

Chapter 3: Climate Change and Hydrology in the Blue Mountains

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Current and Historical Climate in the Blue Mountains

The dominant influences on climatic patterns in the Pacific Northwest are the Pacific Ocean and the Cascade Range. The diurnal temperature range is higher east of the Cascade crest, further inland from the Pacific Ocean. More precipitation falls west of the Cascade Mountains crest, and a strong rain shadow greatly reduces precipitation east of the crest. The southern portion of the Blue Mountains, including the Strawberry subrange, is in the rain shadow of the Cascade Mountains and is predominantly influenced by Great Basin climatic patterns, resulting in warmer and drier conditions. In the northern Blue Mountains, maritime air flows through the Columbia River Gorge, resulting in higher precipitation and more moderate temperature variations (Western Regional Climate Center 2015).

It is important to establish a baseline of historical climate in the Blue Mountains before considering future change. The Blue Mountains area aligns closely with the National Climatic Data Center's (NCDC) Northeast Oregon climate division (Oregon Climate Division 8), which is the area for which we consider current and historical climate. Note that the NCDC information consists of low-elevation climate data, and high-elevation climate patterns may differ from those at low elevations (Luce et al. 2013). The topography of the Blue Mountains results in orographically enhanced local precipitation totals despite being in the lee of the

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Human influence on the climate is clear and changes in the climate are already being realized across the Pacific Northwest.

Cascade Range. The regional annual average precipitation is 44 cm (20th century average), with greater amounts in higher elevation areas in the region. The surrounding Columbia River Plateau and High Desert see less precipitation on an annual basis. The temperatures in the Blue Mountains are slightly cooler than those of the entire region; regionally averaged mean annual temperature is about 7.5 °C (1901 to 2000 average), with colder temperatures at higher elevations.

Human influence on the climate is clear (IPCC 2013), and changes in the climate are already being realized across the Pacific Northwest, where temperatures have warmed by a statistically significant amount. Mean annual temperature in northeast Oregon increased by 0.06 °C per decade between 1895 and 2013, consistent with the overall temperature trend of the entire Pacific Northwest (USDC NOAA National Centers for Environmental Information 2015) (fig. 3.1). Only 3 years have been below the 20th century annual average temperature of 7.5 °C since 1990 (fig. 3.1). Precipitation in the Pacific Northwest is still dominated by interannual variability, such as the El Niño Southern Oscillation (ENSO) (Mote et al. 2013). However, the Blue Mountains region does not exhibit a clear precipitation signal in terms of ENSO phase in the winter months (USDC NOAA Climate Prediction Center 2015). Although there is no detectable or significant precipitation trend in the region, the last 30 years were generally drier than the 20th century average, but with a few very wet years in the mid-late 1990s; the preceding decades (1940 through 1980) were much wetter than recent years (fig 3.2).

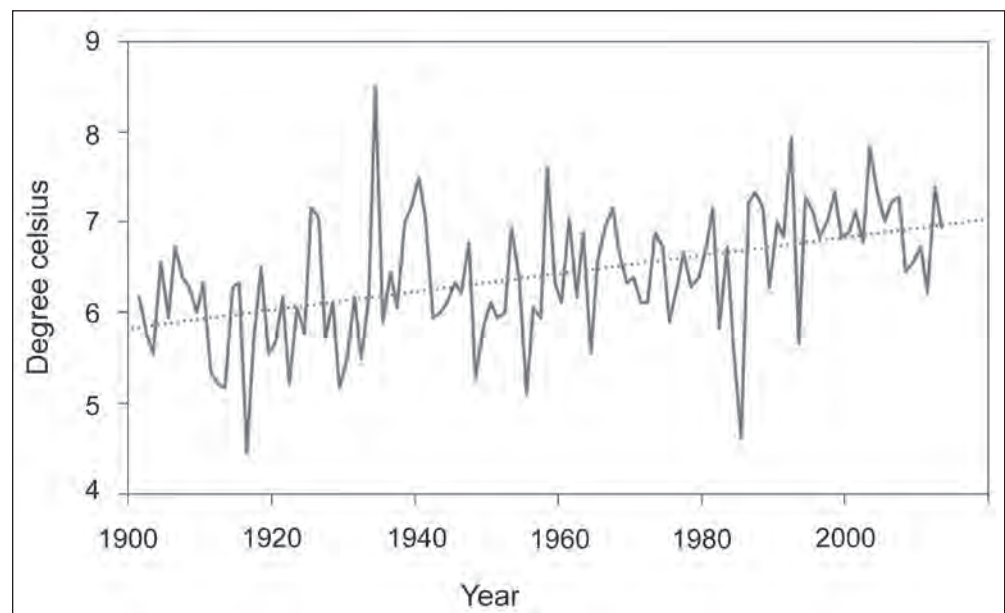


Figure 3.1—Annual historical temperature for Oregon Climate Division 8. Data are from the National Oceanic and Atmospheric Administration National Centers for Environmental Information (<http://www.ncdc.noaa.gov/cag/time-series/us>).

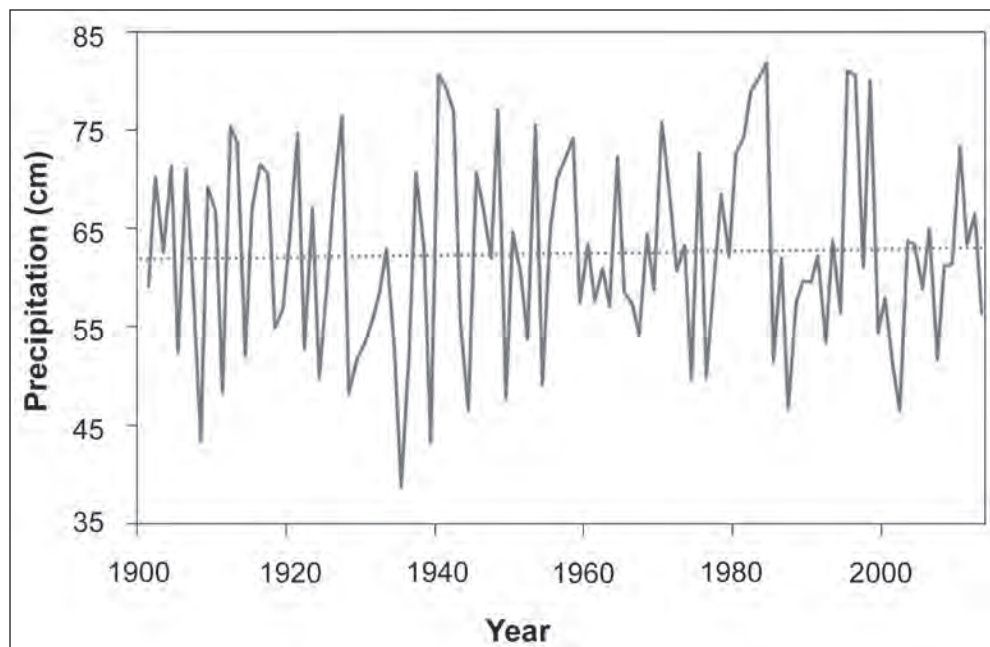


Figure 3.2—Annual historical precipitation for Oregon Climate Division 8. Data are from the National Oceanic and Atmospheric Administration National Centers for Environmental Information (<http://www.ncdc.noaa.gov/cag/time-series/us>).

Future Climate Projections for the Pacific Northwest

Complex global climate models (GCMs) begin to answer questions about future climate. Climate modeling is mostly conducted at global to regional scales because of the computational power required to run GCMs. The disparity between the scale of GCM output and information needs for regional to subregional climate change planning presents some challenges. However, we consider the projections for the Pacific Northwest region relevant for planning in the Blue Mountains; variations in monthly and annual temperature are highly correlated across the Pacific Northwest region.

A number of modeling groups around the world have developed and run GCM simulations, which project future global climate under different future scenarios. The Coupled Model Intercomparison Project (CMIP) is a coordinated experiment involving many of these modeling groups worldwide, offering many simulations for scientists to assess the range of future climate projections for the globe. The latest CMIP experiment is the fifth phase of the project, referred to as CMIP5. Simulations of future climate are driven by Representative Concentration Pathways (RCPs), a departure from the last CMIP experiment, somewhat confusingly titled CMIP3. CMIP3 relied on the Special Report on Emissions Scenarios to drive model projections of future climate (Nakićenović and Swart 2000). The RCPs do not

define greenhouse gas emissions, but instead define future concentrations of greenhouse gases, aerosols, and chemically active gases. The RCPs encompass the range of current estimates regarding the evolution of radiative forcing, or the assumed rate of extra energy entering the climate system throughout the 21st century and beyond (van Vuuren et al. 2011). Although the models are run at the global scale, the following projections are for the Pacific Northwest (generally Oregon, Washington, Idaho, and western Montana) and are based on the analysis described in Mote et al. (2013). More information on CMIP can be found at <http://cmip-pcmdi.llnl.gov/index.html>.

