1. Climate change in Oregon's land and marine environments

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Summary and Knowledge Gaps

The human race is profoundly altering the composition of Earth's atmosphere, chiefly by burning fossil fuels, and there is strong evidence that these changes are responsible for much of the global increase in temperature since the mid-20th century. A recent (2001-2009) leveling off of global temperature trends can be understood as an interaction between continuing increases of greenhouse gases and a slight decline in solar output connected with the 11-year solar cycle, and does not indicate that global warming has ceased permanently.

Attribution - that is, formally understanding causes - of changes in regional climate is difficult owing to the large variability, relative to the signal, on such spatial scales. Nonetheless, the warming observed in the Pacific Northwest during the 20th century (1.5°F, 0.8°C) is roughly the same as that expected with rising greenhouse gases.

Future regional climate change will likely be marked by:

- Increases in temperature of around 0.3°C (0.5°F) per decade, which could be lower if global greenhouse gas emissions are lower than expected
- An accentuated warming and drying in summer
- Increased frequency of extreme daily precipitation
- A northward shift in the storm track and slightly fewer but more intense storms

There is little indication yet from global models that patterns of climate variability such as El Niño-Southern Oscillation or North Pacific variability will change substantially in the future.

Oregon's relatively well-monitored coastal waters have exhibited increases in wind-driven coastal upwelling since 1948. Since subsurface monitoring began in the 1960s, scientists have observed a warming and freshening of subsurface waters over the continental shelf, the continental slope, and offshore, contributing 3 cm (1.2") of sea level rise. Since 1975, the concentration of subsurface dissolved oxygen along the Oregon coast has decreased and during some recent summers, oxygen concentrations near the coastal ocean floor have sometimes been nearly or fully depleted. Links between these observed changes and human influences on the climate system have not been established.

Future changes in Oregon's coastal ocean are likely to include substantial increases in water temperatures, far surpassing natural variability. Although most work indicates negligible

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future changes in the strength of coastal upwelling, the acidity of freshly upwelled waters will continue to increase as it has in recent decades. This increase in corrosive effects will likely impact some estuarine shellfish species in the next few decades.

1.1 Overview of Global Climate Change

Climate on Earth has changed profoundly during the planet's history, but the pace, scope, and cause of recent changes are unprecedented during the period of human existence. Furthermore, these changes arrive at a time when a burgeoning human population has built infrastructure and expectations around the climate of the past. In this section we examine natural climate changes in the remote and the recent past, and contrast it with the current, human-caused changes of the past few decades.

1.1.1 Climate Change in the Distant and Recent Past

1.1.1.1 Before 2.6 million years ago

Over geologic time scales, Earth's climate has ranged from much warmer than today to much cooler. At one extreme, during the latter part of the Proterozoic (around 700 million years ago), ice sheets may have extended to near the equator (MacDonald 2010). At the other extreme, during the Cretaceous (100 million years ago), frost-sensitive plants occurred within the Arctic Circle. These extremes resulted from complex interactions between a changing geography (the location, extent, and topography of continents and oceans), atmospheric composition, orbital geometry (for example, the tilt of Earth's axis and the eccentricity of its orbit), and solar luminosity. Earth scientists understand this climate history and the various mechanisms operating in the climate system with decreasing detail the further back in time this history is studied. This summary will be limited to the Cenozoic era, the most recent 65.5 million years that follows the catastrophic extinctions that are linked to a meteor impact.

The temperature record of the Cenozoic was constructed by compiling records of oxygen isotopes measured on carbonate organisms in ocean sediment cores (Figure 1.1; Zachos et al., 2001). The values of these isotopes reflect the deep-sea temperature, but during times of continental glaciations the record mainly reflects the volume of ice sheets on land. The early Cenozoic is marked by deep-sea temperatures of at least 12°C warmer than today, peaking in the early Eocene. At 55 million years ago, a "brief" several-thousand year period of higher temperatures (Paleocene-Eocene Thermal Maximum) is marked by significant changes in the distribution of species and marks the beginning of many mammal lineages (Gingerich, 2006). Current evidence points to a large release of methane from ocean sediments as the cause of this warmth. Other greenhouse gases clearly were important for maintaining the overall warmth of the early Cenozoic. While it is difficult to reconstruct atmospheric CO_2 concentration from ocean-sediment evidence, the most recent studies show broad agreement between two approaches. These studies suggest CO_2 concentrations were as much as 15 times higher than pre-industrial concentrations.

After 50 million years ago, the Cenozoic is marked by a long-term global cooling trend that was interrupted by several distinct events (Figure 1.1A). Two leading hypotheses for this cooling are establishment of gateways in oceanic circulation that isolated Antarctica in its southern position and the drawdown of CO_2. The first hypothesis is supported by the timing of the onset of cooling and glaciation in the Oligocene. The second hypothesis is supported by recent reconstructions of CO_2 (Figure 1.1A), especially declines in CO_2 at the start of the Pleistocene (Jansen et al., 2007). Weathering of silicate rocks consumes carbonic acids that are derived from

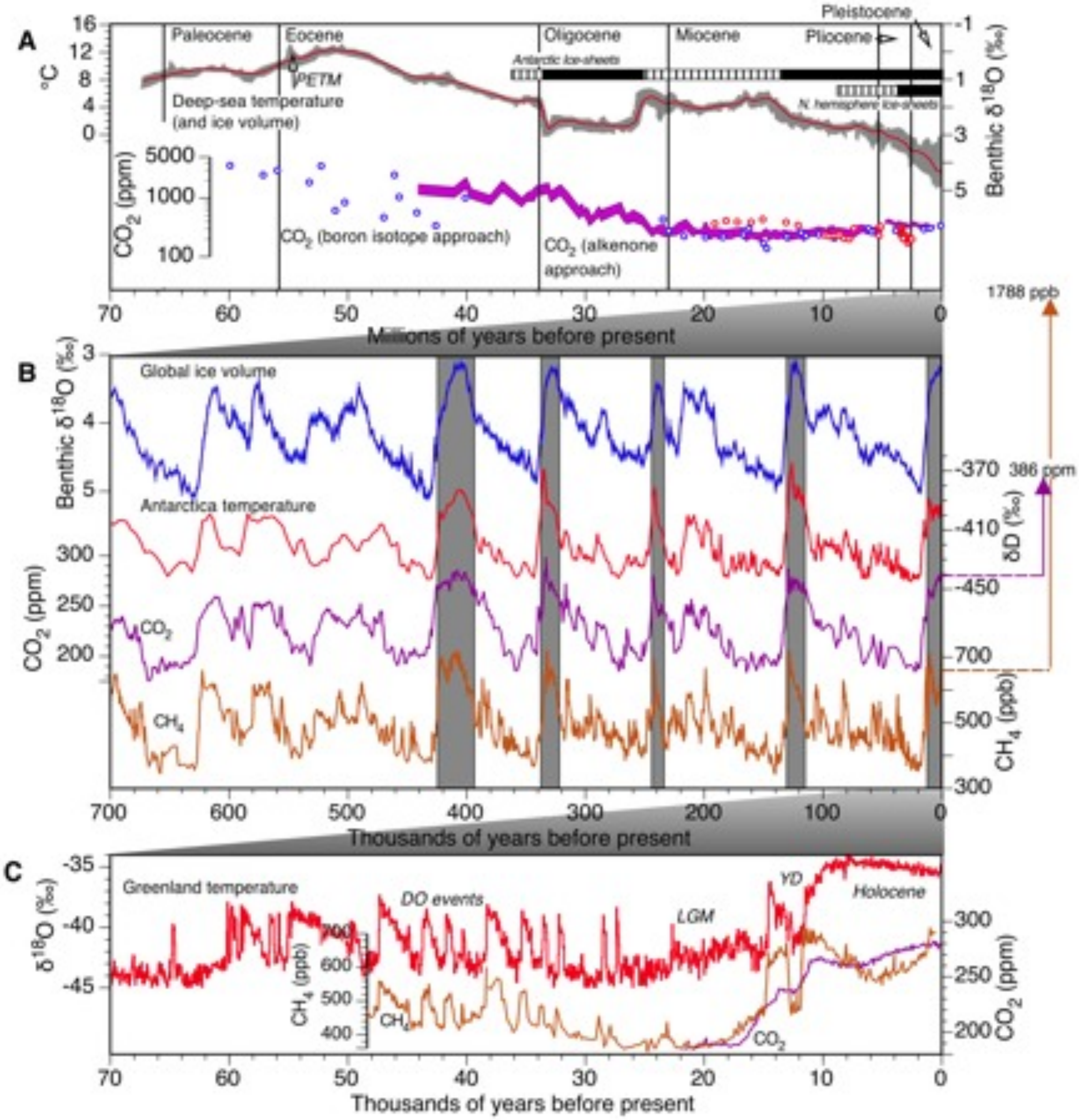

Figure 1.1 Global climate and greenhouse gas records over three time scales. **A.** Deep-sea temperature over the Cenozoic determined from stable oxygen isotope measurements from several ocean sediment cores (Zachos et al. 2001). The $\delta_{18}O$ values are increased by the presence of continental ice sheets, and thus the temperature scale only refers to periods without significant glaciation. The red line a smoothed curve fit and gray background shows the range of measurements. CO2 concentration through the Cenozoic is difficult to measure, and thus several studies are shown here. The boron-isotope approach is shown as blue circles (Pearson and Palmer 2000). A more recent study (red circles) shows a much stronger correlation of CO_2 with events in the Cenozoic (Tripati et al. 2009). An alternative approach using alkenones (purple line) shows changes in the Oligocene but not during late-Cenozoic climate change (Pagani et al. 2005). **B.** Changes in global ice volume estimate from ocean sediment cores (Liseicki and Raymo 2005), and deuterium (δD, a proxy for local temperature), CO_2 and CH_4 concentrations from the EPICA ice core (Petit et al., 1999; Monnin et al. 2001; EPICA community members, 2004; Spahni et al., 2005; Siegenthaler et al., 2005; Jensen et al., 2004). Gray bars denote warm interglacial periods between glacial periods marked by cooling and fluctuating temperatures. Current CO_2 and CH_4 levels, plotted on the same scale as the ice-core data, are shown to the right. **C.** The past 70,000 years of temperatures on the Greenland ice sheet (North Greenland Ice Core Project members. 2004), showing distinct Dansgaard-Oeschger events, the last glacial maximum (LGM), the Younger Dryas (YD) and the Holocene (past 11,600 years). A high-resolution CH_4 record shows that the abrupt temperature fluctuations during the glacial period had global influence (Flückiger et al. 2004; EPICA Community Members 2006). High resolution CO_2 since the LGM is from Monnin et al. (2004).

atmospheric CO_2; thus, increased weathering from the uplift of the Himalaya and Tibet Plateau, which began 55 million years ago, could increase movement of CO_2 from the atmosphere to carbonates in ocean sediments (Garzione, 2008).

1.1.1.2 Glacial and interglacial climate changes

Beginning 2.6 million years ago (the Pleistocene), global climate completed its transition from a "greenhouse" world that lacks ice sheets to an "ice-house" world marked by cyclic glaciations in the northern hemisphere. Causes of the onset of the glacial cycles remains a major area of research, and leading hypotheses involve increased delivery of moisture to North America (to build snowpack) and decreases in summer temperatures (to reduce snowmelt). In contrast, the general pacing of the glacial cycles and the feedbacks in the climate system are better understood. The detailed climate records from ocean sediments and ice sheets reveal dramatic changes in climate and greenhouse gas concentrations that suggest tight linkages among components of the climate system (Figure 1.1B).

During the glacial cycles of the Pleistocene, the three major controls of Earth's climate were incoming solar radiation (*insolation*), ice extent, and greenhouse gases. Slow, progressive changes over time in three features of Earth's orbit (orbital shape, axial tilt, and seasonal precession of the perihelion) have combined to affect the latitudinal distribution and seasonal cycle of insolation. The Milankovitch theory of orbital control of onset of glaciations states that times of ice sheet growth occur when northern hemisphere summer insolation is low, a pattern that is broadly consistent with the timing of glacial cycles. During extensive glaciation, ice sheets reflect shortwave solar radiation back to space, affecting both the global balance of incoming and outgoing radiation as well as modifying timing of past glacial cycles

The EPICA ice core from Antarctica reveals a close correspondence of CO_2 concentration and the global ice volume as estimated from ocean sediment cores (blue and purple lines in Figure 1.1B). The proposed causes of the 90 ppm drop in CO_2 during glacial periods include moving atmospheric CO_2 into the ocean, by changes in solubility of CO_2 in the ocean, increased biological productivity in the ocean surface waters, and changes in ocean circulation. In contrast, methane (CH_4) concentration is closely correlated with the northern hemisphere incoming solar radiation, which affects the strength of monsoons and thus the extent of wetlands (that produce methane) in subtropical climates (Ruddiman, 2006). Both gases have roles in the climate system as forcing ice sheet response and as positive feedbacks to changes in the ice sheet already underway. At cycles of around 23,000 years, the response of ice volume follows changes in CO2 within a few thousand years, suggesting ice sheets are responding to greenhouse gas forcing (and other factors). However, for cycles of 41,000 years (the orbital tilt cycle) both gases have immediate response to changing ice sheet extent, providing a powerful rapid positive feedback to changes in ice sheets.

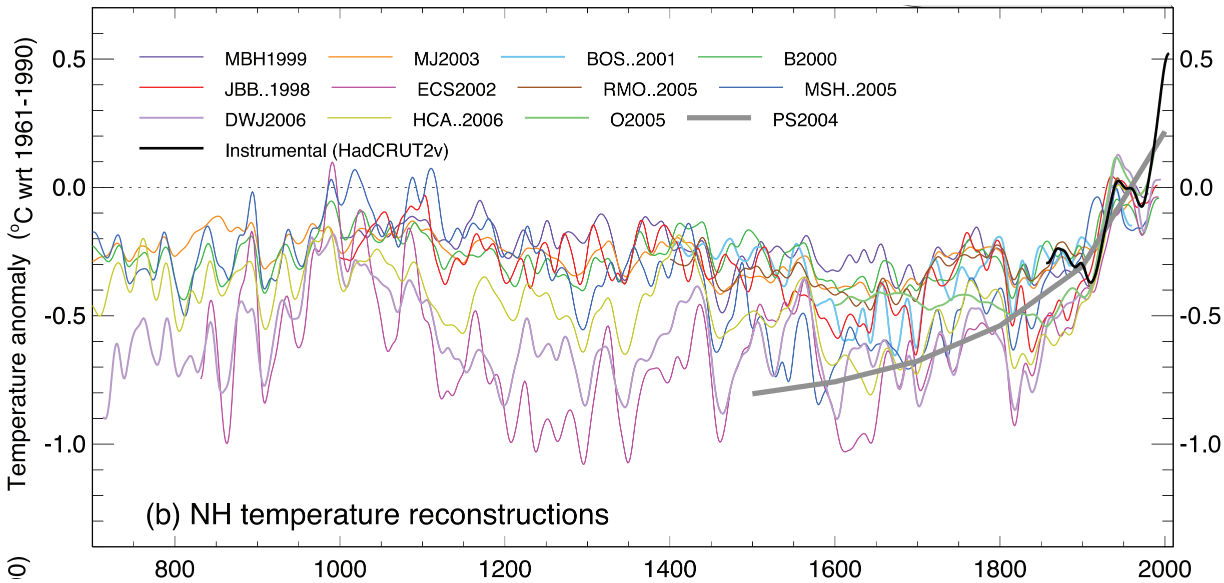

Figure 1.2 Records of NH temperature variation during the last 1,300 years. Reconstructions using multiple climate proxy records and the HadCRUT2v instrumental temperature record in black. All series have been smoothed with a Gaussian-weighted filter to remove fluctuations on timescales less than 30 years; smoothed values are obtained up to both ends of each record by extending the records with the mean of the adjacent existing values. All temperatures represent anomalies (°C) from the 1961 to 1990 mean. Figure 6.10 from IPCC (Jansen et al. 2007).

1.1.1.3 Deglacial and Holocene climate (the past 20,000 years)

The Last Glacial Maximum (LGM), when the maximum extent of the North American and European ice sheets occurred, occurred near the end of the last glacial period, at 21,000 years ago. Increased summer insolation after that time was sufficient to reduce ice sheet extent. Near the most rapid phase of this deglaciation, at 12,900 years ago, a distinct reversal to a cold climate called the Younger Dryas began and lasted until 11,600 years ago. The causes of this event are unknown, but some data suggest a relationship to a change in ocean circulation patterns in the northern Atlantic initiated by a meltwater pulse, while others have proposed a comet impact as the initiator of the meltwater pulse. The effects of the Younger Dryas period are observed throughout the northern hemisphere, though they are most distinct in the North Atlantic.

The Holocene refers to roughly the last 11,600 years, during which time human civilization developed. Following the Younger Dryas period, the early Holocene was marked by an increase in northern-hemisphere summer insolation and a decrease in northern-hemisphere winter insolation. During these strongly seasonal climates, pollen records show more extensive fire and dry-adapted vegetation in most of North America (except those areas affected by increased monsoonal precipitation). At around 8400 years ago, a major collapse of the Laurentide (North American) ice sheet resulted in a large outburst of fresh water, likely causing a century-scale cold period (at 8200 years ago) by slowing down the conveyor circulation in the North Atlantic. The latter half of the Holocene in North America is marked by a general cooling trend, mirroring the trend in summer insolation.

Tree-ring and other proxy sources of year-to-year variations allow scientists to infer climatic conditions in the last ~2000 years. Using a combination of proxy data and statistical algorithms, a number of researchers have estimated the northern hemisphere average temperatures; errors associated with the sparseness of the proxy records diminish over time. Much of the region around the North Atlantic tended to be warmer than the long-term average in the years 1000 to 1250 (first called the "Medieval Warm Period" and now commonly called the "Medieval Climatic Anomaly") but colder than average from 1400 to 1830 (the "Little Ice Age"). These anomalies were generally not as strong or synchronous in other parts of the world, and consequently hemispheric reconstructions indicate a difference of at most 1°C (1.8°F) between the warmest decade of the Medieval Climatic Anomaly and the coolest period of the Little Ice Age, and most reconstructions indicate a substantially smaller difference (Figure 1.2). The potential forcings of these climate fluctuations are still being studied, including CO_2 concentration, solar variability, and major volcanic eruptions that increase atmospheric reflectivity and reduce solar radiation reaching the surface. From the coldest period of the 19th century until 1900, hemispheric temperatures rose, by various estimates between 0 and 0.7°C (1.3°F), compared with 0.7°C (1.3°F) warming during the 20th century.

1.1.1.4 The instrumental period[1]

One might surmise that the advent of thermometers would provide a perfect measurement of a location's temperatures, and that temperatures could simply be averaged to provide the globally averaged temperature. But because these observations originally were intended simply

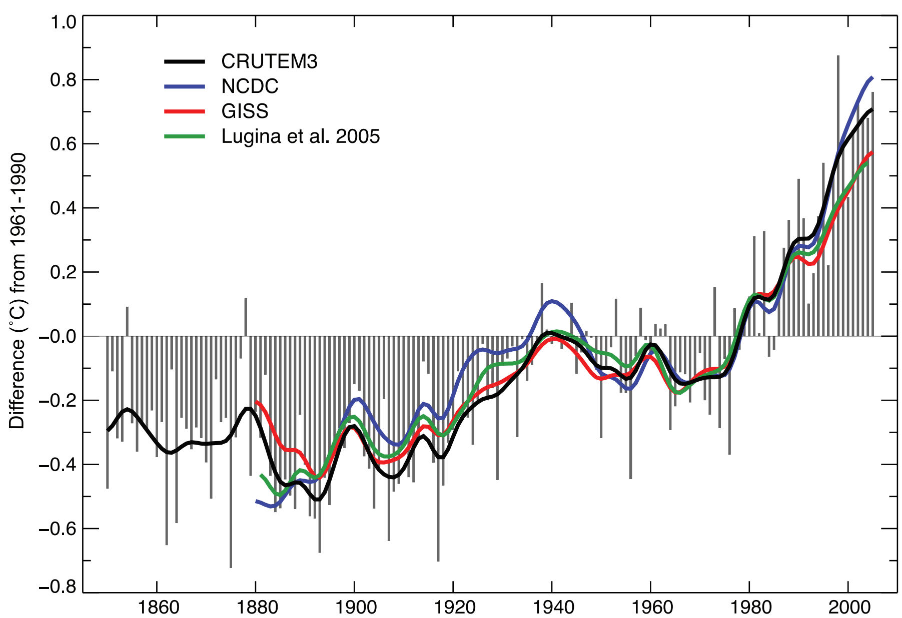

Figure 1.3 Annual anomalies of global land-surface air temperature (°C), 1850 to 2005, relative to the 1961 to 1990 mean for CRUTEM3 updated from Brohan et al., (2006). The smooth curves show decadal variations (see Appendix 3.A). The black curve from CRUTEM3 is compared with those from NCDC (Smith and Reynolds, 2005; blue), GISS (Hansen et al., 2001; red) and Lugina et al., (2005; green). From IPCC, (2007) Figure 3.1.

[1] Unless otherwise noted, the material in this section comes from IPCC chapter 3, Trenberth et al., 2007.

to record the day's weather conditions, most records of temperature or precipitation are affected by non-climatic influences: stations moved, instruments changed, the time of observation changed. Consequently it is a substantial scientific effort simply to identify and remove these non-climatic influences from each observational record and to combine point observations into area-averaged data, and to account for data gaps and other uncertainties.

Several research groups have undertaken these calculations and although the details differ slightly, all agree on several main features of global climate. The global land average temperature (Figure 1.3) fluctuated but showed little trend until about the 1920s, rose about 0.3–0.4°C (0.5-0.7°F) between the 1920s and 1940s, declined about 0.1–0.2°C (0.2-0.4°F) between the 1940s and the 1960s, and then rose 0.8–1.0°C (1.4-1.8°F) between the 1960s and 2005. The linear trend in global land average temperature is 0.8±0.2°C (1.4±0.4°F) over the 1901-2005 period according to the various reconstructions.

The climate system has changed in ways consistent with observed warming. The average temperature of the global oceans from the surface to a depth of 3 km (1.9 mi) has increased in the last 40 years, as the oceans have absorbed more than 80% of the additional heat energy added to the climate. Sea level rose 7.5 cm (3.0") between 1961 and 2003; nearly half of that rise was in the last ten years. Ice losses from the ice sheets of Greenland and Antarctica likely contributed to the accelerated rise of the last ten years. Since 1979, when reliable satellite-based global measurements began, the amount of water vapor in the atmosphere has increased in locations and quantities consistent with the extra water vapor that warmer air can hold. Because water vapor is a powerful greenhouse gas, and its abundance is primarily controlled by ocean surface temperatures, water vapor provides a substantial "positive feedback," accentuating other factors that influence surface temperature.

In summary, past changes in global climate have been substantial, but recent decades have been warmer than at any time in roughly 120,000 years.

1.1.2 Understanding Causes of Climate Change

Scientific assessments continue to underscore that scientific understanding of both the observed changes in global climate and their causes remains strong:

> Most of the [global] warming over the last several decades can be attributed to human activities that release carbon dioxide (CO_2) and other heat-trapping greenhouse gases (GHGs) into the atmosphere. The burning of fossil fuels—coal, oil, and natural gas—for energy is the single largest human driver of climate change (NRC 2010).

1.1.2.1 Drivers of climate change

To a close approximation, net energy input from the Sun - mostly in the form of visible light - is balanced by radiation of infrared (heat) energy back to space, averaged over the globe. "Drivers" or "forcings" of climate refer to natural or artificial processes that change some aspect of the climate by altering this energy balance. These forcings include the amount of solar radiation received at the top of the atmosphere (as affected by direct solar output and the Earth's orbital fluctuations), things that change the reflectivity (albedo) of the planet, such as particles in the atmosphere, and things that affect the efficiency of infrared energy loss to space.

These latter include both clouds and certain trace gases that absorb outgoing infrared energy and are commonly called greenhouse gases. In order of global importance, these greenhouse gases include water vapor, carbon dioxide (CO_2), methane (CH_4), ozone, chlorofluorocarbons (CFCs), of which CFC-12 dominates, nitrous oxide, and dozens of others. Human activities in the Industrial Era have substantially increased the quantity of all of these gases, and some (the CFCs among them) are entirely man-made. Carbon dioxide alone is responsible for about 63% of the total forcing by long-lived greenhouse gases (Forster et al., 2007).

White or grey particles in the atmosphere from smoke, smog, haze and volcanic ash scatter enough incoming sunlight that they offset a substantial fraction of the greenhouse gas warming. Particles also cause additional clouds to form, which adds to the cooling effect, as does the tendency for these clouds to be brighter than natural clouds. Both types of forcing, together, offset up to approximately one-half of the greenhouse gas warming, although confidence in the amount of this offset is low owing to difficulties in measuring the relevant quantities.

Changes in the sun's energy output and volcanic eruptions are the most important natural external forcings of climate. (Fluctuations in water vapor, surface albedo related to vegetation or snow cover, and other factors are considered internal responses of the climate system.) Solar changes may be partly responsible for the cool period in the 16th to 18th centuries (Figure 1.2) and for the warming early in the 20th century, but observations from satellites of solar output since late 1978 demonstrates that solar changes cannot be responsible for the large increase in global temperatures during the last 32 years: solar output has not increased over that period, but has fluctuated with the roughly 11-year solar cycle whose amplitude is about 0.1%. Since the solar cycle is presently just past a minimum, solar output is actually slightly lower than it was in 1978 (Lean and Rind, 2009).

A rare type of volcanic eruption — a very powerful tropical eruption — can cool the Earth for one to two years. Most volcanic eruptions briefly pollute the troposphere, the layer of the atmosphere with weather, up to about 10km (6 miles) above the surface in midlatitudes and 16 km (10 miles) in the tropics. Besides ash, which quickly falls out, volcanic emissions include sulfur dioxide, which (given enough time) turns into sulfuric acid particles that reflect sunlight. But the particles quickly attract water vapor, form clouds, and precipitate out. Some eruptions reach the stratosphere but in middle and high latitudes stratospheric air is gradually sinking and the volcanic emissions are pushed into the troposphere within a month or two. The most effective volcanic eruptions that cool the Earth are tropical volcanic eruptions of sufficient force to reach the stratosphere, in the latitudes where stratospheric air is rising and hence can suspend the reflective particles.

1.2.2 Understanding the factors in climate change

An easily understood approach to separating natural and human influences on climate change is to perform simulations of past and present climate with human influences included, and then with human influences excluded (Figure 1.4). By comparing observed temperatures and temperatures simulated with only natural influences (lower panel of Figure 1.4), marked differences emerge after about 1960. But simulations with human influences included match quite well the observed record, including drops in temperature associated with the major tropical volcanic eruptions indicated in the figure.

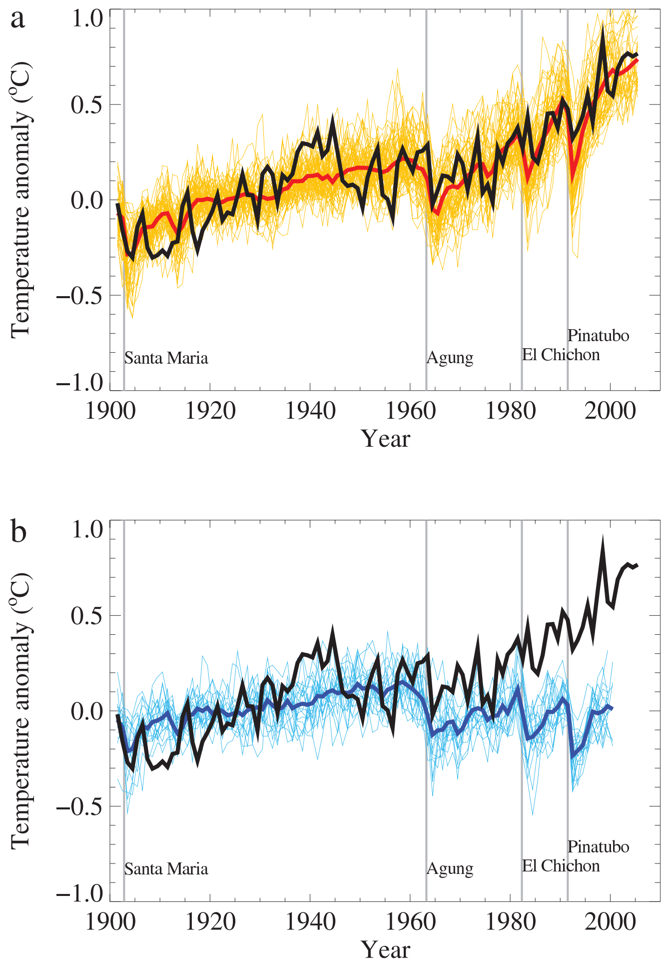

Figure 1.4 Comparison between global mean surface temperature anomalies (°C) from observations (black) and AOGCM simulations forced with (a) both anthropogenic and natural forcings and (b) natural forcings only. All data are shown as global mean temperature anomalies relative to the period 1901 to 1950, as observed and, in (a) as obtained from 58 simulations produced by 14 models with both anthropogenic and natural forcings. The multi-model ensemble mean is shown as a thick red curve and individual simulations are shown as thin yellow curves. Vertical grey lines indicate the timing of major volcanic events. The simulated global mean temperature anomalies in (b) are from 19 simulations produced by five models with natural forcings only. The multi-model ensemble mean is shown as a thick blue curve and individual simulations are shown as thin blue curves. From IPCC (Hegerl et al. 2007), Figure 9.5.

A more sophisticated approach to separating natural and human influences is to compare some aspect of the pattern of change in space and time with the pattern generated by climate model simulations. This pattern-matching or "fingerprinting" approach determines statistically whether the pattern could occur by chance or whether it is consistent with the forcing in question. Using this approach, a human influence on climate has been detected in global mean temperatures, in precipitation averaged in latitude bands, in atmospheric pressure, and in other fields (Hegerl et al., 2007).

Most of the research attributing recent climate change to specific causes has used global climate models (e.g., Figure 1.4). Lean and Rind (2009) instead use empirical data to diagnose four main factors influencing changes in global temperature in the last 30 years. Three are natural: the El Niño/Southern Oscillation, ENSO; cooling by volcanic particles; and energy output of the Sun.

The fourth factor is human influences, a steady and persistent rise. Lean and Rind's (2009) empirical approach explains the recent level period in global temperatures as the result of a competition between the waning phase of the solar cycle and the slow growth of human influences, and suggests that global temperature will resume its increase from 2010 to 2015, followed by another level period. Of course, a large tropical volcanic eruption or ENSO event could change the details.

In short, several lines of evidence including basic physics point to the rising concentration of greenhouse gases as the cause of substantial global warming since about 1950.

1.1.3 Future Global Climate Change

1.1.3.1 Tools and driving scenarios

One approach to understanding future global climate change would be to find a past analog: a period when conditions were similar to what is expected in, say, 2050. However, in the period of most detailed past climate information - the 700,000 years for which data from Antarctic ice cores exist - has no precedent for a climate with greenhouse gas forcing as high as it is today, let alone as high as it will be by 2050. Measurements of CO_2 in Antarctic ice cores never exceeded 300 parts per million, compared with almost 390 in 2010. Indirect measurements suggest that the last time CO_2 exceeded 300 ppm was at least several million years ago (Royer, 2006). Without a clear example from the past, we are left with two approaches to estimating global climate in a future with much higher levels of greenhouse gases:

(1) Using <u>observations</u>, estimate the climate sensitivity: the response of global mean temperature to a doubling of CO_2. Considering a wide range of studies, the sensitivity lies in the range from 2°C - 4.5°C (3.6°-8.1°F) with a best estimate of 3°C (5.4°F) for a doubling of CO_2, according to the IPCC (Hegerl et al. 2007). Given that CO_2 has already increased about 40% above pre-industrial concentrations, the likelihood of doubling during this century is fairly high, so the globally averaged increase in temperature during this century will almost certainly exceed 2°C, absent large increases in efforts to reduce global emissions.

(2) Using <u>physically based models</u> of the ocean, atmosphere, land, and ice, calculate the future climate. These global climate models (GCMs) have been developed by modeling groups in many nations. The Intergovernmental Panel on Climate Change (IPCC) coordinated a common set of simulations that used 21 models (Randall et al., 2007). One of many advantages of this approach is that these models estimate changes in climate in far more detail than only global mean temperature.

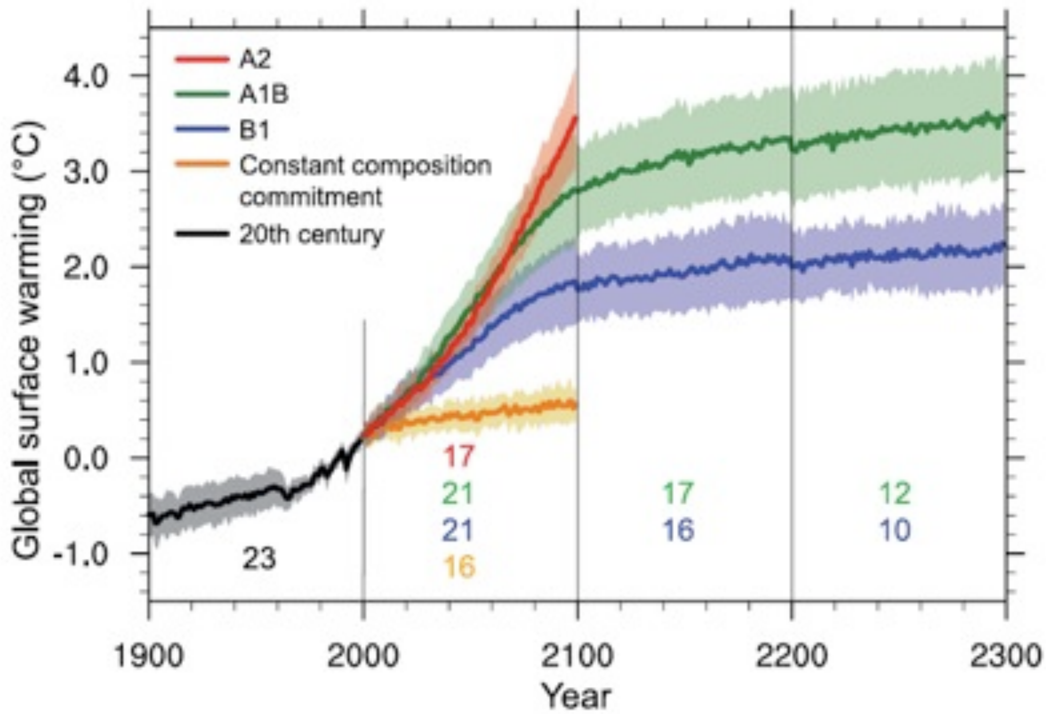

Figure 1.5. Simulated global mean temperature response to both the 20th century radiative forcing (1900-2000), to three scenarios of radiative forcing in the 21st century, and for three assumptions of constant CO_2: for the 21st century at year 2000 levels ("constant commitment"), and for the 22nd and 23rd centuries at the constant values reached in year 2100 under scenarios A1B and B1. Heavy curves show the multi-model means, shading denotes the ±1 standard deviation range of individual model annual means. Numbers in each century denote the number of contributing models. From Meehl et al., (2007).

Simulations of climate over the 21st century (Figure 1.5) require modelers to project the forcing of climate by radiation, especially the warming due to greenhouse gases - CO_2, methane, and a few others - and the cooling caused by atmospheric particles. In the 1990s, the IPCC produced more than 40 socio-economic scenarios (SRES; Nakićenović and Swart 2000) that generated concentrations of the leading greenhouse gases and reflective atmospheric particles. Of these, six scenarios were extensively discussed, and three of these were chosen by modeling groups for forcing the global climate models: the scenarios B1, A1B, and A2. All scenarios have similar climate forcing factors until about 2020 because CO_2 molecules last so long in the atmosphere (more than 50 years) that the CO_2 concentration in the whole atmosphere changes very slowly after a change in emissions. Of the chosen three, scenario A2 has the highest climate forcing by the year 2100, but before the middle of this century, none of the scenarios is consistently the highest. Another scenario, A1FI (not in the chosen three, and not used by most modeling groups) has even higher climate forcing than any of the chosen three by 2100. Their forecasts of CO_2 concentration for the year 2100 are 549, 717, 856 and over 1100 ppm in scenarios B1, A1B, A2, and A1FI, respectively (from 2 to 3.5 times the pre-Industrial value). Actual fossil fuel emissions of CO_2 since 2001 exceeded all but one of the six SRES scenarios (Myrhe et al., 2009) even though a few countries began to limit their emissions of greenhouse gases.

1.1.3.2 Projected global temperature changes

In the IPCC Fourth Assessment Report (AR4), Meehl et al., (2007) summarize projections of future climate change from the full suite of AR4 GCMs. Figure 1.5 shows the global mean surface temperature simulated by these GCMs. The figure shows global mean temperature rising about 0.7°C in the 20th century, as observed, along with dips in global mean temperature associated with major tropical volcanic eruptions. For the 21st century, note that models estimate an additional 0.2°C of "committed" warming even with constant CO_2. This underscores the point that the climate system — especially the ocean — is still catching up to the forcing already in place, and that climate change will continue (although at a considerably reduced rate) even after CO_2 is stabilized in the atmosphere.

1.1.3.3 Changes in other aspects of global climate

We summarize here some key aspects of global climate described by Meehl et al., (2007). Globally averaged precipitation increases slightly in 21st century simulations, around 1.4% per degree C (0.8% per degree F), owing to the enhanced water-holding capacity of warmer air. The global hydrologic cycle speeds up, with precipitation generally increasing in areas with above-average precipitation (tropics and mid to high latitudes) and decreasing in the subtropics, around 20-30° latitude in both hemispheres. Models produce some consistent changes in cloud patterns, with most models showing reductions over low to middle latitude land areas.

Changes in atmospheric circulation in the models include a stronger Hadley circulation (rising in the tropics, sinking in the subtropics, and return flow as trade winds), which is linked to the pattern of global precipitation change; and a slight poleward shift and intensification of the storm tracks (e.g., Yin, 2005). The latter change can be linked to global changes in the troposphere including the equator-to-pole temperature gradient in the middle-troposphere.

1.2. Climate Change in Oregon and the Northwest

1.2.1 Pacific Northwest Climate in the Past

Since the last glacial maximum 21,000 years ago, Oregon's climate has fluctuated greatly over a wide range of timescales. Transitioning from the glacial climate regime to the modern climate regime took thousands of years, and there are fluctuations on timescales of centuries, decades, and year to year, as well as abrupt changes in climate averages and variability.

Paleoclimatologists use a variety of methods to reconstruct these fluctuations and to attribute the fluctuations to factors that are known to affect the climate. In Oregon, there is great potential to study climate history over thousands of years because alpine ice fields (which locally eliminate the geologic sediment records) covered only a small portion of the state during the glacial maximum. In addition, many tree species reach great ages, allowing the study of past climate from annual growth rates.

The three major controls of Earth's climate - incoming solar radiation (insolation), ice extent, and greenhouse gases - have changed dramatically from the glacial maximum to the present. Slow progressive changes over time in three features of Earth's orbit (orbital shape, tilt of its axis, and seasonal precession of the perihelion, or closest earth-sun distance), combine to affect the latitudinal distribution and seasonal cycle of insolation. During the last glacial maximum, extensive ice sheets, which reflect short-wave solar radiation back to space, affected both the global balance of incoming and outgoing radiation as well as modifying the locations of high and low pressure systems, the location of jet streams, and routes of moisture to continental interiors. Greenhouse gases, including carbon dioxide and methane, were lower during the glacial maximum, thus increasing the amount of outgoing longwave radiation escaping from Earth's surface and atmosphere to space.

Insights into the climatic changes of the past come from both climate models and from a myriad of geological records. The discussion below summarizes the millennial-scale patterns for which there is strong agreement between the climate simulations and the data, as well as highlights recent findings of shorter-term and abrupt changes in climate.

1.2.1.1 The Last Glacial Maximum: 21,000 years ago

During the height of the last glaciation, the seasonal pattern of insolation was similar to the present day; extensive ice sheets were located to the north and atmospheric CO_2 was as low as 180 ppm (65% of preindustrial levels). The high albedo and high elevation of the ice sheets produced very cold air, resulting in a large high pressure system and associated anticyclonic (clockwise) winds, which were especially intense during winter. The anticyclonic winds spun off the ice sheets producing an east-to-west wind in Oregon, deflecting to the south an onshore flow of moist air. Lower greenhouse gas concentrations also affected the global energy balance.

Oregon's climate was significantly colder and drier than present. The latest paleoclimate reconstructions and models indicate that mean annual temperature was as much as 10°C (18°F) colder in eastern Oregon and about 5°C (9°F) colder in western Oregon. This is consistent with the occurrence of winds from the east during the glacial maximum affecting eastern Oregon more than coastal Oregon. Glaciers along the High Cascades were extensive and merged into continuous ice fields. Cooler temperatures resulted in much less evaporation than today, increasing depths of lakes in the closed basins of central and southeastern Oregon (Figure 1.6). The depths of these basin lakes fluctuated throughout the glacial periods (peaking before the glacial maximum), likely reflecting fluctuations in the strength of the glacial anticyclone and the degree to which the jet stream and moisture were deflected to the south. Similarly, the emplacement of dunes along the Oregon coast (which extended farther west with lower sea levels) occurred during periods before the glacial maximum when the glacial anticyclone was weak, allowing strong winds from the west to move sand inland.

While few glacial maximum pollen records have been collected for Oregon, those available suggest that the Coast Range mountains supported a park-like landscape of trees and meadow, somewhat similar to the colder and drier forests at higher elevation in the eastern Cascades and Rocky Mountains today, while eastern Oregon was marked by many more drought-adapted shrubs near the forest/shrubland border.

