maintain aspen community types. The aspen stands located within the high and very high vulnerability areas will not be recommended for aspen regeneration treatments unless a site visit indicates available water could be sustained in a warmer, drier climate.

2. Pinyon-juniper encroachment into shrublands and grasslands is occurring because of a lack of fire. Removal of the pinyon and juniper is recommended in areas that currently have mountain big sagebrush in the understory, with the objective of increasing sagebrush cover. Removal would be accomplished through mechanical treatments (mastication) and lop-and-scatter. Very high to low soil vulnerability ratings were found across the project area where pinyon-juniper exists. Soils that were rated as very high to high in vulnerability are areas with shallow soils and low available water holding capacity. Pinyon-juniper may decrease in dominance in these areas with a reduction in available soil water due to climate change. Removal of pinyon and juniper may make these areas more susceptible to invasion by nonnative species such as cheatgrass. In areas rated as having low to moderate vulnerability to climate change, the ability to manage for restoration of shrublands may be higher as soils are expected to retain soil moisture longer. Site visits will be used to verify potential for management.

Summary and Conclusions

Climate change will alter fundamental physical processes in the IAP region, including hydrological processes and soil processes. Changes in physical processes will in turn affect biological processes, including soil microbial processes and vegetation growth and development. These physical and biological effects of climate change are complex and will be highly variable across the IAP region.

Warming temperatures will reduce snowpack accumulation and advance snowmelt timing in the region. Despite mixed signals from precipitation and temperature changes in the historical record, future temperature changes are expected to be higher than historical temperature trends, and future precipitation declines are expected to be less pronounced—and increased precipitation is possible. Earlier streamflow center of timing is expected over much of the region, and summer low flows are expected to be lower. Total water yields may decrease due to increased ET, but precipitation amounts are uncertain. Increasing precipitation could outweigh ET effects on total water yields, but decreasing precipitation could substantially exacerbate declines in annual water yield and low flows. The frequency and extent of midwinter flooding are expected to increase. Flood

magnitudes are also expected to increase because rain-on-snow-driven peakflows will become more common.

Places with seasonally intermittent snowpacks are likely to see snow more rarely. Some mid- to low-elevation seasonal snowpacks are likely to become intermittent. Higher-elevation snowpacks may or may not undergo substantial changes in April 1 SWE, snow residence time, or center of melt timing, depending on precipitation outcomes. In warmer locations, temperature-dependent changes are relatively robust even if precipitation increases. In colder locations, a precipitation increase within the range of projected possibilities could cancel or overwhelm the effects of even a relatively large temperature change. Alternatively, a precipitation decrease could exacerbate projected temperature-related declines.

Glacier accumulation zones are at some of the highest elevations of the region, so they may respond positively if precipitation increases. Annual dynamics of mass balance with respect to input and output suggest that the equilibrium line (demarcating places where annual snow does not completely melt each summer) will increase in elevation, regardless of precipitation; where that elevation falls on each glacier will influence glacier response. Most glaciers will be reduced in volume and area and may become small enough to prevent movement. If climate at higher elevations becomes both warmer and drier, glaciers are unlikely to persist.

Groundwater recharge is likely to decrease in the southern portion of the IAP region, but changes in recharge remain uncertain throughout the region given limited understanding of mountain recharge processes and groundwater flow in mountain blocks. Groundwater recharge has been examined in only a few locations, and little is known about groundwater recharge processes in many watersheds. Higher minimum temperatures will reduce the longevity of snowpack, and decrease the length of time aquifer recharge can occur, potentially leading to faster runoff and less groundwater recharge. Some watersheds will be shifting from snow-dominated to rain-dominated, which may result in declines in groundwater recharge. Because many biological processes are temperature dependent, climate-induced changes in groundwater temperature may negatively affect aquatic communities. But because the thermal regime of groundwater systems is less dependent on air temperature patterns than on surface waters, the effects of rising air temperatures are likely to be less pronounced in groundwater discharges. Plant species in GDEs can respond to even slight changes in water table elevation, and shifts in composition of both vascular and bryophyte species could occur with lowered water tables.

Soil temperature and moisture are the primary drivers of change for all soil processes. The magnitude of projected change in soils with climate change is variable and depends on existing soil resources and existing climate. An increase in soil temperature will generally produce an increase in soil biological activity and soil respiration. In the current semi-arid soils of the IAP region, an increase in soil temperature

without an increase in soil moisture is likely to result in reduced biological activity, increased respiration, and decreased potential to store carbon. In the colder and wetter areas of the IAP region, an increase in soil temperature may lead to greater biological activity and to longer growing seasons if soil moisture is not limiting. Soils derived from coarse-textured parent material will transfer heat more efficiently down into the soil profile than will fine-textured soils. The heat transfer downward can affect soil processes and even groundwater temperatures. Fine-textured soils are capable of storing water longer in the soil profile, providing a buffer to warming and higher water demands by plants.

Changes in soil temperature and moisture will affect carbon and nitrogen cycles. Changes to carbon and nitrogen cycles may include an increase or reduction in cycling rates or storage of carbon and nitrogen. Soil organic carbon may contain more than three times as much carbon as is found in the atmosphere or terrestrial vegetation, and it supports many soil processes and functions. These include providing nutrients for plants, binding soil particles together and thereby maintaining structure, providing an energy source for microbes, increasing water infiltration and retention, and providing cation/anion exchange for retention of ions and nutrients. Climate change will affect SOC and ultimately the functions and processes supported by SOC. Most of the soils in the IAP region can sequester additional carbon if soil temperatures decrease and soil moisture increases. However, most climate models suggest warmer soil temperatures and various soil moisture changes. The warming temperatures without additional moisture may reduce SOC and capability of soils to store carbon.

Changes to SOC with climate change can cause changes to several soil properties that are directly tied to the amount of SOC. These include soil structure, bulk density, and soil porosity. These soil properties affect water infiltration, rooting depth, soil erosional losses, and water holding capacity. These properties are potential indicators that can be used to determine the effects of climate change and where management changes may be needed to adapt to a changing soil environment.

Soil organisms perform many functions including decomposition and nutrient cycling. The effects of climate change on soil biology are mixed. Warming of the soil may result in greater microbial activity, releasing more carbon to the atmosphere through increased decomposition. Warming of the soil may also result in slowed microbial growth and less carbon being released through respiration. Greater soil biodiversity is expected to increase soil resilience to changing climate.

Potential effects to soil chemical properties with climate change are linked to other biological and physical changes in the soil, all of which are driven by the soil temperature and moisture inputs. Salinization, acidification, pH, and cation exchange capacity are soil processes and properties that will change with changes to climate. In general, the lower-elevation, drier shrubland and grassland soils are

more vulnerable to changes in soil chemical processes and properties.

There are many potential management actions to decrease vulnerabilities to climate change. The information in this chapter was used as the basis for development of climate change adaptation strategies and tactics for water use, GDEs, and soils (Appendix 4, Chapter 14).

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