For the Blue Mountains vulnerability assessment, following Mote et al. (2013), we considered two of the CMIP5 scenarios: RCP 4.5 (significant reduction in global greenhouse gases and climate stabilization by year 2100) and RCP 8.5 (increasing greenhouse gases to the end of the 21st century). For the Pacific Northwest, every GCM shows an increase in temperatures in the future, with differences depending on global greenhouse gas emissions (fig. 3.3). There is no plausible future climate scenario from any GCM in which the Pacific Northwest cools in future decades. For the 2041–70 period, models project warming of 1.1 °C to 4.7 °C compared to 1970–99, with the lower end possible only if global greenhouse gas emissions are

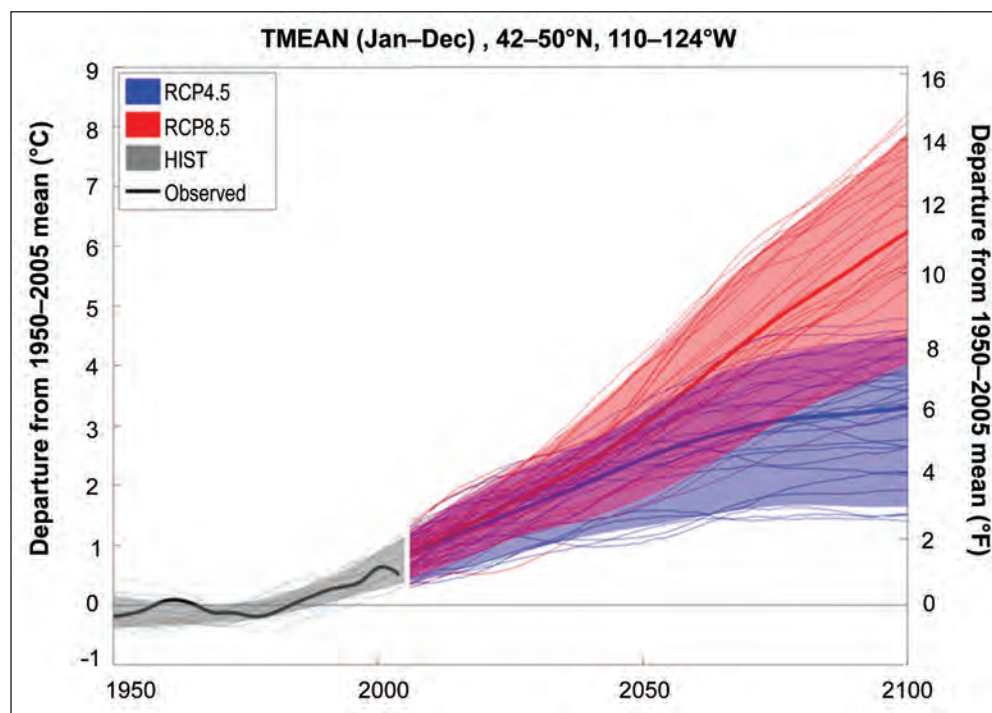


Figure 3.3—Observed (1950–2011) and simulated (1950–2100) Pacific Northwest regional mean temperature for selected global climate models under Representative Concentration Pathways (RCP) 4.5 and 8.5. The gray (historical [HIST]), red (RCP 8.5), and blue (RCP 4.5) envelopes represent the range of model projections. Each individual narrow line represents a model projection, and bold lines are means of the model projections. Figure from Mote et al. (2013).

significantly reduced (RCP 4.5 scenario). Through about 2040, both RCP 4.5 and 8.5 show a similar amount of warming; regional temperatures beyond 2040 depend on future global greenhouse gas emissions. Similar to annual temperature, all models are in agreement that each season will be warmer in the future, with the largest amount of warming occurring in the summer (table 3.1). In each season, RCP 8.5 projects warmer temperatures than RCP 4.5 (Dalton et al. 2013).

Projections for future annual precipitation do not display as clear of a signal as those for temperature; annual precipitation projections range from wetter to drier, and projections for future annual precipitation indicated small trends compared to natural year-to-year variability. Averaging all the model outputs for annual precipitation, the projected future precipitation is close to no change from historical, with a wide range of projections. There is some indication and greater model agreement that summers will be drier in the future, although summers in the Pacific Northwest are already quite dry (table 3.2) (Mote et al. 2013).

Hydrologic Processes in the Blue Mountains

Climate change will likely affect physical hydrological processes and resource values influenced by hydrological processes, including water use, infrastructure, and fish. Specifically, climate change will affect the amount, timing, and type of precipitation, and timing and rate of snowmelt (Luce et al. 2012, 2013; Safeeq et al. 2013), which will affect snowpack volumes (Hamlet et al. 2005), streamflows (Hidalgo et al. 2009, Mantua et al. 2010), and stream temperatures (Isaak et al. 2012, Luce et al. 2014b). Changes in the amount and timing of precipitation will

Climate change will likely affect physical hydrological processes and resource values influenced by hydrological processes, including water use, infrastructure, and fish.

Table 3.1—Summary of changes in temperature projections for the Representative Concentration Pathways (RCPs) 4.5 and 8.5 annually and by season for the Pacific Northwest for the change from historical (1950–1999) to mid-21st century (2041–2070)^a

	Annual		Winter (DJF)		Spring (MAM)		Summer (JJA)		Autumn (SON)	
RCP	4.5	8.5	4.5	8.5	4.5	8.5	4.5	8.5	4.5	8.5
	°C									
Maximum	3.7	4.7	4.0	5.1	4.1	4.6	4.1	5.2	3.2	4.6
Mean	2.4	3.2	2.5	3.2	2.4	3.0	2.6	3.6	2.2	3.1
Minimum	1.1	1.7	0.9	1.3	0.5	1.0	1.3	1.9	0.8	1.6

DJF = December, January, February; MAM = March, April, May; JJA = June, July, August; SON = September, October, November.

^a Values are for the maximum model projection, multimodel mean, and minimum model projection.

Table 3.2—Summary of changes in precipitation projections for Representative Concentration Pathways (RCP) 4.5 and 8.5 emission scenarios annually and by season for the Pacific Northwest for the change from historical (1950–1999) to mid-21st century (2041–2070)^a

	Annual		Winter (DJF)		Spring (MAM)		Summer (JJA)		Autumn (SON)	
RCP	4.5	8.5	4.5	8.5	4.5	8.5	4.5	8.5	4.5	8.5
<i>Percent change</i>										
Maximum	10.1	13.4	16.3	19.8	18.8	26.6	18.0	12.4	13.1	12.3
Mean	2.8	3.2	5.4	7.2	4.3	6.5	-5.6	-7.5	3.2	1.5
Minimum	-4.3	-4.7	-5.6	-10.6	-6.8	-10.6	-33.6	-27.8	-8.5	-11.0

DJF = December, January, February; MAM = March, April, May; JJA = June, July, August; SON = September, October, November.

^a Values are for the maximum model projection, multimodel mean, and minimum projection.

also affect vegetation (chapters 6 and 7), which will further alter water supplies (Adams et al. 2011). Although climate change effects on vegetation will likely be important, they are not considered in the hydrological projections in this chapter. Here we describe hydrologic processes and regimes in the Blue Mountains, historical trends in hydrologic parameters (snowpack, peak streamflow, low streamflow, and stream temperatures), and projected effects of climate change on those hydrologic parameters (box 3.1).

Some of the streamflow simulations shown in this report were generated by the Variable Infiltration Capacity (VIC) model (Liang et al. 1994) using GCMs from the Intergovernmental Panel on Climate Change (IPCC 2013) AR4 assessment to project future climates (Elsner et al. 2009). The VIC projections were prepared from an ensemble of GCM models that had the best match with observations in the historical period (see Littell et al. [2011] for details). Projections for the “2040s” cover an average from 2030 to 2059, and the “2080s” cover 2070 to 2099. Historical metrics were based on the period 1977 through 1997 (Wenger et al. 2010). The VIC data were computed on a 1/16th-degree (about 6-km) grid to produce daily flow data that were further analyzed for metrics important to aquatic ecology (Wenger et al. 2010, 2011b).