1.2.1.2 The Late Glacial: 21,000 to 11,600 years ago.

A period of deglaciation between 21,000 and 11,600 years ago, driven by changes in insolation, was marked by decreasing ice sheet extent, increasing sea level, and increasing atmospheric CO_2, resulting in a complex series of climate changes differing greatly across North America. As the ice sheets retreated north into what is now Canada, the glacial anticyclone weakened

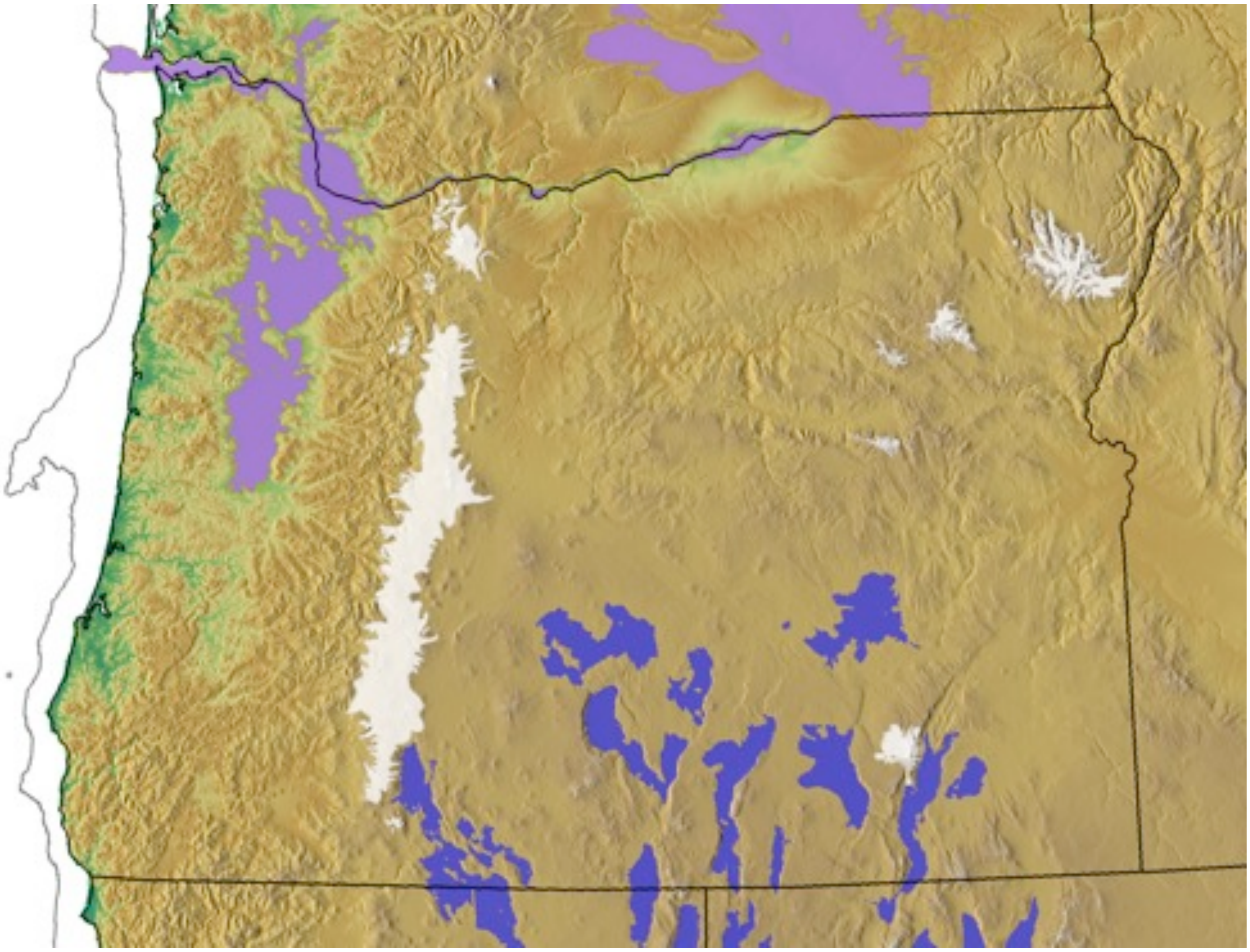

Figure 1.6 The glacial periods left distinctive fingerprints across Oregon. During the Last Glacial Maximum, sea level was approximately 120 m (400 ft) lower than today and the Pacific Coast was 20 - 60 km (13 -38 mi) further to the west (shown as gray line). The "pluvial" lakes of central and southeast Oregon (dark blue) were 70 m (230 ft) deeper than at present for most of the late Pleistocene (and deeper during the early Pleistocene). At roughly 17,000 years ago, glaciers (white) descending from the mountains reached their maximum extent. These glaciers were restricted to the Cascades, Steens, Aldrich, Greenhorn, Strawberry, and Wallowa Mountains. From 20,000 to 15,000 years ago, dozens of Missoula Flood events (purple), originating from glacial lake Missoula where the Clark Fork River was ice-dammed in northern Idaho, swept across eastern Washington and backed into the Willamette Valley, forming glacial Lake Allison to a surface elevation of 120 m (400 feet) above modern sea level. Glacier extent used in this map was compiled from the Atlas of Oregon (Loy et al., 2001) and various other data sources.

resulting in increased onshore wind flow. A large increase in moisture occurred 17,000 years ago causing advances of alpine glaciers throughout the Pacific Northwest, marking the time when alpine glaciers descended to elevations as low as 1000 m (1600 ft) in the western Cascades, 2100 m (3400 ft) in the Wallowas, and 3250 m (5200 ft) on Steens Mountain. At the same time, glacial Lake Missoula breached its ice dam dozens of times resulting in cataclysmic floods down the Columbia River, which backed into the Willamette Valley. These floods recurred as the Cordilleran ice sheet re-advanced to form new ice dams on the Clark Fork River, only to burst and send forth another massive flood. The last Missoula Flood occurred around 15,000 years ago and the alpine glaciers retreated around 14,000 years ago leaving vestiges on a few of the highest peaks. Nearly all lake basins in mid to high elevations of Oregon became ice-free at that time. The few lake-sediment records from unglaciated areas show an abrupt warming 14,000 years ago with the replacement of subalpine habitats by more productive forests of Douglas-fir and alder.

The Younger Dryas period between 12,900 and 11,600 years ago, has been identified to various degrees in the Pacific Northwest. The inconsistency of response likely reflects a lower magnitude of the event than occurred in the Atlantic, and the wide climatic tolerance of the recently-established vegetation at the various study sites. For example, some records show that the Younger Dryas-like climate event in the Northwest slightly lagged events in the North Atlantic (Mathewes et al. 1993). Nevertheless, some well-dated records from speleothems (stalagmites) in Oregon Caves, as well as lake sediments, are broadly synchronous with the Younger Dryas, but the magnitude of the cooling event in the Pacific Northwest is yet unclear (Vacco et al. 2005).

1.2.1.3 The Holocene: 11,600 years ago to present

Around 11,600 years before present, ice sheets retreated rapidly across North America, insolation was at its peak in seasonality (8% more insolation in the summer and 8% less in the winter compared to today), and CO_2 was at levels typical of the preindustrial period (280 parts per million, vs about 390 ppm in 2010 and 180 ppm during the glacial maxima). Greenland ice cores indicate that the beginning of the Holocene period was marked by an abrupt increase in temperature in a period of less than 5 years.

In the Pacific Northwest, this abrupt warming was observed at several sites in Washington and British Columbia. That the record of this warming is less than clear in many sediment records in Oregon may reflect the facts that (a) warm-adapted vegetation was well established before the Younger Dryas period and (b) many species (e.g., Douglas-fir) tolerate a broad range of climate and their abundances on the landscape hence do not closely follow the climatic changes through the Younger Dryas period.

The early Holocene was marked throughout the Pacific Northwest by hotter summers and increased droughts and forest fires. The increased summer insolation during the Holocene may have led to an intensified Pacific Subtropical High pressure system, which created warm, stable, dry air to its east (i.e., the Pacific Northwest). In western Oregon and Washington, increased summer warmth and wildfires led to widespread Douglas-fir and alder forests, species that are adapted to fire by reproducing rapidly in burned areas (Sea and Whitlock 1995). Prehistoric insect remains recovered from lake sediments in southern British Columbia suggest early Holocene summer temperatures at 3°C (5.4°F) warmer than present, in agreement with temperature reconstructions based on early Holocene pollen data from southeast Oregon (Walker and Pellatt 2003; Minckley et al. 2007). There is mounting evidence, however, that the early Holocene was not uniformly warm and dry, but was marked by distinct century-scale periods of increased moisture (Heine 1998). The remnants of the waning ice sheet to the north may have still been influencing the jet stream across western North America.

An event during the early Holocene deserving special mention is the eruption of Mount Mazama that created Crater Lake 7,600 years ago. Ash deposits from this eruption blanketed the Northwest, especially east of the Cascade crest. This ash layer created seed beds for today's forests, and established the patterns of future forest growth. In some areas, extremely thick ash deposits led to "tephra plains" that remain very dry and sparsely vegetated today. In other areas, ash deposits thickened soils, weathered into clays, and may have allowed for greater water retention. In still other areas, forests were successfully established on steep rocky slopes only when tephra deposits were laid down (Gavin et al., 2001).

As summer insolation decreased through the middle Holocene, cooler and moister summers resulted in lower fire occurrence and the establishment of the dense, deep-shade tolerant vegetation currently typical in western Oregon and Washington. In Washington, 6000 years before present marks the onset of forests resembling today's old growth Douglas-fir forests (Brubaker 1991). In Oregon, evidence of this transition is less distinct but suggests a progressive increase in moisture until modern forests became established around 4000 to 3000 years ago (Whitlock 1992).

This latter part of the Holocene is termed the "neoglacial" because many alpine glaciers began advancement downslope about 4000 years ago. Between 5000 to 4000 years ago, dunes became more established along the Oregon coast indicative of intensified onshore winds. Later in the neoglacial, many glacial advances were synchronous across the West, including events at 3300 and 2400 years ago. Of all the glacial advances during the Holocene, almost without exception the largest was a series of Little Ice Age glacier advances from 1350 to 1850 AD.

1.2.1.4 The instrumental period: 1850 to the present

For reasons noted in Section 1.1.4, instrumental records of temperature must be carefully treated to remove non-climatic influences. By about 1920, enough stations in the US Historical Climate Network (Karl et al., 1990) were in place to analyze regionally averaged changes in temperature and precipitation (Mote, 2003) and over the 1920 - 2000 period, they indicated a warming for the Pacific Northwest of 0.8°C (1.5°F)/century, and almost every one of the individual trends is positive (Figure 1.7). Two examples are shown, both with trends around 2.0°F/century and with periods of record 1903 - 2006. However, throughout the instrumental record, regionally averaged precipitation has fluctuated substantially.

Understanding the causes of these trends and fluctuations remains an active area of research. The fluctuations in annual mean temperature and precipitation are partly related to atmospheric variability over the Pacific Ocean (Section 2.2). Mote (2003) estimated that Pacific variability could explain about 10% of the temperature trend over the 1920 - 1995 period. Formal detection and attribution studies like those described in Section 1.2.2 have not been performed for regions as small as the Pacific Northwest, but the analysis of Bonfils et al., (2008) finds a human influence on temperature of the mountainous West.

Other aspects of climate, though perhaps more relevant for society, have received less attention from researchers than warming has. Trends in extreme precipitation are ambiguous. Groisman et al. (2004) examine regionally averaged trends in number of days greater than the 99th and 99.7th percentile of daily precipitation; for the Pacific Northwest, over the 1908 - 2000 period, trends are not statistically significant in any season. Kunkel et al. (2003) examine precipitation extremes averaged over the continental US for a range of definitions (1- , 5-, and 20-year return period; and 1- 5-, 10- and 30-day precipitation totals), and note that all the time series had a similar shape with high values during the late 19th and early 20th centuries, lower values from the 1920s to 1970s, and then increasing; for most definitions of extremes, the recent maximum was larger than the earlier maximum, but combined with the results of Groisman et al. (2004) it is clear that the recent increase in extremes happened mainly in the eastern third of the country, not in the West. Madsen and Figdor (2007) examine station trends in the Northwest and find a statistically significant *decrease* in extreme precipitation in Oregon over the 1948 - 2006 period. Rosenberg et al., (2010) construct regionally averaged probability distributions from hourly station data, normalized by each station's long-term mean, for 1956-30 and 1981-2005 in Washington State and the Portland, OR, area. For the Portland area stations, the extreme 1-hour precipitation increased across the probability distribution, whereas extreme 24-hour storms decreased slightly for the 99th percentile and increased substantially at all higher percentiles.

1.2.2 Patterns of Climate Variability Influencing the Northwest

Variations of climate include variations across the landscape — spatial patterns — and variations in time — temporal patterns. Spatial patterns of climate in Oregon and the entire Pacific Northwest are strongly influenced by the north-south mountain ranges, chiefly the Cascades but also the coast range and Blue-Wallowa mountains of northeast Oregon. The effects of mountains on precipitation are clear in Figure 1.8, constructed with the PRISM (Parameter-elevation Regressions on Independent Slopes Model) approach to geospatial mapping (Daly et al., 2004) using observations and statistical relationships between terrain and precipitation. The western slopes of the Coast Range and the Cascades are very wet, with many places estimated to receive over 250 cm (100 inches) of precipitation per year. Gradients in precipitation can be quite sharp, with differences of a factor of 10 in less than 32 km (20 miles) near Bend (labeled in Figure 1.8) for example.

Temporal patterns of climate variability in the Northwest are strongly influeced by variations over the Pacific Ocean, chiefly El Niño/Southern Oscillation (ENSO). ENSO involves linked variations in the tropical Pacific Ocean and overlying atmosphere. Most of the time, the warmest water lies north of Australia and the presence of the warm water draws warm moist air, which forms thunderstorms. Hence, the warmest water coincides with heavy precipitation. The air rising in thunderstorms is part of an equator-to-subtropics circulation called the Hadley Circulation, which is part of the global energy cycle and affects atmospheric circulation throughout the globe.

Before an El Niño event, something happens to disrupt the normal distribution of sea surface temperature, winds, and precipitation. Both the warm water and the heavy precipitation move eastward, with warm water anomalies appearing along the equator as far as the South American coast. (In fact, the name El Niño, for "the [Christ] child" was given centuries ago by fishermen who noticed the periodic disruption of the productive fisheries by warm water near Christmas). A typical El Niño event begins during northern hemisphere summer or fall, peaks around late December with warm water anomalies of 1°C or more along the equator, and then fades during northern hemisphere spring, often followed by an accentuated return to normal conditions, called La Niña as an antonym of El Niño. On average, El Niño events occur once per four years, but they have occurred in successive years.

During the El Niño phase of ENSO, the wintertime jet stream tends to split, with warmer air flowing into the Northwest and Alaska, and a southern branch of the jet stream directing unusually frequent and heavy storms toward southern California. During El Niño winter and spring, Oregon's climate is slightly more likely than usual to be warm and dry. The effect is more pronounced farther north into British Columbia.

One manifestation of ENSO in the North Pacific has been termed the Pacific Decadal Oscillation (PDO), so named because in 20th century records, variations in north Pacific sea surface temperature (SST) patterns appear to have phases lasting 20 - 30 years (Mantua et al., 1997). However, paleo reconstructions of the PDO using tree rings (e.g., Gedalof et al., 2002) indicate a similar behavior of the PDO from the mid-18th to early 19th century, then very different behavior in the succeeding 100 years. Also, after 1998 the PDO index has shown no evidence of decadal persistence. In addition, Newman et al., (2003) show that the best statistical model of the PDO treats it not as a distinct pattern of variation independent of ENSO, but simply a slow North Pacific response to ENSO forcing. Furthermore, linear trends over periods of a few decades can be affected by the phases of ENSO and PDO.

1.2.3 Future climate change in the Pacific Northwest

1.2.3.1 Model evaluations

The global climate models used in the IPCC (2007) assessment report were examined for the Pacific Northwest[2] by Mote and Salathé (2010). They compare observed temperature and precipitation with the simulated regional temperature and precipitation in the 20th century, including the annual averages, the seasonal cycle, and the trends. They also compare the temperature, precipitation, and sea level pressure patterns over a much larger region including most of the North Pacific Ocean and North America. See Randall et al. (2007) for a list of the climate model references, attributes, and abbreviations.

The mean temperature produced by the set of all models is about 1.8°C (3.2°F) cooler than observed, while the seasonal cycle of temperature was close to what was observed (within 1°C, 1.8°F in one observational dataset). All models produce the observed contrast between wet winters and dry summers. However a few produce summers only slightly drier than the winters, and for every model, the annual precipitation is considerably higher than observed. Mote and Salathé do not diagnose a reason for this wet bias of the models. Comparing each model's annual cycle with observations and calculating root-mean-square difference to rank the models, the "best" five models for temperature are, with one exception (GISS_ER), different from the best five models for precipitation.

Mote and Salathé also evaluate the models' linear trend in temperature over the 20th century (see Figure 1.9). On regional scales, temperature trends are influenced more by atmospheric circulation than by greenhouse gas forcing; still, eight of the models simulate a warming in the Northwest for the period 1900–2000 within 0.2°C (0.4°F) of the observed warming of +0.8°C (1.4°F) during that period. In both observations and models, precipitation fluctuates much more than temperature; indeed, there is little evidence that observed precipitation (globally or at these latitudes) responded to greenhouse gas forcing in the 20th century (Zhang et al., 2007).

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[2] The region of analysis is a rectangle bounded by latitudes from 41.5° to 49.5° North, and longitudes from 124° to 111° West.

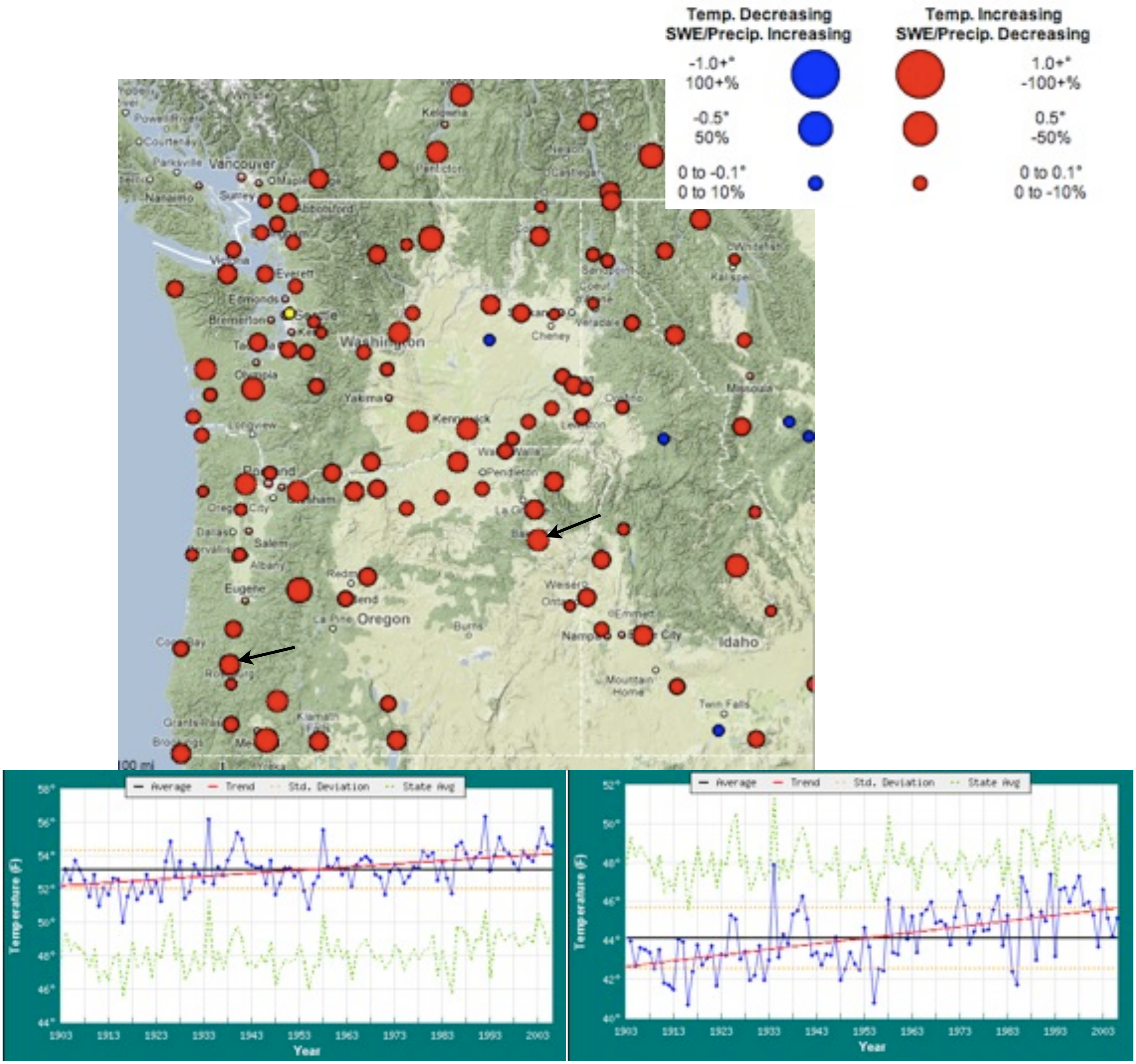

Figure 1.7 (top) Linear trends in annual mean temperature at US Historical Climate Network stations in the Northwest. Red means warming, blue cooling, yellow no trend; size proportional to warming trend (°F/decade). **(bottom)** Annual mean temperatures at two stations in Oregon indicated by arrows on the map: Drain (left) and Baker City (right) overlaid with the linear trends (red), the mean (solid black) and standard deviation (dotted orange), and statewide mean (dotted green). Figures created using a utility at the Office of Washington State Climatologist, climate.washington.edu.

1.2.3.2 Projected changes in annual temperature and precipitation

Mote and Salathé (2010) combine the model results for future periods using "reliability averaging of ensembles," a technique which gives more weight to models that perform better and that yield results closer to the average of all the models. We refer to these results as

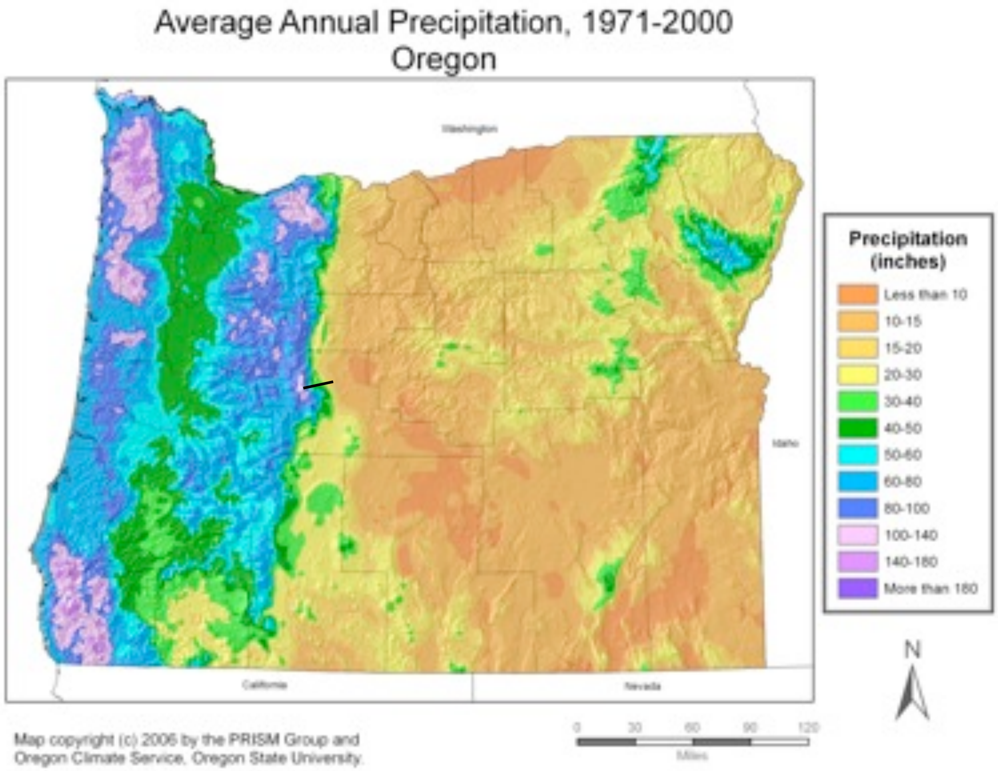

Figure 1.8 Map of annually averaged precipitation formed using geospatial mapping techniques and monthly mean station observations (Daly et al. 1994, 2007). Black line near Bend highlights one region of very sharp gradient in precipitation.

weighted multi-model means.

Figure 1.10 shows the projected model-average temperature and precipitation for the Pacific Northwest, for all simulations in the B1 and A1B scenarios, from 1900 to 2100. By the 2080s, the models suggest a change in mean temperature of 3.4°C (6.1°F) for the A1B, and 2.5°C (4.5°F) for the B1. Considering the full range produced by all the models across the two scenarios, the range of warming is 1.5–5.8°C (2.7–10.4°F); other IPCC emission scenarios would produce more warming by 2100, but B1 produces the least warming.

The observed trend in regional mean temperature is statistically significant for the 20th century, because the change exceeds what would be expected from a time series with no trend but having the same amount of year-to-year natural variability (Mote, 2003). Likewise, the projected future trends are substantially greater than the trend observed in the 20th century, even for the scenarios having the very lowest temperature changes.

For changes in precipitation, model results do not show very clear trends (bottom panel of Figure 1.10). The vast majority of climate models project increases in average annual precipitation in the northern third of North America, but decreases in the southern third (Christensen et al., 2007); the Pacific Northwest lies in the ambiguous area in between, where some models project increases and others project decreases in precipitation. The simulated multi-model average annual precipitation for the Pacific Northwest is practically unchanged through the 20th and 21st century, although individual models project changes of as much as 10% below or 20% above late 20th century mean precipitation by the 2080s.

1.2.3.3 Projected changes in seasonal temperatures and precipitation

Seasonal changes of climate often have greater impacts than changes in annual average climate. Figure 1.11 depicts the projected change in temperature and precipitation for each season, for three future decades, the 2020s, 2040s, and 2080s in comparison to the 1980s. For both scenarios B1 and A1B, regional warming is projected to be largest in summer. In most seasons B1 has the lowest projected change and A1B the highest, but this will not always be true in the 2020s when the climate forcing of the two scenarios becomes very similar. The most consistent changes in precipitation appear in the summertime, with a large majority of models (68–90% of them) projecting decreases; the multi-model average value reaching –14% by the 2080s. Some models forecast reductions of as much as 20 to 40% in summer precipitation, though these large percentages really only translate to 3 to 6 cm (1.2 - 2.4″) of water depth over the summer season. This is 3 to 6% of the all-model average value for the annual average in the 20th century. While this is a small amount in hydrological terms, summer precipitation and its associated cloudiness nonetheless strongly impacts evaporation, and hence urban water use (Palmer and Hahn, 2002) and forest fires (McKenzie et al., 2004) in the Pacific Northwest.

In contrast to summer, a majority of models project increases in precipitation in the winter. The multi-model-mean annual mean precipitation increases +8% (about 3 cm, 1.2″ of water) by the 2080s in the A1B scenario, which is still small in comparison to the year-to-year variability. And although some of the models suggest modest reductions in fall or winter precipitation, others suggest very large increases (up to 42%). Changes of this magnitude would substantially alter regional hydrology and ecosystems.

For other important aspects of climate, less research has been done about past and likely future changes. Rosenberg et al. (2010) examine the changes in extreme precipitation in daily outputs of two global models and one regional model; for Portland, small (2 - 10%) and generally not significant increases are found for most definitions of extreme return period and duration.

1.3. Past and Future Changes in Properties of the Coastal Ocean

Oregon's marine environment (Figure 1.12) is influenced by the open North Pacific Ocean, and by the atmosphere especially over the continental shelf and slope. Oregon's coastal ocean has strong spatial variations, vertically and in offshore and alongshore directions. Long-term observations of currents and water properties are scarce, but after fifty years of oceanographic research at Oregon State University, it is clear that Oregon's coastal ocean is highly variable over time, responding to winds and tides, to heating and cooling, and to rainfall and river discharge. The short-term, seasonal and interannual variability all tend to be greatest near shore, over the continental shelf and slope. In spite of this variable background, and the paucity of historical time series, there is growing evidence of substantial changes in Oregon's marine environment, some of which can be linked to global climate change.

New monitoring programs already in place will provide more abundant time series, but these and other new programs will need to be continued for many years to monitor future change. High-quality long-term observations are required for testing and evaluating models that make predictions; forecasts unconstrained by observations have much less value. Oregon's

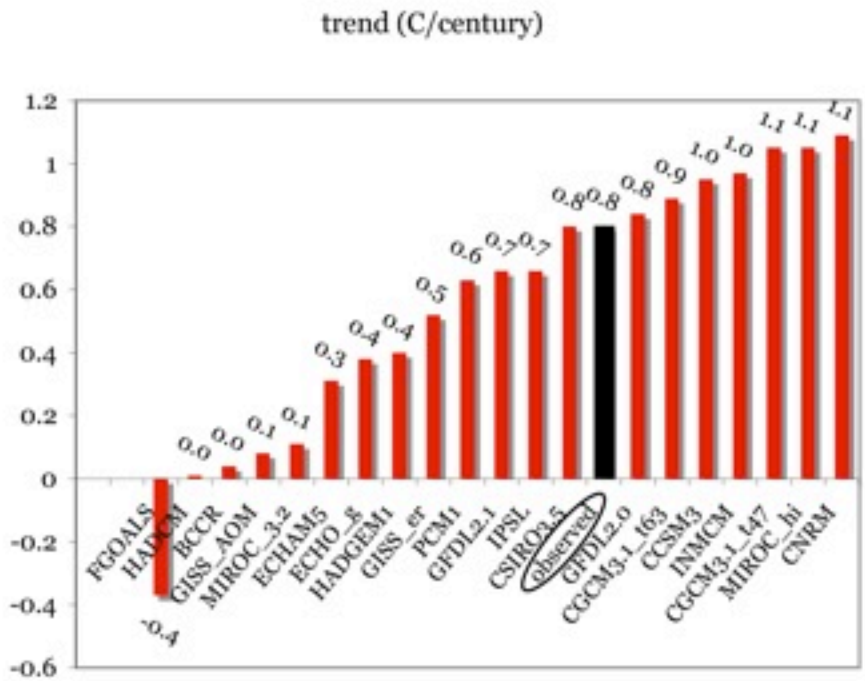

Figure 1.9 Trend in each model's annual mean temperature for the Pacific NorthWest over the 20th century, and the observed trend (Mote 2003 updated). The observed trend is close to the median trend of the models.

monitoring of coastal waters needs to be continued to provide baseline and to monitor future changes.

1.3.1 Observed (past) changes in Oregon's marine environment

In order to provide some context for understanding how the marine environment could change in the future, we describe the types of observations available and the characterization that they give of variability and change in Oregon's coastal waters over the past 10 - 50 years. We also describe the physical processes that contribute to the seasonal and interannual variability.

1.3.1.1 Local Wind Forcing

Much of the variability over Oregon's continental shelf is very closely related to local winds, which exert drag on the ocean surface layer. Combined with Earth's rotation, this drag can push the surface layer either toward or away from shore, as well as along the coast in either direction; the amount of water transported (the "Ekman transport") is directly proportional to the wind stress. The sea surface stays nearly horizontal: when surface waters move offshore, they are replaced by water from deeper levels, in a process called upwelling; when surface waters move onshore, they are pushed downward ("downwelling"). The vertical velocity is too weak to be measured directly, but theory shows that the vertical transport of ocean water can be estimated from measurements of wind speed and direction. If the wind is uniform, upwelling or downwelling occurs in a narrow coastal strip whose width is 20 km (12 mi) or less. Deeper waters are colder, saltier and denser than surface waters; they are also richer in nutrients and poorer in dissolved oxygen.

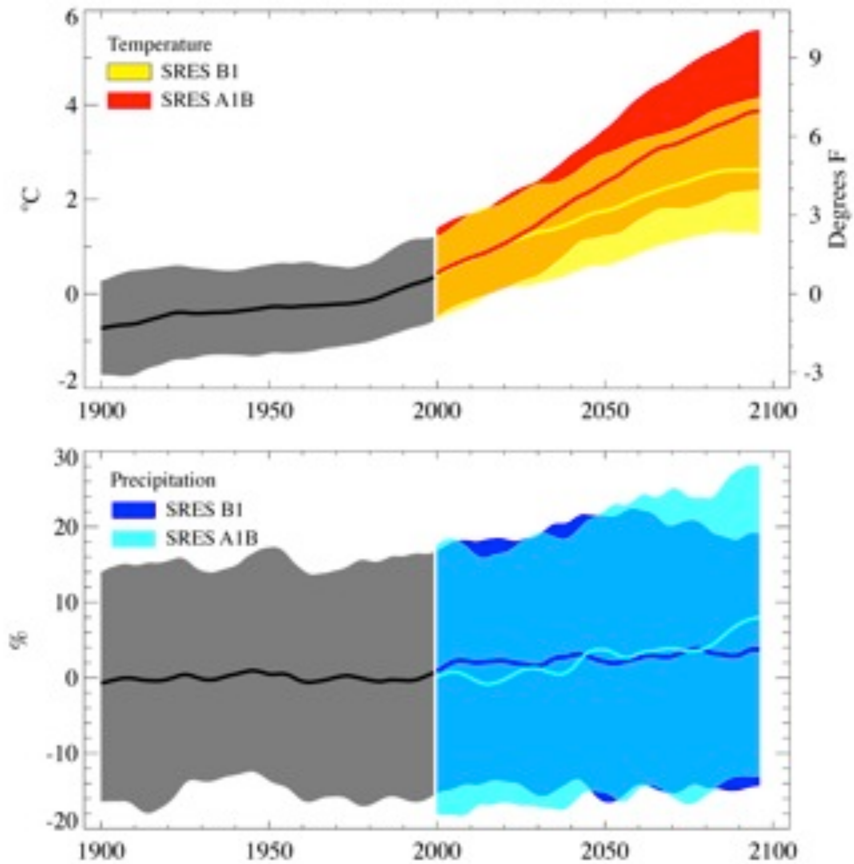

Figure 1.10 Smoothed traces in temperature (top) and precipitation (bottom) for the 20th and 21st century model simulations for the PNW, relative to the 1970 - 99 mean. The heavy smooth curve for each scenario is the weighted multi-model mean value, calculated for each year and then smoothed. The top and bottom bounds of the shaded area are the 5th and 95th percentiles of annual values (in a running 10-year window) from the ~20 simulations, smoothed in the same manner as the mean value. Mean warming rates for the 21st century differ substantially between the two SRES scenarios after 2020, whereas for precipitation the range is much wider than the trend and there is little difference between scenarios. From Mote and Salathé (2010).

When winds blow southward along the Oregon coast, as they generally do in summer, they cause warm, fresh (low-salinity), nutrient-depleted, but oxygen-rich surface waters to move offshore, and they bring cold, salty, dense, nutrient-rich, oxygen-poor waters up to the surface inshore. The high-gradient region between these very different offshore and inshore waters is called the "upwelling front." After sustained summer upwelling, waters near the surface over the inner shelf have properties similar to those observed at depths of 200 – 250 m (660 - 820 ft) offshore; the resulting density gradients cause a southward current whose speed is greatest at the surface and decreases with depth. When the local wind blows northward along the coast, as it generally does in winter, surface waters move toward the coast, and warm, fresh, nutrient-poor, offshore waters flood the nearshore region and move downward over the inner shelf. After sustained winter downwelling, the coastal current is northward; its speed tends to decrease with depth. In both seasons, alongshore currents are much stronger than the onshore/offshore currents of the upwelling/downwelling circulation.

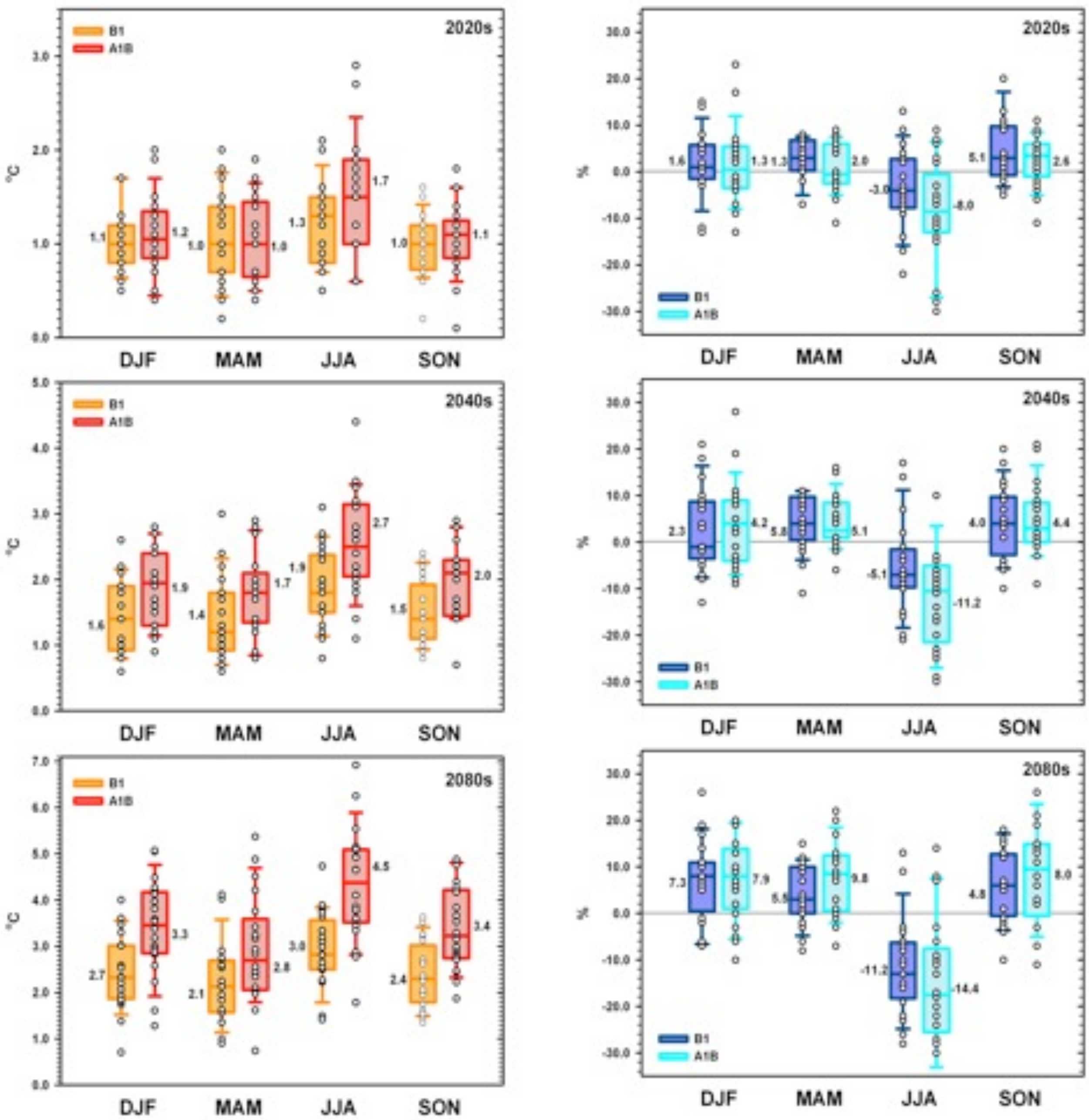

Figure 1.11 Range (lowest to highest) of projected changes in temperature (left) and precipitation (right) for each season (DJF=winter, etc.), relative to the 1970 - 99 mean. In each pair of box- and-whiskers, the left one is for SRES scenario B1 and the right is A1B; circles are individual model values. Box-and-whiskers plots indicate 10th and 90th percentiles (whiskers), 25th and 75th percentiles (box ends), and median (solid middle bar) for each season and scenario. Printed values are the weighted multi-model mean of all GCMs for the season and scenario. From Mote and Salathé [2010].

The transition from the downwelling regime to the upwelling regime typically occurs in early spring. Regardless of season, current fluctuations with periods of days-to-weeks are highly correlated with alongshore wind stress, as is the position of the summer upwelling front. Some interannual variability in shelf currents and temperature can also be explained by local longshore wind stress: for example, unusually strong northward winds in the El Niño winters of 1983 and 1998 enhanced the northward coastal current (Kosro, 2002); unusually late arrival of northerly winds in the spring of 2005 delayed the onset of upwelling by more than a month (Pierce et al., 2006).

Because there are no multi-decade observations of currents or water temperatures from the Oregon shelf, we use three estimates of the alongshore wind stress as valuable indicators of the intensity of upwelling. First, we use winds measured at Buoy 46050, about 37 km (20 nautical miles) west of Newport; these hourly measurements began in 1985, and data are available from the National Oceanic and Atmospheric Administration (NOAA). This time series is not yet long enough to determine a long-term trend, but data indicate that the average intensity of upwelling in each year from 2005 to 2008 was stronger than the 20-year average of 1985 – 2005.

Second, we use wind stress data from the National Center for Environmental Prediction (NCEP, Kalnay et al., 1996), available since 1948 at a spatial resolution of 2 degrees. Though they do not actually resolve spatial scales smaller than 1000 km (625 mi; Milliff et al. 2004), their time series are well correlated with winds measured at buoys in the North East Pacific (Ladd and Bond, 2002). Daily wind stress values were used to determine the dates of onset and cessation of seasonal upwelling, and to calculate the average and variance of the alongshore wind stress during each upwelling season. The seasonal average has no significant trend, but the variance has increased significantly over the last 50 years (Figure 1.13), by about 35% at 45°N and by about 50% at 41°N.

Third, monthly values of the Coastal Upwelling Index at 45°N, 125°W and at 42°N, 125°W provided by the Pacific Fisheries Environmental Laboratory (http://www.pfeg.noaa.gov).

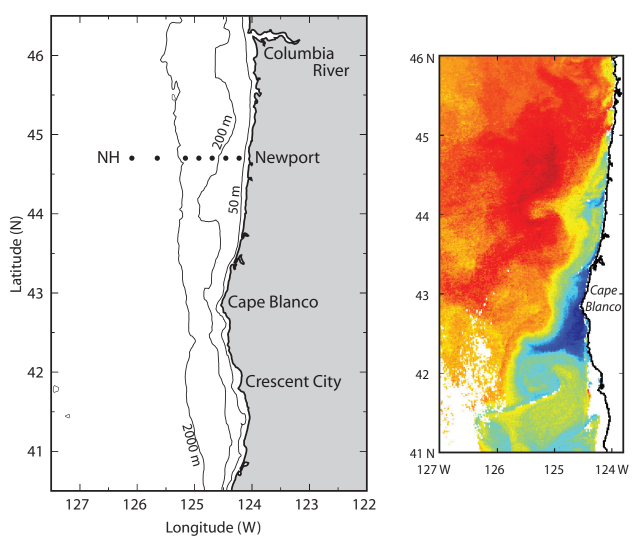

Figure 1.12. Maps of the coastal ocean off Oregon: (a) left, stations of the Newport Hydrographic Line (dots), and a few bottom contours: 2000 m (6600) ft at the foot of the continental margin, 200 m (660 ft) at the shelf edge, and 50 m (160 ft) over the inner shelf; (b) right, satellite image of sea surface temperature for 5 July 1999 (from Huyer et al., 2005; white areas represent clouds or fog, red indicates warmer waters and blue cooler).

These values are derived from the Fleet Numerical Meteorology and Oceanography Center monthly average pressure fields using a three-degree mesh. Averaging the June, July, August and September (JJAS) values together yields an annual estimate of the intensity of upwelling. The average JJAS index at this location increased over the past 50 years, particularly off southern Oregon (Figure 1.14), though much of the trend is due to a recent decade of strong winds (1995 - 2005). The slow variations in the 11-year running average do not seem to be correlated with the Pacific Decadal Oscillation (Mantua et al., 1997; but see Section 2.2).