Snowpack

Effects of climate change on snowpack in watersheds of the Pacific Northwest can be broadly distinguished by mid-winter temperatures in each basin (Hamlet and Lettenmaier 2007). Rain-dominated basins are above freezing most of the time in winter, and snow accumulation is minimal (<10 percent of October through March precipitation). At a relatively coarse time scale, rain-dominated basins typically

Box 3.1—Summary of climate change effects on hydrology in the Blue Mountains

Broad-scale climate change effect

Warming temperatures will lead to decreased snowpack accumulation and earlier melt out, resulting in shifts in timing and magnitude of streamflow and decreased summer soil moisture. Both peak and low flows may be affected.

Habitat, ecosystem function, or species

Changes in streamflow timing and magnitude could potentially affect all aquatic species and riparian vegetation through either increased or decreased peak flows or decreased summer streamflows. Changes in soil moisture could potentially affect most terrestrial vegetation.

Current condition, existing stressors

Vegetation water stress varies annually under the current climatic regime but may become more pronounced in the future. Topography influences precipitation. Historical land use still affects some sites.

Sensitivity to climatic variability and change

Because of the fundamental role of water in all ecosystems, changes in availability, timing, and volume of water will have ramifications for most terrestrial and aquatic ecosystems.

Expected effects of climate change

The most pronounced changes in snow/streamflow in the Blue Mountains are likely to occur in headwater basins of the Wallowa Mountains, notably the higher elevation radial drainages out of the Eagle Cap Wilderness, with other large changes occurring in the more northerly sections of the Umatilla and Wallowa-Whitman National Forests along the Oregon-Washington border.

Adaptive capacity

Variable. Key challenges posed by changes in hydrology include increased water stress to vegetation with consequences for fire, mortality, growth, etc. Hydrological changes to streamflow are likely to affect some fish species more than others (see chapter 5).

Geographic locations most vulnerable

Mid elevations; areas where snow is not persistent (e.g., Northern Blue Mountains, margins of Wallowa, Elkhorns, Greenhorn, and Strawberry Mountains)

Box 3.1 continued**Risk assessment**

Potential magnitude of climate change effects

- For those regions determined to be sensitive, as identified in the hydrological assessment products
- Moderate magnitude by 2040
 - High magnitude by 2080

Likelihood of climate change effects

- For those regions determined to be sensitive, as identified in the hydrological assessment products
- Moderate likelihood by 2040
 - High likelihood by 2080

have one broad peak in streamflows in the winter that coincides with the regional winter peak in precipitation. However, at a finer time scale, rain-dominated basins may display multiple peaks in streamflow that coincide with individual storms or rain events. Mixed rain and snow (also called “transient” or “transitional”) basins can collect substantial snowpack in winter (10 to 40 percent of October through March precipitation), but are typically only a few degrees below freezing on average in mid-winter. Mixed rain-and-snow basins typically have multiple seasonal streamflow peaks, with one primary peak in late autumn caused by rain, and another in late spring caused by snowmelt. Snowmelt-dominated basins are relatively cold in winter and capture a larger percentage (>40 percent) of their October through March precipitation as snow. Snowmelt-dominated basins typically have relatively low flows through winter and a period of streamflow peaks in spring that coincides with seasonal snowmelt.

Increasing temperatures in the Pacific Northwest over the past 50 years have led to more precipitation falling as rain rather than snow, earlier snowmelt (Hamlet et al. 2007, Stewart et al. 2005), and reduced spring snowpack (Barnett et al. 2008, Hamlet et al. 2005, Mote 2003, Mote et al. 2005). Snowpack in the Pacific Northwest is expected to be sensitive to future temperature increases with changing climate. In response to warming, shifts from snowmelt-dominant to mixed rain-and-snow basins, and from mixed rain-and-snow to rain-dominant basins are projected by the 2040s in the Pacific Northwest (Tohver et al. 2014).

Kramer² developed a snowpack sensitivity map for the Pacific Northwest using data from the Snow Data Assimilation System (SNODAS) (NOHRSC 2004). SNODAS snow water equivalent (SWE) data from 2003 to 2012 were used to characterize the sensitivity of snowpack to climate variability (table 3.3). Luce et al. (2014a) also evaluated snow sensitivity to climate at Snowpack Telemetry sites in the Pacific Northwest, using a spatial analog method to make April 1 SWE projections under a future climate-warming scenario of 3 °C warmer than the past 20 years (expected by around 2050 for the RCP 8.5 scenario [fig. 3.3]). An analysis of snow cover data from strongly contrasting years gives some insight about potential sensitivity of late-season snowpack to a changing climate (see footnote 2) (table 3.3). Results of both studies suggest that there will likely be future declines in snowpack persistence and April 1 SWE throughout the Pacific Northwest, with the largest declines in mid-elevation and wetter locations.

In the Blue Mountains, large areas could lose all or significant portions of April 1 SWE under a 3 °C temperature increase (expected by around 2050 for the RCP 8.5 scenario [fig. 3.3]) (fig. 3.4). Results indicate that snowpack sensitivity is relatively high in the Strawberry Mountains, Monument Rock Wilderness, Wenaha-Tucannon Wilderness, and at mid-elevations in the North Fork John Day, Eagle Cap Wilderness, and Hells Canyon Wilderness (fig. 3.4). Snowpack sensitivity is lower at high elevations in the Wallowa Mountains (Eagle Cap Wilderness), Greenhorn Mountains (North Fork John Day Wilderness), and Hells Canyon Wilderness Area. However, snowpack loss may still be significant (40 to 100 percent loss) in some of these areas (Luce et al. 2014a).

There will likely be future declines in snowpack persistence and April 1 SWE throughout the Pacific Northwest.

² **Kramer, M.G.** Unpublished data. On file with: M.G. Kramer, University of Florida, Gainesville, FL 30503.

Table 3.3—Snowpack sensitivity definitions used in chapter 3 (fig. 3.4)

Sensitivity class	Definition
Persistent—least sensitive	Timing of peak snowmelt differed by >30 days between the warmest, driest year and coldest, wettest year in >30 percent of the subwatershed.
Persistent—more sensitive	Timing of peak snowmelt in the warmest, driest year (2003, El Niño year) occurred >30 days earlier than the coldest, wettest year (2011, La Niña year) in >50 percent of the subwatershed.
Ephemeral snow	April 1 snow water equivalent was <3.8 cm during dry years (no snow) and >3.8 cm during wet years (snow cover) in >80 percent of the subwatershed.

Source: Kramer, M.G. Unpublished data. On file with: M.G. Kramer, University of Florida, Gainesville, FL 30503.