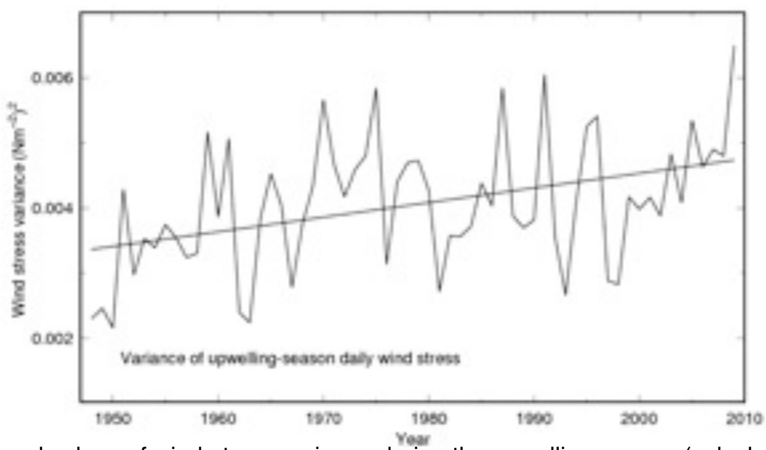

Figure 1.13. Annual values of wind-stress variance during the upwelling season (calculated from daily NCEP/NCAR reanalysis wind-stress data see text for details).

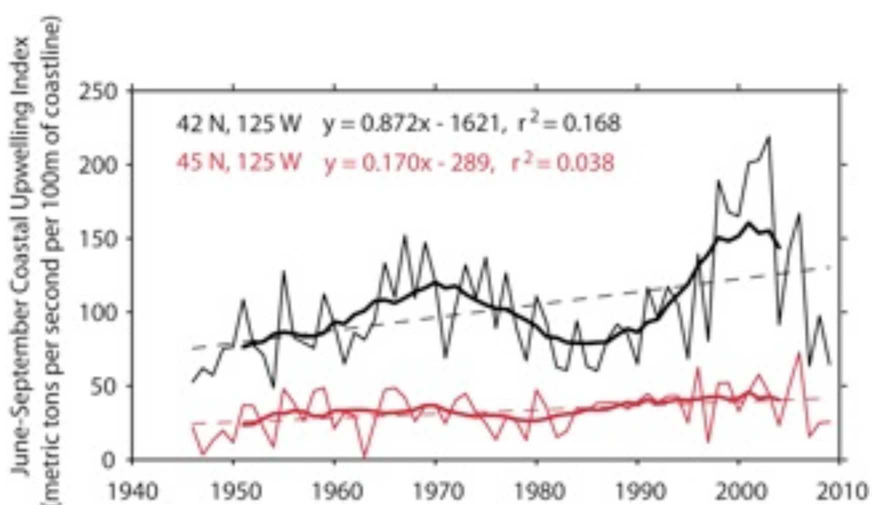

Figure 1.14. Time series of the June-September average monthly Coastal Upwelling Index at 45ºN, 125ºW and 42ºN, 125ºW. Data are provided by Pacific Fisheries Environmental Laboratory. The heavy curves show values of a centered 11-year running average. At these locations, index values of 50, 100 and 200 correspond to wind stress magnitudes of about 0.05, 0.10 and 0.20 N m^{-2}, respectively.

1.3.1.2 Basin-scale wind forcing

As well as inshore upwelling or downwelling along the coastline, there can also be upwelling or downwelling offshore where the wind field has sufficient "curl" (curvature or shear), as it does off southern Oregon (e.g., Chelton et al., 2007). This curl results both from the topographic interaction of the wind blowing past Cape Blanco (Perlin et al., 2004) and from air-sea interactions around the upwelling front: under the same overlying wind speed and direction, the wind stress at the sea surface is higher over warm water than over cold water. As a result of this positive wind stress curl, the cold summer upwelling domain off southern Oregon extends at least twice as far offshore as it does off Newport (Figure 1.12; Huyer et al., 2005; Springer et al., 2009).

Large-scale atmospheric variations over the North Pacific Ocean also affect Oregon's coastal ocean by changing the strength of large-scale upstream currents. For example, the North Pacific Current was stronger than normal in 2002 (Strub and James, 2003) and it brought unusually cold, nutrient-rich water to Oregon (Wheeler et al., 2003). The strength of these large-scale currents can be estimated from gradients of sea surface height measured by satellite altimeters (Strub and James, 2003). The altimeter record is not yet long enough to provide reliable estimates of a long-term trend in the large-scale currents. Various indices of the large-scale atmospheric patterns such as the Pacific Decadal Oscillation (Mantua et al., 1997) have proven to be of limited value for monitoring large-scale currents.

Because fluctuations in current and sea level propagate along ocean margins, Oregon's coastal ocean can be affected by changes in the remote winds over the western equatorial Pacific. For example, during both the 1982 - 83 and the 1997 - 98 El Niño, higher sea-surface temperatures and higher sea levels were observed off Oregon even before local wind patterns became anomalous (Huyer and Smith, 1985; Huyer et al., 2002). These current and temperature anomalies propagate very quickly along the west coast, and can arrive here within a month or so after signals are observed on the equator. Equatorial anomalies in the Pacific are monitored by an array of moorings; data are displayed in real time (http://www.pmel.noaa.gov/tao/).

1.3.1.3 Freshwater input

Surface salinity in Oregon's coastal ocean is strongly affected by fresh-water discharge from the land, with two types of seasonal cycles: through rain-dominant coastal rivers and streams in winter (Austin and Barth, 2002), and through the snowmelt-dominant Columbia River in summer (Huyer et al., 2007). In winter, the downwelling circulation pushes the fresh water from coastal rivers toward shore, forming a narrow lens over the inner shelf, and enhancing the onshore pressure gradient and associated northward current; this northward current can reach speeds of 1 m s^{-1} (2.2 mph; Austin and Barth, 2002). In summer, the upwelling circulation pushes the diluted freshwater discharge from the Columbia River out to sea where it is carried southward by the coastal currents (Barnes et al., 1972; Rivas and Samelson, 2010). The dilute waters of the Columbia River plume are less dense than surrounding ocean waters, and the inshore boundary of the plume tends to coincide with the upwelling front, enhancing both the density gradient and the intensity of the southward coastal current.

The seasonal cycle of the Columbia River discharge has been modified significantly by major dams and deliberate management: peak discharge historically occurred in late spring, but now occurs in autumn (Sherwood et al., 1990). The annual average discharge of the Columbia River shows large interannual variability and some interdecadal variability, but no significant long-term trend between 1928 and 2009. In contrast, the average May-through-July discharge has decreased by about 30% between 1928 and 2009 (Figure 1.15); these changes result from a combination of dam construction and reservoir management, and climate variability and change. Future climate-related reductions in summer flow in snowmelt-dominated rivers like the Columbia are likely (see Chapter 3). With less summer discharge, we would expect the Columbia River plume to be less intense and its inshore boundary adjacent to the upwelling front to be more diffuse.

1.3.1.4 Temperature, salinity, and dissolved oxygen

For more than a decade, 1961 - 1971, water temperature and salinity were measured at intervals of 1 - 3 months at a set of standard stations along a line extending west from Newport to a point 305 km (165 nm) offshore (Figure 1.12). Seasonal sampling of the stations within 190 km (100 nm) of shore resumed for about six years during 1997 - 2003, with limited additional sampling in 2004 and 2005 (Huyer et al., 2007). Both periods also include measurements of dissolved oxygen, though these are less plentiful. Data from the station farthest offshore that was sampled in both periods show temperature decreasing with depth and salinity increasing with depth. The surface layer of the sampled ocean water is thin, warm, quite fresh, and saturated with oxygen. Waters with salinities less than 33.8 represent surface waters of the Subarctic Pacific brought into Oregon's coastal waters by the prevailing California Current; those with salinities less than 32.5 have been locally diluted, usually by discharge from the Columbia River (Barnes et al., 1972). Waters at depths of 300 - 800 m (980 - 2600 ft) with salinities between 33.9 and 34.2 are influenced both by "North Pacific Intermediate Water" which originates in the northwest Pacific (Talley, 1993) and by "equatorial water" which is brought into Oregon's coastal waters by the California Undercurrent which flows northward along the continental slope (Pierce et al., 2000).

Differences between the averages over the two sampling periods (1961 - 1971 and 1997 - 2003, an interval of about 35 years) show some significant long-term changes (Figure 1.16). The surface layer has warmed at a rate of about 0.3˚C (0.5°F) per decade; the layer between 200 and 500 m (660 - 1640 ft) has warmed at a rate of about 0.04°C (0.07°F) per decade. Salinity has decreased in the layer between 500 and 800 m (1600 - 2600 ft) at a rate of 0.006±0.004 per decade; this may reflect the large-scale freshening of North Pacific Intermediate Water (Bindoff et al., 2007). Density has decreased in the surface layer and in the layer between 400 and 600 m at -0.007±0.006 kg m^{-3} per decade. Integrated over the upper 500 m of the ocean, the observed change in density corresponds to a sea level rise about 3 cm (1.2") in about 35 years.

The concentration of dissolved oxygen has decreased significantly at all depths between 200 and 1000 m (660 - 3300 ft) since 1961-1971.

For stations over the continental shelf and slope, Huyer et al., (2007) calculate long-term temperature and salinity differences separately for winter and summer. They find winter temperatures to be higher during 1998 - 2003 than 1961 - 71 but because of high variability between individual winters, the difference is not statistically significant. Similarly they find no significant difference in the average winter salinities. They did, however, find significant warming and freshening in the summer-season averages (Figure 1.17). The largest temperature change (>2°C, 3.2°F) occurs in the thermocline (the layer of steep temperature gradient) which lies 10 - 20 m (33 - 66 ft) deeper in 1997-2005 than in 1961 - 1971. The layer with salinities between 32.5 and 33.8 lies about 20 m (66 ft) deeper now than in 1961 - 1971. The large confidence intervals of the salinity difference in the top 20 m (66 ft) reflect very high variability in the position of the Columbia River plume as a result of day-to-day and week-to-week fluctuations in the wind stress. At NH-35, over the continental slope, 65 km (35 nm) west of Newport, the dissolved oxygen concentration at 200 m (660 ft) has been decreasing at a rate of 0.63 ± 0.27 μmol/kg/yr (Figure 1.18). Inshore, the NH-line shelf has summer oxygen decreases of 1.8-2.0 μmol/kg/yr in near-bottom waters with densities of 1025.8 - 1026.3 kg/m^3. The decrease in oxygen concentration is larger inshore than at the slope station; this could be related to the increased intensity and variability of wind-driven upwelling (Figure 1.14) which could have caused an increase in primary production and respiration over the shelf.

In recent years, hypoxic waters with very low dissolved oxygen concentrations (less than 1.4 ml/l) have been observed near the bottom on the mid to inner shelf during the upwelling season (Chan et al., 2008; Barth et al., 2010; Adams et al., 2010). Hypoxia seems to be especially prevalent in the region of Stonewall and Heceta Banks. Minimum values are often found over the mid to inner shelf (50 - 100 m water depth), reflecting the effect of local biological production and respiration. The size of the hypoxic zone increases over the upwelling season, reaching its maximum extent in mid to late summer.

The NH-line is now being sampled by means of autonomous vehicles (or "gliders"; Erofeev et al., 2010). Sampling extends 90 km (56 mi) offshore from the point at which the water is 30m (100 ft) deep on the inner shelf to 125.1ºW over the continental slope, and vertically from the sea surface to a maximum depth of about 200 m (660 ft). They measure dissolved oxygen as well as temperature and salinity. Routine operations began in 2006, and over 110,000 vertical profiles along 25,000 km of track have now been collected (Erofeev et al., 2010) and the challenge of grafting them to the historical NH-line data sets is being solved (Flink et al., 2010). Glider sampling will continue for the foreseeable future, and we expect that it will show continued warming and further depletion of dissolved oxygen during the summer upwelling season.

1.3.1.5 Nutrients and Acidification

Nutrient concentrations in the upper ocean off Oregon tend to mirror concentrations of dissolved oxygen: high in deep water and very low in the surface water except at times and locations of active or recent upwelling. Both nutrients and oxygen are strongly affected by local primary production: growing phytoplankton absorb nitrate, silicate and phosphate from ambient waters while releasing dissolved oxygen, which may reach high levels of supersaturation during a plankton bloom. Respiration of organic matter absorbs oxygen from ambient waters, and releases nutrients and carbon dioxide. Surface concentrations of nitrate during the summer upwelling season vary from non-detectable to a maximum of about 35 μM, depending on the fluctuations between upwelling and relaxation events; nitrate is usually depleted before phosphate and silicate (Wheeler et al., 2003). The highest concentrations of inshore surface nitrate during the upwelling season are about the same as those observed at depths of 250 - 300m offshore in winter. The width of the coastal strip with elevated nitrate levels is much greater off southern Oregon than it is off central Oregon because of the stronger upwelling there (Huyer et al., 2005). Nutrient concentrations vary interannually, with lower concentrations during El Niño (Corwith and Wheeler, 2002) and higher concentrations during the Subarctic invasion of 2002 (Wheeler et al., 2003). We do not have historical nutrient data of sufficient quality and quantity to estimate long-term trends in nutrient concentrations.

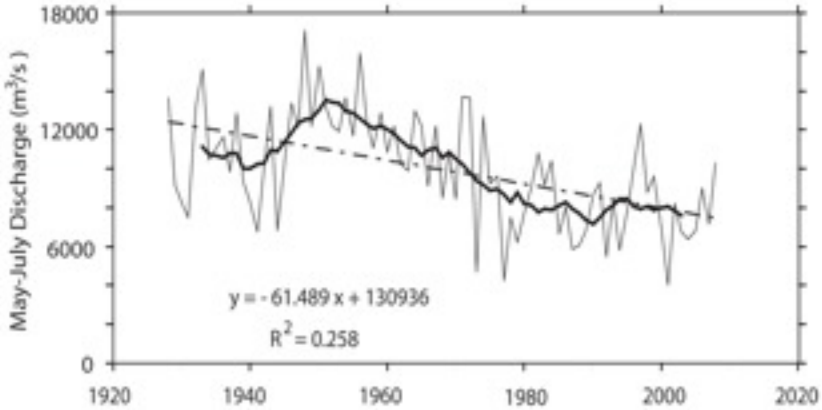

Figure 1.15. Time series of fresh-water discharge from the Columbia River for the late spring and summer months of May through July. The time series of monthly values was collated from USGS sources (as described by Smith et al., 2001) and updated with recent values from USGS Station 14246900 (http://waterdata.usgs.gov/usa/nwis/). The heavy curve shows a centered 11-year running average.

The acidity of seawater depends on the concentration of dissolved carbon dioxide, and it is therefore especially vulnerable to ocean absorption of elevated carbon dioxide levels in the atmosphere. Dissolved carbon dioxide also arises from respiration: thus ocean acidity tends to be highest when and where oxygen concentrations are lowest. Ocean acidity affects the solubility and availability of carbonate ions necessary for the formation of calcium carbonate shells and skeletons of many marine organisms: organisms are potentially vulnerable wherever (and whenever) the seawater saturation of aragonite or calcite is less than 100% (both consist of calcium carbonate; aragonite is more soluble than calcite). Most of the surface ocean is presently supersaturated for aragonite, while the deep ocean is undersaturated; the boundary between them is called the 'saturation horizon' (Bindoff et al., 2007). The saturation horizon is especially shallow in the Northeast Pacific Ocean, where it lies less than 300 m (980 ft) below the sea surface (Feely et al., 2008); scientists estimate that the saturation horizon has moved 50 to 100 m (160 - 330 ft) toward the surface since 1750 (Bindoff et al., 2007). A very recent (May-June 2007) survey of the western continental margin of North America shows that the aragonite saturation horizon lies at a depth of less than 300 m (980 ft) at offshore locations, but less than 100 m (330 ft) over the continental shelf, and even to the sea surface during strong upwelling (Feely 2008; Juranek et al., 2009). Thus Oregon shelf waters are already potentially corrosive to species that form aragonite shells. As the ocean continues to absorb carbon dioxide from the atmosphere, the saturation horizon is certain to rise, and corrosive effects will increase.

Units of measure: dissolved substances in ocean water
Salinity or saltiness: unitless numbers on the "practical salinity scale" which is almost equal to parts per thousand
Dissolved gases: milliliters per liter (ml/l) is equivalent to 1.4 parts per thousand by mass
Nitrate micromolar (μM)= 14 ppm N

1.3.1.6 Summary

Observations of Oregon's coastal waters, mostly by OSU oceanographers, over the past 50 years show an environment that varies tremendously from season to season and from year to year and is influenced by both local and remote processes. Many substantial changes have been observed during this period, including a substantial warming and freshening of the surface

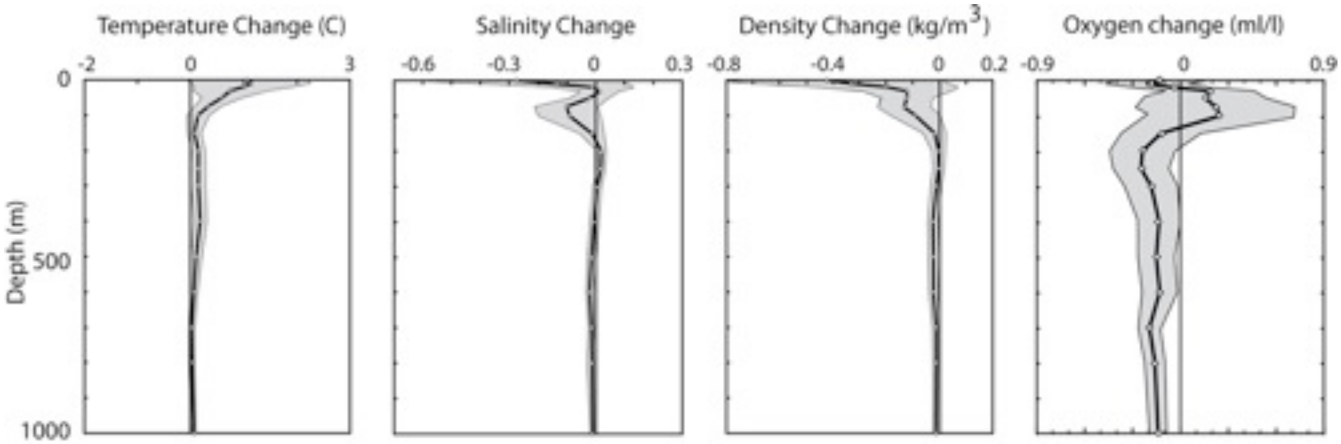

Figure 1.16. Differences between average temperature, salinity (parts per thousand), density and dissolved oxygen profiles at NH-85 (85 nm west of Newport) for two periods: 1997-2005 (~38 samples) minus 1961 - 1971 (~75 samples), with 95% confidence limits.

layer year-round and a reduction in dissolved oxygen. Recent hypoxic events represent a scientific surprise and their cause, and possible links to larger climatic changes driven by human activity, still unknown. Interpretation of the causes of these changes, especially vis-à-vis the human contribution, is hampered by the inadequate quantification of year-to-year and especially decade-to-decade variability which could easily be mistaken for a linear trend related to global climate change.

1.3.2 Future changes in Oregon's marine environment

We have made modest progress in estimating recent trends in Oregon's marine environment, but we have only a very limited ability to predict future changes. Besides assuming persistence of the trends already observed, our principal tool for predicting how future climate will affect the marine environment is the set of global models of the coupled ocean-atmosphere system discussed in Section 1.3 above.

In these models, simulated future changes in the mean surface wind are very small over the Pacific Northwest, especially the alongshore summertime winds that drive coastal upwelling (Mote and Mantua, 2002; Mote and Salathé, 2010). Figure 1.19a shows no significant difference between model estimates of alongshore wind stress for the 1960 - 1999 period and three SRES scenarios for the 2030 - 2059 period. The simulated wind stress is somewhat too weak (-0.03 N m^{-2} vs -0.05 N m^{-2} at 45°N and -0.1 N m^{-2} at 42°N; Risien and Chelton, 2008), but the global models provide the only quantitative prediction available.

Each of the 20 coupled global models discussed above has an ocean model with data points spaced more closely than the atmospheric model, and each model simulates sea surface temperature (SST). However, the modeled SST of Oregon's coastal waters is quite different from the observed SST, especially in the seasonal cycle, because the ocean model is still too coarse to represent the complex oceanic processes over the continental margin. Figure 1.19b shows the simulated mean annual cycle for the 1970 - 1999 and 2030 - 2059 periods for coastal grid points between 46°N and 49°N latitude. The modeled increase in SST is about +1.2°C (2.2°F), somewhat less than for the land areas (+2.0°C, 3.6°F), but a significant change compared to the typical interannual variability of the coastal ocean. Note that the simulated seasonal cycle

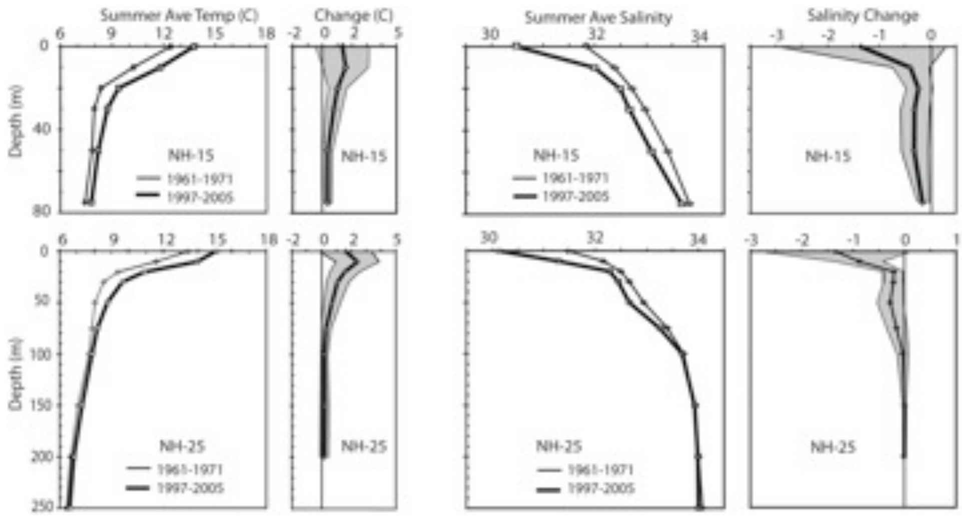

Figure 1.17 Summer-average temperature and salinity profiles at two stations on the NH-line calculated separately for two periods: 1961 - 1971 and 1997 - 2005, and the difference between them (with 95% confidence limits). NH-15 and NH-25 are 15 and 25 nautical miles west of Newport.

for 1970-1999 does not adequately represent the observed temperature of waters over the inner continental shelf, which are likely to be as cool as 8 - 12°C (46 - 54°F) in summer (see Figure 1.12); it more adequately represents surface waters about 100 km offshore. The forecast increase of about 1.2°C is also likely to apply to offshore waters. Note that this modeled increase is less than the summertime increase observed in recent decades (Figure 1.17).

1.3.3 Mean Sea Level

Globally, sea surface elevation rises when land ice melts, increasing the amount of water in the sea, and also when the ocean temperatures rise (due to thermal expansion). Gradients in sea surface elevation are associated with ocean currents – this is why satellite altimeters can be used to study the ocean circulation. Along the west coast of each continent, including North America, summertime winds blowing from higher latitudes pull water offshore and water must rise up from the depths to replace it. This coastal upwelling not only affects the properties of the surface water as explained in section 1.3.2, it also affects the height of the coastal ocean waters. There is approximately 0.5m (19”) difference in mean sea level between winter (higher) and summer (lower), owing to the wind-driven ocean circulation.

These variations are smaller than the tides and are usually noticeable only at times of high or low tide, and they have little effect on the marine environment except in the intertidal zone. The surface elevation measured by coastal tide gages is relative to the land, which itself may be moving slowly upward (e.g., from glacial rebound) or downward (from tectonic subduction). Global sea level rise, and impacts to the Oregon Coast, are explained in Chapter 6.

1.4. Outlook and knowledge gaps

The climatic and marine trends described above are based on available observations. In many cases, the observing networks or sampling frequency have suffered declines in recent decades,

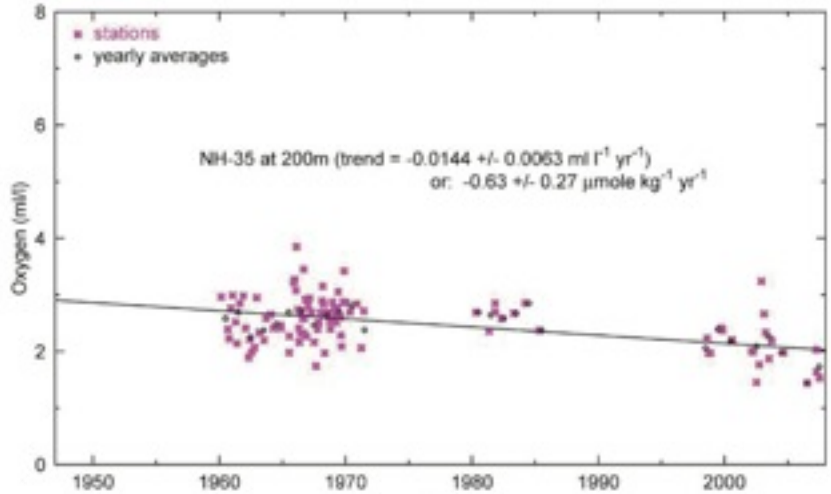

Figure 1.18 Dissolved oxygen concentration in ml/l at 200 m depth at NH-35 (35 nm west of Newport, Oregon) plotted as a function of time.

and scientists' ability to monitor our changing environment and place those changes in a longer-term context is constrained by these declines. While clever statistical analysis and data rescue are expanding both the availability and applicability of past observations (e.g., by digitizing daily records at weather stations previously available only in monthly means), observations and monitoring for the future will require vigorous and sustained effort. The Climate Reference Network is a good start: a nationwide network of 120 high-quality climate stations with multiply redundant instruments and a large buffer of land cover unlikely to change in future decades. Preserving legacy stations and marine sampling capabilities will be important for placing new observations in context with historical observations.

Labor-intensive oceanic observations initiated in the 1960s have recently been augmented by unmanned gliders, long-term moorings, and coastal radar arrays to provide more frequent and continuous sampling of critical properties of Oregon's coastal ocean. These observations will help understand both the natural and man-made variations of the temperature, salinity, dissolved oxygen, nutrients and currents, all of which contribute to the ecological and economic productivity of Oregon's coastal waters.

As this Assessment is being written in 2010, results from a new generation of climate models will become available from modeling groups around the world to support the assessment activities of the IPCC Fifth Assessment report, which scheduled for release in 2013. Enhanced regional modeling capability through a citizen science effort called regional climateprediction.net, hosted by OCCRI, will also be available in 2011. These thousands of simulations of regional climate at 25 km (16 mi) spatial resolution will provide an unprecedented combination of statistical and spatial detail for the western US.

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2. Climate change in Oregon: defining the problem and its causes

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Summary and Knowledge Gaps

Human activities can affect Earth's climate and particularly its temperature by two fundamental processes: first by changing atmospheric composition which causes a greater trapping of the Earth's heat, and second by changing the amount of sunlight reflected or scattered by the atmosphere or the Earth's surface. The first increases the natural greenhouse effect, caused mostly by water vapor and carbon dioxide along with a few non-CO_2 greenhouse gases, and leads to global warming. The second causes the albedo to change which can either increase or decrease Earth's temperature. Most of what is driving global warming comes from the direct emissions of carbon dioxide, methane, nitrous oxide and other source gases released by human activities related to the production of energy and food. The climatic effect of such emissions is increased by water vapor feedback, which means as global warming occurs due to the increase of the greenhouse gases mentioned above, the amount of water that is held in the atmosphere increases, causing a further increase of temperature. There are many other feedbacks that are less well understood. Additionally, various gases emitted by human activities produce ozone in the lower atmosphere, which also adds to global warming.

Changes of albedo due to human activities can either exacerbate or ameliorate net global warming depending on a complex set of conditions that drives such changes. The albedo effect is thought to be small at present, but its future trend is not well understood. Earth's albedo is determined by a large number of disparate elements in the environment - dominated by clouds but also including aerosols and surface elements such as forests, ice caps, and oceans. Human activities that release gases such as sulfur dioxide can increase the albedo through the formation of sulfate aerosols in the atmosphere producing a cooling effect, while the direct release of black carbon aerosol (soot) can increase global warming under the right circumstances, or melt glacial and polar ice further decreasing the albedo.

The sources of Oregon's greenhouse gas emissions can be broadly listed as energy, agriculture, industrial processes, and waste management (Figure 2.1). Energy, particularly electricity consumption and transportation, is the largest source of greenhouse gas emissions in the state. Emissions associated with the consumption of electricity have been between 20 - 24 MMTCO2e (million metric tons of CO_2 equivalent) per year in the last decade, representing about 33% of

Oregon's total emissions. The transportation sector represents about 37% of greenhouse gas

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emissions in Oregon, ranging from about 21 to 25 MMTCO2e over the last twenty years.

In light of Oregon's small population, metrics other than total emissions should be used to assess Oregon's contribution to climate change. The simplest of these is per capita emissions, which provides a method to compare the carbon footprint of Oregonians to national and international norms. Oregon's per capita emissions of 18 MTCO2e in 2005 was the eleventh lowest of all U.S. states, or about twenty percent lower than the national average (24 MTCO2e). However, compared to developed countries, Oregon's per capita emissions rank quite high. Compared to the 39 industrialized countries with reported inventories in 2005, Oregon emissions are fifth highest, and are nearly double the European Community average. Relative to the global average, Oregon's per capita emissions are almost three times higher. One contributing factor to Europe's low emissions seems to be high population densities; a correlation also observed in U.S. states.

Ultimately, global scale reduction of the climate warming will require a global scale agreement and implementation. A single state such as Oregon, by itself, will not be able to alter the course of global warming. However, once there is a concerted effort to implement a global policy, each state or similar municipality, will need to tailor their contribution to what causes emissions and global warming from their region.

2.1 The Causes of Climate Change

Human activities can affect Earth's climate, and particularly its temperature, by two fundamental processes: first by changing atmospheric composition, which causes an increased trapping of the Earth's heat, and second by changing how much sunlight is reflected or scattered by the atmosphere or the Earth's surface. The first increases the natural greenhouse effect, caused mostly by water vapor and carbon dioxide along with a few key non-CO_2 greenhouse gases, and leads to global warming. The second causes the albedo or the reflectivity of the earth to change, which depending on how it is affected, can either increase or decrease the surface temperature. Most of what is driving global warming and the resulting climatic changes comes from the direct emissions of carbon dioxide, methane, nitrous oxide and other source gases released by human activities mostly related to the production of energy and food. The climatic effect of such emissions is increased by water vapor feedback, which means that as the earth warms due to the increased levels of the greenhouse gases the amount of water vapor held in the atmosphere increases causing a further increase of temperature. There are many other atmospheric feedbacks that are less well understood. Additionally, various gases emitted by human activities produce ozone in the lower atmosphere, also adding to global warming.

Changes of albedo due to human activities can either exacerbate or ameliorate net global warming depending on a complex set of conditions that drive such changes. The albedo effect is thought to be small at present but its future trend is not known. The Earth's albedo is determined by a large number of disparate elements in the environment; dominated by clouds but also including aerosols and surface features such as forests, ice caps, and oceans. Human activities that release gases such as sulfur dioxide can add sulfate aerosol in the atmosphere that in turn increases the albedo causing a cooling effect, or the release of soot aerosol (black carbon) that can, under the right circumstances, increase global warming, or melt glacial and polar ice and further decrease the albedo. Land use changes such as re-forestation can reduce the albedo and may offset some of the gains of carbon sequestration.

In terms of environmental policy to reduce climatic change, the first target is reductions in the source gases. When concerns about global warming first developed, it seemed that controlling carbon dioxide might solve the problem. In time however, it became clear that such a policy, as difficult as it is was to implement, may not be enough. It was shown that although the non-CO_2 greenhouse gases are individually not as abundant as carbon dioxide, there are several of them and they are twenty to several thousand times more effective at causing global warming for each kilogram added to the atmosphere when compared with each added kilogram of carbon dioxide (the global warming potential is used as such an index). Therefore, reductions of these gases were included under a more comprehensive plan to control global warming in the Kyoto Protocol. Now, we know that there are yet other drivers of climate change that may require scrutiny such as aerosols and albedo altering mechanisms affected by human activities. How all these elements can be integrated into a policy that is both effective and economical is a difficult challenge for legislators, policy makers and scientists.

Ultimately, global scale reduction of warming trends will require a global scale agreement and implementation. A single state such as Oregon, by itself, will not be able to alter the course of global warming. However, once there is a concerted effort to implement a global warming policy, each state or similar municipality will need to tailor their strategies to what causes emissions of greenhouse gases from their region.

In view of the preceding discussion, we classify the drivers of climate change into three categories. First, we consider greenhouse gases and black carbon emissions as the main global warming elements that can be managed directly. On the second level are the global warming effects of ozone changes and the potential cooling effects of sulfur emissions arising from precursors rather than direct emissions. And finally, there are potential changes of albedo due to land use change and other indirect effects. As we implement direct policies to mitigate climate change it is important to remain cognizant of these other complex interactions that may undo our efforts, or though less likely, may allow us to relax our targets and meet the same goals.

In this chapter we focus on the emissions of greenhouse gases, since the first policies to reduce global warming will most likely be based to a large extent on controlling these emissions. To understand the drivers of future climate change for Oregon we write the annual anthropogenic emissions of a greenhouse gas in terms of basic factors that drive emissions. As an example, consider Oregon's annual emissions (E) in tons of carbon dioxide emitted per year from energy use. These can be written as a product of three more or less independently changeable variables. These are the emission factor (K = tons CO_2 emitted / BTU of energy used, in this example), the per capita consumption rate (C = BTU used per person per year) and the population (P = number of people). This formula $E = K \times C \times P$ can be written for any greenhouse gas and any activity recognizing that it is the product of the emission factor (K), the per capita activity rate (C), and the total activity (P). We will use this general framework later to discuss emissions from various sources in Oregon and each of these variables will take on specific meanings related to the source and activity under consideration. To continue our example, the total emissions of carbon dioxide from Oregon then would be the sum of these estimates from all activities that result in emissions of carbon dioxide. To get an estimate of the total effect of Oregon's emissions on the climate, we have to make similar estimates for the other greenhouse gases such as methane and nitrous oxide. Often the other greenhouse gas emissions are multiplied by a factor such as the global warming potential, which is usually quite large, to arrive at the number of tons of carbon dioxide that if emitted, would have a similar climatic impact as the emissions of the non-CO2 greenhouse gas. This way, we can come up with a single number (tons) of equivalent carbon dioxide emissions from Oregon for a given year. Clearly this number will change from year to year since any or all of the components change in time (K, C, and P). Most of our control strategies focus on reducing the emission factor, and to some extent on reducing per capita consumption, which could reduce affluence.

While this is a universal formula that can be used to estimate emissions from states, or countries, or even the whole world, the components that are likely to change are quite different on these various geographical scales. On a global scale, for example, the emissions calculated from such a formula have been, and still are driven most by changes in per capita consumption rates (C) and much less by population increases contrary to what most people think. China in recent times is a good example of emissions rising sharply as the per capita consumption has risen so that now China's carbon dioxide emissions exceed those of the United States. For many of our states, including Oregon, the population is already very affluent compared to world standards. There is a saturation effect whereby people are not likely to consume more per capita than they already are. Therefore for Oregon the major driver of future climate change may well be the expected changes in population about which very little can be done. Thus our projections of future greenhouse gas emissions from Oregon depend heavily on expected population increases that are driven not by the natural increases but rather by people moving in or out of the state. The effect of population changes can be normalized by targeting a reduction in per capita emission rates rather than in the absolute tons of greenhouse gases emitted from Oregon over the coming decades.

In this chapter we examine Oregon's place in the climate change issue and how the state may approach its role in national and global policies to limit greenhouse gases and other causes of global climate change with a mixture of mitigation and adaptation. To this end, we describe first the emissions of greenhouse gases from various sectors representing different relationships with per capita consumption and population. The present estimates are based on inventory methods that are subject to considerable uncertainty which can impede the success of policies aimed at reducing emissions. To guard against such an occurrence, a validation component is necessary as has been advocated by a recent report from the National Academy of Sciences (2010). We will examine validation case studies and methods that exist and hold promise for future policy. We tie together these findings in the final section where we deal with what emissions we expect from Oregon and how these are justified by the discussion of the factors specific to Oregon. In addition, we discuss some issues and novel strategies that are of particular importance for policies to limit emissions from small regions such as Oregon.

2.1.1. Emissions and Trends of Greenhouse Gases by Sector

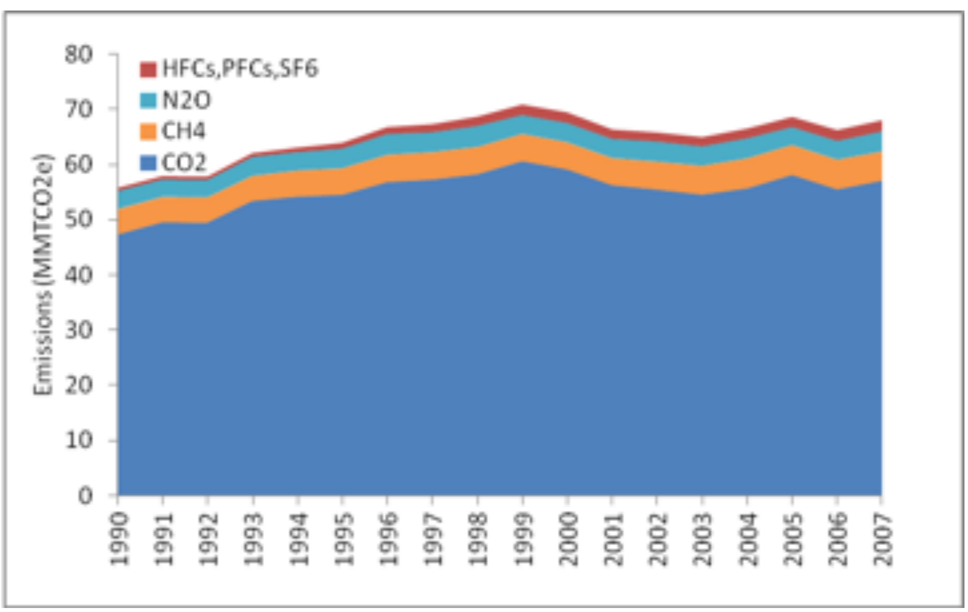

Figure 2.1 Oregon's greenhouse gas emissions by gas. Emissions are reported as million metric tons of equivalent carbon dioxide (MMTCO2e). Emissions of non-CO2 greenhouse gases are scaled by their respective global warming potentials.

Greenhouse gas emissions can be categorized either by gas species (for example carbon dioxide, methane, and nitrous oxide) or by economic sector and human activity (energy use, electricity generation, agriculture, etc.). There are advantages to each classification depending on the planned use of the inventory. For scientific assessment of a greenhouse gas, the former is required to delineate the sources and sinks of the gas. For policy guidance, the latter approach is typically used in recognition that economic controls are typically applied to specific sectors of activity and not to any one particular gas. In this approach, emissions of non-CO2 greenhouse gases are scaled by their global warming potentials to produce equivalent carbon dioxide emissions (CO2e). Here we report emissions by sector but include a figure of Oregon's greenhouse gas emissions by gas species for reference (Figure 2.1). In 2005 carbon dioxide contributed 86% of total emission, while methane, nitrous oxide, and the high GWP gases respectively contributed 7%, 4%, 3% of the total (GWP is the Global Warming Potential as defined in Forster et al. (2007).

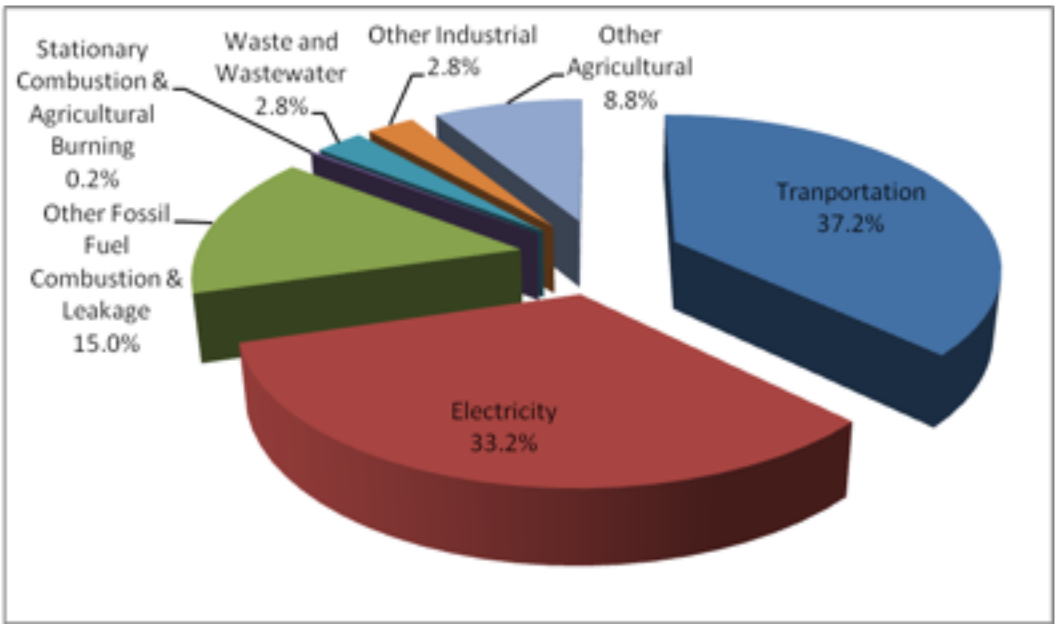

Figure 2.2 Sources of greenhouse gases in Oregon. Data is averaged over 2003 - 2007, Oregon Greenhouse Gas Inventory, 2010.

The sources of Oregon's greenhouse gas emissions can be broadly listed as energy, agriculture, industrial processes, and waste management (Figure 2.2). The energy sector consists of a number of important sub-sectors including electricity use and transportation. Unlike some greenhouse gas inventories, emissions from electricity use are kept separate here and are not distributed according to residential, commercial, or industrial use. Rather, greenhouse gas emissions listed for these sectors are only from the direct combustion of fossil fuels. We feel this method is more relevant for policymakers as it clearly identifies the impact of changes to Oregon's mix of electricity suppliers.

2.1.2 Energy Use

2.1.2.1 Electricity

Greenhouse gas emissions from electricity can be estimated based on either production or consumption. The U.S. EPA (2009) reports production-based emissions since their chief concern is to develop a national inventory, and tracking emissions where they are produced rather than where the electricity is consumed is simpler. To evaluate the role of states in the national emissions, it makes more sense to track electricity consumption rather than production. This method reduces emissions for energy exporting states and increases emissions for importers.

Oregon's mix of electricity sources is nearly evenly divided between coal and hydropower. Emissions associated with the consumption of electricity have historically been one of the largest sources of greenhouse gases in Oregon, between 20 to 25 MMTCO2e (million metric tons of CO2 equivalent) per year in the last decade, representing about 33% of Oregon's total greenhouse gas emissions. Oregon tracks these emissions by tracing back the power used by Oregonians to its sources. In most years Oregon has imported anywhere between one and twenty percent of its electricity. In good water years when hydropower is abundant, Oregon exports electricity. As most of Oregon's imported electricity is generated by coal-fired plants, the switch from a production-based to a consumption-based inventory increases Oregon's emissions. About two-thirds of Oregon’s emissions from electricity use come from out-of-state energy supply. Emissions from electricity use was one of the fastest growing sources of greenhouse gas emissions from 1990 to 2007, though much of this increase was due to the shutdown of the Trojan nuclear plant in the early 1990s and the subsequent replacement of its lost production with coal-generated electricity (Oregon, State Reports 2008, 2009, 2010).

Emissions from electricity consumption are highest for residential use, followed by the commercial and industrial sectors (9, 7, and 6 MMTCO2e respectively for 2003-2007average use in round numbers). Though industrial emissions have remained the same or decreased from 1990 to 2007, residential and commercial emissions increased by 45% and 55%. These large increases are likely driven primarily by population growth.

2.1.2.2 Transportation

The transportation sector has been one of the single largest sources of greenhouse gas emission in Oregon and throughout the Pacific Northwest, over the last twenty years. Emission estimates are derived from fuel use for Oregon provided by the U.S. Department of Energy for the state, and not aggregated from vehicle use statistics. Because transportation fuels are taxed, these data are expected to be accurate (Oregon Department of Transportation, 2009).

Burning fossil fuels in automobiles and other vehicles used to transport goods and people leads to emissions of several greenhouse gases. Carbon dioxide is the main emission and methane and nitrous oxide are emitted in lesser quantities. Accounting for emissions from the transportation sector should also include full life cycle costs, specifically emissions from extracting, producing and distributing transportation fuels and in manufacturing, distributing and disposing of vehicles themselves. The Transportation Research Board Special Report (TRBS, 2008) provides a comprehensive review of emissions from the transportation sector.

Worldwide estimates show that transportation related greenhouse gas emissions account for about 20% of total global greenhouse gas emissions. All combustion engine transportation modes emit greenhouse gases including rail, buses, airplanes, trucks, ships, passenger vehicles, light duty trucks and motorcycles. Light-duty vehicles such as passenger cars, pickup trucks, mini-vans and sport utility vehicles account for 60 percent of all emissions (TRBS, 2008). However, emissions from the air transport (12.4%) and truck freight (18.4%) are significant and are projected to grow faster than emissions from light duty vehicles.

Emissions of carbon dioxide and other greenhouse gases from the transportation sector follow the same formula as discussed earlier: $Ei = Ki \times Ci \times Pi$. Here "i" represents the vehicle type (cars, trucks, rail etc); Ki is the emission factor for type "i" vehicle (Tons GHG emitted / BTU consumed); Ci is the energy consumed by or the efficiency of this type of vehicle (BTU consumed / mile traveled) and Pi is the number of miles travelled/year by type "i" vehicles. The total emissions is the sum over all the types of vehicles represented by the index "i." The primary driver of transportation sector emissions historically has been the level of transportation activity measured by vehicle miles traveled by people and transportation of goods. Improvements in vehicle fuel efficiency that reduce emissions have been overwhelmed by increased activity and by switches to more energy intensive modes, such as from freight rail to trucks.

Gross total emissions from all sectors of the Oregon economy amount to about 67 million metric tons of carbon dioxide (CO_2) equivalent for the year 2005. This tabulation is based on the consumption by Oregonians, regardless of where the energy comes from. The transportation sector accounts for about 37% of the greenhouse gas emissions in Oregon. This is a larger percentage than the national average because Oregon's total energy use sector includes hydropower resources, which do not result in greenhouse gas emissions and thus reduce emissions from sectors other than transportation.

Next, we examine each of the factors affecting emissions from transportation: the energy efficiency of vehicles, carbon content of fuels or emission factor, and the miles driven.

Vehicle Efficiency : Vehicle efficiency has a major effect on greenhouse gas emissions. There is both political and technical uncertainty in our ability to predict future vehicle efficiency, however, the technologies for reducing fuel consumption are well known. They include engine efficiency technology, transmission technologies, reductions in vehicle weight and improved aerodynamics. Even when vehicle efficiency is improved, greenhouse gas reductions may not be realized because of driver behavior, the adoption rate for the technology as fleets turnover, or decisions to use such technologies.

Carbon Content of Fuel: Motor vehicle fuels vary in the amount of greenhouse gases they produce for the energy they contain, especially when emissions for producing and transporting the fuels to the end users are included in the calculation. The greenhouse gas production of different fuels is known as the carbon intensity of the fuel. It is measured as the grams of greenhouse gas, expressed as CO_2 equivalents, per mega joule of energy contained in the fuel (or another energy unit). For example, the California Air Resources Board has calculated that gasoline emits about 96 grams of greenhouse gas per mega joule. In comparison, the emissions from biodiesel produced from waste cooking oil are about 14 grams of greenhouse gas per mega joule. Therefore, greenhouse gases emissions can be reduced by lowering the carbon intensity of transport fuels.

Efficiency of the Transportation Network: The operational efficiency of the transportation network affects the efficiency of vehicles. Vehicle travel speeds affect light vehicle fuel efficiency. Figure 2.3 shows that vehicle fuel efficiency and the speed at which maximum efficiency is achieved varies among different models of vehicles. In general, fuel efficiency declines at speeds below 25 mph and over 60 mph.

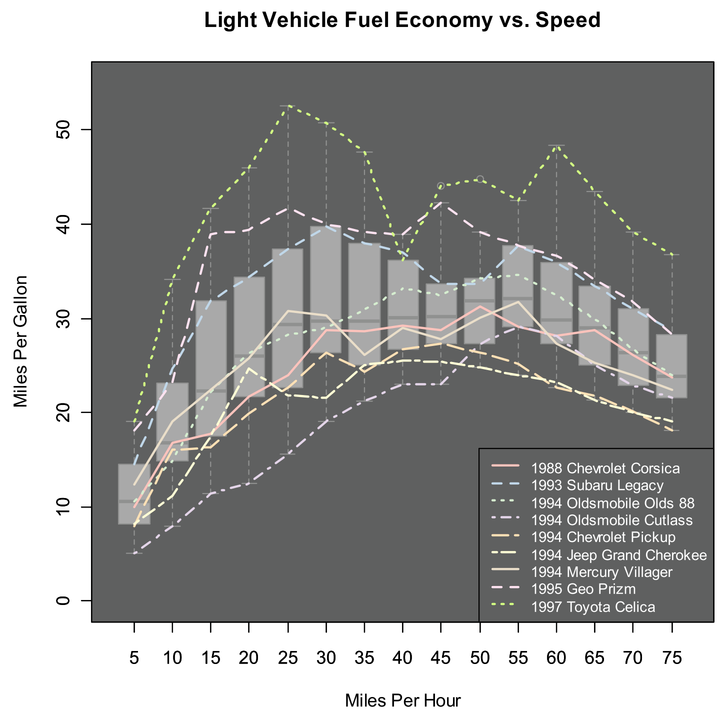

Figure 2.3 Light vehicle fuel economy vs. steady state speed (Davis et al., Transportation Energy Data Book, ORNL, 2009, Table 4.28)

Fuel economy is also adversely affected by vehicle deceleration and acceleration. Deceleration wastes energy as the vehicle slows down; acceleration uses more energy than constant speed travel of the same distance; stopping wastes fuel in idling as well. The design and operation of roads affects vehicle speeds and the amount of acceleration, deceleration and stopping of cars. Optimization of traffic signal spacing and coordination of traffic signals allows for optimization of traffic progression and minimizes stopping at traffic signals. Freeway management strategies, such as incident management and ramp metering, also smooth out traffic flow and improve vehicle fuel efficiency thus reducing greenhouse gas emissions.

While increasing speeds can improve vehicle fuel efficiency, it may also increase the amount of vehicle travel. The distances that people travel are affected by travel speeds and times. In the absence of other influences, travel distances increase as travel speeds increase and travel times decrease. The net effect on greenhouse gas emissions will be the result of complex interactions and cannot be determined without modeling the specific situation.

Improving transportation system efficiency can also be important for reducing greenhouse gas emissions from public transportation. Urban transit buses running with low occupancies produce more greenhouse gas emissions per passenger mile of travel than a high efficiency single-occupant automobile (M.J. Bradley & Assoc., 2007; Univ. of MN, 2008).
Amount of Driving (VMT): It is estimated that Vehicle Miles Traveled (VMT) in the United States will increase by about 60% between 2005 and 2030 (Ewing et al., 2007). The miles traveled have been growing at a relatively slow rate in Oregon, and the Oregon Transportation Plan forecasts an increase of 40 percent over that time. The growth of miles traveled in Oregon began to level off and per capita miles travelled began declining even before fuel prices started their rapid rise in recent years, Figure 2.4. The miles traveled per capita in the Portland metropolitan area started declining in the mid-1990s (Lomax and Schrank, 2009). Meanwhile the per capita miles travelled for other large metropolitan areas have continued to grow.

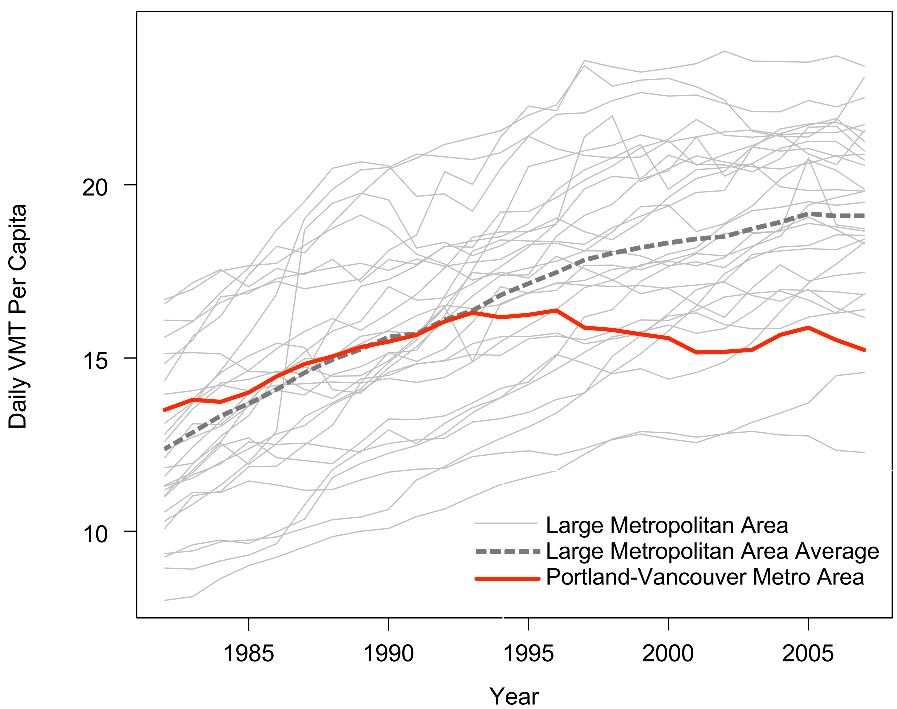

Figure 2.4: Daily Vehicle Miles Traveled (VMT) per capita on major roads in large metropolitan areas, 1982 - 2007 (Texas Transportation Institute Urban Mobility Report database.) Large Metropolitan areas have populations from 1 - 3 million in 2005 or later. Major road daily VMT per capita is calculated by adding annual average dailyVMT for freeways and arterial roads, then dividing by the metropolitan area population.

Total motor fuel consumption in Oregon has increased by only 0.25 percent from 1999 to 2007, while the population of the state increased by 10.4 percent. The net effect has been a 9.2 percent decrease in consumption per capita (see Oregon Dept. of Transportation Data website). Some of the leveling off in state highway miles driven may be due to the shifting of a portion of traffic away from state highways to local roads as a result of rising congestion. These trends, if they can be maintained will make it easier for Oregon's overall strategy to limit greenhouse gas emissions to agreed upon targets.

2.1.2.3 Other combustion: residential, commercial, and industrial

Most of the energy use in this category comes from the burning of natural gas and other fossil fuels to heat homes and businesses. Collectively, emissions from this energy use are about 17% of the state total in Oregon. Emissions from the residential and industrial sectors rose rapidly from 1990 to 2005 (29% and 26% increase respectively) and are among the fastest growing sectors. Although the flows of natural gas are well understood, the data on the use of other heating fuels, especially fuel oil and propane, are highly uncertain. In addition, the blurring of residential and commercial uses in many areas makes it difficult to differentiate uses between these two sectors. Data aggregated between the two sources is more reliable.

2.1.2.4 Waste management

Waste streams are managed with a variety of methods including incineration, wastewater treatment plants, and landfills. The incineration of waste is a source of carbon dioxide, some of it may be of fossil origin, while wastewater treatment and landfills are sources of both methane and nitrous oxide. Municipal and industrial landfills are the second largest source of methane after enteric fermentation (digestion of carbon food sources in ruminant animals). In many anaerobic systems, that is, where methane producing bacteria grow in the absence of oxygen, a significant fraction of the methane that is produced is oxidized by other bacteria before it can be released to the atmosphere. Overall, emissions from waste contribute only about 3% to Oregon's total greenhouse gases, but they have increased by 24% since 1990.

2.2.1.5 Agriculture

Emissions of the greenhouse gases methane and nitrous oxide from agriculture in Oregon were estimated to be about 7% of the state's annual greenhouse emissions (Oregon Dept. of Energy, 2004). Carbon dioxide emitted from harvest, field burning, or pruning is not counted as an addition to the atmosphere, as it is considered part of the annual cycle of CO2 when crops re-grow in spring.

For methane, the agricultural emissions come from ruminant livestock, particularly cattle. Ruminants ferment cellulose in their stomachs, producing methane as a by-product of digestion. Emissions vary depending on the age of the animal and the digestibility of the feed (Johnson et al., 2000). State agricultural data, submitted to the federal government, are used to estimate these agricultural greenhouse gas emissions. There are many known sources of uncertainty in these estimates making the estimates potentially inaccurate.

The application of fertilizers for agriculture, and to a lesser degree, for residences and commercial buildings, is a source of nitrous oxide to the atmosphere. By analyzing the data on the fertilizer use in Oregon it is possible to estimate emissions. Additional sources of nitrous oxide and methane come from animal manure, agricultural biomass burning and wildfires Estimates from these sources at this time small, less than 1% of the agricultural source for biomass burning, while manure management is about 9% of the total agricultural source.

2.1.2.6 Industrial sources

Manufacturing and industrial facilities emit greenhouse gases through a number of chemical processes used to create products. These sources include cement and lime manufacturing which generates carbon dioxide; aluminum, and semiconductor production which release small amounts of high global warming potential gases such as the perflurocarbons and sulfur hexafluoride. Because in the past there has been no requirement to report greenhouse gas emissions from these types of facilities, estimates of industrial emissions are extremely uncertain, mostly derived from taking population-based proportionate shares of national emission estimates and applying them to Oregon. The trend from industrial emissions is thought to be high, with emissions increasing by over 50% since 1990. Still they contribute less than 5%of Oregon's emissions. Starting in 2009 Oregon began requiring reporting of emissions from many industrial sources, which will lead to improved emissions data for many sources.

2.2. Oregon Emissions in Perspective

In the previous section we delineated Oregon's greenhouse gas emissions according to economic sectors. This provides an important perspective on what the sources of Oregon's emissions are and what activities may be targeted for mitigation. Next we will place Oregon's emissions into perspective with regard to both national and global emissions. It comes as no surprise that Oregon contributes only a small fraction to national and global emissions. In 2005 Oregon's emissions were 1% of U.S. national emissions and 0.2% of global emissions. Even if policies in Oregon were to generate dramatic reductions in emissions, these would have imperceptible impacts on global climate change. Global emissions will determine the climatic impacts that Oregon will experience and to which the state will need to adapt.

In light of Oregon's small population, metrics other than total emissions should be used to assess Oregon's contribution to climate change. The simplest of these is per capita emissions, which provides a method to compare the carbon footprint of Oregonians to national and international norms (Figure 2.5). Oregon's per capita emissions of 18 MTCO2e in 2005 were the eleventh lowest of all U.S. states or about 25% lower than the national average (24 MTCO2e). However, compared to developed countries, Oregon's per capita emissions rank quite high. Of the 39 (industrialized) countries with reported inventories in 2005, Oregon's emissions are fifth highest, and are nearly double the European Community average. Relative to the global average, Oregon's per capita emissions are almost three times higher. One contributing factor to Europe's low emissions seems to be high population densities; a correlation also observed in U.S. states as we discuss below.

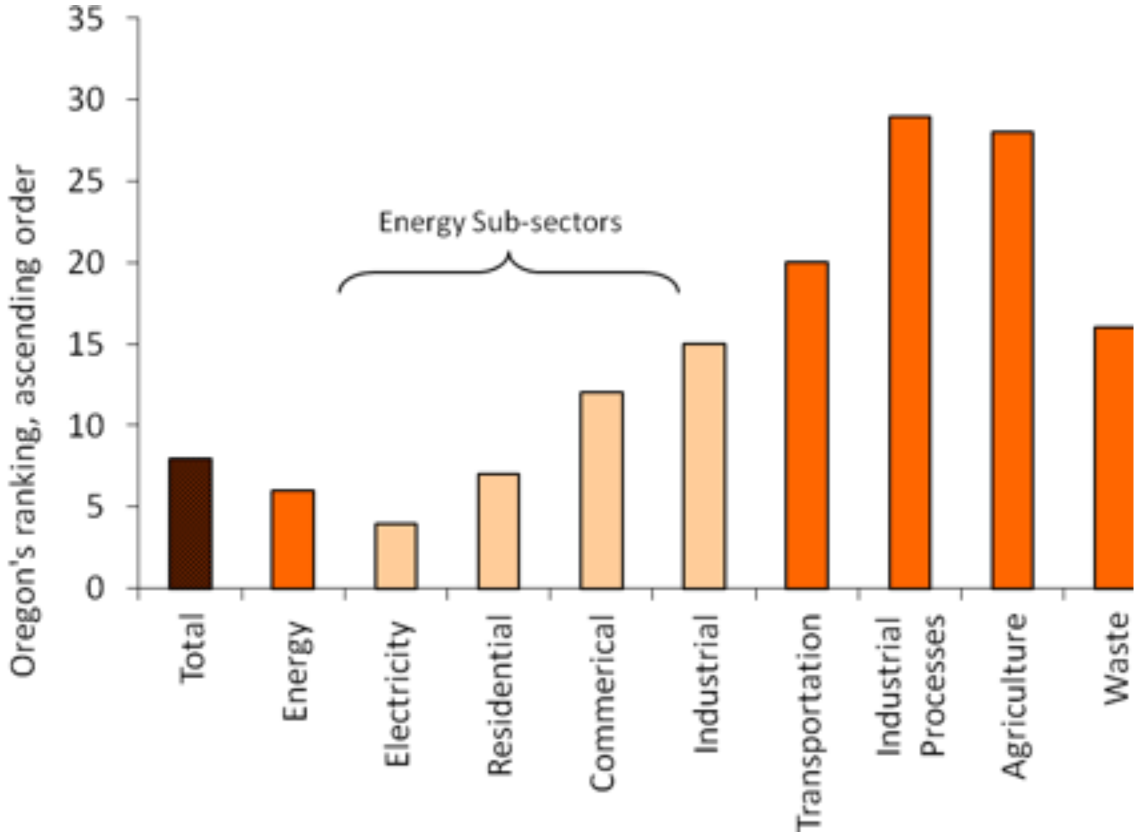

Figure 2.6 Oregon's ranking of per capita emissions by sector relative to other U.S. states, data year 2005. Ranking is based on data in ascending order. (That is lowest emission is 1; highest emissions per capita is 50.) Emissions by sector data from World Resources Institute Climate Analysis Indicators Tool (CAIT US Version 4.0, 2010), except for the electricity sector data which was derived from the U.S. EPA (2010).

Why are Oregon's emissions low compared to other U.S. states? Answers here will help policymakers identify and target sectors that may be amenable to future emission reductions. Figure 2.6 shows Oregon's ranking relative to all states in sector-based per capita emissions. A low ranking means that Oregon has low per capita emissions in that sector relative to other states. Of the four major economic sectors, Oregon's total per capita emissions are driven primarily by energy. In energy use Oregon ranks eleventh lowest over all states, and this ranking in turn is driven primarily by electricity generation where Oregon ranks thirteenth lowest in the country. Within the framework of our equation E / P (per capita emissions) = K × C this ranking is based on a combination of per capita electricity consumption (C) and the emission factor (K) for Oregon's electricity consumption, or a combination of both. In 2008 Oregon had the fourteenth highest per capita consumption rate of electricity (U.S. Energy Information Administration, 2010) and the eleventh lowest emission factor (tons CO2e per MWh) among states (US EPA, 2010), which in large part is due to the abundance of hydroelectric power sources in the region that provide Oregon's electricity.

In other sectors, Oregon's ranking is closer to the national average. Per capita emissions from industrial processes and agriculture rank 21st and 28th respectively, while Oregon's per capita transportation emissions are ranked 21st nationally. Oregon's residential emissions also rank low relative to national norms (eighth). Residential energy usage is dominated by fossil fuel combustion used for heating. The low emissions are in part due to Oregon's mild natural climate with below average number of degree heating days, ranking twenty-ninth in the country.

Finally, one of the most useful ways to assess Oregon's greenhouse gas emissions is by the carbon intensity of its economy. The carbon intensity reflects the amount of greenhouse gases emitted per one dollar of gross state product (GSP) produced. It is calculated simply by dividing the state's Oregon's greenhouse gas emissions by its GSP. A state planning to lower its emissions, while keeping economic output high, would aim to reduce this number.

Oregon emits 460 metric tons of CO2e per million dollars of GSP (2000 data) ranking favorably among other states. Oregon is the eleventh least carbon intensive economy in the country, but is the most carbon intensive of the contiguous Pacific coast states. The other states with low carbon economies are primarily small New England states plus the District of Columbia. The economies of the Pacific coast states benefit from low carbon electricity sources, but the eastern seaboard states rely primarily on carbon-intensive coal for electricity, which means their low carbon emissions must be due to other factors. These states are highly urbanized and have among the highest population densities in the country. In Figure 2.7 we plot carbon intensities as a function of state urbanization. Here urbanization is defined according to the US Census Bureau (U.S. Census Bureau, 2010).

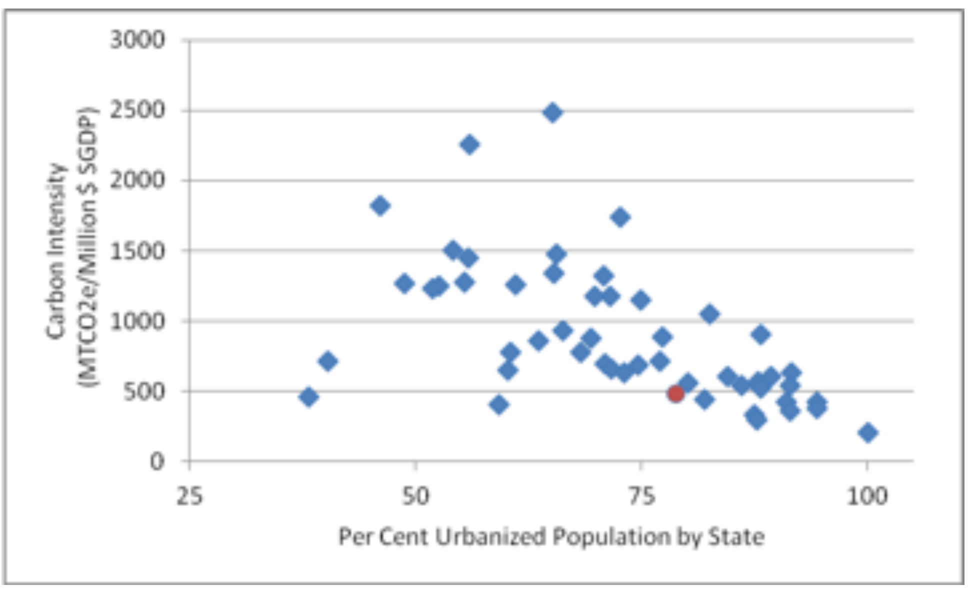

Figure 2.7 The economic carbon intensity is plotted for all states relative to their urbanization. Oregon is highlighted in red. Urbanization is defined by US Census Bureau 2000 Census of Population and Housing, http://www.census.gov; state emissions data for 2000 from the US EPA, http://www.epa.gov; SGDP (State Gross Domestic Product) data for year 2000 from the US Dept. of Commerce Bureau of Economic Analysis. http://www.bea.gov.

A negative correlation exists between carbon intensity and urbanization; generally as urbanization increases, carbon intensity decreases. This correlation likely reflects underlying causal relations between the two variables. For example, as population urbanizes, per capita miles driven likely decrease due to shorter commutes and proximity to public transportation. Agricultural activities, which are carbon intensive on account of fuels, fertilizers, and transportation, also likely decrease with urbanization. Oregon's population is 79% urbanized. Of states with similar urbanization (e.g., Delaware, Ohio, Pennsylvania), Oregon carbon intensity is lowest.

In summary, Oregon has low per capita emissions compared to other states while maintaining a relatively high gross state product, producing an economy that is the eleventh lowest in carbon intensity. Oregon benefits from being close to hydropower generation and a mild climate keeps heating emissions low. In addition Oregon has an above average rate of urbanization, reducing per capita miles driven and other factors that generate greenhouse gases. Partially offsetting these factors is Oregon's large agricultural sector, which tends to be carbon intensive.

2.3. Uncertainty and Validation of Emissions

2.3.1 Uncertainty and Validation

Greenhouse gas emissions can be estimated by several independent means. An emissions inventory is a common method by which estimates are made for each process that releases a greenhouse gas to the atmosphere. These estimates are based on an emission factor and an extrapolation factor. For example, we may have estimates based on measurements of the average number of grams of CO_2 emitted per mile of driving. We can multiply this by an estimate of the number of miles per year that people drive in Oregon (or the per capita miles driven times the population, to put it in the same form as $E = K \times C \times P$ discussed earlier , where the combination $C \times P$ is the number of miles driven per year). The accuracy of both these factors is questionable and may lead to systematic under or over estimates of emissions. In addition to the problem of accuracy just described, there may be substantial year-to-year variations in these factors depending on environmental or economic conditions, hence an estimate for one year may not be reliable for another.

Most regulations for air and other forms of pollution require measurements and testing to determine whether compliance with regulations has been achieved. For mitigating climate change also, methods to validate emissions inventories must be made an integral part of the policy. Without validation we will neither know whether our estimates are reliable and accurate, nor will we know whether our policies to reduce emissions are working.

The space and timescales over which greenhouse gas inventories are taken also have a major effect on the accuracy of the estimates. Generally, the smaller the space and timescales are, the less reliable the numbers are likely to be. This is partly due to the fact that the actual emissions tend to vary more from one small space to another. Since we cannot measure the emission factors from each small region, applying averaged emission factors based on data from distant locations is likely to produce large uncertainties in the estimates. Over large areas the average emission factors are more reliable as deviations tend to cancel out. For very large scales such as over a hemisphere or the global scale, there are potential constraints that can be used to check the accuracy of emission inventories. For instance, observations of excess concentration seen in the northern hemisphere compared with the south puts limits on how much emissions may be emitted from the north (Khalil and Rasmussen, 1981). In some cases, such as for methane or nitrous oxide, the total global annual emissions can be constrained by estimating how many tons of these gases are destroyed annually by known atmospheric or terrestrial processes. This has to match the amount we emit and the change of concentration that is observed on the global scale. Such constraints are not directly applicable to smaller spatial scales such as a state like Oregon, but they suggest ways to achieve the same end. In recent times, satellite observations of greenhouse gases show promise for future validation studies for small sized regions.

Although there is no substitute for an emission inventory, there are methods by which we can verify both whether our estimate is accurate and whether our regulations are causing the desired reduction in greenhouse gas emissions. Increasingly this validation is done by direct measurements of greenhouse gases in or near large sources such as urban areas. Existing or new techniques will have to be included in policies to control greenhouse gas emissions from Oregon to validate whether the policies are working (National Academy of Sciences, 2010).

2.3.2 Case Studies

In Oregon two studies dealing with greenhouse gases have shown promise for validation of changes. The first was by Khalil and Rasmussen (2004) who reported evidence of substantial decreases in emissions of ozone depleting compounds from the Portland area and nearby regions, while greenhouse gases did not show a similar trend. In the second and ongoing study Rice and others are looking at CO2 concentrations and the 13C isotopic composition of CO2 from three regional monitoring stations in Multnomah County (Rice and Bostrum, 2010). The study takes advantage of seasonal prevailing winds from the Columbia Gorge to track the relative concentrations of CO2 between a rural site at the edge of Portland, a downtown site, and an urban residential site. Early results from summer and fall measurements indicate enhanced CO2 at the urban sites is almost entirely from automotive sources. Data collection continues and will provide evaluation of variations in urban CO2 throughout a full calendar year. Sources should vary with season, e.g., increased use of natural gas and oil for heating in winter months. Second, more extensive data collection and analysis of the carbon isotopic composition of CO2 in Portland should help differentiate between important CO2 sources and their temporal trends. Isotopes are a new and state of the science method for looking at the origins of greenhouse gases in various environments.

2.3.3 The Role of Aerosols

Atmospheric particles, either liquid or solid (aerosols), can play a significant role in climate change, particularly at regional scales of the size of Oregon. Atmospheric aerosols affect the Earth's surface temperature by scattering and absorbing radiation and altering cloud properties, such as how long clouds persist. The direct effect of aerosols is scattering and absorption of radiation. The indirect effects are typically separated into modification of cloud albedo (reflectivity) through a change in the number of cloud droplets per cubic meter of air, and droplet size for a fixed cloud liquid water content (Twomey, 1977), and the second indirect effect is modification of cloud liquid water content, cloud height, and cloud lifetime (Albrecht 1989). Further effects of aerosols have been proposed (Hansen, 1997; Koren, 2008; Isaksen et al., 2009)

Generally aerosols with significant fractions of dust or black carbon absorb the Sun's radiation resulting in positive radiative forcing that causes warming of the atmosphere and the surface, while aerosols with significant fractions of other constituents (sea salt, sulfates, nitrates, organics) are non-absorbing and cause a cooling of the surface. Non- or weakly absorbing aerosols dominate most latitudes (Ramanthan et al., 2001), and thus produce the overall net-cooling effect; however, absorbing aerosols may play an important role at high altitudes and at local scales (Charlson et al., 1992).

At present it is not known how aerosols are affecting Oregon's climate and to what extent their emissions in the future can mitigate the climatic change that may occur here. The prevailing conditions with substantial annual rainfall and many sources of organic aerosols suggest that the aerosol effects may be important for Oregon climate in the future.

2.4. The Challenge of Meeting Climate Targets

2.4.1 Future Emissions

2.4.1.1 Social drivers of future greenhouse gas emission levels

The impacts of climate change in Oregon will depend on the extent of global climate change. The extent of global climate change in turn will be determined by emissions of greenhouse gases and the interaction of those emissions with the natural climate system. What factors, then, will determine the amount of future greenhouse gas emissions emitted in Oregon? As discussed earlier, there are three elements that determine emissions (E), these are the emission factor (K), per capita consumption rate (C), and population (P). When emissions are considered for the entire state across economic sectors and activities we can use carbon intensity (MTCO2e per million dollars) as the emission factor K, and GSP per capita as the consumption rate C. We can then examine the patterns of these three drivers for Oregon in the past, present, and future. We consider three time periods for which we have economic and emission data, 1990-1999, 2000 - 2005, and 2010 - 2020. For the latter period we use forecasts of economic activity and population.

We look at how much of the percent change in emissions $\%\Delta E$ over each of these periods is driven by the percent changes in the three factors ($\%\Delta K$, $\%\Delta C$ and $\%\Delta P$) For example, we define the percent change for the first period as E(in 1999) – E(in 1990) divided by E(in 1990) and multiplied by 100% to convert to percent. The same is done for the other variables K, C and P and for the other time periods. Table 2.1 lists these changes for the three time periods.

%Changes of Variable	*1990-1999*	*2000-2005*	*2010-2020*
Population (Δ P)	16 %	6 %	13 %
GSP per capita (Δ C)	30 %	6 %	8 %
Carbon intensity (Δ K)	-17 %	-10 %	??
Emissions (ΔE)	25 %	1 %	??

Table 2.1 Changes in emissions and drivers. Values are percentage changes over the time period indicated.

Delineating the drivers of emissions in this fashion reveals why Oregon's emissions are changing and what to expect in the future. During the 1990s Oregon reduced the carbon intensity of its economy by 17%. In spite of this decrease, Oregon's emissions still grew by 25%. The reasons are clearly seen; population grew by 16% and consumption or gross state product per capita increased by 30% thus overwhelming the gains in carbon intensity. Thus despite a double digit decrease in the gross emission factor, emissions still rose by double digits.

During the early part of the last decade Oregon reduced its carbon intensity by 10%, but this reduction was again offset by increasing population and consumption. By the year 2020 Oregon's population is expected to increase by 13%. If we assume Oregon's economy will grow at ~2.5% per year, a rate that seems likely to continue once the current recession ends (http://www.census.gov/compendia/statab/2010/tables/10s0655.xls), then per capita GSP will increase ~8% by 2020. If the carbon intensity of the economy remains constant over the next period, then we should expect Oregon's emissions to increase by about 20%. This highlights the challenge Oregon faces to reduce its future emissions. Since most governments appear disinclined toward policies that curb population growth or economic growth, emissions will grow at a rate that is the product of population growth and growth in per capita income: the rate will continue to increase barring changes in available technologies or consumer and producer choices among those technologies. Past growth trends suggest this upward pressure on emissions will be about 20% per decade in Oregon. Given this, state policy-makers face the daunting task of developing polices that reduce carbon intensity (i.e., emissions per dollar of GSP) by about 2 to 3% per year just to avert increases in CO_2e emission levels. Generating decreases in emissions will require policies that reduce carbon intensity at rates greater than those 2% to 3% per year levels.

This analysis raises an important question about what metric Oregon should use to assess its future greenhouse gas emissions. Policies regulating emissions will likely try to reduce emission factors (K). For example, higher efficiency vehicles and increased production of renewable energy would lower the emission factor for the transportation and energy sectors respectively. Even if Oregon successfully reduces emission factors across sectors through mitigation policies, these reductions will be offset by population and economic growth. Of the population growth from 2000 to 2020, it is estimated that 63% will be due to net migration into the state. This added growth will put greater pressure on actions designed to reduce emission factors and may derail efforts to reduce total overall statewide emissions by target dates.

Oregon's per capita emissions have been constant or declining over the past decades. This suggests that Oregon may wish to explore the use of a modified per capita emissions metric to assess its emissions goal instead of total emissions. The basic idea is that the targets for the state should be set by managing the per capita emission rate rather than the total emissions. This way Oregon will not be punished for net migration into the state as other states are rewarded for net migration out. This new metric would use a modified form of population that would be updated annually only by net migration and not natural growth. Oregon's per capita emissions would be calculated using this modified population to normalize for net migration into or out of the state. For example 68% of Oregon's population growth by 2020 is expected to be from net migration into the state. Thus natural population growth by 2020 is only about 4%. Using the modified metric would require Oregon to reduce its carbon intensity by a more manageable 12% rather than 21% to keep emissions constant. If this metric were adopted by all states, then national inventory targets could still be met and states would have a fairer allocation of responsibility towards the success of their mitigation policies.

2.4.3 Policy Considerations

There are two ideas that we want to bring up in closing. The first is that policies to reduce greenhouse gas emissions at the state level and perhaps even on the national scale can be set as an average over several years to a decade with flexibility in year to year emissions. Or what amounts to the same thing, a target can be set for aggregate emissions over a 10 year period, with flexibility on how much is emitted in each of the years of the target period. As long as emissions can be maintained on average at target levels within the specified time period, which may be 5 year or 10 years, it would be acceptable to allow increases in emissions over some years to accommodate unexpected opportunities for economic or industrial growth. Climate change responds very slowly to changes of emissions and thus maintaining an average emission rate on target over a decade or perhaps an even longer period can produce nearly the same results as trying to maintain the same emissions every year. This is unlike the case for urban air pollution where violations of a standard on even one day can lead to harmful effects.

The second matter relates to the actions that Oregon can take given the complex and uneven responses from the rest of the states and other countries of the world. Even if policies in Oregon were to generate dramatic emissions reductions, these would not insure that Oregonians will get manageable or modest climate change in the future unless similar policies were adopted by people in other states and countries. Because both the emissions that cause climate change and their undesirable impacts are global in scale, the costs of action to reduce emissions will be borne by those taking such actions but the benefits of those actions will accrue to everyone. Similar situations have been called the "Tragedy of the Commons" (Hardin, 1968) and are known to create strong incentives for people to take less action than is necessary to remedy the problem. Indeed, arguments are likely to be made that Oregonians should not adopt policies that may inhibit local economic growth unless we can be assured that those policies will be matched by other governments around the world, thereby actually reducing overall global emissions and stemming the effects of climate change. The flexibility of being able to relax emissions for a few years at a time while maintaining emissions at the target on average can reduce the burden of taking action to reduce global warming. Whatever actions are taken in Oregon we should be clear that Oregon's efforts to reduce emissions of greenhouse gases will not directly and immediately reduce the impacts of global climate change that we will experience in Oregon or affect the costs of adapting to climatic change.

2.5 Key points

The main points we have discussed are the following:

- Higher population density can in some circumstances lower per capita greenhouse gas emissions. We have discussed this effect and related indices of how to put Oregon in context of global change.

- We argue that state goals should be set as limits on per capita emissions rather than total state-wide emissions. This would reward states that make reductions in emissions and attract economic growth rather than states where emissions may be reduced by migration or economic downturns.

- We have put Oregon greenhouse gas emissions in the context of other states and the world and argued that emissions from energy use should be evaluated by per capita consumption rather than the more readily available data on per capita production by state.

- State and country policies can set targets as emissions over a decade, or possibly longer, rather than for each year. This would allow flexibility in controlling emissions within the decadal period which could be relaxed over a few years because the state is seeing a growth of an important industry or has economic opportunities as long as the 10 year aggregate emissions goal is met. Since climate does not respond rapidly to year by year changes of emissions, there is no need for policy to limit emissions on an annual basis. This is unlike the case for air pollution where the impact is immediate and hence increases of emissions even above daily targets are not permitted.

- Direct validation of emissions reductions is necessary to know whether our policies are working or not. Hence validation has to be made an integral part of policies to control greenhouse gas emissions.

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3. Climate Change and Freshwater Resources in Oregon

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Summary and Knowledge Gaps

Climate change will affect various sectors of water resources in Oregon in the 21st century. The observed trends in streamflow show significant declines in September flow and, although not significant, increases in March flow in many transient rain-snow basins. These streamflow trends are associated with rising temperature and coincident declines in snowpack in spring in the latter half of the 20th century. While there are no distinct trends in high precipitation events, such events are associated with climate variability such as El Niño/Southern Oscillation (ENSO) and Pacific Decadal Oscillation (PDO). Effects of ENSO and PDO are more pronounced at the beginning and end of the wet season in the Willamette River basin.

The amount and seasonality of water supply is projected to shift as the distribution of precipitation changes and temperatures rise. Higher summer air temperatures accompanied by reduced precipitation are projected to increase evapotranspiration and decrease stream flow in summer. Although there are no distinct spatial patterns of changes in precipitation and temperature across the State in the 21st century (uniform increase in temperature across the region), significant regional variations do exist. The magnitude of change depends on the importance of snow in the current water budget, so projected changes are greater for mountainous regions than for low-elevation areas. Transient rain-snow basins, such as those in the Western Cascade basins, are projected to be more sensitive to these changes in precipitation and temperature. The high Cascade basins that are primarily fed by deep groundwater systems could sustain low flow during summer months. Basins in the east of the Cascades are projected to have low summer flow in a distant future as groundwater recharge declines over time. April 1 snow water equivalent (SWE) will decline and the center timing of runoff will become earlier in transient rain-snow basins as snowpack is projected to decline consistently in the 21st century.

These model projections should be viewed with caution for several reasons when considering climate change impacts on water supply in Oregon. First, this chapter shows that few consistent trends in runoff are apparent in streamflow records from Oregon; instead, the direction and magnitude of change in streamflow varies by season, by basin size, and among ecoregions in Oregon. Second, observed streamflow trends (e.g., declining flows in summer, or

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in September) may be explained by factors that may not be directly related to global climate change. For example, recent low-flow years are attributable to low precipitation years (especially 2001 and 2005) and perhaps to interannual variations in snowpack associated with cyclical variation in ocean temperatures, while long-term decreases in summer flows are attributable to the combination of summer precipitation decline and increasing water withdrawals for consumptive use. Third, model projections do not account for possible resilience and adaptations in natural ecosystems that may alter water use and lead to smaller than expected changes in streamflow. More work is needed on vegetation responses to climate variability, the interactions between soil water and vegetation, and the relationship between streamflow and precipitation (runoff ratios) in large basins.

Water quality is also projected to change with rising air temperature and seasonal shifts in flow availability. Water temperature is projected to rise as air temperature increases in the 21st century, particularly in urban streams where natural riparian vegetation is typically lacking. A decline in summer stream flow will exacerbate water temperature increases, because the low volume of water will be heated up more quickly than during times with larger instream flows. Changes in water temperature can have significant implications for stream ecology and salmon habitat in many Oregon streams. Lower order streams in transient rain-snow basins and in semi-arid eastern Oregon will be the most vulnerable to rising summer air temperature and diminished low flow. Sediment and phosphorus loads are expected to increase in winter as winter flow is projected to rise. Conservation-oriented urban development could potentially reduce storm runoff amount and subsequent sediment and pollutant loads, providing potential opportunity for local adaptation to climate change. At the basin scale, new dam or reservoir operation rules might be required to maintain environmental flows in summer. The complex interactions among changing hydrology, watershed biogeochemistry, and land management need further investigation.