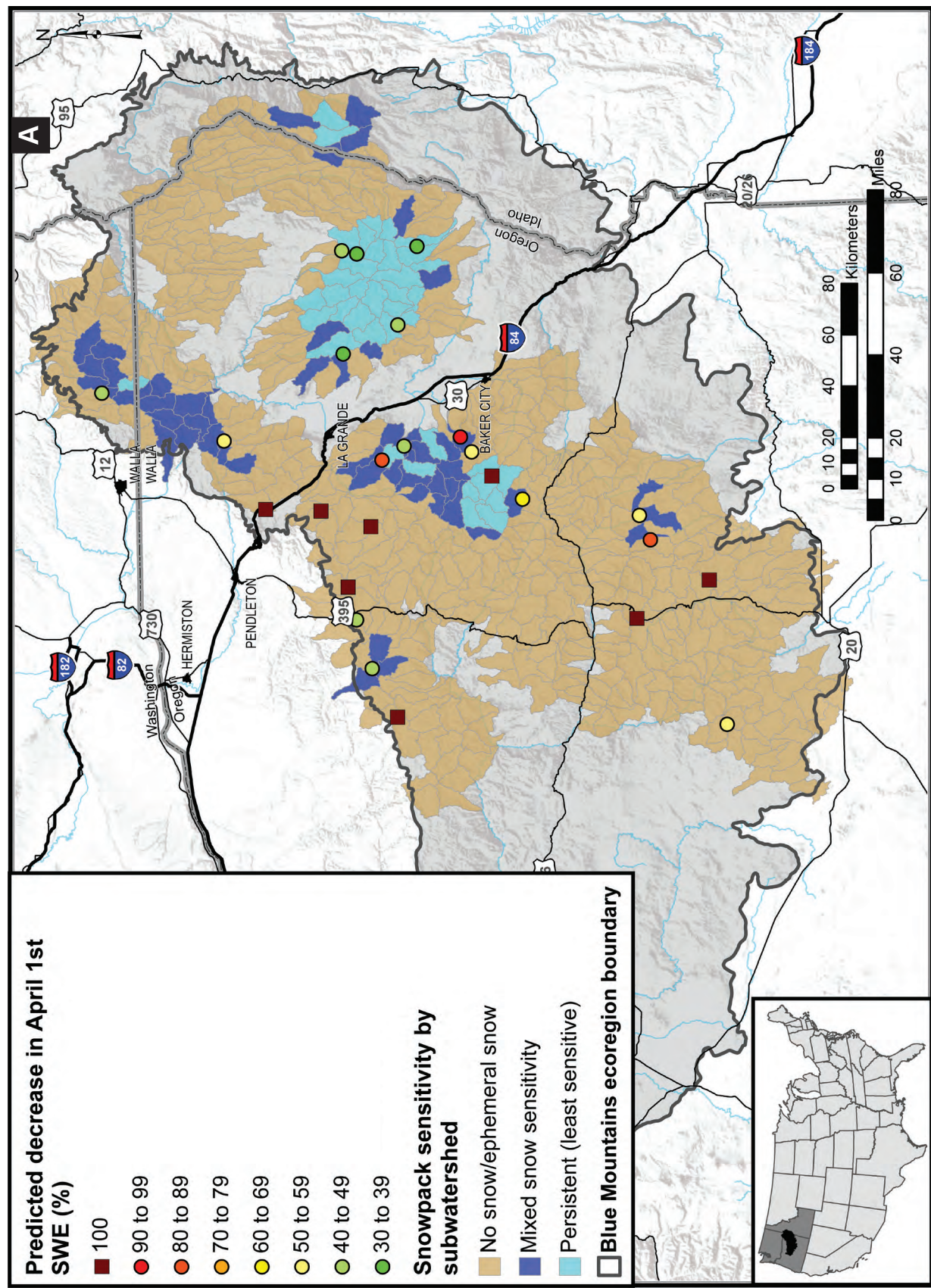


Figure 3.4—Projected change in snow water equivalent (SWE) (a) and mean snow residence time (b) for a 3 °C increase in temperature in the Blue Mountains. Point data are projections at Snowpack Telemetry (SNOTEL) stations from Luce et al. (2014). Snowpack sensitivity classes (the same in both figures) reflect the amount of shift in snowmelt timing seen in two contrasting historical years (Kramer [see footnote 2]; summarized in table 3.3).

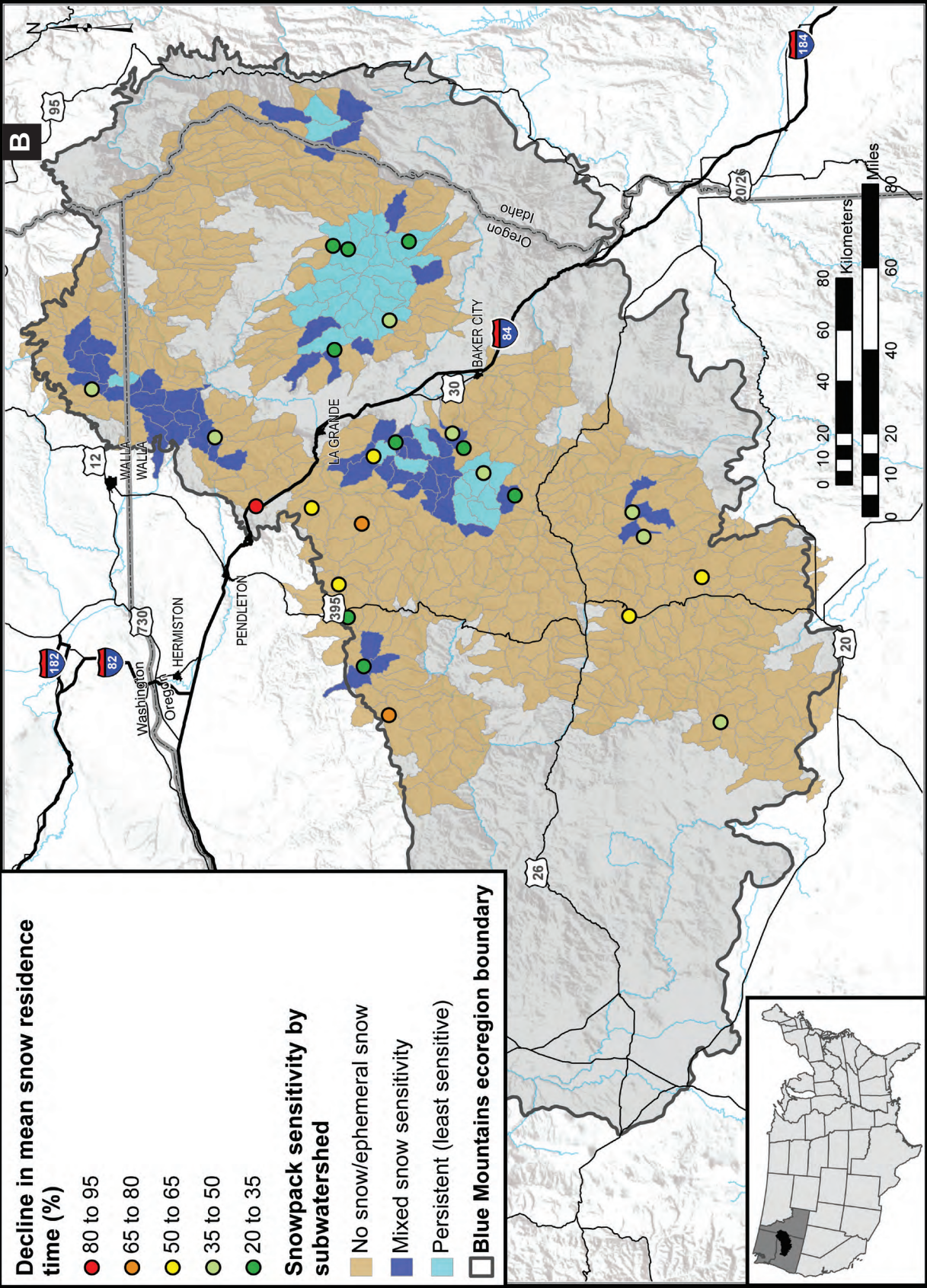


Figure 3.4 continued

Similarly, the VIC model was used to project up to 100 percent loss of April 1 SWE in parts of the Blue Mountains by the 2080s (Hamlet et al. 2013). This study also projected that most of the watersheds in the Blue Mountains that were historically classified as mixed rain and snow will become rain dominant by the 2080s. These watersheds will likely receive more rain and less snow in the winter months.

Peak Flows

Flooding regimes in the Pacific Northwest are sensitive to precipitation intensity, temperature effects on freezing elevation (which determines whether precipitation falls as rain or snow), and the effects of temperature and precipitation change on seasonal snow dynamics (Hamlet and Lettenmaier 2007, Tohver et al. 2014). Floods in the Pacific Northwest typically occur during the autumn and winter because of heavy rainfall (sometimes combined with melting snow) or in spring because of unusually heavy snowpack and rapid snowmelt (Hamlet and Lettenmaier 2007, Sumioka et al. 1998). Summer thunderstorms can also cause local flooding and mass wasting, particularly after wildfire (e.g., Cannon et al. 2010, Istanbuloglu et al. 2004, Luce et al. 2012, Moody and Martin 2009).

Flooding can be exacerbated by rain-on-snow (ROS) events, which are contingent on the windspeed, air temperature, absolute humidity, intensity of precipitation, elevation of the freezing line, and existing snowpack when storms happen (Eiriksson et al. 2013, Harr 1986, Marks et al. 1998, McCabe et al. 2007). Warming affects future flood risk from ROS events differently depending on the importance of these events as a driver of flooding in different basins under the current climate. As temperatures warm, the ROS zone, an elevation band below which there is rarely snow and above which there is rarely rain, will likely shift upwards in elevation. This upward shift in the ROS zone will tend to strongly increase flooding in basins where the current ROS zone is low in the basin (with a large snow collection area above). In contrast, in basins in which the ROS zone is higher in the basin, the upward shift in the ROS zone may only modestly increase the fractional contributing basin area with ROS or potentially shrink the relative contribution of ROS.