As shown in the Portland water consumption case study (Section 3.6), when other demand factors are held constant, increases in temperature alone result in higher demands for peak season water. While demand during winter months is expected to remain constant, research on urban water demand suggests temperature-induced water consumption, particularly among single family residential (SFR) households. These impacts are also evident at multiple scales, including the household, neighborhood, and region. At the regional scale, urban land uses have different water demands, and will have varying impacts on water demand. Overall, single-family residential land use is the largest consumer of water. At the neighborhood scale, the density of urban development helps predict future water use, where higher density residential developments have lower per capita water demand. Finally, at the household scale, a coupling of structural attributes (e.g., building and lot area) and temperatures affect water demand. High-density housing developments with smaller homes could limit the growth of residential water demand relative to other water uses in the region.

Uncertainty is still high in projecting future changes in runoff, water quality, and urban water demand in Oregon. While the main source of uncertainty stems from the choice of global circulation models, additional sources of uncertainty include GHG emission scenarios, downscaling methods, hydrologic model structure and parameterization, and impact assessment methods. Multi-ensemble models that take into account all sources of uncertainty with different weights might provide a means of quantifying different sources of uncertainties. Communicating uncertainty to water resource decision makers is another challenge for adaptive water resource management in a changing climate. While a more sophisticated hydrologic impact assessment model is yet to be developed, climate adaptation strategies can be implemented at multiple spatial scales.

Land use planning may be helpful in meeting the future water needs of the State. Currently, land use and water resource management agencies have limited coordination in their responsibilities. Zoning and public involvement can be instrumental to improving the coordination between land and water management agencies. Zoning can be used to link types of future development that include a combination of infill, expansion, connecting existing developments, with explicit identification of water demands on different land uses in the region. To date, few plans have explicitly included dimensions of integrated land and water management. Outreach and education campaigns can help inform the public about the relationship between water demand and supply, but can also assist in adapting to a future with increasingly limited resources. The details of those plans and the precise nature of the outreach and education campaigns will require further investigation, and will likely be part of the second assessment of Oregon's water resources.

3.1 Introduction

The hydrology and water resources of Oregon are sensitive to changes in precipitation and temperature, but the rate of change varies across basins with different topographical, geological, and ecological characteristics. According to the Fourth Assessment Report (AR4), many Oregon streams will experience higher winter flow and reduced summer flows as temperature rises and the variability of precipitation increases. In addition, various human activities, especially land cover modification and dam or reservoir operations, have modified the hydrologic regime of many Oregon streams since the late 1800s. Understanding the complex interactions among climate systems, terrestrial systems, and human systems is essential to predicting future changes in water resources and implementing sustainable water resource management in Oregon.

For this first statewide climate impact assessment, we have both initiated some new research studies using downscaled climate change simulations, and compiled existing relevant studies, putting them into the context of climate change impact assessment. While most studies rely on empirical statistical data analysis using observed data, some case studies use downscaled global circulation model results combined with hydrologic simulation models for climate change impact assessment (e.g., Graves and Chang, 2007; Franczyk and Chang, 2009; Chang et al., 2010a; Praskievicz and Chang, 2011; Chang and Jung, 2010; Jung et al., in review). Others use synthetic climate change scenarios (e.g., Tague et al., 2008 and Tague and Gran,t 2009). Based on these case studies and the best available information, we attempt to assess the current status of Oregon water resources and identify emerging water issues under the stress of climate change.

This chapter is composed of six main sections. Section 3.2 assesses observed variability and trends in various components of hydrology (e.g., snow water equivalent, glacier mass balance, extreme hydrologic events) in selected Oregon river basins. Section 3.2 assesses future changes in surface water hydrology including spatial and temporal variations of runoff, snow water equivalent and uncertainty in projecting future runoff. Section 3.3 describes future projections of surface water, methods of downscaling for hydrologic impact assessment, trends in future precipitation and temperature in the 21st century, and uncertainty associated with climate impact assessments. Section 3.4 examines potential changes in groundwater systems and their contribution to streamflow under future climate change scenarios. Section 3.5 investigates possible changes in water quality with a focus on water temperature and nutrients. Section 3.6 describes case studies of Portland and Hillsboro municipal water demand associated with climate variability. Section 3.7 discusses water infrastructure management, including urban water demand management and dam operation. The concluding section offers a concise summary of the main findings of this water resources impact assessment and discusses possible future research directions.

3.2. Observed Variability and Trends (Historic Perspective)

Streamflow in Oregon is highly variable in space, and over multiple time scales. It varies in space according to elevation, topography, geology, and basin area, and varies seasonally according to the amount of precipitation, relative proportions of rain and snow, topography, geology and vegetation. Streamflow also fluctuates on interannual time scales. Changes in snowpack accumulation and melt from climate warming are expected to influence streamflow, but these effects will be more pronounced where streamflow patterns are controlled by snowmelt. Glacial melt and retreat also may affect streamflow, but only in very small, high-elevation basins, and this effect will diminish as basin size increases. Streamflow also depends on the human-controlled factors of vegetation cover, urbanization, and river regulation (e.g. by dams); changes in these factors have significantly altered streamflow in the past century. Therefore, it is extremely challenging to disentangle long-term trends in streamflow from temporal variability. It is also easy to mistakenly attribute observed trends to climate, when they may be due to flow regulation (dams) or land use changes.

3.2.1 Annual and Seasonal Surface Flow and Variability

3.2.1.1 Water budget

The water budget indicates the potential mechanisms and magnitude of various hypothesized streamflow changes in response to climate variability. The conceptual water budget for Oregon watersheds involves multiple components, including precipitation, cloudwater interception, canopy evaporation, transpiration, snow storage, and snowmelt. Climate change and variability potentially affect all these components.

3.2.1.2 Spatial patterns of annual runoff

The spatial patterns of runoff in Oregon pose significant challenges for detecting climate change effects on historical streamflow. Most of Oregon is forested, strongly influencing runoff patterns through evapotranspiration, which may exceed 50% of precipitation. Precipitation is orographic, and highest in mountains and in western Oregon (see Chapter 1). Because precipitation is concentrated in mountainous areas and in western Oregon, large drainage basins in the eastern two-thirds of Oregon produce much less streamflow than the Willamette and coastal basins (see Figure 3.1 for major river basins in Oregon). Additionally, the lower Columbia, Willamette, and Oregon Coastal watersheds produce higher peak flows and water yields than the eastern Oregon watersheds, although they are partially covered by forests. The highest peakflows occur in southern and northern coastal Oregon, reflecting the high precipitation and steep drainages (see Figure 3.2).

Figure 3.1 Major river basins in Oregon

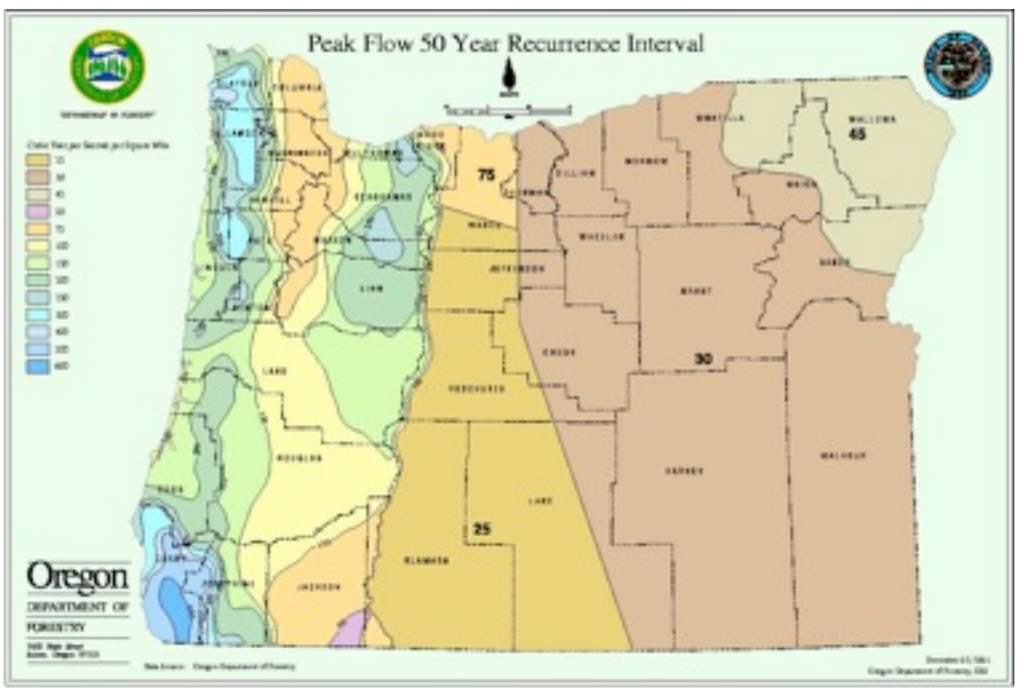

Figure 3.2 Peak flow 50 year recurrence interval (source: http://www.odf.state.or.us/atlas/maps/peakfl75.jpg).

The seasonal cycle of the Columbia River discharge has already been modified significantly by major dams and deliberate management: peak discharge formerly occurred in late spring, but now occurs in autumn (Sherwood et al., 1990). The annual average discharge shows large interannual variability and some interdecadal variability, but no significant long-term trend between 1928 and 2009 (Figure 3.3). In contrast, the average May-through-July discharge has decreased by about 30%; most of this decrease occurred between 1950 and 1990, as a result of management for flood control, irrigation and hydroelectric power. In recent years, concern for salmon smolt survival has led to increased spillage over the dams in spring and early summer; if this concern continues to prevail, the summer discharge might recover or at least stabilize.

Most populated areas occur in the lower reaches of large basins, but upstream dams and land use regulate streamflow at downstream gages. Basins above dams provide records that are unaffected by dams, but these watersheds are mostly in areas of low population density, and streamflow in these basins has been affected by forest harvest and other land use changes over the past century. The construction of dams for flood control and irrigation in the middle part of the 20th century throughout much of western Oregon greatly diminished peak discharges and altered the seasonal pattern of discharge in large basins, such as the Willamette River. A hydrologic simulation model of the natural flow regime (Figure 3.4) illustrates the impact of dams between 1977 and 2008 for three stations in the Willamette River: late summer flow is augmented by water released from dams.

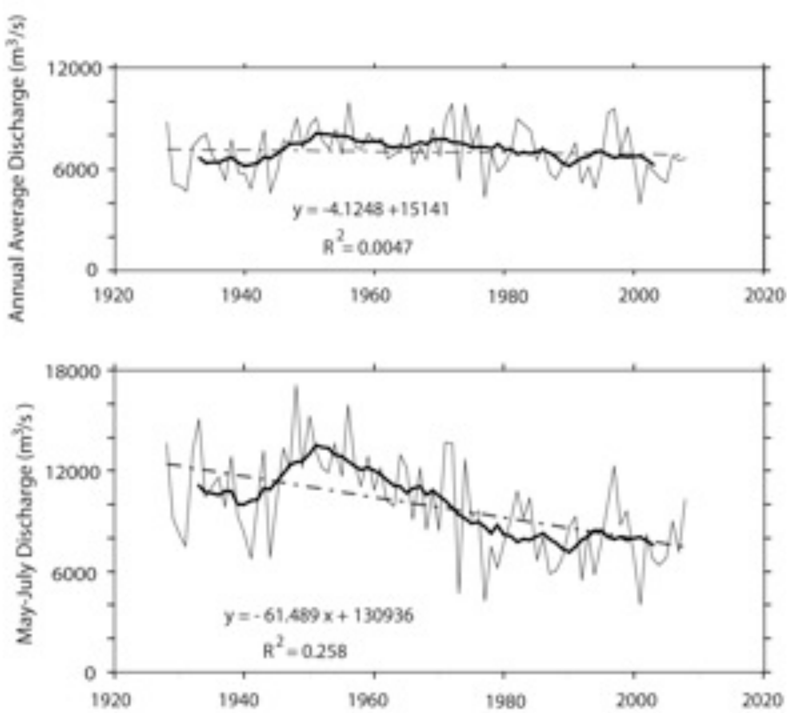

Figure 3.3 Annual average and May – July discharge in the lower Columbia River, 1928-2009

Forest harvest significantly and persistently increased winter and spring water yields in small watersheds of western Oregon (Jones and Post, 2004), and also altered peak discharges of at least small peaks, and (arguably) large peaks in small and intermediate basins (Jones and Grant, 1996, Thomas and Megahan, 1998, Beschta et al., 2000, Jones ,2000, Grant et al., 2008).

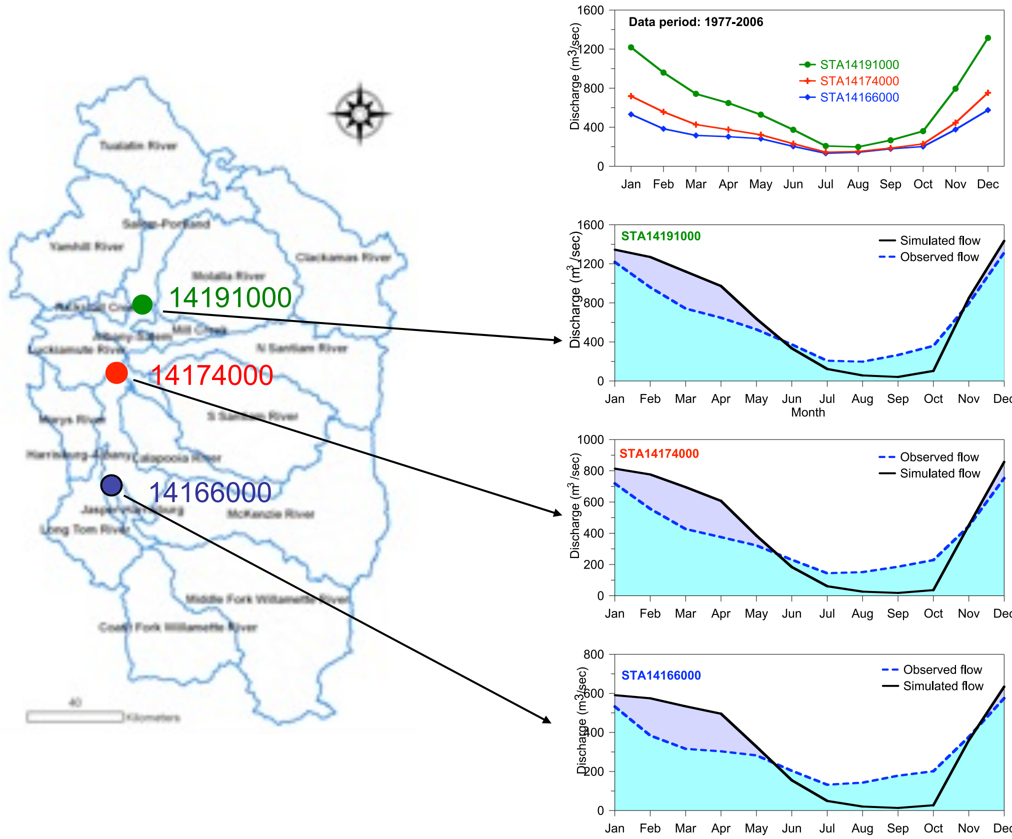

Figure 3.4 Comparison of the observed flow and the PRMS simulated flow for the three monitoring stations in the Willamette River (Chang et al., 2009)

3.2.1.3 Temporal patterns of streamflow

Seasonal patterns of runoff vary across Oregon depending on precipitation type (rain vs. snow), basin size, topography, and geology. Runoff in Oregon is strongly seasonal: over 75% of streamflow occurs in the six months of October to April (Willamette River, John Day River, mean monthly discharge, Oct to Sep). In small basins on highly weathered old volcanic rocks in western Oregon streamflow varies even more by season. In contrast, streamflow from basins on recent, porous lavas of the High Cascades (e.g., Clear Creek) have low seasonal variability because deep groundwater augments summer low flows (Tague et al., 2008; Chang and Jung, 2010). In contrast, flow in the western Cascades (e.g., Lookout Creek), primarily fed by shallow subsurface flow, diminishes rapidly during dry summer season.

Figure 3.5 Comparison of late summer streamflow in Clear Creek (groundwater fed) and Lookout Creek (shallow subsurface-fed). Photo credit: Chang.

Figure 3.6 illustrates monthly hydrographs for six representative basins in Oregon. They are located in different hydrologic and ecoregions, which reflect different climate and vegetation regimes. Basins A (coastal basin) and B (Willamette Valley) are primarily fed by rainfall, while flow in basin C (Hood River) is contributed by a mix of rain and snowfall, and basins east of the Cascade Range (D,E, and F) are fed by snowmelt (Fig. 7a). Basins A and B have a rainfall-dominated peak in December, basin C has a rainfall-dominated peak in December and a snowmelt-dominated peak in April, and basins D, E, and F have a single snowmelt-dominated peak in late winter and spring (Figure 3.7b). Total annual runoff amounts in basins in eastern Oregon, which received much less precipitation, are much smaller than those in the Valley or coastal areas. Geology also controls the timing and amount of runoff in the Deschutes basin (Figure 3.7b-d).

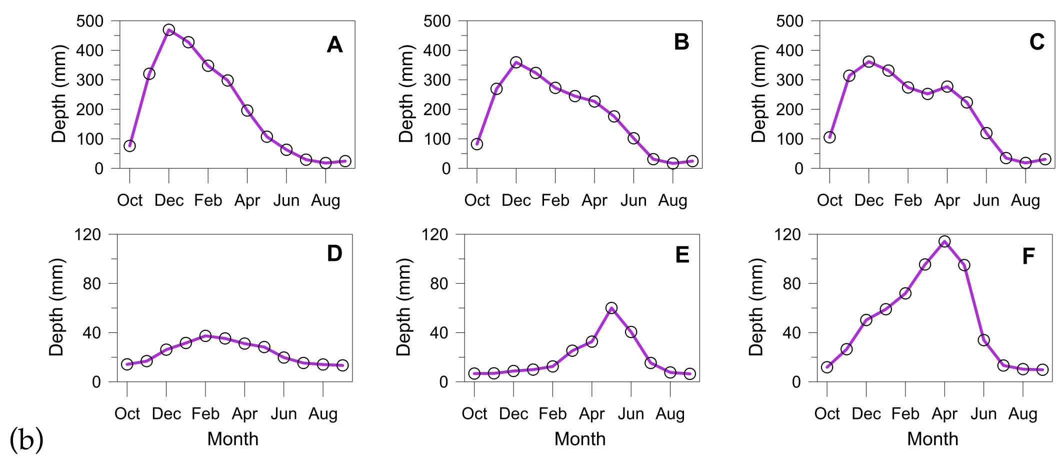

Figure 3.6 Monthly mean runoff for annual total runoff and the ratio of summer flow to annual flow (Source: Chang et al. in preparation). A = Wilson River near Tillamook; B = Little North Santiam River near Mehama; C = Blazed Alder Creek near Rhododendron; D = Warm Springs River near Kahneeta Hot Springs; E = Donner und Blitzen River near French Glen; F = Umatilla River above Meacham Creek near Gibbon.

3.2.2 Trends in annual and seasonal flow

Observed interannual trends in annual discharge in very large basins can be seen from 100-year records at the Willamette River (Salem) vs. John Day (McDonald). The lowest streamflow in 100 years of record was 1977 on both the west and east sides of the Oregon Cascades (Figure 3.7). The wettest years were in the early 1970s on the west side, and early 1980s (ENSO) on the east side. On the west side, 2001 and 2005 were among the six lowest-ranked streamflow years, but these were not unusually lowflow years on the east side.

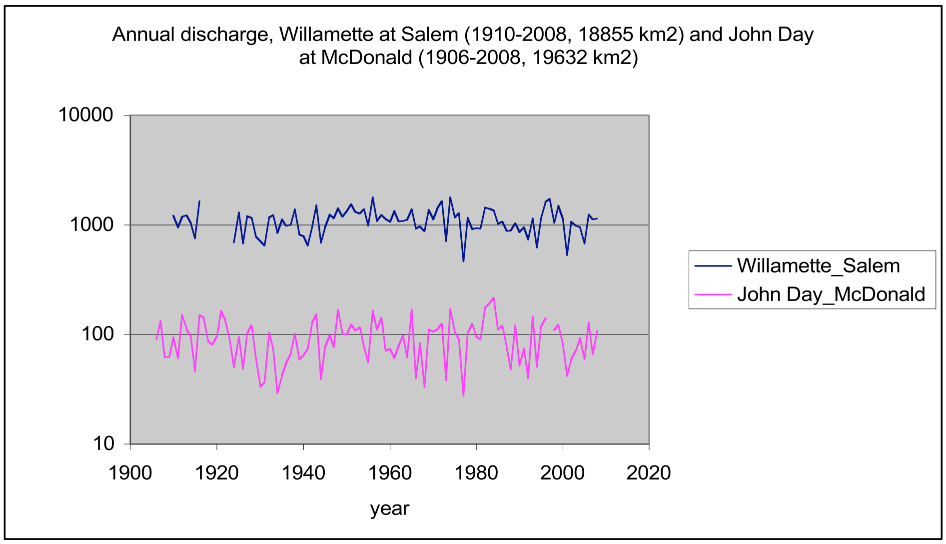

Figure 3.7 Annual discharge 1906 - 2008 on the west side (Willamette) and east side (John Day) of the Cascade Range in Oregon

Lins and Slack (1999) found decreases in streamflow in the Pacific Northwest streams, particularly in low flow regimes during the 20th century. Subsequent studies in the PNW also show declining streamflow trends (Hamlet et al., 2007; Stewart et al., 2004, 2005; Barnett et al., 2008). Similarly, Luce and Holden (2009) found significant decreases in the magnitudes of the lowest 25% of streamflow years over the period 1948-2005 in Idaho, Washington, and Oregon, and speculate that on the east side of the Cascades these declines may be due to declining precipitation. However, precipitation is not declining in the central western Cascades of Oregon (Jones, unpublished data from the Andrews Forest, and PRISM maps/data from C. Daly). More work is needed to relate streamflow trends to precipitation in large basins.

Warming air temperatures are expected to shorten snowpack duration and speed snowmelt timing, resulting in earlier peak annual streamflow. Based on a study of the western United States, Stewart et al., (2004, 2005) found that peak streamflow timing now comes one to four weeks earlier than it did in the the middle of the 20th century, and attribute this change to earlier spring snowmelt. However, the temporal center of mass of snowmelt-dominated streams in Oregon historically occurs in March, whereas the western basins most affected by warming are those with peak streamflow in April to June. Oregon streams in this study mostly experienced shifts of <10 days, and most of these streams were in eastern and southeastern Oregon (Stewart et al., 2004, 2005).

To assess climate variability influences on streamflow in Oregon, we selected USGS and Oregon Department of Water Resources stream gauging stations that have more than 30 years of record and have not been affected by upstream dams or significant diversions. Thirty stations in Oregon meet such criteria and are analyzed for trends; 21 were analyzed for years 1958 - 2008, and 9 with shorter records were analyzed for years 1975 - 2008. The Mann-Kendall's test was used to detect the direction and significance of trend in each station. While summer flow declined in over two thirds of the stations during the study period, spring flow increased in one third of the study stations. Twenty-five stations exhibit declining trends in mean annual flow, while only 4 of the 25 stations show significant trends (Table 3.1) (Chang et al. in preparation). September flow declined significantly at most of the studied stations, while March flow increased significantly for only two stations (see Figure 3.8). Decreasing September precipitation appears to be responsible for the declines in September flow.

Table 3.1. The number of positive and negative trend stations for 30 stream gauging stations (21 with period 1958-2008, 9 with period 1975-2008). Numbers in parenthesis show statistically significant trend stations ($P < 0.05$).

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
+	9 (0)	1 (0)	14 (1)	17 (0)	7 (0)	11 (0)	6 (0)	4 (0)	3 (0)	4 (0)	7 (0)	13 (0)	5 (0)
–	21 (1)	29 (2)	16 (2)	13 (3)	23 (5)	19 (1)	24 (2)	26 (9)	27 (16)	26 (3)	23 (2)	17 (2)	25 (4)

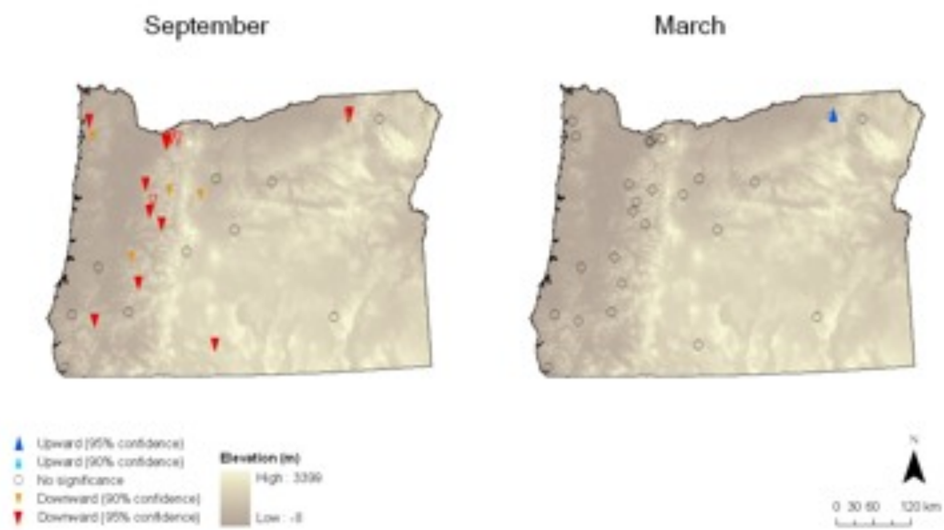

Figure 3.8 Trends in average runoff for March and September for 30 stream gauging stations. Numbers in parenthesis show statistically significant trend stations ($P < 0.05$).

The Umatilla River above Meacham Creek (USGS station number 14020000) illustrates a case of increasing flows in early spring, and declining flows in September over the period 1958 and 2007 (see Figure 3.9). The increase in March streamflow may be due to earlier spring snowmelt. The interannual variability of September flow also declined during the study period.

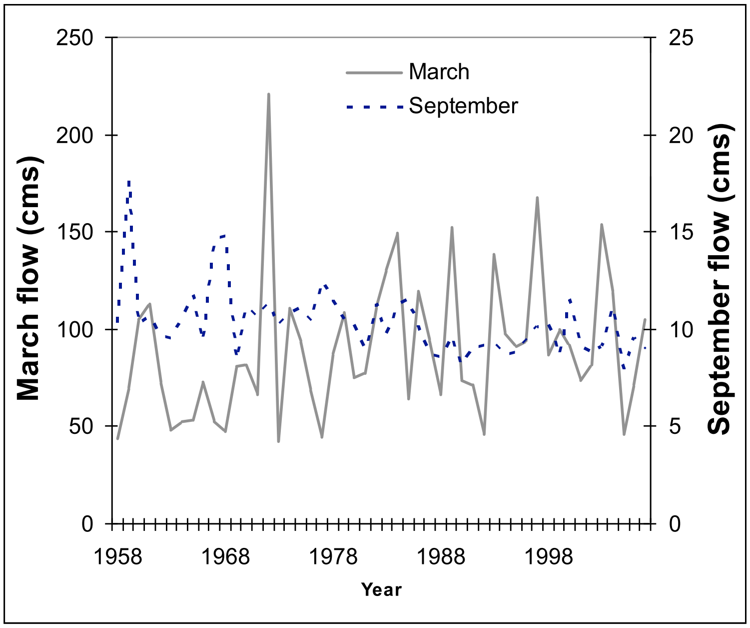

Figure 3.9 Trends in March and September flow for Umatilla River above Meacham Creek, near Gibbon

In the Upper Klamath River basin, dry season (April to September) and summer streamflow (July to September) declined 16%, and 38%, respectively during the period between 1961-2009 (Mayer and Naman 2010). This decline is closely associated with decline in April 1st snowpack, which decreased approximately 40% during the same study period for snowcourse sites located below 1820 m elevation.

Streamflow trends vary according to the underlying geology and the importance of snow in the annual hydrograph. In the Cascade Range of western Oregon, Jefferson et al., (2006) found that relative streamflow in August (i.e., August streamflow as a proportion of annual flow) decreased significantly over the past century in two snow-dominated basins, but not in two rain-dominated basins. Basins draining the High Cascades (Clear Lake and McKenzie River) experienced significant declines in August streamflow from the early 1920s (McKenzie River) or early 1950s (Clear Lake). However, basins draining the highly-weathered western Cascades (South Santiam, Smith River) did not experience declines in relative streamflow in August over similar periods of record.

In small, undisturbed forested basins, runoff ratios and baseflow have declined significantly during spring, but they have not changed during summer or winter in the Andrews Experimental Forest in the Willamette basin, over the period 1952 - 2006 (Moore, 2010: Figure

3.10). These patterns suggest that declining spring streamflow is explained by increasing air temperatures and corresponding declines in snowpack accumulation and spring melt, as well as increased evapotranspiration from increased spring air temperatures. Corresponding increases in winter rain have not produced detectable increases in winter runoff in these small basins, either because the increase in rain is relatively small compared to interannual variability, or because warming temperatures have increased photosynthesis and transpiration in winter, mitigating any effect of increased ratio of winter rain to snow. Declining spring discharges also were not associated with declining summer discharges, either because the decline in spring runoff is not sufficiently large to influence summer soil moisture storage and runoff, or because dominant conifer trees are adapted to intra- and interannual variations in moisture availability and adjust transpiration accordingly (Moore, 2010).

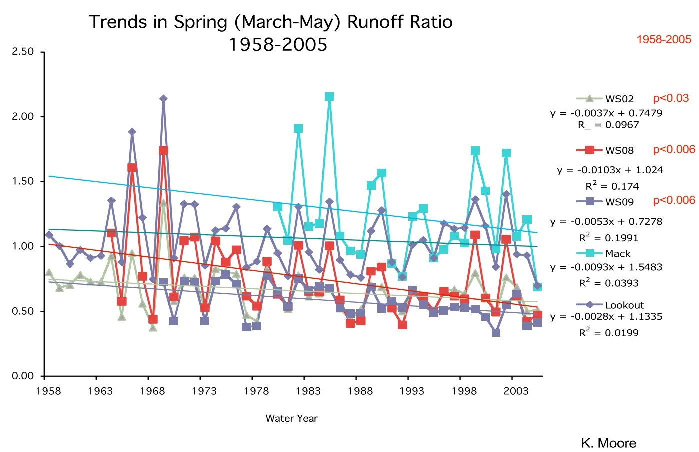

Figure 3.10 Declining spring runoff ratios from small, forested reference basins in the Andrews Forest, western Oregon.

In the Portland metropolitan area, there are no significant trends in annual mean flow between 1950 and 2000 regardless of urban development during the study period, suggesting that shift in climate regime may have masked the urban influence on hydrology, although urban streams show the flashiness and dryness (Chang 2007).

Overall, despite apparent increases in spring air temperatures and corresponding decreases in snowpacks, few consistent trends are apparent in long-term streamflow records.

3.2.3 Trend in Snow Water Equivalences

The timing of streamflow depends on snowpack size and the timing of melt in much of the western US, including many parts of Oregon. Annual precipitation in western Oregon is high (above 2500 mm in mountainous areas), but 70 - 80% of this precipitation occurs in winter (November to April). Hence, summer streamflows are dependent upon snowmelt. Therefore, climate warming effects on snowpacks may reduce streamflow during spring and summer periods, when water yield is limited.

Analysis of historic data show that warmer temperatures at higher elevations result in a shift in the form of precipitation toward more rain and less snow. Significant declines in snow water equivalent (SWE) in the Pacific Northwest and a shift from snow to rain coinciding with increases in temperature since the 1950s are well documented (Mote, 2003; Mote et al., 2005; Knowles et al., 2006), and this change has been related to trends in hydrologic response (Mayer and Naman, 2009).

Throughout the intermountain West, current analyses of projected climate change impacts predict that rising temperatures will diminish snowpacks, and these studies predict future summer water shortages (Folland et al., 2001; Service, 2004). Knowles and Cayan (2002) predict that the April to July fraction of total annual flow will be reduced by 30% in the Sierras by 2060 as a result of reduced snow accumulation and earlier melt. More recent climate simulations taking different greenhouse gas emission pathways into account predict future snowpack reductions of 30 – 90% (Hayhoe et al., 2004).

Snowpacks in the Pacific Northwest are expected to be particularly sensitive to warming. Climate models predict continued winter warming of 0.2 to 0.6°C per decade in the Pacific Northwest (Mote and Salathé 2010), and Cascade snowpacks are projected to be less than half of what they are today by 2050 (Leung **et al., 2**004). Lower elevations of the Cascade Ranges, for example, are predicted to exhibit the greatest differences in the timing and magnitude of snowmelt (Hayhoe et. al., 2004; Payne et. al, 2004). Because snow in much of the Cascades accumulates close to the melting point, future warming would mean that large areas could shift from a snow-dominated to a rain-dominated winter precipitation regime (Nolin and Daly, 2006), potentially increasing winter peak flows and reducing summer low flows as discussed above in Section 3.2.2.

In response to temperature increases in the Pacific Northwest (Cayan et al., 2001; Regonda et al., 2005), snowpacks in western North America have declined over the past 50 years (Mote et al., 2005). Using measurements of April 1 snow water equivalent (SWE) dating back to 1950, Mote et al. (2005) noted that the Pacific Northwest has experienced the largest declines in snowpacks in the western United States. A similar decline in April 1 snow water equivalent has been identified in the Clackamas River basin of Oregon between 1948 and 2000 (Graves and Chang, 2007). This change can be primarily attributed to an increase in winter temperatures (Mote, 2003; Mote et al. 2005; Barnett et al. 2008).

Some large portions of the mountains of Oregon may lose their snowpack, converting the hydrograph from a snowmelt to a rain-dominated pattern. Knowles et al. (2006) documented a significant trend towards increased rainfall and decreased snowfall (corrected for changes in precipitation) over the western United States from 1949-2004. The Pacific Northwest demonstrated a strong connection between Pacific Decadal Oscillation and temperature for days on which precipitation occurred. However, longer-term temperature trends also appear to be responsible for the shift from snowfall to rainfall. Most watersheds on the western slope of the Oregon Cascades encompass elevations that receive winter precipitation as a mixture of rain and snow. These watersheds have complex winter hydrographs that are dependent on the distribution of rain and snow during individual events, which in turn is controlled by storm temperatures and catchment hypsometry. Snow cover typically accumulates at temperatures close to the melting point, and thus is at risk from climate warming because temperature affects both the rate of snowmelt and the phase of precipitation. With a projected 2°C winter warming by mid-century, 9200 km^2 of currently snow-covered area in the Pacific Northwest would receive winter rainfall instead (Nolin and Daly, 2006).

Regional climate models predict that Pacific Northwest summers will become hotter and drier over the next century (Christensen et al., 2007), exacerbating existing stresses. Tague et al. (2008) used a hydro-ecological model named RHESS to examine the influence of geology on Cascadian streamflow response to warming scenarios. Their model showed that warmer temperatures resulted in greater reductions in August discharge and annual minimum flows for the High Cascades than the Western Cascade watershed, both in terms of absolute volumes and normalized by drainage area. The Western Cascade streams, however, showed greater relative reductions in these summer streamflow metrics. Model results illustrate that differences between the responses of the two sites were primarily due to differences in groundwater flow, as manifested in drainage efficiency of the watersheds. Spatial differences in recharge characteristics and the timing of snow accumulation and melt were shown to be important, but secondary, in terms of explaining responses at the two sites.

3.2.4 Trend in Glacier Mass Balance

Glacier runoff contributions to streamflow provide critical water supply in many mountainous regions (e.g. Singh and Singh, 2001; Barnett et al., 2005). Historical records and future climate projections point to the loss of midlatitude glaciers throughout the world (Oerlemans, 2005; Lemke, 2007), resulting in significant changes to both total annual and summer streamflow downstream (Chen and Ohmura, 1990; Barnett et al., 2005; Hock et al., 2005; Juen et al., 2007). Glacier runoff supplies fresh water to numerous communities in throughout the world and is highly sensitive to changes in temperature (Chen and Ohmura, 1990). Warmer temperatures cause increased glacial melt but as glaciers recede, their potential contributions to water supplies are diminished (Barnett et al., 2005; Hock et al., 2005). Glaciers also moderate intra- and inter-annual flow variability by storing water in the form of ice during years of high precipitation and releasing melt water during seasons and years of high temperature (Fountain and Tangborn, 1985). The hydrologic properties of glaciated watersheds differ from glacier-free watersheds in several ways. Glaciers release an estimated two to ten times more water than neighboring catchments of equal area and altitudes in the United States (Mayo, 1984). Runoff variability in glaciated watersheds is controlled primarily by surface energy fluxes whereas runoff variability in glacier-free watersheds is dominated by precipitation patterns (Jansson et al., 2003). There is a lag effect caused by glacial storage and the delayed networking of englacial and subglacial conduits (Jansson et al., 2003) such that runoff from glacier melt is delayed until later in the summer, when other contributions to streamflow are much reduced. Glacier melt decreases streamflow variation, bolsters late season runoff, and is especially important in drought years (Fountain and Tangborn, 1985). Under negative mass balance conditions, glaciers discharge a greater volume of water than is input in the form of precipitation and this “excess discharge” can be substantial, even for watersheds having less than 15% glacier coverage (Lambrecht and Mayer, 2009).

In the northwestern United States, glaciers diminished throughout the 20th century and model simulations suggest this trend will continue through the next 100 years (Dyurgerov and Meier, 2000; Hall and Fagre, 2003). Recent studies document that Mount Hood's glaciers have decreased in length as much as 61% over the past century (Lillquist and Walker, 2006). Coe Glacier has diminished at a rate 27% slower than that of the Eliot in the last century (Jackson, 2007), and we estimate that by about 2057 its area will be about 61% of its present day area. On a regional basis, temperatures are expected to increase by a range of 1.1 – 6.4°C in the next 100 years (Lemke et al., 2007). Nolin et al. (in review) showed that for the Upper Middle Fork Hood River, 74% of late summer streamflow is derived directly from glacier melt, most of which goes to irrigation of high value crops in the Hood River Valley. Their model simulations indicate that, while increased temperature leads to more rapid glacier melt and therefore increased streamflow, glacier recession ultimately overcomes this effect, leading to substantial declines in streamflow. These results show that the disappearance of Mount Hood's glaciers will likely result in the loss of about 27% of total late summer discharge in the Upper Middle Fork Hood River.

Glaciers in Oregon, like much of the west (e.g., Nylen, 2004; Hoffman et al., 2007) have been receding since the start of the last century when observations first began (Lillquist and Walker, 2006; Jackson and Fountain, 2007). The glaciers rapidly retreated since about 1910, slowed and advanced during the 1960s to middle 1970s before retreating again in the early 1980s. Since the late 1990s glacier retreat has accelerated. Between 1900 and 2004, the glaciers in Oregon have lost about 40% of their area. Some glaciers have lost as much as 60%. No glaciers are advancing in Oregon. (details about glacier change in Oregon: http://glaciers.research.pdx.edu/states/oregon.php)

Generally speaking, glaciers respond to variations in snow accumulation which nourish the glacier and to variations in summer air temperatures which cause melt. No long term trends in precipitation exist but summer air temperatures have been warming. Consequently, the shrinkage of glaciers in Oregon is due to warming air temperatures (Jackson and Fountain, 2007). This supports other work regarding the thinning of seasonal snowpack in Oregon (Mote, et al., 2005; Nolin and Daly, 2006). We expect that as the climate continues to warm, the glaciers will continue to recede.

Glaciers are locally important contributors to water supplies, and their contributions are important for augmenting summer lowflows. However, the area of glacial cover is very small, and the proportion of total water yield in Oregon that originates from glaciers is extremely small.

3.2.5 Trend in Extreme Hydrologic Events

In an analysis of climate change impacts for the State of Washington, Rosenberg et al. (2009a) found that peak flows and total annual precipitation have decreased over time while the magnitude of large, low frequency events of all durations has increased in some areas. In this study, trends in high flow (top 5% daily flow) and low flows (low 5% daily flow) were examined for the same 30 gaging stations in Oregon that were used in section 3.2.1. There are no significant trends in high flow for most of the stations examined (not shown). The average of the driest 5% of years, however, decreased at 25 sites, and 12 of these trends were significant (see Figure 3.11). Seventeen out of the 21 sites with longer records showed decreases in the average of the driest 5% of years, and the M-K test showed that 9 of these negative trends were significant. Most of the stations that exhibit significant negative trends are located in high elevations, suggesting that diminishing spring snow covers and consequent low summer flows may explain the declines in low flows in those stations. However, increased water use by young forest plantations, which were established during the period of streamflow record in these basins, also may be a factor (e.g., Perry, 2007). Only two stations show a significant increase in low flow.

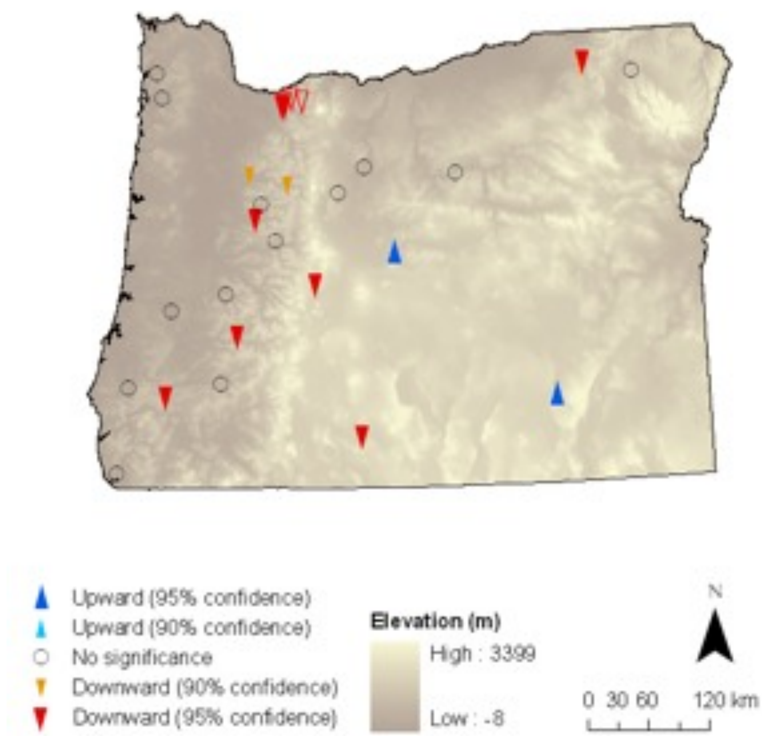

Figure 3.11 Trends in the average flow for the driest 5% of years, with significance determined by the M-K test (Source: Chang et al., in preparation).

The largest peak flow events in Oregon are produced by rain-on-snow events, when warm rain and winds contribute to rapid snowmelt (Harr, 1981, 1986). The Cascade Range of Oregon produces the highest 1% of floods on record in >500 km^2 basins in the United States, because large storms produce sustained rainfall and sometimes snowmelt for multiple days over broad areas of mountain ranges (O'Connor and Costa, 2004). Changes in forest cover significantly and persistently increase peak discharges in forested basins upstream of dams in western Oregon, especially for small events (<1-yr return intervals), but also, arguably, of large events (>1-yr return periods) (Jones and Grant, 1996; Thomas and Megahan 1998; Jones, 2000; Beschta et al., 2000, Grant et al., 2008).

If extreme rain-on-snow events are sensitive to the area of simultaneous snowmelt, climate warming could have a range of effects on extreme floods. Hamlet and Lettenmaier (2007) speculated that a climate warming-induced reduction in snow-covered area could reduce flood risk, but an increase in the effective basin area contributing to runoff from rainfall could increase flood risk. Climate-warming effects on extreme rain-on-snow floods are likely to depend on changes in atmospheric circulation and air mass behavior. Extreme rain-on-snow events occur when a rare sequence of marine polar air masses is followed by marine tropical air masses, creating simultaneous melt in large snow-covered areas and producing large effective contributing areas and extreme floods.

The probability of the sequence of events leading to an extreme rain-on-snow flood is already very low, and will only be affected by climate change if climate change alters (1) the occurrence of widespread snowpacks or (2) the energy of warm, wet tropical air masses.

3.2.6 Relation Between Climate Variability (ENSO & PDO) and Hydrology

This seasonal variability of heavy rainfall has implications for the quantity and quality of water resources in the Willamette River Valley.

In the western Cascades of Oregon, winter air temperatures, April snowpack, and winter streamflow are strongly related to the Pacific Decadal Oscillation. Over the period 1958 - 2007, in years with positive PDO (warm ocean temperatures near the coast of Oregon), air temperatures were significantly higher than average, snowpacks were significantly lower than average, and winter streamflow was significantly lower than average (Jones, in preparation).

These relationships—lower than average December and April streamflow in years with warm ocean conditions (WP/EP) – also are apparent in streamflow from basins in the Willamette Valley (Figure 13). The relation between wintertime precipitation intensity, as measured by both simple intensity and number of heavy precipitation days per year, and climate variability as measured by different phases of ENSO and PDO, shows some mixed results for the eight stations in the Willamette basin between 1972 and 2006. While the relation between ENSO phase and precipitation intensity is generally negative in November and positive in April, the relation between PDO and intensity is generally negative and strongest in January and March. These varying seasonal associations with ENSO/PDO phase may be associated with the Willamette Valley's location in the transitional zone between positive and negative El Niño response and to the moderating effects of out-of-phase ENSO/PDO (Praskievicz and Chang, 2009a). Figure 3.12 , illustrates the relation between different phases of ENSO/PDO and streamflow variability in April and December for four stations (A, B, D, F) shown in previous Figure 3.6. As shown in this Figure, December streamflow is high during the cool phase of PDO and La Niña years. In April, three sites show lowest median flow during the warm phase of PDO and El Niño years.

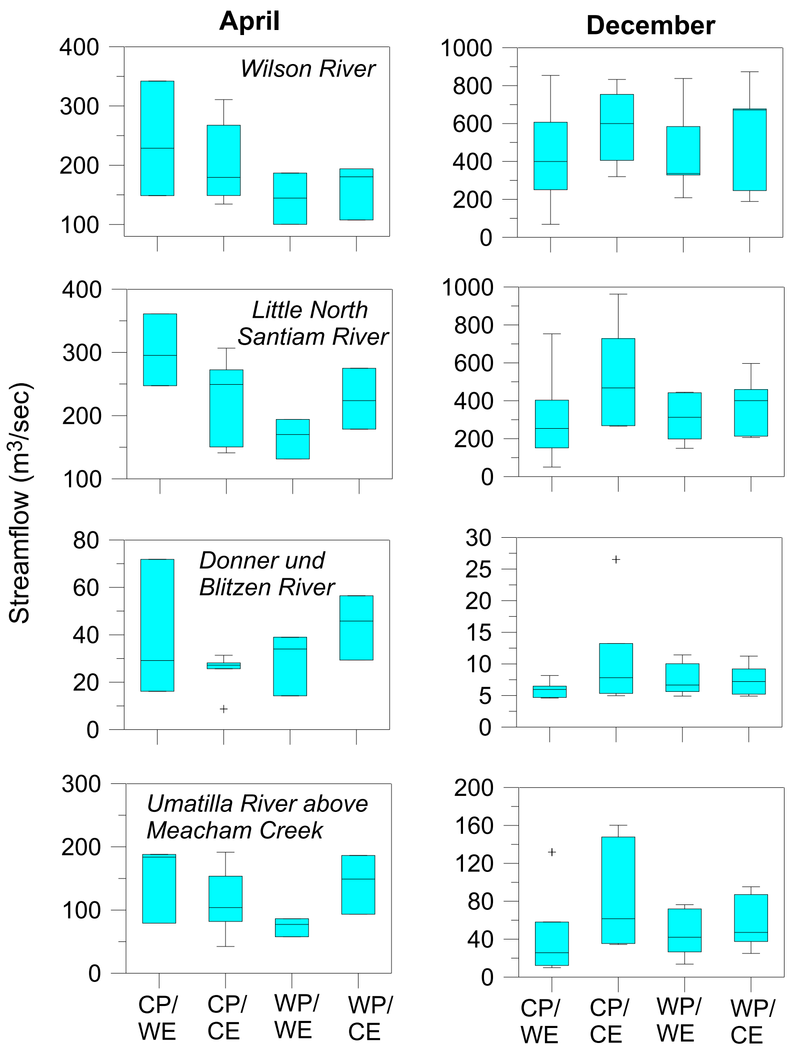

Figure 3.12 Differences in April and December streamflow among four combinations of warm and cold PDO/ ENSO phases for 4 USGS stations. Box and whiskers represent all years in category. CP/WE = cold PDO and warm ENSO; CP/CE = cold PDO and cold ENSO; WP/WE = warm PDO and warm ENSO; WP/CE = warm PDO and cold ENSO.

3.2.6 Climate Variability and Water Resources

The overall effect of climate variability on water resources in Oregon depends on hydrologic mechanisms operating at multiple spatial and temporal scales (snow water storage and melt, evapotranspiration). Three aspects of Oregon geography and hydrology will critically determine whether climate change effects exceed interannual variability of climate: (1) the extent of basin area affected by changes in snow water storage and snowmelt, which is associated with basin size, topography, and geology, (2) ecosystem adaptation and resilience to climate variability and trends, and (3) the relative magnitude and rate of climate-induced changes compared to historical effects of anthropogenic activities, such as dams and land use change on the magnitude and timing of streamflow.

In general, we expect the following.

- Climate change effects on streamflow will be largest close to melting glaciers or in seasonal snow zones, and decline in increasingly large basins as the snow-affected zone decreases as a proportion of contributing area.

- Climate change effects may be mitigated by ecosystem adaptations to climate variability, such as increased water uptake by vegetation during winter, which could offset predicted increases in rain:snow and winter discharge, and decreased water uptake by vegetation during summer, which could offset predicted declines in summer discharge.

- Historical effects of land use change (forest harvest, forest expansion after fire suppression) and dam management (winter water storage and summer releases) may be larger than as-yet-observed streamflow responses to climate change.

These issues will be discussed in more detail in section 3.3.

3.3. Projected Future Changes in Surface Water Hydrology

Future changes in surface water hydrology will depend on a range of factors. Hydrologic and climate models have been used to explore a range of possible outcomes from expected climate changes. Most of the model efforts have focused on the first three of these four hypothesized mechanisms for streamflow response to climate change.

- Increased air temperatures lead to decreases in the ratio of snow to rain, which decreases snow water equivalent (water stored in snowpacks), which decreases the snowmelt contribution to runoff in the spring;

- Decreased spring runoff carries over into summer, leading to decreased summer streamflows;

- Increased air temperatures decrease the ratio of snow to rain, which increases winter streamflow; and

- Increased air temperatures increase evapotranspiration and decrease spring and summer streamflow.

The overall effect of these mechanisms on water resources depends on (1) the extent of basin area affected by changes in snow water storage and snowmelt, (2) ecosystem adaptation and resilience to climate variability and trends, and (3) the relative magnitude of climate-induced changes compared to historical effects of dams and land use change on streamflow.

Future changes in climate factors that will affect streamflow include changes in peak flows, summer low flows, and seasonal water yield. These future changes will depend on future precipitation (not predicted to change much) and future temperature and its effects on snow storage and ET. Future temperatures are expected to reduce snow-covered area in much of Oregon, especially in the forested mountains. Reductions in snowpack are expected to increase winter flows (higher rain:snow) and reduce water delivery in spring and summer. However, these streamflow changes will also depend on forest vegetation response to future warming. For example, if future warming increases photosynthesis and respiration in the winter, that may offset some of the expected future increases in winter water yield and peak flows. Also, if future warming and reduced summer streamflows enhance hydrologic drought, drought-adapted conifers may be able to compensate by reducing ET, which in turn may offset some of the expected future declines in summer lowflows. Such changes in flow regime will have significant economic impacts for basin-wide water uses (Franczky and Chang ,2007). Future changes in runoff will also be affected by land use changes, which should be factored into future climate change impact studies (Praskievicz and Chang, 2009b). The spatial variability of the current water use patterns could then be factored into adaptive water resource management in a changing climate (Franczyk and Chang, 2009).

A variety of studies have developed quantitative estimates of the expected future impacts on surface water hydrology associated with climate change for hydrologic systems in the Western US. The hydrological response to projected future shifts in climate conditions has been described in parts of northwestern Oregon by Graves and Chang (2007) and Franczyk and Chang (2009), in the Oregon Cascades by Tague et al. (2008) and Tague and Grant (2009), and in California by Dettinger et al. (2004), Hanson and Dettinger (2005), and Dettinger and Earman (2007a). All of these studies show that the general hydrologic response to warming and the resulting reduction in the ratio of snow to rain will be increased winter runoff, earlier snow melt, and diminished spring and summer runoff. An analysis of the hydrologic response to climate change in the upper Deschutes Basin based on ensemble GCM predictions coupled to hydrologic models shows similar changes in runoff (Waibel et al., 2009)

This chapter summarizes key findings of the available literature and suggests research directions that may serve to increase the capacity to adapt to changing future conditions in Oregon.

3.3.1 Changes in snow water equivalent

Hydrologic systems in Oregon are relatively sensitive to changing climate, in large part because of the presence of seasonal snowpack. The snowpack develops in the mountains each winter, storing water through the period, and releasing it during spring, as air temperatures increase. This spring melt is channeled through a system of storage reservoirs, which are operated to both reduce downstream flooding and to provide water supply across much of the state over the relatively dry summer months. The amount of water stored as snow and the timing of melt depend very directly on spring air temperatures. As outlined earlier in this report, a wide variety of research has evaluated trends in historic snowpack data, with an emerging consensus that the snowpack throughout the West has experienced measureable declines over the period for which measurements are available.

The trajectory with which these observed changes will continue into the future is of particular interest to managers and stakeholders, particularly in light of the projected increases in air temperatures which have consistently arisen from global climate change research. The impact of projected future climate on freshwater resources is most frequently evaluated through the use of modeling. While results from these modeling-based impact studies vary depending upon the particular area of study and the study methods, a number of common themes emerge from studies developed in snowmelt dominated systems in the western US. The most significant result is that the warm snowpack that exists throughout the Washington and Oregon Cascade mountains is particularly vulnerable to commonly projected increases in winter temperatures. Peak stream flow volumes, which characterize snowmelt peaks in snow-dominated watersheds, are commonly used to summarize and assess the snowpack dynamics.

In one such study, Stewart et al. (2004) evaluated changes in the timing of peakflows under Business as Usual (BAU) emissions using PCM (Parallel Climate Model) and the VIC (Variable Infiltration Capacity) model of hydrology. They focused on the centroid of yearly stream flow as an indicator of snowmelt, and consistently projected statistically significant earlier peak runoff values in Washington. Supporting this work, Barnett et al. (2005) suggest that in the snowmelt dominated regions of the Western US, spring peak streamflows are likely to consistently occur up to a month earlier by 2050. Using PCM and a BAU scenario, Dettinger et al. (2004) employed the PRMS hydrologic model to provide more direct estimates of the future snowpack. They found that for the American River basin, in California, the average April 1 snowpack will approach 15 percent of historical values at the end of the 21st century. More recent modeling work by Elsner et al. (2010) also supports the notion of continuing snowpack decreases, in this case in Washington, suggesting statewide decreases in April 1 snow pack of 27 - 29% by 2020,

37 - 44% by 2040 and up to a 65% decrease by 2080. They focus on two emissions scenarios and used both the VIC and DHSVM (Distributed Hydrology Soil Vegetation Model) hydrological models.

In the Willamette River basin, the ratio of April 1st SWE to Precipitation (SWE/P) declined substantially from the reference period of 1970-1999 under two GHG emission scenarios with a greater reduction in the 2080s. The decline in the ratio is most pronounced under the high emission A1B scenario (see Figure 3.13). This is a combined result of increase in precipitation falling as rainfall in winter and earlier snowmelt caused by rising temperature. Snowmelt, estimated by the PRMS model, was projected to decrease gradually over time. For example, models of the upper Mckenzie River sub-basins indicate a decrease in snowmelt of up to -52% for the 2040s and up to -78% for the 2080s relative to the reference period, 1960 - 1989 (Chang and Jung, 2010).

3.3.2 Spatial and Temporal Variations of Changes in Runoff

Available research consistently projects reductions in winter snowpack in the Northwest US, as well as earlier runoff in snowmelt dominated basins, yet many regions in this area do not develop a winter snowpack and are rainfall dominated. While many of these lower elevation basins are influenced by higher elevation, snowmelt dominated areas, the runoff response is characterized by a wider range of variables, including potential changes in precipitation and groundwater contributions.

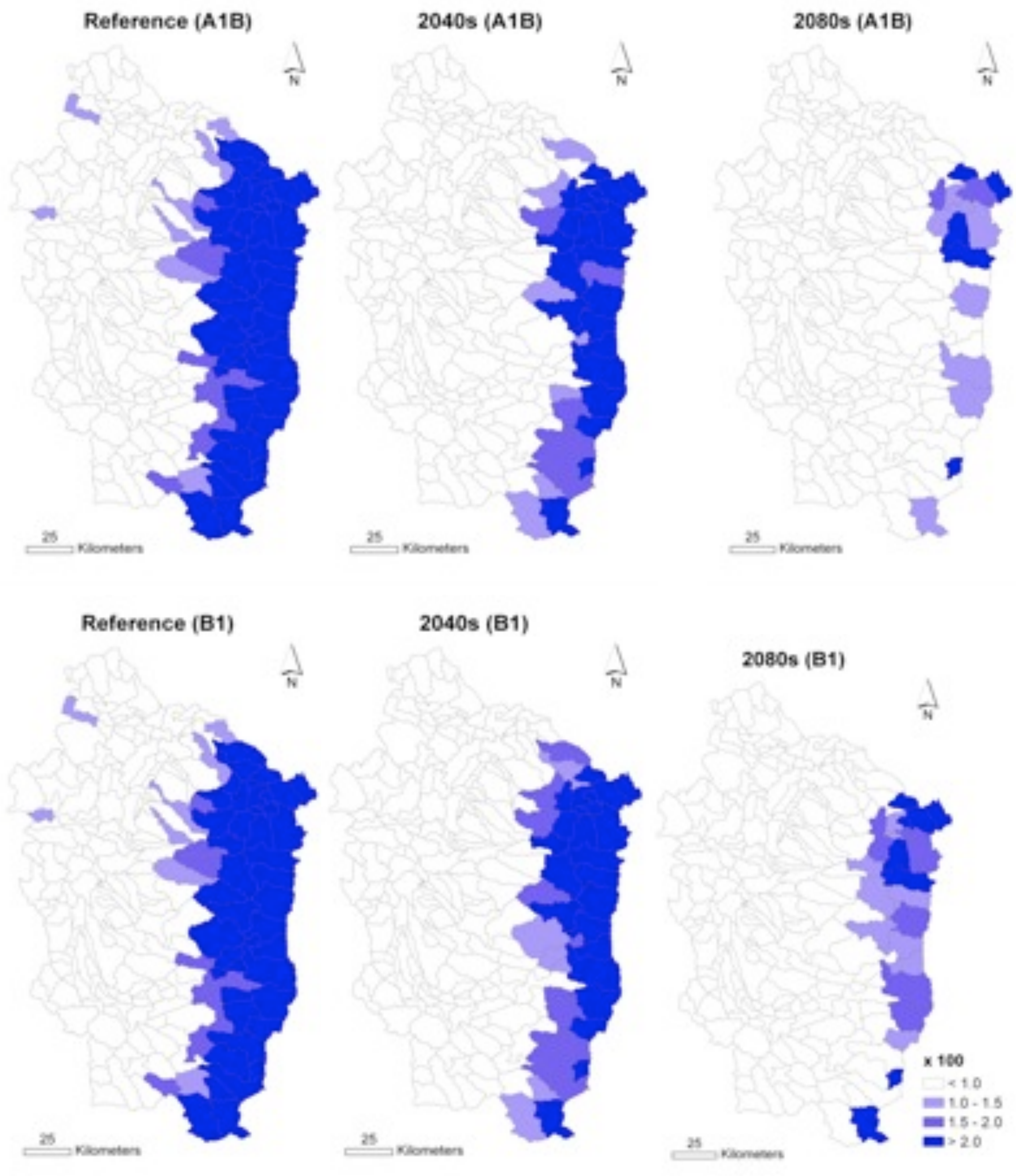

Figure 3.13 Ensemble mean changes (averaged over eight GCMs) in SWE/P in the Willamette River basin for reference, the 2040s, and the 2080s by GHG emission scenario. The ratio is multiplied by 100 for representation, Source: Chang and Jung 2010.

As the seasonal distribution of precipitation changes and temperature rises, watershed hydrology is likely to be modified at multiple spatial and temporal scales. Latitude and elevation control the sensitivity of particular regions to a changing climate, primarily because of the strong relationship between these factors and the presence of winter snowpack. While seasonal runoff will be more affected by increases in temperature in snow-melt dominated basins (Graves and Chang, 2007), in rainfall-dominated basins it will be more affected by changes in precipitation (Franczyk and Chang, 2009a).

In the Willamette River basin, the complex topography and geology also partially control the sensitivity of each sub-basin response to changes in climate. In the Western Cascades, snow-water equivalent is predicted to decline and peak runoff is predicted to occur earlier by the 2080s. In the High Cascades, with relatively gentle slope and young volcanic rocks, summer runoff may be sustained by existing large groundwater reservoirs (Tague and Grant, 2009; Jefferson et al., 2006). However, the uncertainty of projected future runoff is high, particularly for the High Cascade basins where groundwater is a big component of the seasonal water cycle (Chang and Jung, 2010).

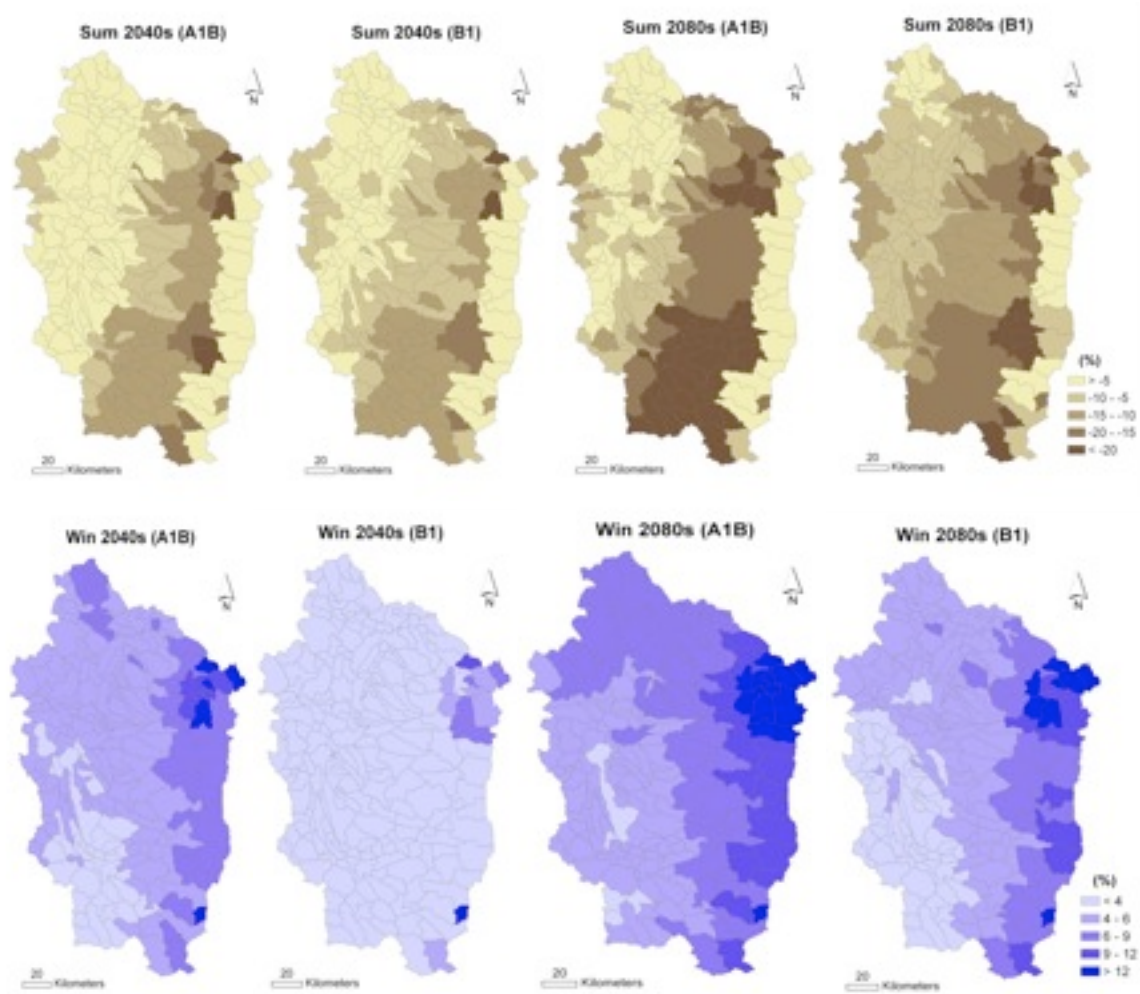

Figure 3.14 Ensemble mean changes in summer runoff (upper panel) and winter runoff (lower panel) for the 2040s and the 2080s by each emission scenario (Source: Chang and Jung, 2010).

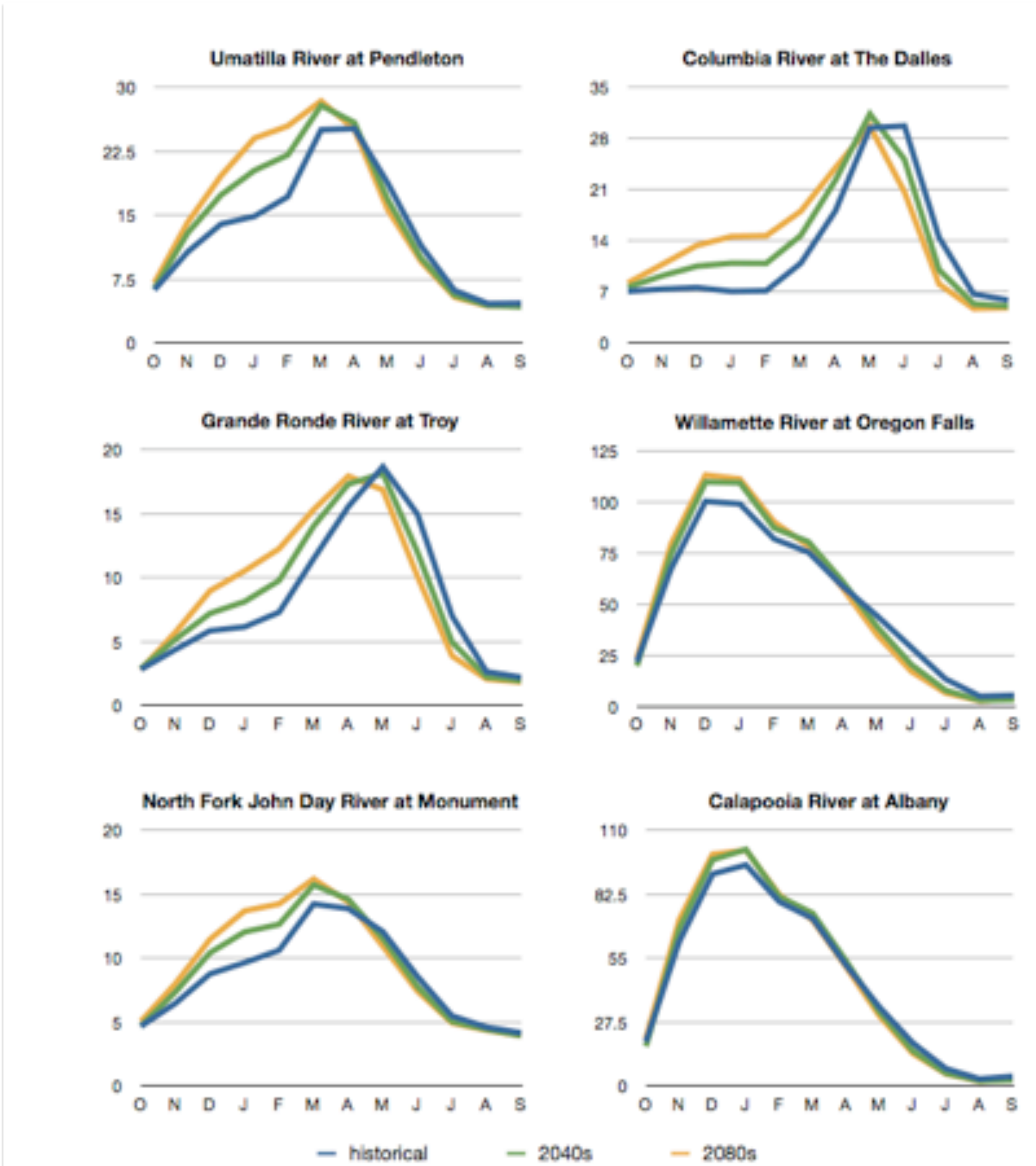

Figure 3.15 Hydrographs of combined flow for selected locations in Oregon using VIC outputs (hybrid delta, ensemble mean of 10 GCMs, A1B scenario) credit Hamlet et al, 2010, Climate Impacts Group, www.hydro.washington.edu/2860.

The University of Washington Climate Impacts Group produced 297 hydrologic scenarios for the Columbia River Basin using the Variable Infiltration Capacity (VIC) model. For this assessment, combined flows were downloaded from the website for the A1B scenario for the historical period, the 2040s and the 2080s. Combined flow is average total runoff and baseflow as an average depth (mm). These hydrographs are constructed using an ensemble average of 10 GCMs that were downscaled using the hybrid delta method. To read more about the project, downscaling techniques, and to download the data, visit the project website at http://www.hydro.washington.edu/2860/.

The six Oregon sites used for this assessment were selected based on two factors: their spatial distribution and their Nash-Sutcliffe efficiency. An effort was made to select sites that were distributed well across the larger Columbia River Basin and the state. Additionally, each site selected for this assessment had a Nash-Sutcliffe efficiency of 0.5 or greater. Nash-Sutcliffe efficiency is used to determine the predictive capabilities of a hydrologic model. The coefficient is a number between 0 and 1. The closer to 1 that the coefficient is, the better the model is at simulating flows. For applications such as these, a N-S efficiency of 0.5 is considered good.

Each hydrograph shows, with varying degrees of magnitude, a shift in streamflow largely consistent with climate projections. In snowmelt-dominated sites such as the Columbia River at the Dalles and the Grande Ronde River at Troy, flows are projected to increase in the winter months and decrease in the summer months through the 21st century. Peak flow shifts earlier into the spring at both sites in this scenario. At sites where the peak flows occur in the wetter winter months (Willamette and Calapooia), flows are projected to increase in the winter and decrease slightly in the summer. Both the Umatilla and North Fork John Day River project a significant increase in winter flows, but only a slight decline in summer flow. Diminished summer flows have implications for many sectors including agriculture, water resources and recreation, among others.

While changes in winter runoff volume and timing are commonly reported as potential impacts of projected temperatures, increased variation in hydrological response is also a potential response. Hamlet and Lettenmaier (2007) simulated runoff response and flood risk to explore impacts of documented (Mote et al., 2005) 20th century temperature increases. They used the VIC hydrological model and while they note significant variability in the flooding response, the results suggest that much of this variability can be constrained by regional differences in midwinter temperatures. In cold areas with a winter snowpack, flood risks have decreased, and in warm rainfall-dominated areas, flood risk appears to have significantly increased. Mote et al. (2003) also project an increased volume and earlier peak of winter runoff due to a lack of snowpack storage and an increase in rainfall (as opposed to snowfall), and also predict decreases in summertime low flows. They attribute the decreases in low flows directly to the lack of projected spring snowmelt peak. While changes in peak flows have significant implications, the associated changes in low flows are of potentially greater consequence in that water use is at a maximum during the summer period, while at the same time water availability is, even in the coolest years, at a minimum. The balance between these two quantities is more fully explored in other sections of the report, and we note here that it is an important component in the distribution runoff pathways, and that modifications to it may have potentially far-reaching effects.

While there are no significant trends in winter precipitation intensity in the Willamette River basin since 1972 (Praskievicz and Chang, 2009a), according to the IPCC fourth assessment report, climate change is likely to bring more extreme hydrologic events such as floods and droughts in the region. Urban areas are particularly vulnerable to these changes as impervious areas do not efficiently absorb storm water and infrastructure is densely concentrated (Chang and Franczyk, 2008). A case study of Portland shows that climate change will bring more frequent storm events with a return period of less than 25 years, which means that nuisance flooding is likely to become more common at road cross-sections that have a history of chronic flooding (Chang et al., 2010a).

It is also important to recognize that these examples focus on basins in the Willamette River where precipitation is relatively abundant. In other regions of the state, different processes control the response (rainfall dominance, groundwater contributions, semi-arid hydrology, etc). Additionally, changes in water use by vegetation along with management-based adaptation strategies would also be important features of any future impacts work. An impact assessment for the state would need to be significantly broadened in scope to capture the large degree of hydrologic variability and possible secondary responses to climate change from vegetation and management expressed statewide.

3.3.3 Interactions Between Climate and Land Surface Hydrology

The interactions between climate system and land surface hydrology are rather complex in the Willamette River basin. While elevation is a primary control of basin runoff, other basin characteristics such as geology and topography also affect basin runoff. Elevation is an important determinant of change in basin runoff because it affects the amount of precipitation falling as snow in winter and the snowmelt rate in spring. While winter runoff change is more sensitive to changes in winter temperature than winter precipitation in high elevations, the relative influence of winter temperature declines with elevation, and winter precipitation becomes more important in projecting future winter runoff in low elevations (< 1000m). This is associated with whether the basin runoff generation is dominated by either rainfall or snowmelt, suggesting that the elevation threshold may be associated with other basin characteristics such as geology. Geology could buffer the sub-basin hydrological response to climate change. Basins in the High Cascade Range with significant groundwater exchange may be less sensitive to changes in climate than those in the Western Cascade Range in the near term 2040s (less than 10% reduction in summer runoff) (Chang and Jung, 2010).

3.3.4 Uncertainty in Projecting Future Runoff

Uncertainty is an inherent component of projections related to future runoff. It is an additive property of the modeled system, with each individual component contributing to the overall predictive uncertainty of the system, and as such the degree of uncertainty associated with each model component should be evaluated as a mechanism to communicate confidence in projected results. There are several sources of uncertainties in climate change impact studies. The GCM structural uncertainty is the main source of uncertainty as recognized in several studies (Praskievicz and Chang, 2009b; Graham et al., 2007). Although many investigations have shown that the highest uncertainty is attributed to the GCM's structure, the other sources are not negligible (Wilby, 2005; 2006; Wood, 2004). Other uncertainty sources include the emission scenarios, the downscaling methods, the hydrologic model structure and the hydrologic model parameters.

The importance of each source of uncertainty depends on hydrologic characteristic of the basin under study (Kay et al., 2009; Prudhomme, 2007). For example, two sub-basins of the Willamette watershed, one rain-snow dominated and one rain-dominated, showed different sensitivities to uncertainty. The snow-dominated basin was more influenced by uncertainty in hydrologic parameterization than the rain-dominated basin, although climate model uncertainty was still the main source of uncertainty in both basins (Chang and Jung, 2010).

To date, many hydrologic models have been used to project the possible consequences of the climate change on streamflow, ranging from simple conceptual lumped models to comprehensive physically-based distributed models – e.g. NWSRFS (Nash, 1991), WatBal (Yates, 1996), macro-scale hydrological model (Arnell, 1999), VIC (Lettenmaier, 1999), MODFLOW (Kirshen, 2002), CATCHMOD (Wilby, 2005), PDM (Kay et al., 2009) and PRMS (Jung, 2010) among others. Most of the hydrological models show reliable and accurate results under historical climate conditions (natural variability). However, they have often projected mixed results in runoff change even under identical climate change conditions. This could be attributed to differences in the models' structures (Bae, 2009; Bloeschl and Montanari, 2010; Kay et al., 2009; Wilby, 2006). Bae et al. (2010) employ three semi-distributed models to investigate uncertainty resulting from hydrologic model selection; they conduct their study using 13 GCMs simulations with 3 emission scenarios. They conclud that monthly and seasonal runoff change simulated by a single hydrological model is within ±10% difference from those of a multi-model ensemble except in the low flow season.

Elsner et al. (2010) acknowledge the need to evaluate the uncertainty around modelderived hydrologic projections. While they do not provide estimates of predictive uncertainty, a sensitivity analysis using simulated runoff to estimate precipitation elasticity, or the fractional change in runoff as compared with a fractional change in precipitation, is developed. They also calculate modeled temperature sensitivity as the percent change in projected runoff for a 1 degree increase in temperature. The sensitivities are estimated for 12 different basins in Washington and represent 2 different hydrologic models. The results indicate low sensitivity across this range of simulations, providing additional quantitative evidence that the uncertainty of climatic projections dominates the overall uncertainty of impact projections.

Hawkins and Sutton (2009) state that the uncertainty in regional climate predictions varies with time but there is no constant positive trend because emission scenario uncertainty increases over time but GCM and internal variability uncertainties decrease. Also, the uncertainties vary across the seasons. In the summer, small changes in flow are much more significant than they would be in the high-flow wet season.

The uncertainty associated with hydrologic model selection was studied over the Tualatin river basin (Najafi et al., in review). Figure 3.16 shows the results of each model based on 8 GCM forcing data that are expressed as bias percentage between the future and reference periods. The figures show the runoff results for the winter (Dec-Jan-Feb) and summer (June, July, August) respectively for the two A1B and B1 emission scenarios. The TM model's (the simplest model) summer result shows the highest uncertainty due to the GCMs compared to the other hydrologic models simulations. The uncertainties vary between models for different emission scenarios and time periods. Therefore, the hydrologic model structural uncertainty would be another considerable uncertainty in addition to the climate model, downscaling and emission scenario as the main sources of uncertainty reported in the literature.

The representation of vegetation response to water availability in most hydrologic models is a further source of model uncertainty in hydrologic projections under climate change scenarios. The dominant tree species in Oregon forests are capable of significant adjustments of transpiration rates in response to environmental factors, and these adaptations are not included in most hydrologic models. In addition, vegetation may undergo succession in response to changes in air temperature and/or soil water availability, further altering vegetation water use and hence streamflow, in ways not currently captured by hydrologic models.

As projections of future impacts of climate change on hydrology continue to develop, it is important to recognize uncertainty; but perhaps more important to recall, as outlined by Bloeschl and Montanari (2010), that while projections of future conditions will always be uncertain, the presence of uncertainty does not indicate a lack of understanding.

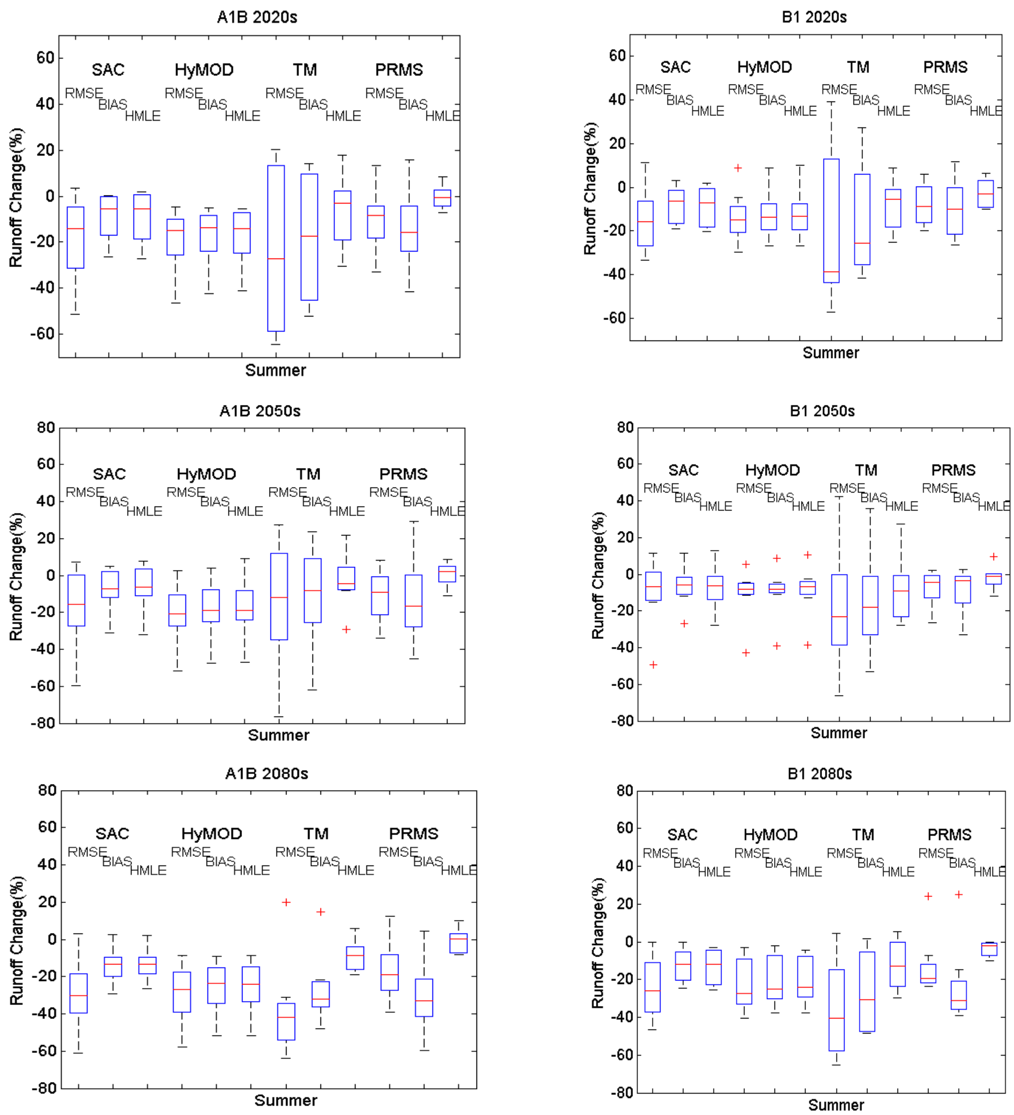

Figure 3.16 Runoff change relative to the reference period obtained by four hydrologic models with different complexities (source, Najafi et al., in review)

3.4 Potential Changes in Groundwater Hydrology

Projected future changes in temperature and precipitation will affect groundwater hydrology. Projected changes in climate will result in alterations of the timing and amount of recharge, increases in evapotranspiration, lowering of heads in boundaries such as streams, lakes, and adjacent aquifers, sea-level rise, and increased pumping demand. Increase pumping demand due to climate change will be exacerbated by population growth. This section presents a brief overview of groundwater hydrology, summarizes ways in which groundwater systems can be affected by projected climate changes, describes the likely response in key geographic settings in Oregon, and suggests directions for future research.

3.4.1 Overview of Groundwater Hydrology

Groundwater originates as precipitation that infiltrates into the ground. The infiltration of rainfall or snow melt into the groundwater system is called recharge. The largest amount of recharge typically occurs in upland areas where precipitation is greatest, although recharge can occur anywhere that precipitation exceeds evapotranspiration and available storage in the root zone. Water percolates downward until it reaches the water table, which generally defines the top of the region in which rock materials are completely saturated with water. In areas where streams are above the water table, leakage from streams can also recharge the groundwater system. Groundwater can also be recharged artificially through deep percolation of irrigation water, particularly in areas irrigated using surface water, and leakage from irrigation canals. Once in a groundwater system, water moves through permeable geologic materials in response to gravitational forces at a rate proportional to the permeability of the material through which it is moving and the hydraulic head gradient (which can be thought of as the slope of the water table). Groundwater eventually discharges back to the surface typically through springs or as diffused seepage to streams, lakes, and wetlands. In areas where the water table is very close to land surface, plants with roots extending to the water table can also be an avenue of discharge. Groundwater can also be removed artificially through wells. Groundwater discharge can occur naturally almost anywhere the water table, or saturated zone, intersects land surface including in uplands, where it supports perennial flow in low-order streams, and in lowlands and major stream valleys.

Groundwater discharge is an important component of streamflow along with surface runoff. Most critically, groundwater is the principal source of streamflow in the late summer and fall when there is little precipitation or snowmelt to supply runoff. In cold regions and at high altitudes, groundwater maintains winter flows critical to many groundwater dependent ecosystems, for example maintaining winter flows over salmonid redds. The portion of streamflow supplied by groundwater is known as baseflow. The proportions of groundwater and runoff in a stream vary seasonally, but also depend on the geology. Streams in areas of very permeable rock with substantial groundwater systems (as is common in parts of the Cascade Range) (figure 3.18) may consist almost entirely of groundwater discharge and have only a small component of runoff. Such streams are termed “groundwater-dominated,” and they are common in young volcanic areas. Groundwater-dominated streams have constant flows with very small seasonal variation (figure 3.18). Streams in relatively impermeable areas such the

Coast Range, in contrast, have a very small component of groundwater and are runoff dominated. Runoff-dominated streams have large seasonal variability and commonly go dry, or nearly so, in the late summer (figure 3.18).

Figure 3.17 Selected physiographic and geographic features in Oregon

Because groundwater systems are recharged by precipitation, they are sensitive to changes in the amount, timing, and form of precipitation. Potential negative responses to climate change could include lower water levels in wells (water table elevations) and reductions in groundwater discharge to streams. Lower water table elevations can reduce the amount of water available for human uses by reducing the saturated thickness of aquifers and the amount of water in storage. Reductions in groundwater discharge to streams limit water available for human uses (such as municipal and agricultural diversions) and the in-stream needs of aquatic ecosystems.