In the latter half of the 20th century, increased temperatures led to earlier runoff timing in snowmelt-dominated and mixed rain-and-snow watersheds across the Western United States (Cayan et al. 2001, Hamlet et al. 2007, Stewart et al. 2005). With future increases in temperature and potentially in amount of precipitation in the winter months, extreme hydrologic events (e.g., those currently rated as having 100-year-recurrence intervals) may become more frequent (Hamlet et al. 2013).

An analysis for the Blue Mountains, using VIC model output from Wenger et al. (2010), projects that flood magnitude will increase in the Wallowa Mountains,

Hells Canyon Wilderness Area, and northeastern portion of the Wallowa-Whitman National Forest by the 2080s, particularly in mid-elevation areas most vulnerable to ROS (fig. 3.5a). The frequency of mid-winter flood events does not change over much of the area (fig. 3.5b, c), but the areas showing the greatest change in flood magnitude (fig. 3.5a) are also showing substantial changes in the frequency of the largest flows in winter (fig. 3.5b, c), a measure of flood seasonality that is important to a number of fall-spawning fish species (Goode et al. 2013; Tonina et al. 2008; Wenger et al. 2011a, 2011b).

Low Flows

As a result of earlier snowmelt and peak streamflows over the last 50 years in the Western United States, spring, early summer, and late-summer flows have been decreasing, and fractions of annual flow occurring earlier in the water year have been increasing (Leppi et al. 2011, Luce and Holden 2009, Safeeq et al. 2013, Stewart et al. 2005). An analysis by Stewart et al. (2005) in eastern Oregon showed some of the largest trends toward decreasing fractional flows from March through June. In addition to decreased summer flows, Luce and Holden (2009) showed declines in some annual streamflow quantiles in the Pacific Northwest between 1948 and 2006; they found decreases in the 25th percentile flow (drought year flows) over the study period, meaning that the driest 25 percent of years have become drier across the Pacific Northwest.

Summer low flows are influenced not only by the timing of snowmelt but also by landscape drainage efficiency, or the inherent geologically mediated efficiency of landscapes in converting recharge (precipitation) into discharge (Safeeq et al. 2013, Tague and Grant 2009). The Blue Mountains, which have moderate groundwater contributions, experienced reduced summer flows of 21 to 28 percent between 1949 and 2010 (Safeeq et al. 2013). Safeeq et al. (2014) developed and applied an analytical framework for characterizing summer streamflow sensitivity to a change in the magnitude (mm mm^{-1}) and timing (mm day^{-1}) of recharge at broad spatial scales (assuming an initial recharge volume of 1 mm). This approach facilitates assessments of relative sensitivities in different locations in a watershed or among watersheds. Sensitivity, in this approach, has a very specific meaning: the amount of summer streamflow (at some defined point during the summer, i.e., July 1 or August 1) change in response to a change in either the amount of water that recharges the aquifer during late winter and early spring, and the timing of that recharge. So magnitude sensitivity relates how much summer discharge will change (in millimeters) with a 1-mm change in amount of recharge, and timing sensitivity

Flows have been decreasing, and fractions of annual flow occurring earlier in the water year have been increasing.

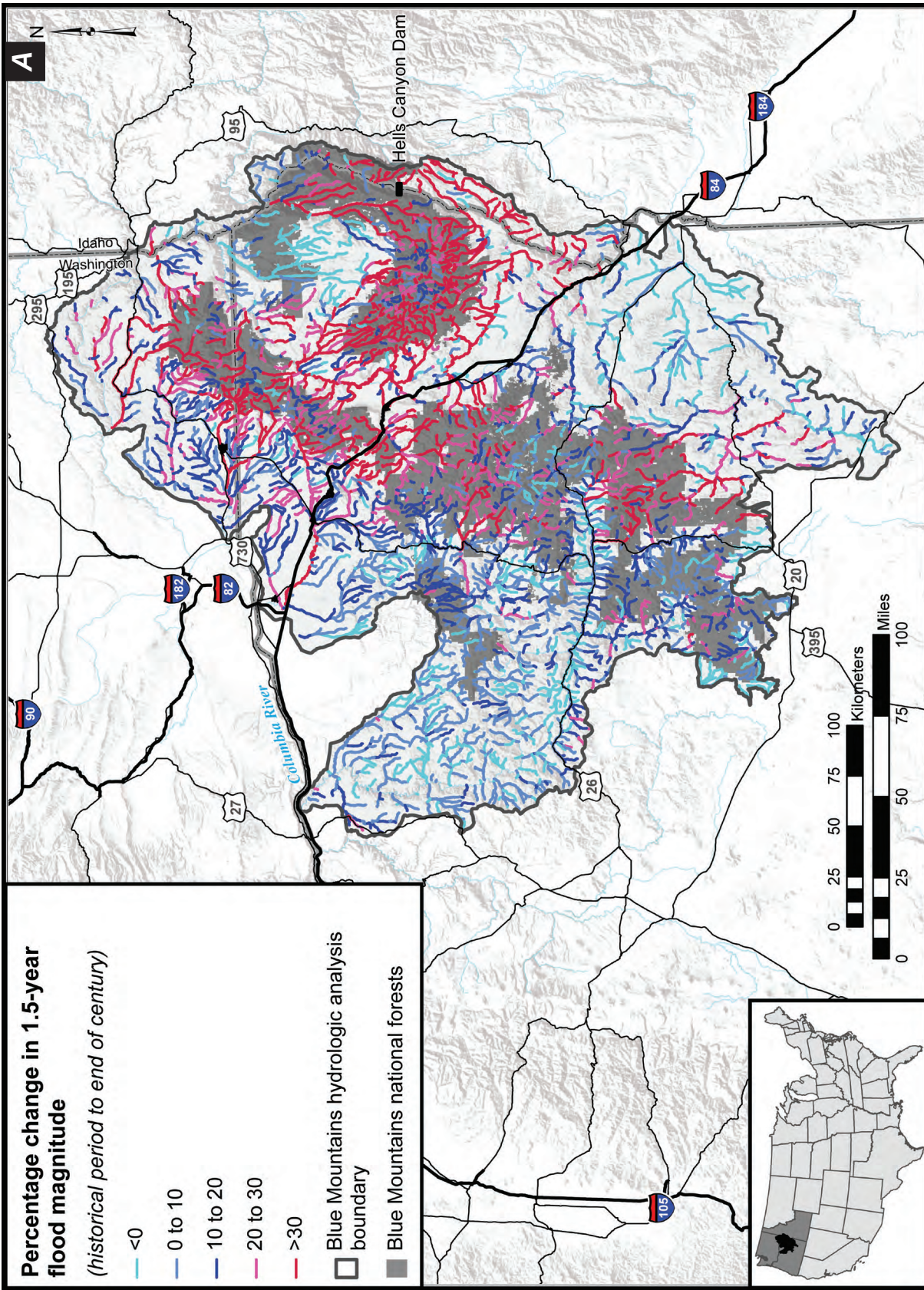


Figure 3.5—(a) Percentage change in the 1.5-year flood magnitude (approximately bankfull) between 2080 and the historical period (1970 to 1999) for the Blue Mountains region; brown lines are state boundaries. (b) Historical mid-winter flooding potential. (c) Projected mid-winter flooding potential for 2080. Flooding potential is shown as the number of days that winter flow is among the highest 5 percent for the year for streams in the Blue Mountains. Comparing the historical (1970–1999) to 2080s model runs shows that although much of the Blue Mountains has frequent mid-winter flooding now, it is rare in some high-elevation areas. In the future, mid-winter flooding is expected to be widespread. All projections are from the Variable Infiltration Capacity (VIC) model, using data from Wenger et al. (2010).