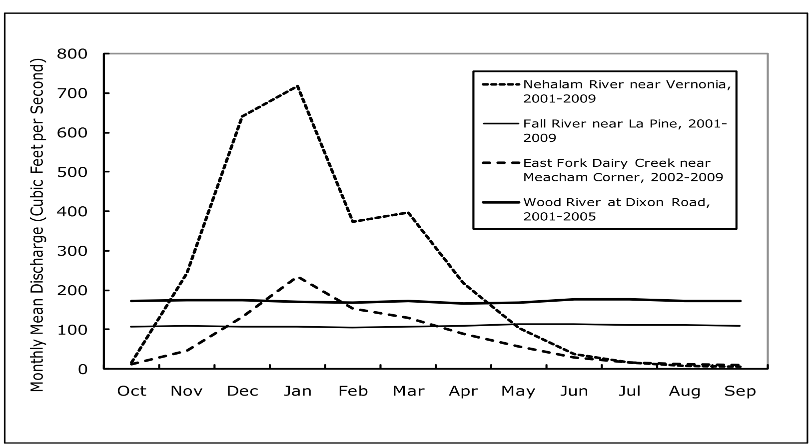

Figure 3.18 Mean monthly discharge of selected runoff dominated streams (dashed lines) and groundwater dominated streams (solid lines) with overlapping periods of record in Oregon. Nehalam River and Dairy Creek data from the U.S. Geological Survey, Fall and Wood River data from the Oregon Water Resources Department.

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Alexander and Palmer (2007) summarized studies of the impacts of climate change on groundwater resources in eight regions in the US and Canada. Most of the studies involved model analyses and incorporated multiple climate models and emission scenarios. Results differ between models, but tend to show that recharge varies in proportion to precipitation. Groundwater systems can be affected by changes in total precipitation, as well as changes in the spatial and seasonal-to-daily distribution of precipitation. Other climate-related factors that are shown to be important include changes in stream stage, increases in evapotranspiration, and increases in groundwater pumping.

3.4.2.1 The response of groundwater systems to climate warming

As surface hydrologic processes respond to climate change, so will groundwater recharge. The same rainfall and snowmelt events that drive runoff also provide groundwater recharge. Most groundwater recharge in Oregon occurs in place, meaning at the location where rainfall or snow melt occurs. Groundwater recharged from anthropogenic sources, such as deep percolation of irrigation water and canal leakage, is usually secondary to in-place recharge at the watershed scale. The close relation between runoff and groundwater recharge is demonstrated by Manga (1997), who successfully uses runoff as a proxy for groundwater recharge in models of groundwater-dominated streams in the Oregon Cascades. This means that changes in the timing of runoff that will occur under warmer climate conditions will result in changes in the timing of groundwater recharge. This is of practical importance where the timing of discharge of groundwater-fed streams is important for reservoir operations or irrigation.

Increased evapotranspiration due to warmer temperatures may also result in reductions in total annual recharge, especially at lower altitudes, as less water will percolate below the rooting depth of plants (Dettinger and Earman, 2007a). Variations in recharge resulting from changes in evapotranspiration are likely to vary geographically. Studies in the Yakima Basin of Washington show a 20% reduction of groundwater recharge (and stream discharge) in some subbasins resulting from a 3.6 °F increase in temperature (J.J. Vaccaro, U.S. Geological Survey, written commun., 2010). Preliminary results from modeling studies in the upper Deschutes Basin in Oregon suggest that total annual basin-wide changes in recharge resulting from warming will be much smaller (M. Scott Waibel, Portland State University, written communication, 2010).

3.4.1.3 The response of groundwater systems to changes precipitation

In general, increases or decreases in total precipitation will be reflected in groundwater recharge. The relation between total precipitation and groundwater can be seen by comparing the cumulative departure from average precipitation and long-term water level trends in wells (figure 3.20). Fluctuations in upland recharge areas can range from several feet to tens of feet in response to, and generally coincident with, decadal drought cycles. This suggests that groundwater levels in many aquifers will reflect any long-term changes in total annual precipitation more or less as they occur. Historic streamflow records show that groundwater discharge to streams also varies with annual precipitation (figure 3.21). If precipitation decreases, groundwater levels will decline and discharge to streams will diminish proportionally. If average annual precipitation increases, groundwater levels will rise and discharge to streams will increase. However, the effects of increased groundwater recharge resulting from the small projected increases in precipitation in Oregon are likely to be offset by other factors such as increases in pumping demand and evapotranspiration resulting from warmer temperatures.

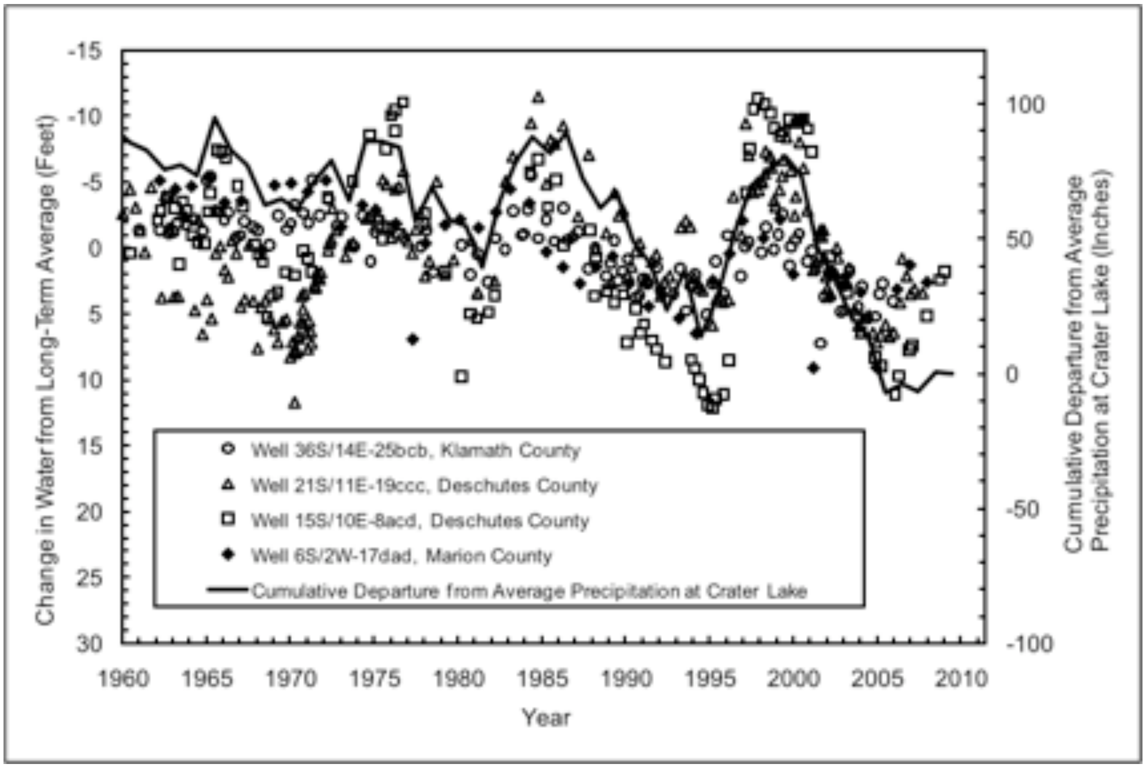


Figure 3.19 Comparison of water level fluctuations in selected wells and cumulative departure from average precipitation at Crater Lake Oregon. Well data from U.S. Geological Survey and Oregon Water Resources Department.

Changes in the seasonal distribution of precipitation will affect seasonal water table fluctuations and of groundwater discharge to springs and streams. If seasonal shifts are sufficiently large and total annual precipitation remains constant, the result could be less annual recharge in some areas. This is because some shallow aquifer systems, such as in parts of the Willamette Valley, fill to capacity and recharge during the winter is already "rejected" and shunted off to streams (Conlon et al., 2005). Increased winter precipitation, therefore, may not result in increased groundwater recharge.

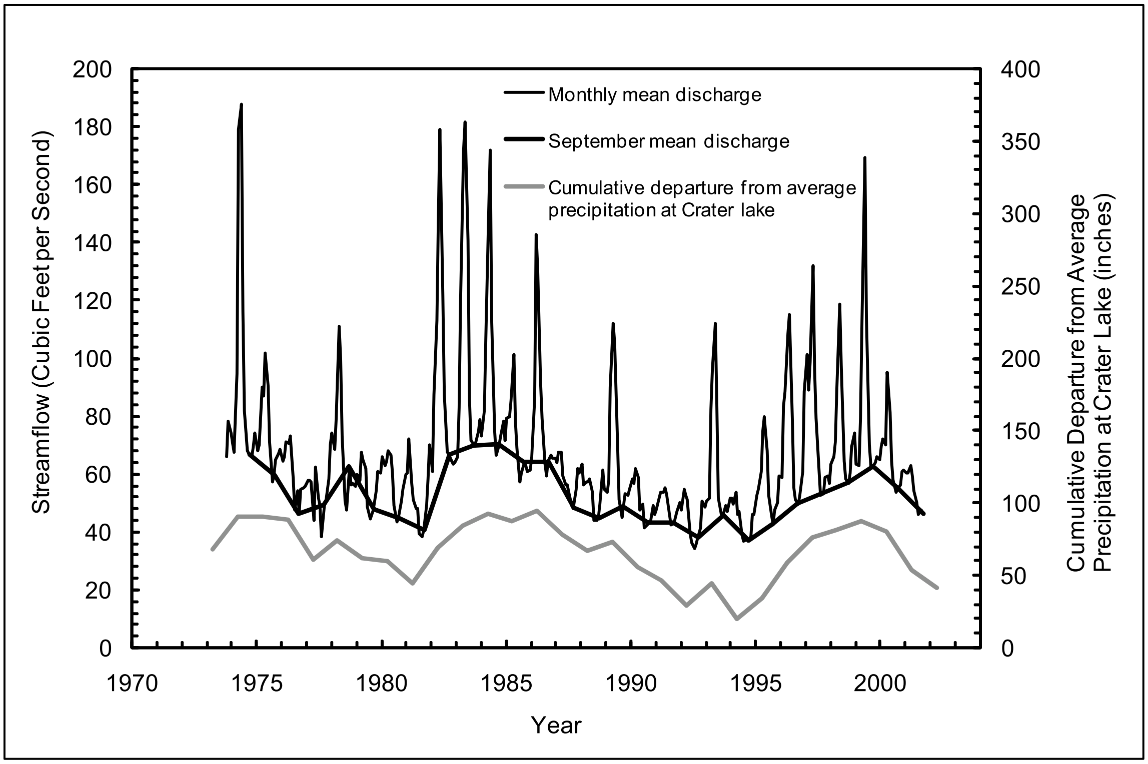


Figure 3.20 Monthly mean discharge of the Williamson River near Lenz, Oregon. Note the relation between the September mean discharge (a proxy for baseflow) and the cumulative departure from average precipitation at Crater Lake.

3.4.2.3 Other factors influencing the response of groundwater to climate change

Groundwater recharge can be affected by factors other than changes in precipitation and evapotranspiration. Studies have also shown that recharge can be affected by changes in the temporal and spatial distribution of frozen ground (Jyrkama and Sykes, 2007) and rainfall intensity (Mileham et al., 2009) resulting from climate change.

In addition to changes in recharge, other factors will influence the way in which groundwater systems respond to climate change. Aquifer systems in alluvial deposits that are in direct hydraulic connection to streams can be affected by the lowering of stream and lake stages resulting from diminished flows. Lower stream and lake stages can result in lower water levels in adjacent aquifers. Reduced streamflow may also result in reduced irrigation diversions and in less artificial recharge from deep percolation of irrigation water and canal seepage. Artificial recharge may also be reduced by conservation measures such as the use of more efficient irrigation methods or lining irrigation canals.

Sea-level rise resulting from climate change also could affect groundwater in coastal regions. Sea level is an important boundary condition affecting water levels in the extensive sand-dune aquifers along the coast such as found near Coos Bay, Reedsport, Florence, and in Clatsop County (Rinella et al., 1980; Brown and Newcomb, 1963; Hampton, 1963; Frank, 1970). A rise in mean sea level will result in a comparable rise in water-table elevations in sand dune aquifers as well as alluvial aquifers hydraulically connected to tidally influenced estuaries. This could result in water-level rises and expansion of groundwater-fed lakes and wetlands in sand dune areas and other low-lying coastal settings. Sea level rise will exacerbate any existing saltwater intrusion problems, as will warming-related increases in groundwater pumping in coastal areas.

Groundwater systems are also susceptible to the effects of increased water demand resulting from climate change. Warmer temperatures typically result in increased groundwater pumping by both municipalities and irrigators. Diminished late season streamflow predicted by most analyses will reduce the surface water available for irrigation which will also increase demand for groundwater. The very small increase in precipitation projected by the ensemble average of climate models for Oregon is unlikely to offset increased pumping due to warmer temperatures and diminished late-season streamflow.

3.4.2 The Influence of Geographic and Geologic Settings on the Response of Groundwater Systems to Climate Change

The response of groundwater systems to climate change will vary with geographic and geologic settings. Principal variables that will influence the response of groundwater systems to climate change include the permeability of the underlying geologic deposits, the geographic setting within the watershed (upland versus lowland aquifers), and the degree to which recharge originates as snow.

The permeability of geologic materials is the most basic factor controlling the presence or absence of groundwater systems and how those systems will respond to climate change. Low permeability units such as the marine sedimentary rocks Coast Range, pre Cenozoic rocks Klamath Mountains, and Mesozoic and early Cenozoic rocks of the Blue Mountains Province in northeastern Oregon, have generally low bedrock permeability and do not host large regional groundwater systems. Aquifers in such areas are largely limited to localized zones of bedrock fractures and alluvial deposits in valley bottoms. The largest impacts to the limited groundwater systems in such settings are likely to result from increased pumping demands and increased ET due to warmer conditions (Loáiciga, 2003; Hanson and Dettinger, 2005; Dettinger and Earman, 2007a). Streams in low permeability areas have very limited baseflow and typically go dry, or nearly so, in late summer and fall (figure 3.19). Late season water needs in such areas are commonly provided by storage in reservoirs that are filled during the winter and spring.

Areas dominated by permeable material such as fractured lava or extensive coarse-grained sedimentary deposits typically contain substantial groundwater systems that can be large enough to be of regional importance (Gannett et al., 2001, 2007; Tague and Grant, 2004). Such areas include the younger (high) Cascade Range, volcanic areas of central and eastern Oregon, and large sedimentary basins such as the Willamette Valley and Portland basin. Most large river basins in Oregon include both high permeability and low permeability areas. An exhaustive analysis of the potential response of groundwater to climate change in geographic settings across the state is beyond the scope of this chapter. Instead, this section discusses the probable response in selected settings to provide general insights. For simplicity of discussion, regional groundwater systems are divided into upland and lowland settings in the following sections. The processes described in upland and lowland settings, such as recharge and discharge, occur at a range of spatial scales.

3.4.3.1 Groundwater systems in upland settings

Upland settings where permeable deposits dominate recharge areas tend to have substantial groundwater systems and a large proportion of groundwater-dominated streams (and associated groundwater dependent ecosystems). The most prominent permeable upland in Oregon is the geologically youngest region of the Cascade Range, often known as the "High" Cascade Range which encompasses the upper parts of the Deschutes, Klamath, and Willamette Basins. The importance of groundwater to the hydrology of basins flanking the Cascades has been recognized for many decades going back to the work of Russell (1905), Meinzer (1927) and Stearns (1929, 1931). More recent work characterizing the importance of groundwater contribution to streams emanating from the Cascades includes that of Grant (1997), Gannett et al. (2001, 2003, 2007), Tague and Grant (2004, 2009), and Jefferson et al. (2006, 2007).

Many permeable upland areas in Oregon, and in the Cascade Range in particular, rely on snowmelt for a large part of their groundwater recharge. As was previously described, warming will result in a shift in the form of precipitation toward less snow and more rain, and consequently more runoff in the winter, earlier (and less) snow melt in the spring, and less runoff during the summer. The timing of groundwater recharge will shift in the same manner as runoff. The shift in timing of recharge will affect groundwater-dominated streams, increasing winter flow and reducing late season flow (Manga, 1997; Tague et al., 2008). Runoff-dominated streams are expected to experience larger changes in the seasonality of flow than groundwater-dominated streams. This is because the groundwater system acts as a reservoir, storing seasonally variable recharge and releasing it to streams at a more constant rate throughout the year. Groundwater discharge from permeable upland areas, therefore, has the potential to moderate the effects of warming to some degree (see, for example, Tague et al., 2008; Tague and Grant, 2009; Chang and Jung 2010; Mayer and Naman 2010). Although groundwater-dominated streams may experience small percentage reductions of late season flow, the changes may be volumetrically large (Tague et al., 2008; Chang and Jung, 2010). In runoff-dominated low-order streams, a small-volume/large-percentage reduction in flows may be catastrophic for some groundwater dependent ecosystems, particularly if those reductions cause perennial streams to become ephemeral.

While the effects of changes in the seasonality of recharge may be moderated by groundwater storage in permeable uplands, such areas will respond to changes in total annual precipitation. Water levels and groundwater discharge to streams will vary in proportion to any long-term increases or decreases in total precipitation. Total recharge likely will also be reduced by increased evapotranspiration under warmer conditions.

3.4.2.2 Groundwater systems in lowland settings

Groundwater recharge in lowland areas (such as the Willamette Valley, Portland Basin, and stream valleys and lake basins in central and eastern Oregon) will respond differently from uplands to warming and to possible changes in the seasonality of precipitation. Moreover, additional climate-related stresses may affect groundwater in lowland areas. Natural recharge in lowland areas is primarily from rain and intermittent snowmelt and does not rely on large, spring snowmelt events. Therefore, changes in the form of precipitation are not likely to have as large an influence on the seasonality of recharge in lowland areas as they will in upland areas. Most climate models do, however, project a shift in the timing of precipitation toward wetter fall and winter seasons and dryer summers (Mote and Salathé, 2010). These shifts will affect seasonal groundwater fluctuations, possibly resulting in lower water levels in wells in the summer which could increase pumping costs and limit well yields. This is an important consideration for irrigation and municipal wells. As mentioned previously, if the shift toward winter precipitation is sufficiently large and total annual precipitation remains the same, total annual recharge could diminish in shallow aquifers that reach capacity and "reject" recharge during the winter.

Groundwater is much less likely to moderate the effects of climate change on streams originating in lowland settings than streams originating in permeable uplands for several reasons. Recharge rates in lowlands areas are generally less because of the smaller amount of precipitation, groundwater discharge to streams from lowland areas is generally less than in upland areas, and groundwater originating in lowlands generally makes up a smaller component of streamflow. For example, groundwater-dominated streams rarely originate in lowland areas except at the bases of uplands.

Groundwater systems in lowland areas are more susceptible to increases in pumping than upland areas. Farms and cities, the largest users of groundwater, tend to be located in lowlands. In addition, some lowland groundwater systems may be susceptible to changes in boundary conditions. Shallow alluvial aquifers in lowland stream valleys tend to be hydraulically connected to streams, and, because of the low head gradients, groundwater levels can be influenced by stream stage over significant areas (for examples in the Willamette Valley see Conlon et al., 2005). Decreases in stream stage resulting from smaller late-season flows could result in proportional water-level declines in hydraulically connected aquifers.

3.4.2.3 Columbia River Basalt Group aquifers

Aquifers in the lava flows of the Columbia River Basalt Group constitute a unique class of aquifers in Oregon. Because of their high transmissivity, Columbia River Basalt Group lavas are productive aquifers. However, these aquifers are susceptible to large pumping-related water level declines due to their low specific storage and recharge that is limited by low vertical permeability. Columbia River Basalt aquifers underlie much of north central Oregon and parts of the Tualatin and northern Willamette valleys.

Poorly-confined aquifers at shallow depths in the Columbia River Basalt Group allow infiltration of water and are often in a state of dynamic equilibrium with climate, and the temporal variations caused by seasonal recharge events and decadal drought cycles are apparent in water-level trends (Conon et al., 2005; Vaccaro et al., 2009; Kenneth E. Lite Jr., Oregon Water Resources Department, oral communication, 2010). Highly confined aquifer systems such as found at depth in the Columbia River Basalt Group, in contrast, have very limited recharge. Consequently, climate signals are often not prominent in water-level data in Columbia River Basalt wells deeper than about 600 feet (Conlon et al., 2005; J.J. Vaccaro, USGS, written communication, 2010). Water level fluctuations in such highly confined systems are usually dominated by anthropogenic influences such as pumping, and, in some areas, head changes due to progressive interconnection of aquifers at different depths by wells (Burns et al., 2009). In an analysis of water level trends in the Yakima Basin in Washington, Vaccaro et al. (2009) show that water level declines are largest in deeper aquifers, indicating diminished recharge with depth. Generally speaking, climate signatures are apparent in upper basalt zones but less so in deeper zones (John Vaccaro, USGS, written communication). Therefore, climate change will affect Columbia River Basalt aquifers to different degrees (or at least with different timing) depending on depth. In deep basalt aquifers, the effects due to changes in recharge are likely to be small in comparison to the effects of increased demand, particularly if average annual precipitation remains largely unchanged.

3.4.3 Strategies for Improving the Understanding of, and Responding to, Changes in Groundwater Hydrology Resulting from Climate Change

Several actions could be taken to improve the understanding of the probable response to groundwater systems in Oregon to climate change and to help in development of adaptive management strategies. Efforts should include additional analysis of historic data to improve understanding the relation between groundwater systems and climate across the state, improved monitoring of groundwater to quantify the current and future response to climate variability, and continued development of hydrologic models to improve understanding of the linkages between climate and groundwater and to improve predictive capabilities.

Considerable information is contained in historic records of groundwater levels and streamflow; these records could be evaluated along with historic meteorological data to improve understanding of the coupling of climate and groundwater. Such an analysis could provide new insights, highlight particularly vulnerable regions or hydrologic settings, and provide useful information for development of new models.

Monitoring of groundwater and streamflow in Oregon has historically been done for specific regulatory or management purposes so present networks, while supplying considerable valuable information regarding the possible effects of climate change, are not optimally suited for that purpose. Hence, water level trends in monitored wells are often dominated by pumping effects that overwhelm the climate signature. The US Geological Survey (USGS) has developed a groundwater climate response network to "portray the effect of climate on groundwater levels in unconfined aquifers or near-surface confined aquifers that are minimally affected by pumping or other anthropogenic stresses" (Cunningham et al., 2007). Of the 500 wells in the national network, eight are in Oregon. It is possible that many of the hundreds of wells presently monitored by the Oregon Water Resources Department (OWRD) and the USGS for other purposes would be suitable for including in a climate response network. It is probable that there are aquifers that are not presently included that should be monitored.

Considerable information on the state of groundwater systems can be provided by monitoring the groundwater discharge to streams and springs using standard stream gaging techniques. Many gaging stations throughout Oregon, operated primarily by the USGS and OWRD, can be used to estimate discharge from some major aquifer systems. These estimates can be compared to climate records as in Figure 3.20 to provide insights into climate/groundwater connections. Most streamflow monitoring, however, is done to assist in operations of dams and reservoirs or for managing irrigation water. There are, therefore, many groundwater systems for which discharge is not monitored. A systematic review of stream gaging networks and large springs in Oregon could identify sites that would provide information on groundwater conditions as well as aquifer systems that are not presently adequately measured.

Developing networks specifically for monitoring changes in groundwater recharge caused by climate change is a topic of growing interest in the western US, however no standard techniques and protocols presently exist. Summaries of existing and emerging techniques for monitoring groundwater recharge are provided by Dettinger and Earman (2007b) and Earman and Dettinger (2008).

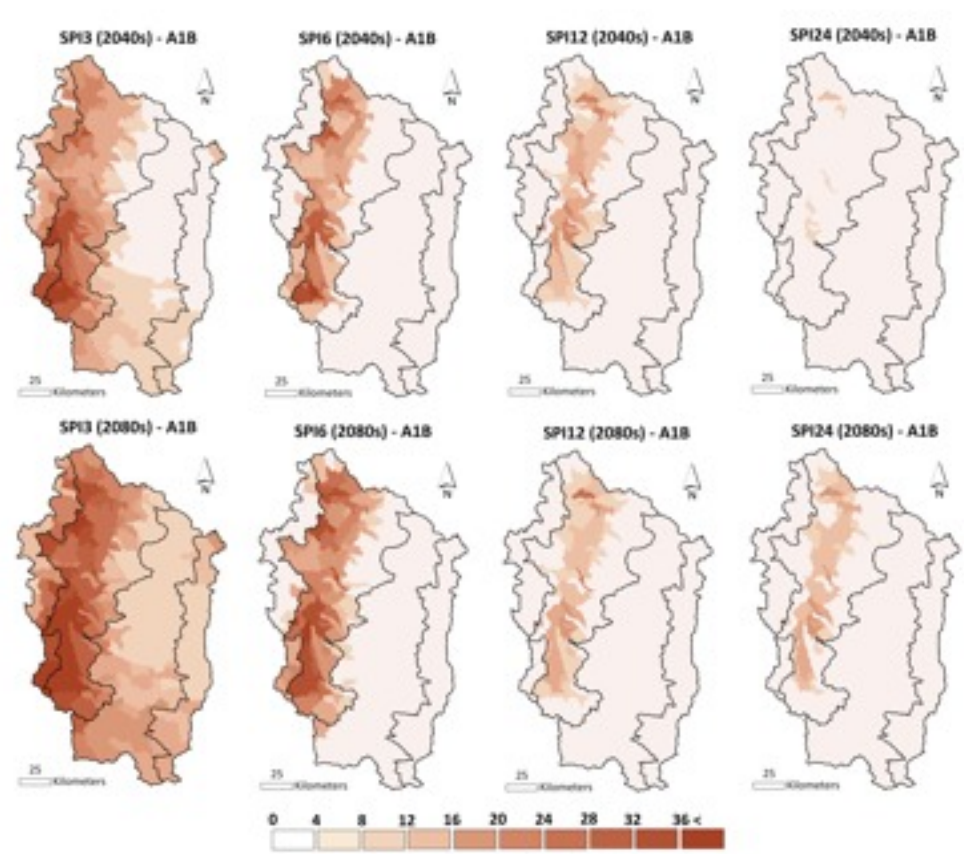
Insight into the probable hydrologic response to the range of projected climate changes in Oregon and the western US has come from modeling studies. Models not only provide important insights, but also predictive capability. Modeling studies of the hydrologic response to climate change in Oregon include work in the northern Willamette Valley by Graves and Chang (2007) and Franczyk and Chang (2009), in the McKenzie River watershed by Tague and Grant (2009), and in the Deschutes Basin by Waibel et al. (2009). Some of these studies incorporate existing hydrologic models, developed for reasons other than climate change research, coupled with downscaled climate model output or other predictions of future climate. Continuation and expansion of modeling efforts will provide additional insights and predictive capability. Emerging techniques for coupled groundwater/surface-water models (such as the USGS GSFLOW code) are particularly promising. Groundwater models exist for a number of major basins in Oregon (for example, Morgan, 1988; Davis-Smith et al., 1988; Morgan and McFarland, 1996; Gannett et al,, 2004) that could be coupled with climate models to gain insights into the hydrologic response of the basins to projected climate change. Coupling hydrologic models with management models using optimization techniques can help identify strategies for resource management under uncertain and changing conditions.

3.4.4 Summary

Groundwater systems in Oregon will be affected by warming and possible changes in the amount and seasonality of precipitation projected by climate models. The principal mechanisms for change in groundwater are expected increases in evapotranspiration, which will decrease groundwater recharge, and increase pumping of water from groundwater wells to compensate for increased evapotranspiration. Secondary mechanisms include small changes in groundwater resulting from small projected changes in precipitation and localized changes in sea level in coastal areas. Responses include changes in the timing, amount, and spatial distribution of recharge, as well as changes in pumping demands and other boundary conditions. The response of groundwater systems will vary among geographic and geologic settings. All groundwater systems are sensitive to changes in the amount and timing of precipitation, from those in humid regions to systems in semiarid regions. Regional groundwater systems in the Cascade Range, important to streamflow in the adjacent basins, may moderate the effects of climate change somewhat but are likely to experience changes in the seasonal distribution of recharge. Lowland groundwater systems are probably most susceptible to increases in groundwater pumping resulting from warmer temperatures. Understanding the linkages between groundwater systems and climate can be improved with expansion of collection and analysis of groundwater data. Continued development of hydrologic models can improve understanding of the likely response of groundwater systems to the range of possible future conditions, and help in development of water management strategies.

Case study: Possible future climatological drought in Willamette River Basin

Drought is a natural hazard that can have severe impacts on regional water sector. The extreme seasonality of precipitation in the Pacific Northwest has induced frequent seasonal water shortage problem, especially at the rain-dominated region in summer. However, more winter rainfall and earlier snowmelt by increasing temperature is likely to increase drought risk at a transient region and snow-dominated region because of reduced water storage for summer use. We investigated possible changes in future climatological drought over the Willamette River Basin. The statistically downscaled 16 climate simulations derived from eight GCMs with two emission scenarios (Chang and Jung, 2010) were used to calculate relative Standardized Precipitation Index (SPI) (Vidal and Wade, 2009; Dubrovsky et al., 2009). The relative SPI can assess the spatial and temporal change of drought frequency at different lasting time scales, 3-, 6-, 12-, 24-month. Multimodel ensemble results projected increase in drought frequency under the A1B and B1 GHG emission scenarios. In particular, short-term 3- and 6-month droughts are likely to increase highly over the Willamette Valley region and the Western Cascade region for the 2080s (2070-2099) due to decreased summer precipitation. Long-term droughts, 12- and 24-month, however, are not projected to change except some of the Willamette Valley region because winter precipitation is projected to increase in these areas.



3.5 Impacts of Climate Variability and Change on Water Quality

3.5.1 Water Temperature

Although numerous studies have examined potential consequences of climate change on river flow (Arnell 2004; Payne et al., 2004), relatively few studies have investigated how river water quality might change in response to warmer air temperature and changing patterns of precipitation distribution (Murdoch et al., 2000, Chang et al., 2001). Water temperature is the most important indicator of stream health; it directly affects the amount of dissolved oxygen in water, which is critical for fish survival. Water temperature also indirectly affects the overall health of streams through its influence on in-stream biogeochemical cycles. In the Pacific Northwest (PNW), summer water temperature is critical for the survival of cold-water species like salmon. Studies have shown that ranges for cold-water fish would be displaced northward with a loss of habitat because cold-water species cannot adapt quickly to abrupt environmental changes (Mohseni et al., 2003).

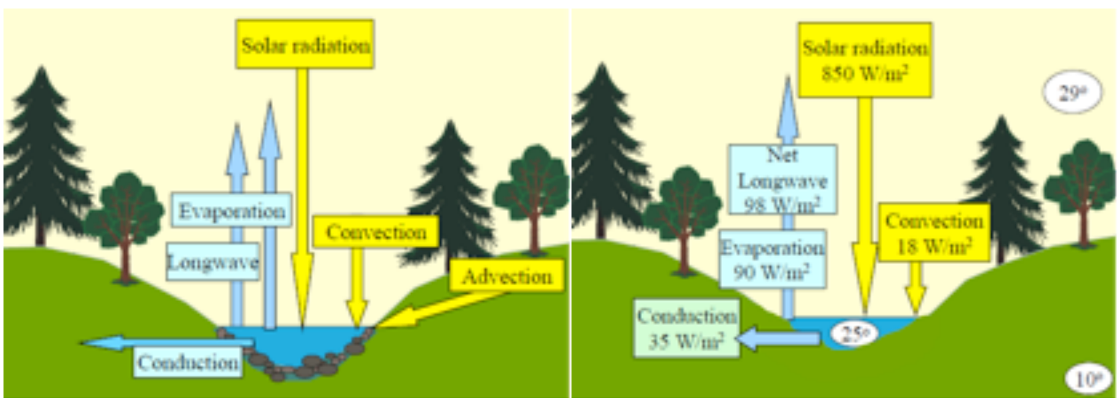


Figure 3.21 Major factors influencing stream temperature and example of heat budget for hot summer day at noon.

Changing climate could affect stream temperature through several mechanisms (Figure 3.21). Although air temperature and water temperature are correlated (Webb, 1996), variation in air temperature alone does not lead to major changes in stream temperature in Oregon streams (Johnson, 2003; Johnson, 2004). The largest effect of climate change on stream temperature may occur indirectly through climate-induced modification of riparian vegetation (although it might take a while), which provides shading for streams, and influences streamflow timing and amount. In the Willamette River basin approximately 35% of runoff comes from the snow pack. This results primarily from increased soil moisture earlier in the spring (Burns et al., 2007). Declining snowpacks could reduce spring and summer streamflow, which thus could increase summer stream temperatures. Low streamflows coincide with maximum summer air temperatures, and if streams become standing pools, temperatures can greatly increase. The degree of temperature change will depend on how much streamflow will decrease in the future and the degree to which stream temperature depends on discharge. Stream discharge could further decline as elevated summer air temperature accelerates the rate of evapotranspiration, which will have harmful effects on freshwater habit of Pacific Salmon (Mantua et al., 2010).

Land use also affects stream temperature (Krause et al., 2004; Johnson and Jones 2000). A reduction of riparian buffers, whether from urban, agricultural or forest land use practices, and the shade they provide leads to increased levels of solar radiation and increased stream temperatures. Urbanized landscapes with high levels of impervious surfaces can absorb more heat energy than rural landscapes, which also increases surface air temperatures. This effect may be more pronounced during the spring and early summer months with runoff flowing over the hot surfaces into streams. These overland flows can cause short-term spikes in water temperature (Nelson et al., 2007).

While there is a growing concern about potential changes in water quality resulting from climate change, only a few studies have investigated this topic in the Pacific Northwest (PNW). Based on an analysis of two PNW streams, Cristea and Burges (2010) showed that stream temperature increases depend on reductions in summer streamflows rather than increases in air temperature. Johnson and Jones (2000) documented a return to preharvest stream temperatures with recovery of riparian vegetation. Tague et al. (2007) examined water temperature variations in a High Cascade and a Western Cascade basin using historical water temperature data. Their results suggest that geology and the source of stream water controls summer water temperature dramatically, with less seasonal variations in water temperature in sub-surface flow- dominated High Cascade watersheds.

Water temperature in Oregon is highest in the summer, with forested headwaters generally having cooler temperatures than larger downstream sites (Figure 3.22). Winters are times of lowest temperatures, and maximum temperatures generally occur in July, August or September, depending on year-to-year weather patterns, discharge trends and timing of shading.

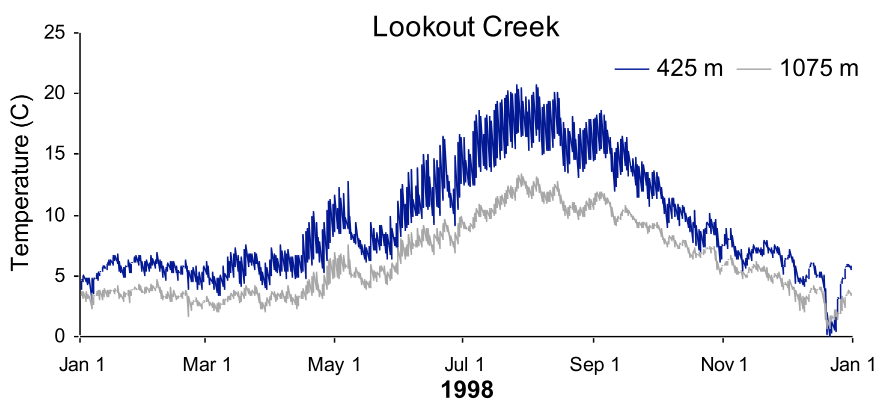


Figure 3.22 Daily maximum and minimum water temperature during 1998 for 1st and 5th order Lookout Creek in the Cascade Mountains.

Figure 3.23 illustrates a general spatial pattern of maximum water temperature during summer (June to September) for selected stations for a period of 1999 and 2008. As shown in this Figure, most basins draining the Cascade Range experience low maximum water temperature, while urban and mixed watersheds (e.g., Johnson Creek and the Tualatin River Basin) exhibit higher than average water temperature. Basins located in southwestern Oregon also experience higher than average water temperature.

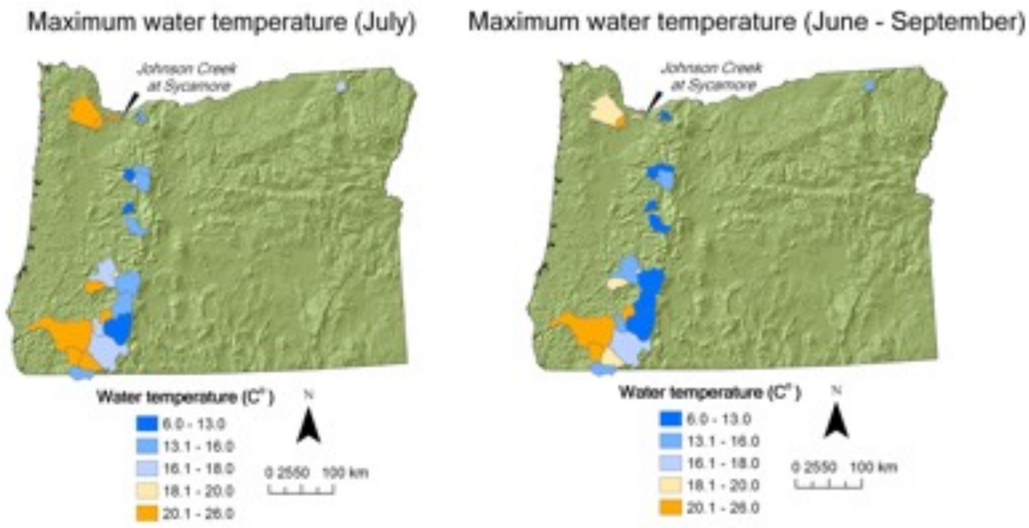


Figure 3.23 Average annual maximum daily water temperatures for July and summer (June to September) for 31 basins in Oregon, 1999-2009.

Long-term trends of water temperature in undisturbed watersheds are rare. Many sites with long-term data have been impacted during the period of record by upstream impoundments or land use changes, which confounds our ability to detect climate related changes. Data from stream gaging stations in the Rogue River basin (Rogue River near Mcleod, station #14337600) and in the Willamette River basin (Blue River at Blue River, station #14162200, and North Santiam River at Niagara, station #14181500) show general increasing trends in maximum water temperature for two of the three stations, particularly in the North Santiam River (Figure 3.25). High variability in August and September water temperature at Blue River appear to be associated with flow regulations in late summer months.

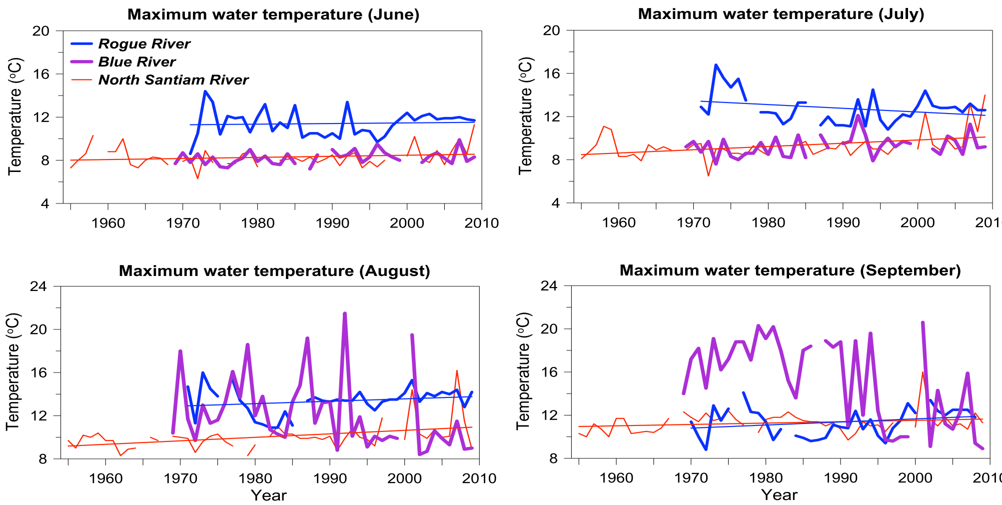


Figure 3.24 Long-term trends of maximum water temperature for Rogue River near Mcleod, Blue River at Blue River, and North Santiam River at Niagara.

Figure 3.25 shows trends over time in maximum water temperature in summer, based on records from 1999 to 2009. Only 6 of 36 stations show increasing trends. According to Mann-Kendall's test, one station located in the Willamette Valley shows a downward trend in summer maximum water temperature. The 7-day average daily maximum temperature, currently used for assessing water temperature threshold for fish habitat (e.g., lethality and migration blockage conditions), increased at 5 stations which are all located in the Portland metropolitan area. As shown in Figure 3.26, the variability of water temperature in Johnson Creek increased over the past 10-year period, suggesting that the stream frequently exceeds the threshold level of 18°C. A comprehensive assessment would be required to determine the degree of disturbance to fish habitat in urban streams.

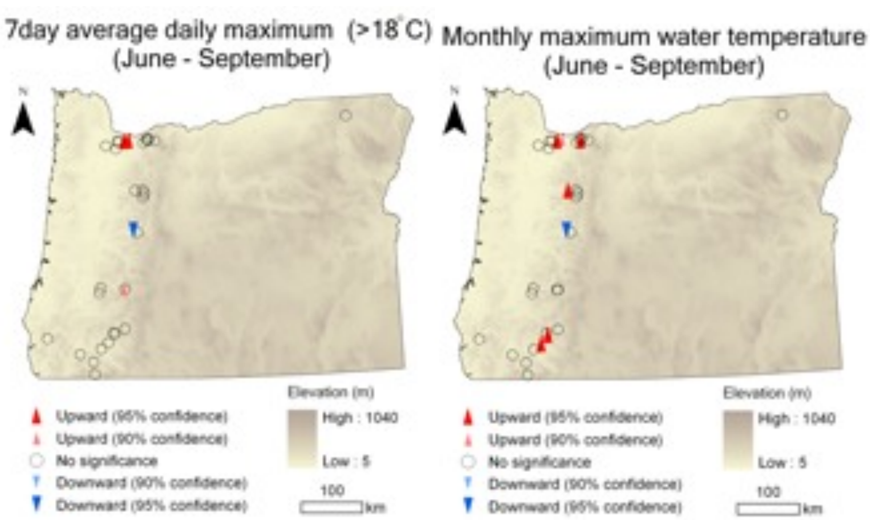


Figure 3.25 Trends in maximum summer water temperature and 7-day average daily maximum temperature for 31 stations in Oregon

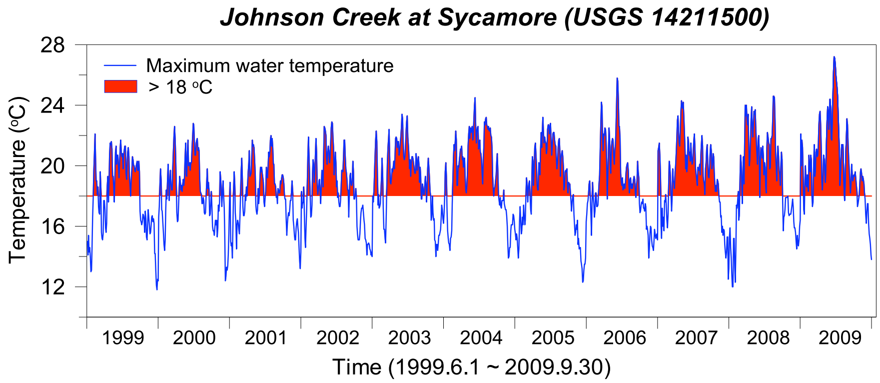


Figure 3.26 Maximum water temperature for Johnson Creek at Sycamore, 1999-2009

A preliminary study of the three streams - Tualatin, Johnson and Clackamas River - in the Portland metropolitan area shows that the degree of landscape disturbance further elevates stream water temperature (Chang and Block 2009). While lagged air temperature can explain approximately 66% of variations in summer water temperature on average, the size of stream, the amount of flow, and the upland hydrological processes also affect water temperature variations (see Figure 3.27).