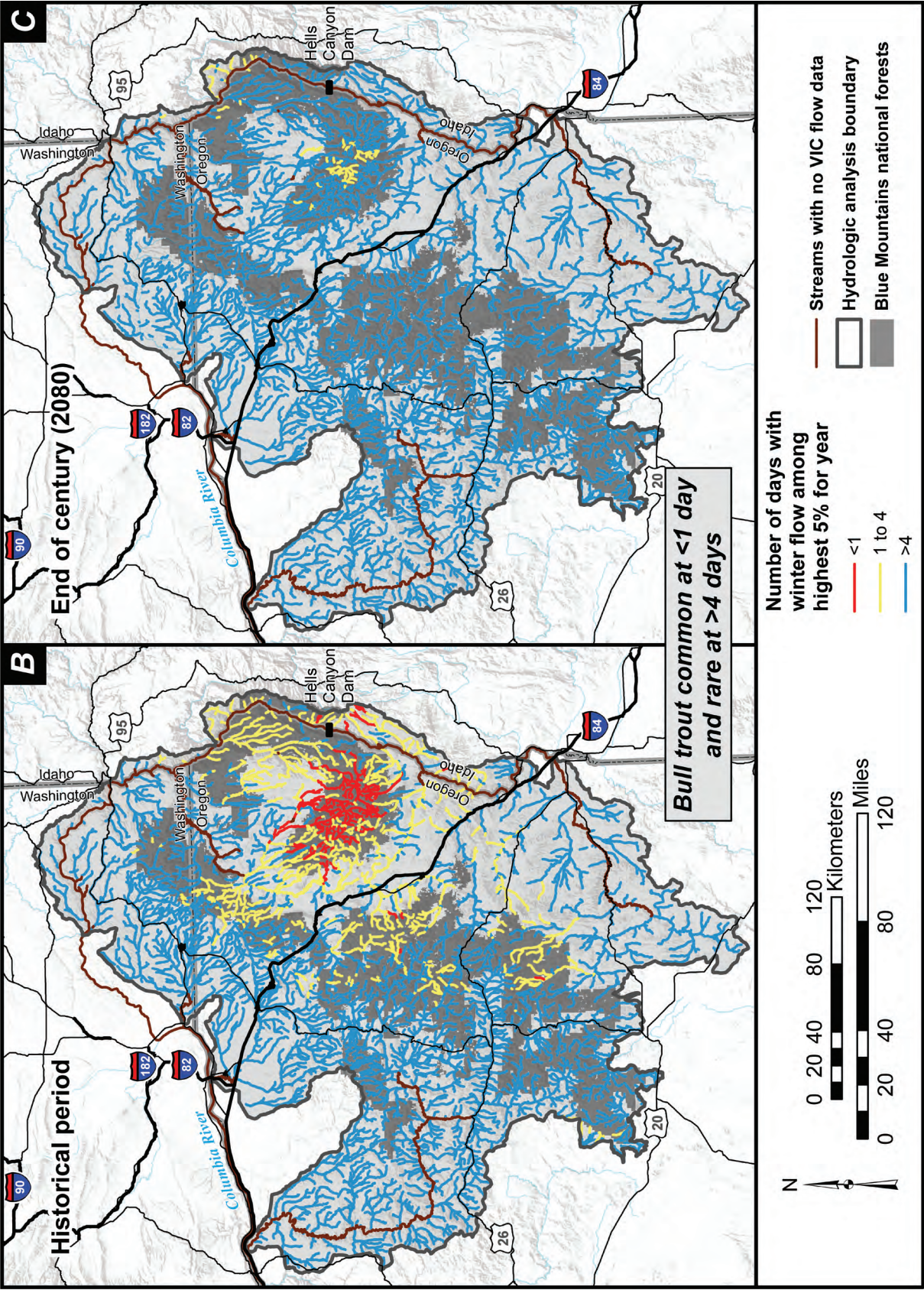


Figure 3.5—(continued).

Snow-dominated regions with late snowmelt, such as the Wallowa Mountains, show relatively high sensitivity.

indicates how much summer discharge will change (in millimeters) with a 1-day change in the timing of recharge. Both metrics make the simplifying assumption that all recharge happens on a particular day, which of course is not the case—recharge happens throughout the rain and snowmelt season. But this approach allows for expressing the intrinsic landscape response to a change in either magnitude or timing of recharge.

Snow-dominated regions with late snowmelt, such as the Wallowa Mountains, show relatively high sensitivity (fig. 3.6), especially early in summer (July), although they are less sensitive than the Cascade and Olympic Mountains. The rest of the Blue Mountains region shows moderate to low sensitivity to changes in the magnitude and timing of snowmelt (fig. 3.6), although sensitivity in the Wallawas is higher in early summer. The level and sensitivity and the spatial extent of highly sensitive areas were shown to diminish over time as summer progresses.

Projections of future low flows using the VIC model (data from Wenger et al. 2010) also show relatively minor decreases in summer streamflow (<10 percent decrease) for 47 percent of perennial streams across the Blue Mountains region by 2080 (fig. 3.7). However, some portions of the region, such as the Wallawas, Greenhorn Mountains, and the Wenaha-Tucannon Wilderness show greater decreases (>30 percent in streamflow by 2080; fig. 3.7).

A direct comparison of the framework developed by Safeeq et al. (2014) and VIC projections for the 2040 time period at the hydrologic unit code 10 watershed scale generally highlights the same portions of the Blue Mountains as being most sensitive to decreases in summer flows as the climate warms in future decades (fig. 3.8). However, the exponential model (Safeeq et al. 2014), which incorporates the role of groundwater, projects larger decreases in summer low flows across the Blue Mountains than the VIC model.

Water Quality

Historical trends in stream temperatures are variable among different studies. Isaak et al. (2010, 2012) found that temperatures at unregulated stream sites closely tracked air temperature trends at nearby weather stations across the Pacific Northwest from 1980 to 2009. Statistically significant stream temperature increases occurred during summer, autumn, and winter, with the highest rates of warming in the summer (reconstructed trend = 0.22 °C per decade). A statistically significant stream cooling trend occurred during the spring season in association with a regional trend towards cooler air temperatures (Abatzoglou et al. 2014, Isaak et al. 2012). Most of the variation in long-term stream temperature trends (80 to 90

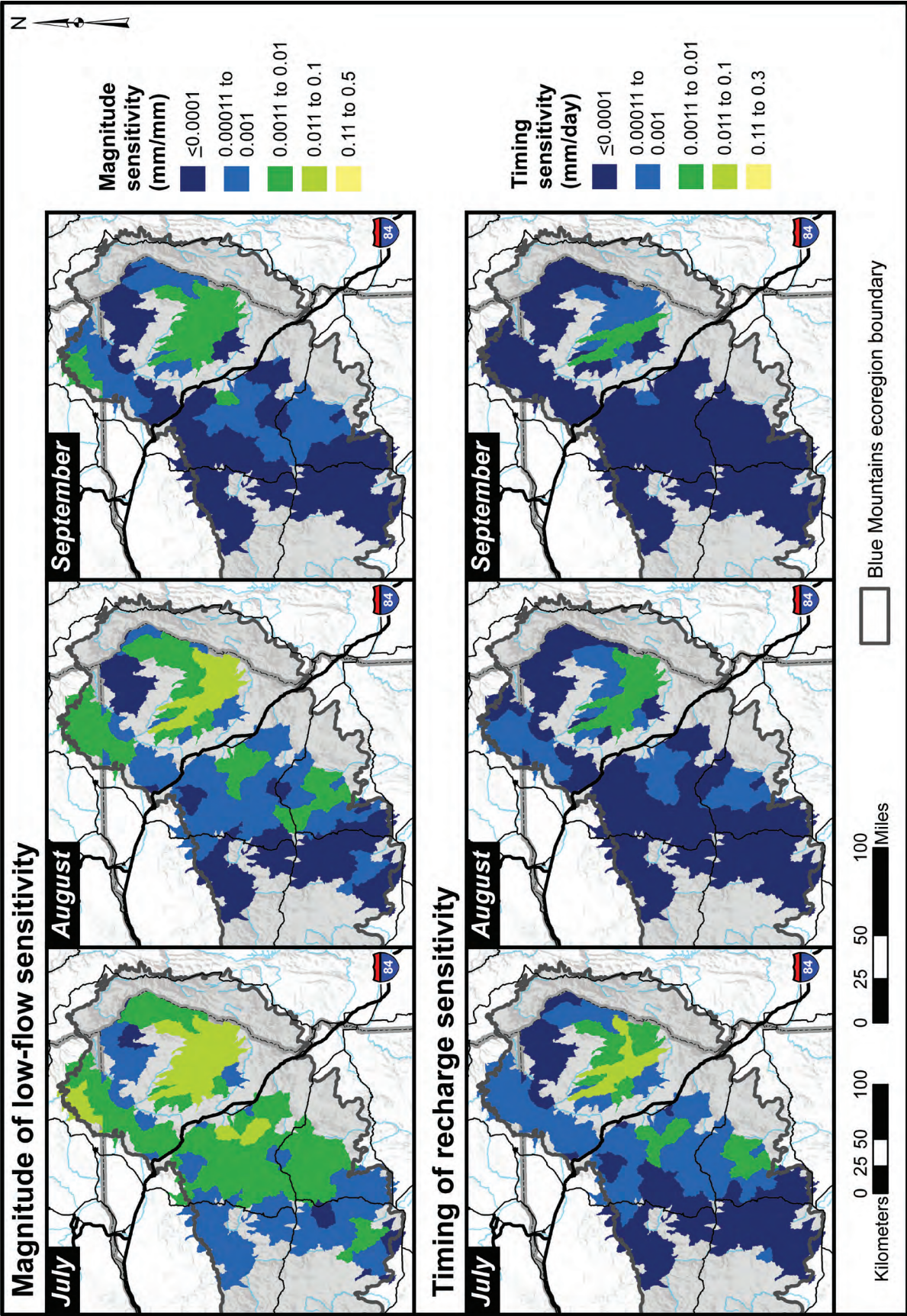


Figure 3.6—Spatial distribution of July, August, and September streamflow sensitivities to a change in magnitude (mm/mm) and timing (mm/day) of recharge from snowmelt or rainfall in the Blue Mountains analysis area. See text for further explanation of streamflow sensitivity. From Safeeq et al. (2014).

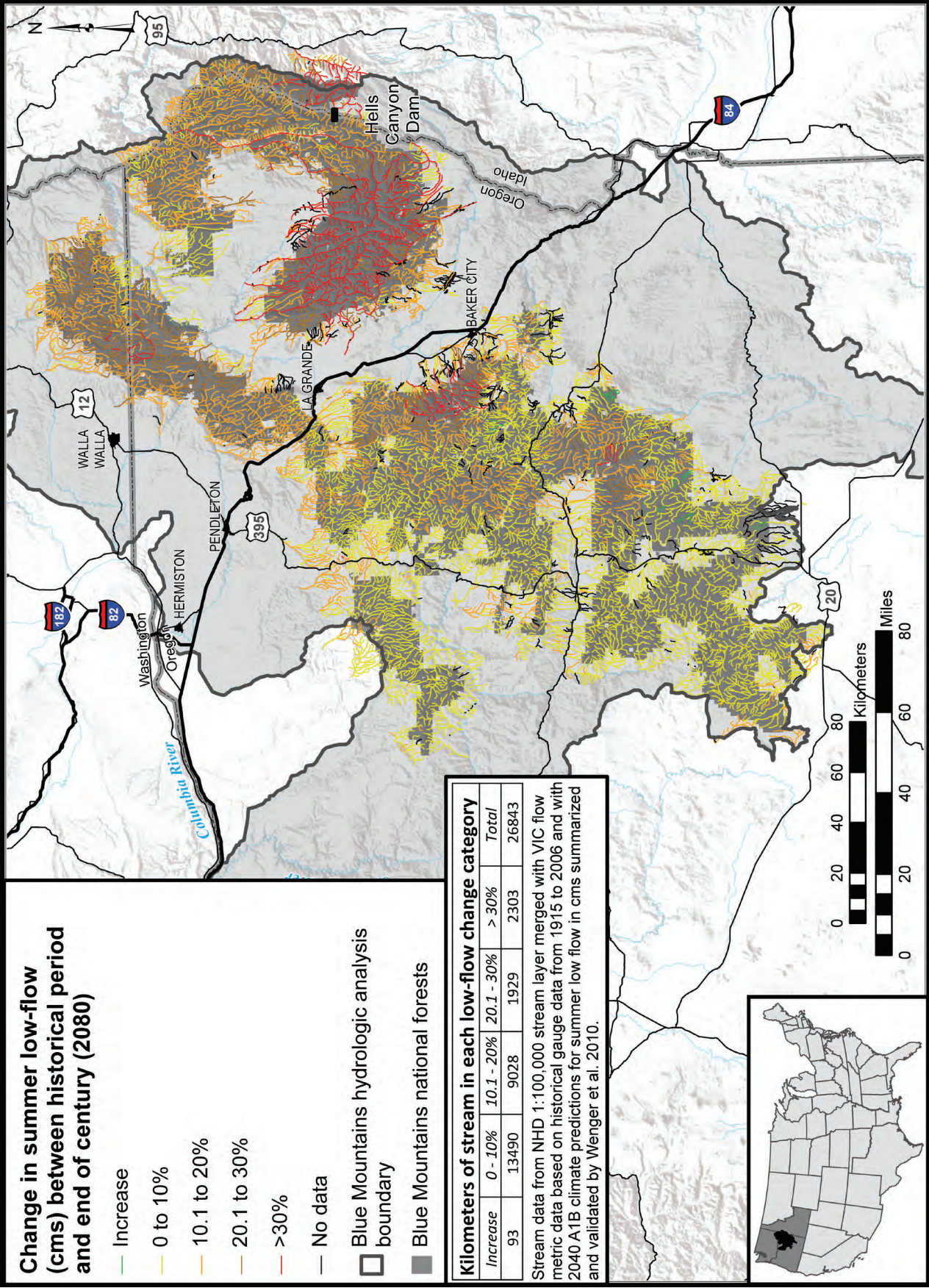


Figure 3.7—Percentage decrease in mean summer streamflow from historical time period (1970–1999) to 2080 for streams in the Blue Mountains region. Projections are from the Variable Infiltration Capacity (VIC) hydrologic model using data from Wenger et al. (2010).

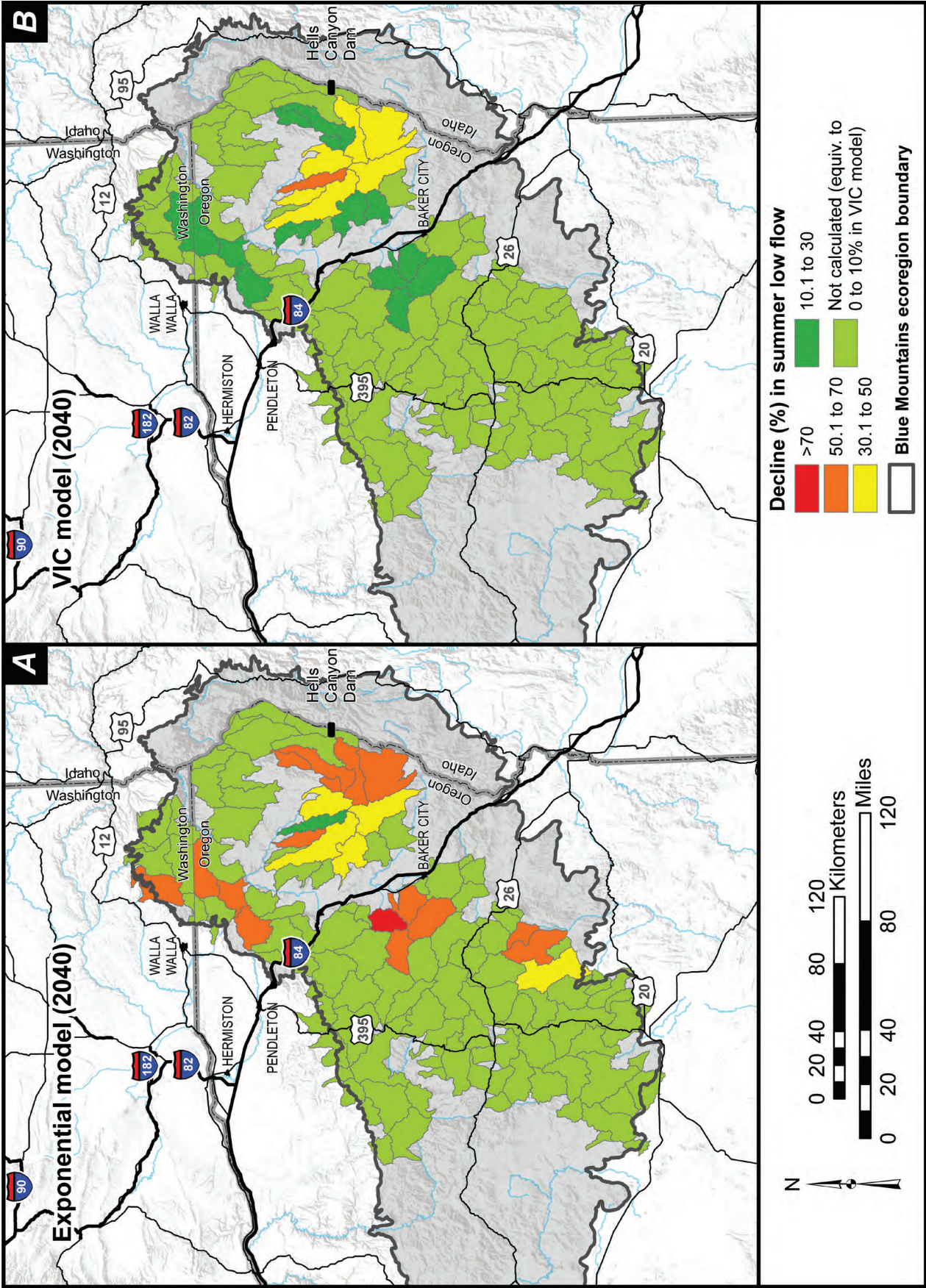


Figure 3.8—Comparison of percentage decline in summer low flows for hydrologic unit code 10 watersheds, calculated as percentage of average daily flow predicted by the Variable Infiltration Capacity (VIC) model (Wenger et al. 2010) and the exponential model (Safeeq et al. 2014) using historical (1915–2006) data and the AIB emission scenario for the 2040s. Low-flow calculations using the exponential model were calculated only when streamflow decline was ≥ 0.01 mm/day.

percent) was explained by air temperature trends and a smaller proportion by discharge trends (10 to 20 percent). Arismendi et al. (2012) examined stream data from a larger number of sites and different periods of record and found variable trends in stream temperature, concluding that stream temperatures have increased at some minimally altered sites in the Pacific Northwest (28 to 44 percent) and decreased at others (22 to 33 percent); no detectable trends were found at the remaining sites. Stream temperature trends were influenced by the length of record, period of record, and location relative to dams, with more warming trends becoming apparent where longer term records were available.

Luce et al. (2014b) analyzed summer stream temperature records from forested streams in the Pacific Northwest and found that cold streams were generally not as sensitive as warm streams to climatic conditions. Thus, temperature in low-elevation, warmer streams (less shade, less cool groundwater inputs) will likely increase the most in the future. These results suggest that these warmer streams in the Blue Mountains are relatively sensitive to climate.

The NorWeST Regional Stream Temperature Database (<http://www.fs.fed.us/rm/boise/AWAE/projects/NorWeST.html>) used extensive stream temperature observations and spatial statistical models to characterize stream temperatures throughout the Blue Mountains under recent historical conditions at a 1-km resolution (Isaak et al. 2015) (fig. 3.9). Future stream temperatures were then projected based on these historical conditions, assessments of past sensitivity to climate, and projections of future climatic conditions. Results project basinwide average August stream temperatures in the Blue Mountains to increase by about 1 °C by 2040 and by nearly 2 °C by 2080 in direct response to climatic conditions (i.e., no consideration of secondary effects, such as increased fire). Warmer streams in the basin will likely warm to a greater degree than cooler ones (Luce et al. 2014b).

Decreasing summer water availability and warming temperatures across the Western United States may contribute to forest mortality in some locations (Adams et al. 2009, Allen et al. 2010, Breshears et al. 2005, Meddens and Hicke 2014, van Mantgem et al. 2009) and increased wildfire extent compared to the mid-20th century (Littell et al. 2009, Westerling et al. 2006). Increased area burned, particularly if fire in riparian areas results in decreased shade over streams, will contribute further to stream temperature increases (Dunham et al. 2007, Isaak et al. 2010). Increases in fire have increased basin-scale sediment yields in some basins (Goode et al. 2012).

Warmer streams in the basin will likely warm to a greater degree than cooler ones.

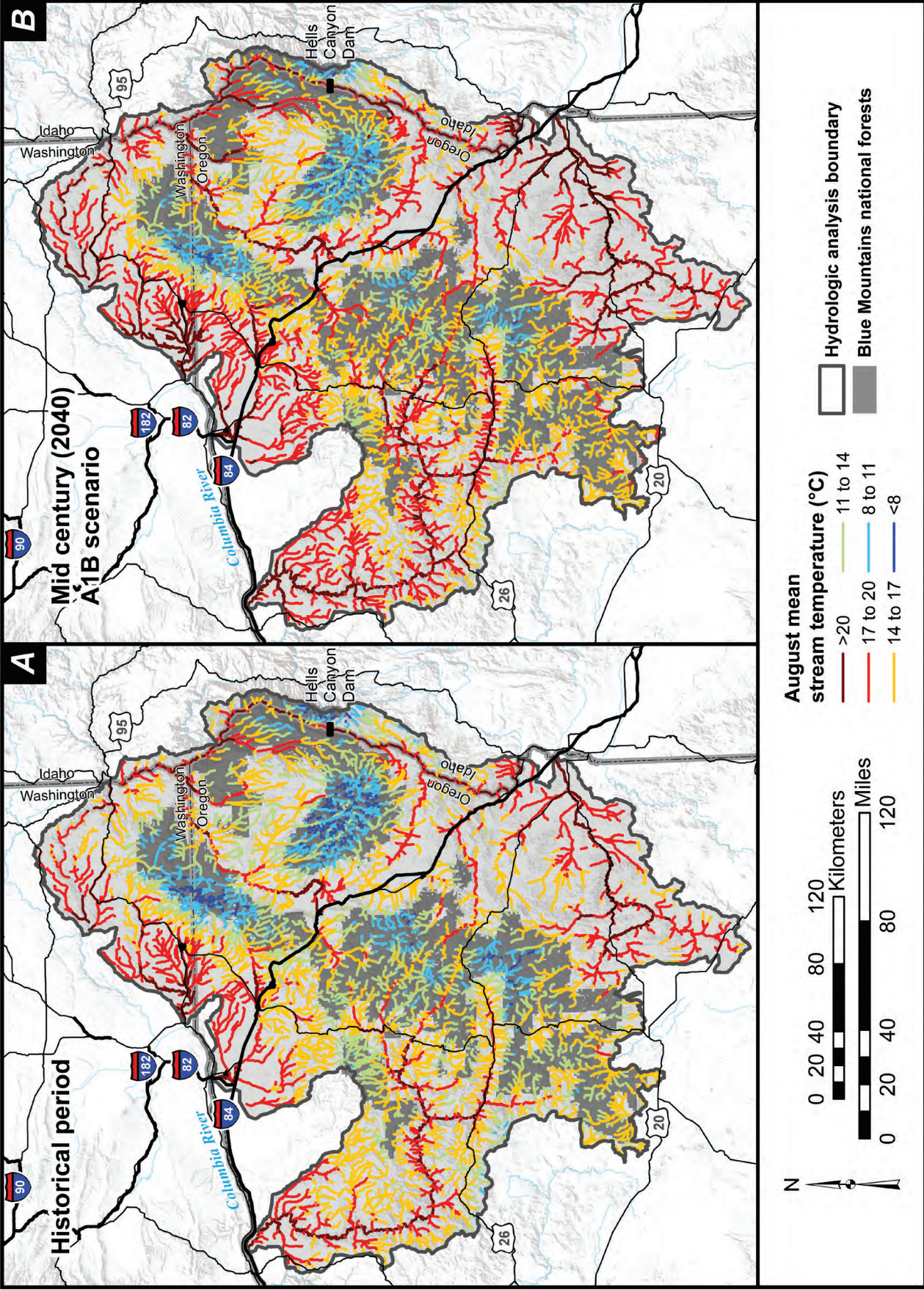


Figure 3.9—Recent historical (1970–1999) (a) and projected future (b) (2040, A1B Scenario) August mean temperatures for streams in the Blue Mountains region. Projections are from NorWeST Regional Stream Temperature Database and Model (<http://www.fs.fed.us/rm/boise/AWAE/projects/NorWeST.html>).

Conclusions

The results and map products discussed in this chapter represent our current best understanding of the likely effects of climate change on key hydrologic processes. Nevertheless, these results should be applied with caution. Key uncertainties include the specific climate trajectories that the Blue Mountains will experience in the future, critical assumptions underlying all models used, and the myriad uncertainties and errors attached to the calibration of each of the models. Resource managers wishing to apply the results of this analysis in forest planning are encouraged to read the primary literature in which the strengths and limitations of different modeling and forecasting approaches are described.

In general, projections of future trends in streamflow and related processes are strongest in characterizing relative sensitivities of different parts of the landscape rather than absolute changes. In other words, the spatial pattern of trends is more robust than projections associated with any particular location. Similarly, more confidence applies to the interpretation of relative, as opposed to absolute, magnitudes of projected changes. Differences in results between modeling approaches, such as the low-flow analysis, should be interpreted as bracketing likely potential changes. Finally, the models used here contain uncertainties related to the quantification of soil, vegetation, and other characteristics used to generate hydrologic dynamics.

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