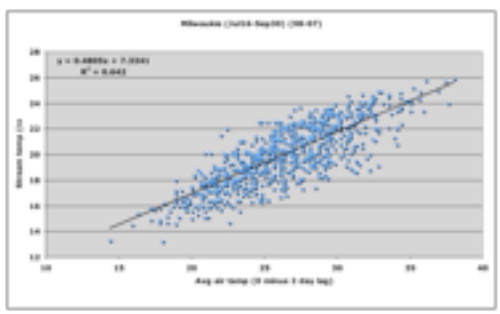


Figure 3.27 Relation between streamflow and water temperature and lagged air temperature and water temperature at Milwaukee in Johnson Creek, 1998 - 2007.

Major controls on stream temperature are riparian vegetation (through shading) and streamflow (which influences heat exchange). Climate warming may increase stream temperatures by reducing riparian vegetation, or by reducing snowpack and spring and summer discharges. Low elevation watersheds in areas of agricultural or urban land use, which are already temperature limited, may be most susceptible to climate-warming-induced increases in stream temperature.

Figure 3.28 shows potential changes in water temperature in the mainstem of the Tualatin River located in the Portland metropolitan area. These results are based on CE-QUAL-W2 simulations for 154 segments in the lower Tualatin River under three combinations of temperature, flow, and riparian scenarios. The nine maps show the number of days that 7-day daily average water temperature exceeds 20˚C between May 15 and October 15 under each scenario. Under the baseline scenario (top left), only the lower segments of the drainage experience water temperature above 20˚C. Under water temperature reduction scenarios, with reductions due to revegetation in tributary riparian zones, only a few immediate segments are affected. Under 5% flow reduction and 1.5˚C air temperature rise scenarios (representing the 2040s), segments with water temperatures in excess of 20˚C for more than 60 days expand to include some upstream areas. Under 10% flow reduction and 3˚C air temperature rise scenarios (which represents the 2070s), they expand further into upstream areas. Riparian vegetation scenarios have the most direct impact on middle segments of the drainage under the highest warming scenario.

In summary, there is little evidence to date of increasing stream temperatures over time in Oregon, except in urban streams, where temperatures may have increased because of cumulative loss of shading from riparian vegetation associated with urban and suburban development. Only a few dozen long-term records of stream temperature exist in Oregon. Existing studies have demonstrated that stream temperatures depend on riparian vegetation cover as well as air temperatures and discharge, which are inversely related. Future changes in stream temperature in response to climate change in Oregon will depend on the degree to which warming results in a reduction of late summer streamflow and how warming influences riparian vegetation. The resulting effects are complex. Warming temperatures may increase late-summer evapotranspiration from riparian vegetation, potentially reducing late summer flow; smaller snowpacks and earlier snowmelt may further reduce late summer streamflow. If streamflow in late summer is reduced, with no changes in riparian vegetation cover, stream temperatures may increase. However, increases in riparian vegetation cover (from stream restoration) could partly counteract these effects. In addition, stream temperature increases from typically forested headwaters, where groundwater contributions are important, to typically agricultural or urban downstream areas, so stream temperatures in downstream areas may be more sensitive than headwaters to future climate-related warming.

3.5.2 Sediment and Nutrients

As sediment and phosphorus loadings typically increase during high flow events, changes in flow variability are expected to alter temporal variability of sediment and phosphorus loadings. A case study in the Tualatin River basin (TRB) of Oregon illustrates that winter sediment nutrient loadings are expected to increase under climate change scenarios as winter flows are projected to increase (Praskievicz and Chang, 2011). Although diminished summer flow is likely to reduce summer nutrient loading, the annual load is expected to increase further with urban development scenarios.

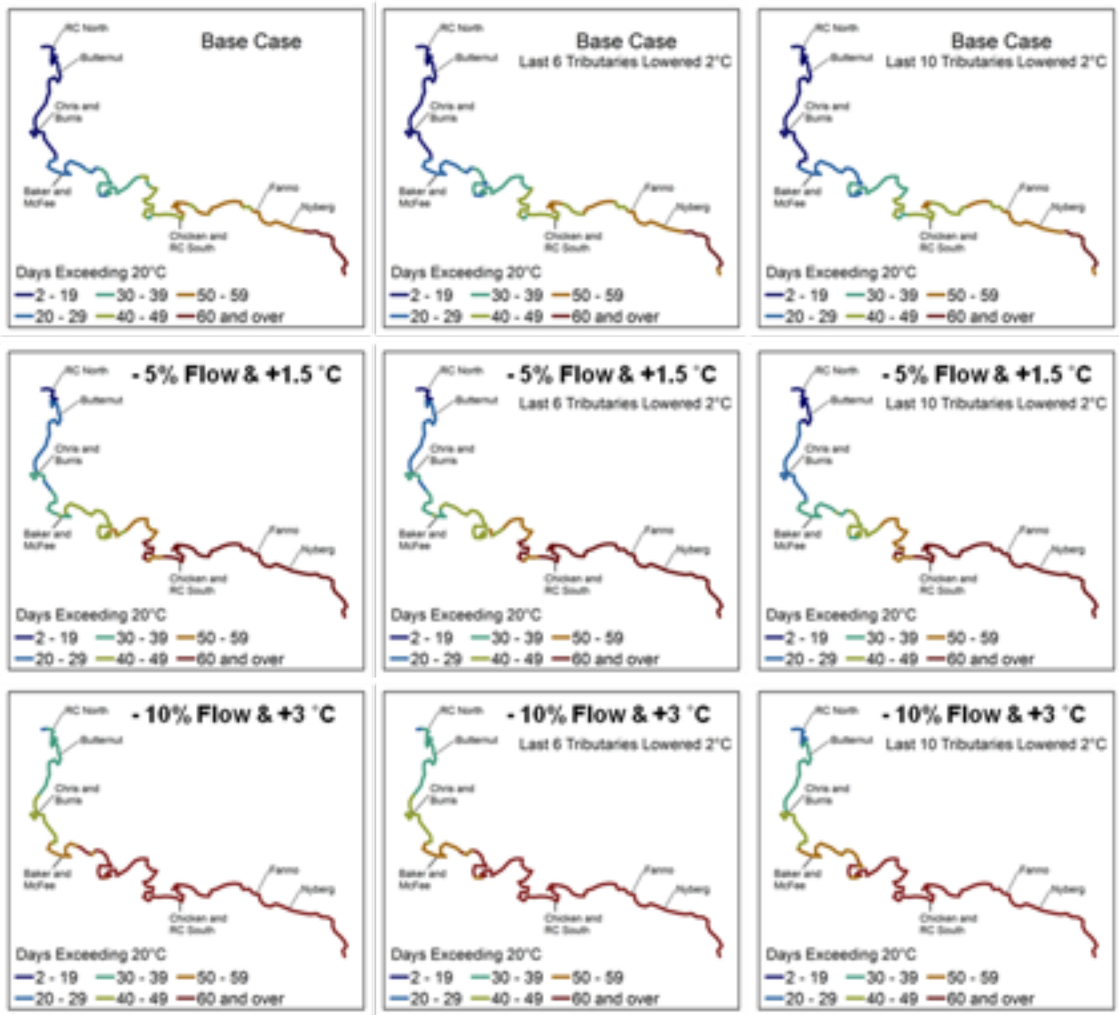


Figure 3.28 Change in water temperature under climate change and tributary riparion vegetation scenarios (Source: Chang and Lawler 2010).

However, conservation-oriented development could reduce erosion and phosphorus loading substantially compared to conventional development. The combination of climate change and urban development scenarios generally produce hydrological and water quality results that track the results from climate change alone, suggesting that the water resource impacts from climate change are more significant than those from land use change in the TRB. The development and conservation scenarios do differ in their hydrological and water quality outcomes, thus representing a potential opportunity for local adaptation to climate change by pursuit of sustainable forms of urban development.

3.6 Impacts of Climate Variability and Change in Water Demand

Municipal water demand patterns have progressively become a greater concern to urban water resource managers due to changes in climate and the expansion of urban areas in many parts of the world during the 20th and early 21st centuries. The recent Intergovernmental Panel on Climate Change report (IPCC) also projected an increase in temperature and spatial and temporal variability of precipitation, which may increase water demand but reduce seasonal water supply (Kundewicz et al., 2007). Although many North American cities have recently implemented conservation measures which have reduced water consumption per capita (Gleick, 2003), growing municipalities located in arid or semi-arid regions or areas prone to drought are increasingly apprehensive about the sustainability of their water resources (Morehouse et al., 2002; Kenney et al., 2008). Even for cities located in relatively temperate climates, such as the Pacific Northwest of North America, potential seasonal changes in runoff due to climate change pose another stress in the sustainability of water resources (Palmer and Hahn,2002; VanRheenen et al., 2003; Palmer et al., 2004; Graves and Chang, 2007). Residential water consumption is a key factor that could affect water availability at the local and regional scale (Gutzler and Nims, 2005; Balling and Gober, 2007).

While previous studies suggest the existence of threshold values of climate variables that affect the sensitivity of urban water consumption, few examined the complex relation between water consumption and climate variables at multiple temporal scales. Water consumption research is typically constrained by a lack of detailed data to draw from; however, a rich dataset of long-term daily water data was available for the preliminary investigation of Portland water consumption. To draw meaningful inferences on water consumption as it relates to climate variability and projected climate change, multi-scale analysis is needed. Multi-scale temporal analyses allow us to project short-term and long-term water demand forecasting based on the fluctuations of climate variables. Water resource managers need not only seasonal climate but also daily weather information as they relate to water supply and demand (Steinemann, 2006).

Here we examined the relationship between urban water consumption and climate variables at daily, monthly, seasonal, and annual scales using 50 years of historical water production data from Portland, Oregon as a case study. Additionally, we also used customer demand monitoring data for a finer temporal analysis at a household level for one specific summer year. This study is a unique investigation concerning the sensitivity of urban water consumption to climate variables as the scale of analysis changes. It will provide useful climate information for urban water resource managers as it relates to water consumption. Urban water managers may be able to use such information to establish proactive plans under increasing pressure from climate change (Ruth et al., 2007; Praskievicz and Chang, 2009c).

3.6.2 Inter-annual Climate Variability and Water Consumption

Water consumption in the Pacific Northwest is highly dependent on weather variations. The annual consumption pattern shows increase in water consumption in warmer and drier months, and in warmer and drier years. Figure 3.29 shows average annual water consumption by retail and wholesale customer classes of the Portland Water Bureau (PWB) along with maximum daily temperature, averaged over the year, measured at the Portland Airport (PDX) weather station. Although consumption is related to population, conservation, land use, and other economic and demographic factors, the inter-annual fluctuations in consumption also are related to maximum air temperature. One exception is the spike in temperature in 1992, which coincided with a dip in consumption. This dip was the result of mandatory curtailment imposed by the PWB during the summer of 1992, when a water shortage occurred due to lack of access to the existing groundwater supply along the Columbia River South Shore, an issue which has now been resolved.

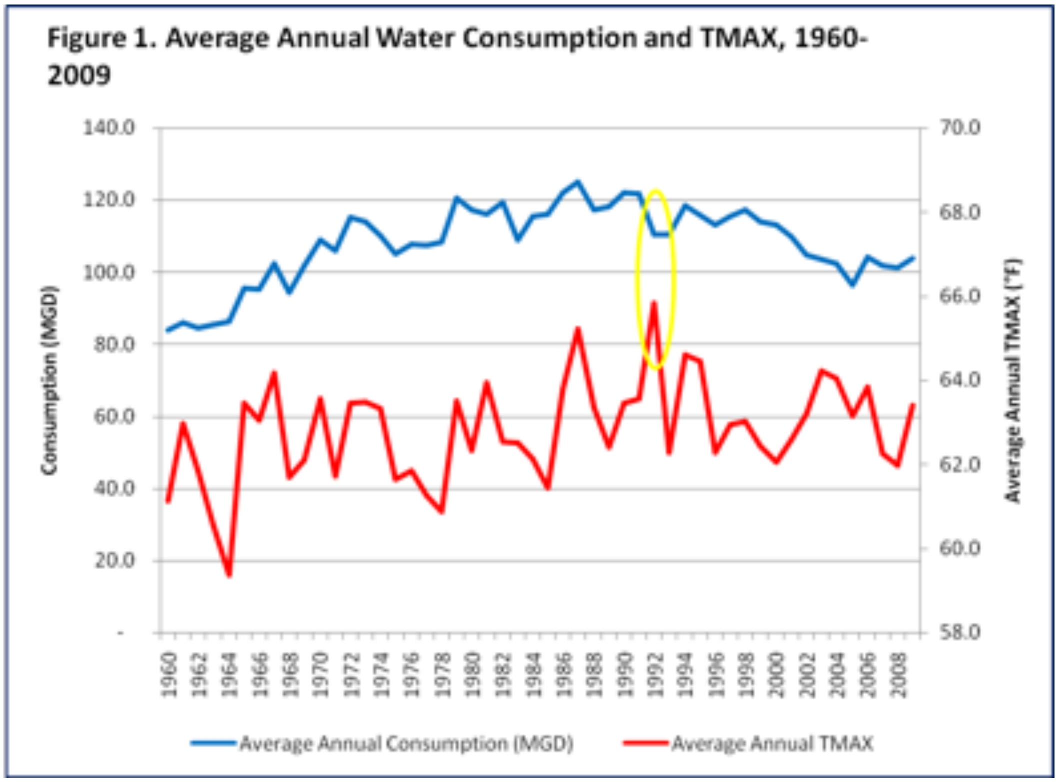


Figure 3.29 The relation between average annual water consumption and average annual maximum temperature.

Consumption has an inverse relationship with total annual precipitation, however, not as strong as that of temperature. Figure 3.30 shows dips in average annual consumption in years that are very wet.

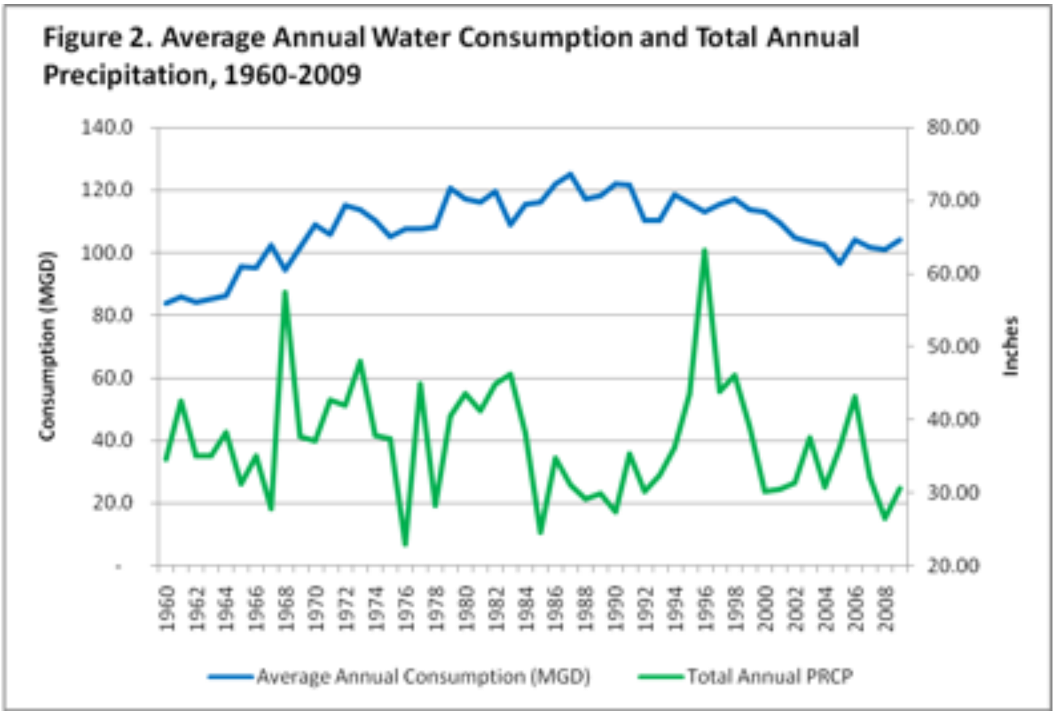


Figure 3.30 Average annual water consumption and total annual precipitation, 1960-2009

3.6.3 Seasonal Climate Variability and Water

Water consumption shows a strong seasonal pattern in the PWB service area. The Bureau recognizes June to September as peak season based on empirical data observations. The annual figures over the 1960 - 2009 period show higher consumption during peak season relative to off-peak in the range of 113% - 176%. Although peak season consumption also depends on economic and demographic factors, it is affected by inter-annual climate variability as well. Figure 3.31 depicts the Peak/Off-Peak ratio and maximum daily temperature, averaged over the year. Again, with the exception of 1992, spikes in the relative peak-off-peak consumption mostly coincide with those of the TMAX.

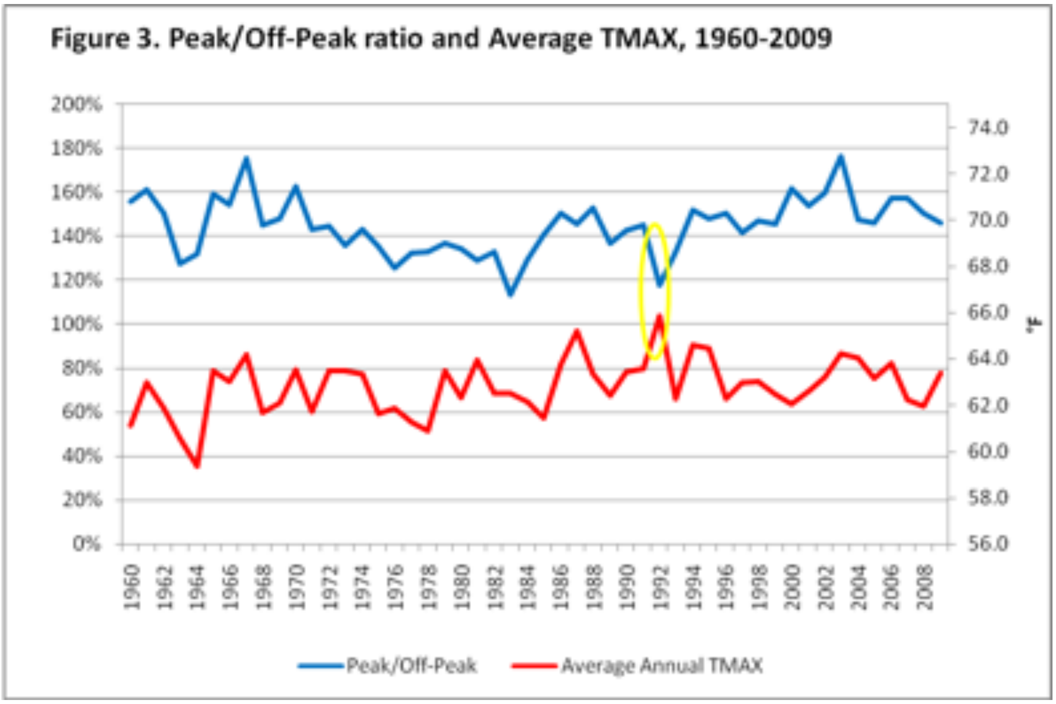


Figure 3.31 Peak/off-peak ratio and average maximum temperature, 1960 - 2009. The peak/offpeak ratio is water consumption in June to September (dry season) divided by water consumption in October to May (wet season).

3.6.4 Daily weather variability and water consumption

Daily fluctuations in consumption are closely related to daily fluctuations in weather. In fact, a simple regression of daily consumption on maximum daily temperature and total daily precipitation shows that 51% of variation in daily water consumption can be explained by total daily temperature and precipitation. Obviously, part of the explanatory power is due to seasonal patterns in both consumption and weather. That is, no matter how hot or cold the summer months are, there will be increases in water consumption due to change in the season. The daily effect of weather above and beyond the seasonal effect can be measured by considering the deviations of daily temperature and precipitation from their historical mean. A more sophisticated regression model, which includes seasonal, economic, demographic, and weather variables in form of deviations from historical means, shows how variations in consumption can be disaggregated to show the effects of these variables. A model developed by PWB shows that daily weather variations explain about 13% of the daily variations in consumption, above and beyond seasonal changes.

There was also wide variation in the average amount of water used by individual households from day to day. Overall, the average daily volume of water used by a household was greater in summer months than in winter months. This corresponds with Portland's cool, wet winters and dry, temperate summers. Average daily household water consumption was also greater on weekend days, which is logical given our sample of residential properties: some residents likely spend much of their weekday time out of the home at work or other locations. Consequently, water-consumptive activities of maintenance (washing clothes and cars) and recreation (gardening and water play) are more likely to take place on weekends.

A multilevel model results suggest that the most important determinants of household water consumption is daily maximum temperature, followed by day of the week (weekend or not), building size and building age. These factors correspond to other previous studies in the same region (Shandas and Parandvash, 2010; Chang et al., 2010). A model suggests that water use increases by 27 liters/household for every 1º C increase with an increase in daily maximum temperature. It requires an 86.5 ft^2 increase in building area to increase water use by an amount equivalent to a 1º C rise in daily max temperature. Larger houses have a greater water consumption than smaller houses on hot days. Thus the larger the house, the greater the increase in water use with rise in daily maximum temperature.

Case study: water demand in the city of Hillsboro: a spatially-explicit assessment

Urban residential water consumption is significantly affected by both interannual seasonal climate variability and periods of drought. In Hillsboro, Oregon, a rapidly growing suburb of Portland, a statistical analysis of single-family residential water records for the period 2004-2007 found that water consumption throughout the entire study area exhibited significant sensitivity to interannual climate variation (House-Peters et al. 2010). Sensitivity to interannual climate variability was manifested as increased household water consumption during the summer season, as compared to the winter season, due to increased external water use for irrigation, pool maintenance and car washing during hot, dry weather. Furthermore, sensitivity to interannual climate variability was spatially heterogeneous throughout the study area. Census blocks displaying the largest magnitude of increased summertime water use, up to 2.2 times greater than winter use, had newer and larger homes, higher property values, and more affluent and well-educated residents. This research also examined water consumption during a drought summer in 2006 when the study area recorded only half as much precipitation as the 30-year mean and exceeded the 30-year mean's average maximum temperature by one degree Celsius (National Climatic Database, station #353908). Although water use across the entire study area did not demonstrate sensitivity to the drought conditions, particular census blocks were highly sensitive to the drought, consuming up to 1.85 times more water for external purposes during the drought summer than an average summer. Interestingly, during the summer characterized by reduced precipitation and higher maximum daily temperatures, external water use was found to be more dependent on physical property characteristics and less dependent on socio-economic characteristics. These results suggest that strategic urban planning and neighborhood design may be able to reduce stress to the water supply system during peak summer demand and future drought episodes.

3.6.5 Conclusions

Statistical analysis of daily water consumption per capita in the studies cited above shows that determining which climate factors are the most influential to consumption per capita is highly dependent on the scale of study. While both precipitation and temperature are significantly associated with water consumption at all scales, the influence of temperature is stronger than that of precipitation on water consumption at the monthly scale. Other hydro-climatic and social behavior variables, such as humidity and social activities, could be also potential factors that affect the variations in water consumption. As soil moisture depends on both precipitation and evaporation, it is important to include humidity as part of water demand modeling, particularly outdoor water use such as lawn irrigation and recreational activities. Changes in lawn irrigation behavior thus can also be an important factor that might influence irrigation water demand (House-Peters et al., 2010). At a daily scale, our multilevel model can provide more nuanced information about the interacting effects of water use, structural attributes, day of the week, and temperature. These results imply how changing temperatures and demographics can lead to development patterns that exacerbate or conserve regional water resources.

This multi-scale analysis of urban water consumption illustrates complex interactions between urban water consumption and climate variables depending on the scale of analysis. It demonstrates what climate information would be useful for short and long-term water consumption forecasting. Urban water resource managers may be able to use such information for establishing proactive water resource management strategies under increasing pressure from potential climate change. While many municipalities in Oregon have prepared water management and conservation plans with supply focus (Bastasch ,2006), now is a time to put climate change into water resources planning at multiple levels.

3.7 Projected and Observed Impacts of Climate Change on Hydrosystems

Oregon is blessed with a varied and ecologically diverse environment; consequently, parts of the state are either too wet or too dry to support many human activities without modifying the natural water supply. These modifications include an extensive series of engineering projects, including reservoirs, dikes and levees, and diversions, to meet a number of sometimes conflicting objectives for Oregon's hydrosystems, including flood control, irrigation and municipal supply, hydropower production, recreation, and recovery of threatened and endangered species. Meeting these objectives in the future is likely to become increasingly difficult as climate and land use change, combined with population growth, alter the demands on and supplies of the water system.

A number of groups have investigated the projected impacts of a changing climate on hydrology and streamflow that represent changes in supply (Udall and Bates, 2007; US Geological Service, 2005; Stewart et al., 2005; Regonda et al., 2005; National Assessment Synthesis Team, 2001; Knowles et al., 2006; Ray, 2008; Baxter and Hauer, 2000; Warren et al. , 1964). Also relevant are projected changes in water demands (NWPCC, 2005; Voisin et al., 2006).

These changes in supply and demand are likely to have important impacts on water infrastructure and the built environment. Regional frequency results indicate that an increased frequency of higher streamflow events (i.e. Figure 3.32) can be expected for most areas of the PNW region, though this pattern is expected to vary spatially (Rosenberg et al., 2009a; Kunkel et al., 1999; Pryor et al., 2009; Madsen and Figdor, 2007). Small basins with a large proportion of their area at the midwinter or transient snow line are likely to be most vulnerable to climate changes (Mote et al., 2003).

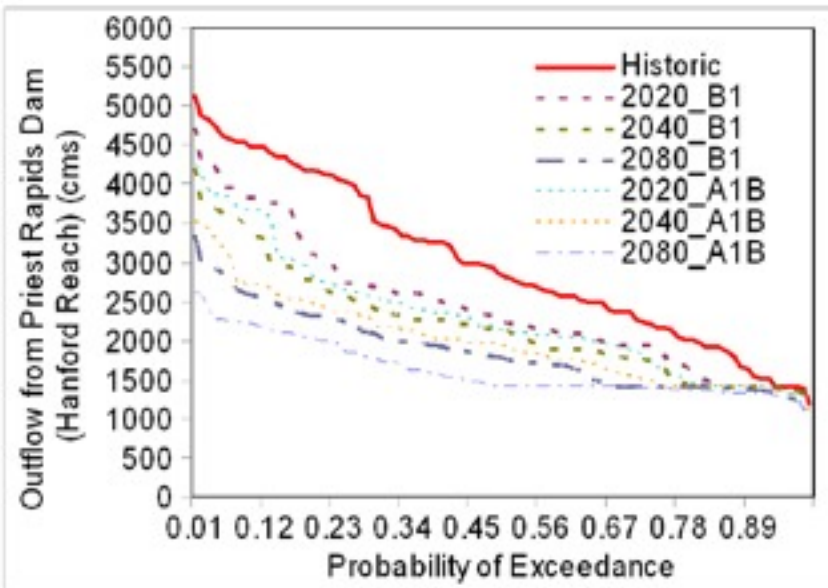


Figure 3.32 Predicted changes in extreme events for the Columbia River Basin. Figure reprinted from Hamlet et al., 2009.

Despite the value of trends projections from GCMs, the variability between GCMs challenges the design of some urban infrastructure (Rosenberg et al. 2009b). As an example, the variability between models represents one characteristic challenge in designing new and retrofit infrastructure (e.g. stormwater) based on the magnitude of the 24-hour extreme. Further, the anticipated changes are likely to be beyond what the current water infrastructure can reliably manage. For example, achieving a reduction of winter flood risk with future increases in peak flows will likely require strengthening dikes and levees, restoring floodplains, improving flood forecasting, changing reservoir management, improving emergency management, and modifying land use policies and flood insurance. With regards to low flows, it is projected that a reduction of summer water will require diversification and development of water supplies, reducing demand, improving efficiency, operational changes at reservoirs, increasing water transfers between users, and increasing drought preparedness (Binder et al., 2009). Thus, maintaining water infrastructure in future climates and land uses is likely to require new design and management approaches to address the potential challenges posed by uncertain changes in both supply and demand.

In this section we further explore these potential impacts of climate and land use changes on water infrastructure in Oregon, leading to questions regarding: How might water management objectives (e.g. hydropower production, flood protection, recreation, municipal and irrigation supply, and instream flows) be impacted by climate change? What indicators and measurements are relevant to evaluating climate change impacts on meeting water management objectives? What adaptation strategies are groups considering and what tradeoffs may be necessary for new water management objectives?

3.7.1 Impacts on Water Infrastructure

A number of climate-related impacts could have important consequences for water infrastructure and management. For example, fish recovery plans may need to be revised, fire risk may increase due to vegetation shifts, infrastructure (e.g. crossings, conduits, landslide risks) could be impacted by increased peak flows, and water and/or reliance on groundwater could increase.

In some areas, impacts to water infrastructure include (also see Miller and Yates, 2007).

- intrusion of bromine/iodine from seawater, leading to problems meeting disinfection by-product rule compliance (AWWA, 2001)
- increasing potential for floods to exceed stormwater systems, leading to water contamination from combined sewer overflows (Ashley et al., 2001)
- increased potential for floods to damage infrastructure (Filion, 2000)
- increased fire frequency leading to increased sediment loads to water treatment plants

Recent work has focused on the development of hydrological projections for the PNW in the context of water resources management (Vano et al. ,2009a and 2009b; Wiley and Palmer, 2008; Trayham, 2007; Polebitski et al., 2008). Together, these studies indicate that shifts in hydrology towards higher flows in fall are likely to variably impact dam operating objectives. For example, Wiley et al., (2008) predicted that April snowpack will decline from 2000 to 2040 and that peak snow accumulation will shift from March to earlier in the year. This is anticipated to result in a decline in fall reservoir storage. In another example, Vano et al., (2009a) used downscaled GCM A1B and B1 emission scenarios as inputs to DHSVM to produce streamflow simulations for the 2020s, 2040s, and 2080s. These results were then input into water resources model run at a daily time step. Their results project a transition from a double peaked hydrograph (December and mid-May snowmelt peak) to a single peak in December (elevated winter runoff). By early April to the end of March, all future scenarios show less water than historical conditions, due to earlier snowmelt. Despite this change, reliability (the ability of the system to meet demands including instream flow and consumptive use) remains at 100% for historical simulations of Seattle, Tacoma, and Everett reservoirs and only drops below 98% for the warmest and driest scenarios (CCSM3 and ECHO_G). However, in their models, reliability decreased markedly if increases in demand are factored into analyses.

Most models agree that instream flow requirements will be increasingly difficult to meet in the more distant future as demands increase and available storage declines. In Puget Sound reservoirs, Vano et al. (2009a) and Wiley and Palmer, (2004) found that instream flows in summer will not be significantly impacted by climate change in the near future as instream flow requirements set by the Habitat Conservation Plan (SPU, 2000) are lower than typical observed flows. However, instream flows are projected to be adversely impacted for the most severe climate scenarios for the 2020s and for A1B and B1 scenarios in the 2040s and 2080s. Without substantial infrastructure changes, tradeoffs are likely to be necessary between hydropower generation, instream flows for fish, reservoir storage, and flood protection. For example, regarding instream flows, the Hamlet et al. (2009) studies suggest that the use of larger reservoir storage will be required to mitigate/supplement summer low-flows for reducing tradeoffs between fish flows and firm energy resources. Along these lines, Payne et al. (2004) found that trade-offs were unavoidable in the winter due to limited reservoir storage level.

It is likely that changing supply and demand of water will require reconsideration of established rule curves to more reliably meet flood protection, instream flows, and hydropower objectives. For example, projections (Figure 3.34) indicate that increased potential for hydropower production exists through the winter months while decreased production is likely through the summer months as we move further into the future. These results are based on simulation of streamflow for the Columbia River Basin (Hamlet et al., 2009) using the VIC hydrologic model (Liang et al., 1994) and methods described by (Elsner et al., 2009). Their objective was to investigate adaptive responses with regard to flood control, using fixed (assuming no change in demands) energy targets for firm and non-firm energy. They found that, for the 2020s, winter hydropower increases 0.5-4% and summer hydropower decreases 9 - 11%, with net decreases of 1 - 4%. By the 2040s, winter production will increase 4% and summer production will decrease 2.5 - 4%. By 2080s, winter production will increase 7 - 10% and summer production will decrease 18 - 21%. The largest reductions in hydropower generation, as compared with 20th century values, are likely to occur July-Sept, coincident with peak seasonal air conditioning loads (Voisin et al., 2006; Westerling et al., 2008).

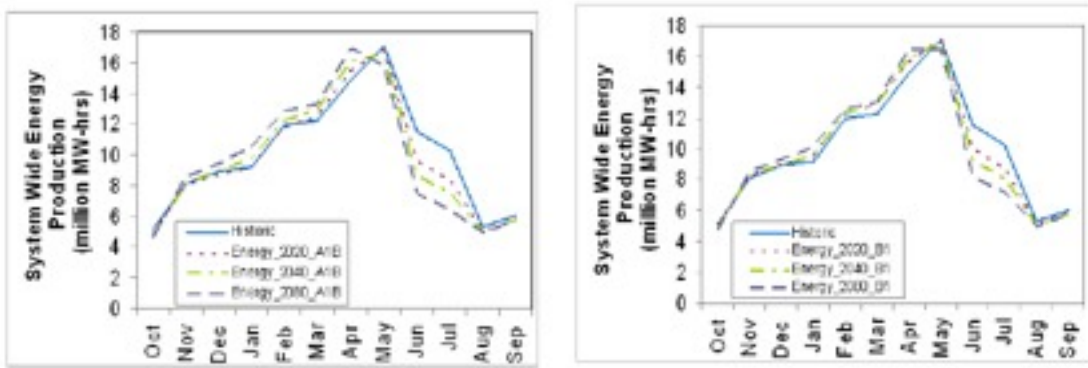


Figure 3.34 Projected Impacts to hydropower production on the Columbia River. Simulated long-term mean, system-wide hydropower production from the Columbia River basin for A1B scenario (top panel) and the B1 scenario (bottom panel). Figure reprinted from Hamlet et al., 2009.

Mote et al. (2003) found that the reliability of energy production remains high in future scenarios but that reliability of flood control diminishes with increased precipitation (Figure 3.34).

In addition to direct changes in flow, synergistic changes in water temperature are likely to influence reservoir operations (Battin et al., 2007; Mantua et al., 2010). Thermal stress periods are predicted to be minimal until 2020s and then subsequently increase (Mantua et al. ,2009). These stress periods will occur during reduced summer flows and are likely to influence both dam operations and infrastructure needs (e.g. temperature control devices). The complex and variable nature of hydrologic, and consequent ecosystem changes, suggests that current strategies and regulations that govern operations of reservoirs in static and uncoordinated ways may inhibit the necessary flexibility required to manage water resources in a changing climate.

3.7.2 Adaptations and Tradeoffs

As agencies and communities adapt to a changing climate, tradeoffs are going to be inevitable. Some of these tradeoffs will be specific to the local hydrogeology and climate while others will be based on social values or existing legal requirements (e.g. ESA). For example, while hydropower is a very inexpensive energy source and accounts for 70% of energy use in the PNW (Hamlet et al., 2009; Binder, 2009) the amount of hydropower generated is controlled by water availability and not demand (Hamlet et al. 2009). Thus, some are considering additional storage (Snover et al., 2007) to meet competing water demands. Indeed, important trade-offs between fish flows and firm energy resources in winter may require the use of much larger reservoir storage (Hamlet et al., 2009; Barnett et al., 2004). In one example, researchers (Payne et al., 2004) examined tradeoffs in alternative reservoir operation strategies. They found that greater storage to capture earlier reservoir refill could help meet instream targets, though at a cost (9 - 35% loss) of firm hydropower production.

3.7.2.1 System optimization

Optimization has been applied to evaluate tradeoffs in operations for multiple-reservoir, multi-purpose hydrosystems for many years (Needham et al., 2000). However, some major challenges exist in basing operations solely on optimization outcomes. Labadie (1997) reviews reservoir optimization models, with discussion on the lag between theoretical developments and implementation. Some of these challenges include establishing relative value or priority among multiple objectives, scaling and time steps/lags (changes in flow now vs. responses of ecosystems vs. other costs/benefits), defining and measuring system health/impairment and response/performance. These factors require a high degree of model complexity, which significantly increasing costs of model development, ease of use, the reliability of its output (Van Lienden and J. Lund, 2004), and the difficulty in addressing deviations and risk. Further, an increase in the frequency of extreme events (floods or droughts) will make it more difficult to meet multiple objectives under an optimized system. For example, reservoirs in the Willamette Basin, designed to meet flood protection, municipal water supply, and Biological Opinions requirements are generally in concert. However, during high flow events, conflict between objectives could occur when, for example, ramping rates required by the Biological Opinions may not be met to satisfy flood protection objectives.

3.7.2.2 Current planning and design practices

Fundamental challenges emerge for water managers when applying the results of GCMs in water infrastructure design. These challenges result from the current design paradigm for water resources engineering, which is based on frequency distributions of precipitation and runoff and assumes stationarity. For example, design of stormwater infrastructure is frequently based on precipitation events of 1-hr to 24-hr duration, and 2 to 25 year return frequency (Osman, 1993). Predictions at this scale are especially important in urban areas, where smaller lag-to-peak and flashy streams are common (Rosenberg et al., 2009). Continuing to design stormwater facilities in this way may require GCMs to be downscaled to hourly precipitation timestep and low spatial resolutions (e.g. ~20km) before input into rainfall-runoff models (Rosenberg et al., 2009), a process that is highly vulnerable to uncertainty. The alternative is the development of a new design paradigm that is not based on frequency distributions. Advantages of various downscaling approaches in the Pacific Northwest have been reviewed (Salathé et al., 2005), and Avise et al., (2006) present some advantages of using regional climate models for various impacts applications. Neither solution has been comprehensively demonstrated or is without uncertainties. Thus, managers will need to consider the resolution of data required for hydrodynamic and management models and either downscale (with a quantitative understanding of uncertainty) or adjust their management models to operate at a coarser scale.

A second challenge regarding engineering design practices is the assumption of stationarity. In addition to key challenges in assuming hydrological stationarity (Milly et al. 2008), nonstationarity in ecological and social drivers is also likely. For example, planning for future energy demands is challenged by unpredictable changes in carbon pricing and policies that may influence where and when power is produced from reservoirs, the availability of alternative energy supplies, the increased importance of managing reservoirs for temperature in both the summer and winter, changes in populations and/or demographics, and the lack of stationarity in ecological systems' responses to changing landscapes and hydrology.

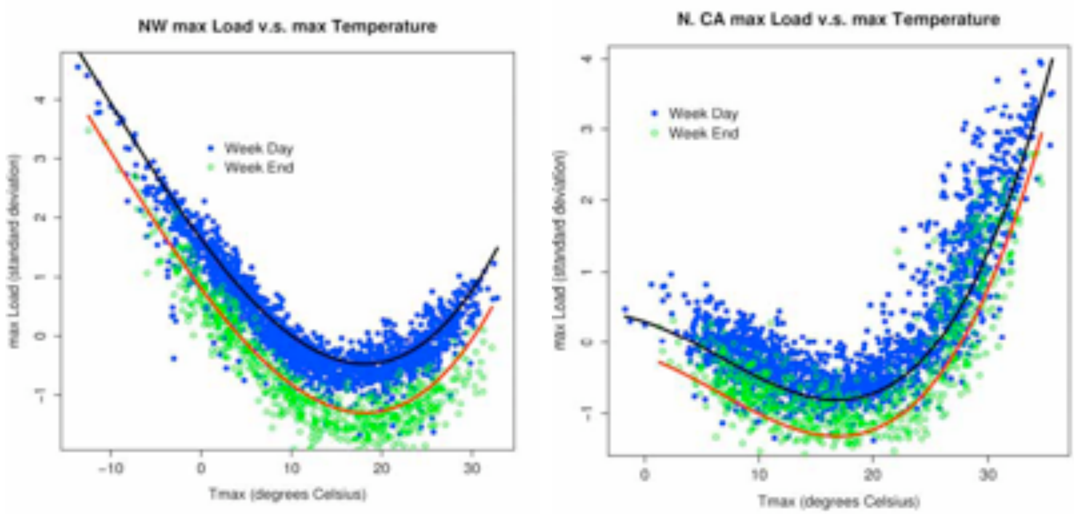


Figure 3.35 Nonlinear relationships between maximum daily temperature and peak energy demand in the PNW and Northern California. (Figures from Westerling et al., 2008).

An example of nonstationarity in social drivers is demonstrated by Hamlet et al. (2009), who investigate the uncertainty in the relationship between daily maximum temperature and peak energy demand in the PNW and northern California. By investigating projected increases in population, heating degree days, cooling degree days, and air conditioning market penetration, this research attempted to address heating and cooling energy demand indices (for residential, not industrial use) for three projected climate periods (2010 - 2039, 2030 - 2059, 2070 - 2099) using IPCC emission scenarios A1B and B1. Their studies projected that peak electricity demand in summer in the PNW would likely remain similar to current demands, but would be higher in CA (Figure 3.35). Their work also emphasized a highly non-linear relationship between demand and maximum temperature. For example, despite warmer temperatures, heating energy demand is projected to increase with increasing population: 22 - 23% for the 2020s, 35 - 42% for 2040s, and 56 - 74% for 2080s. Cooling energy demands will increase by a factor of 2.6 - 3 for the 2020s, 4.6 - 6.5 for 2040s, and 10.8 - 19.5 for 2080s. This work also identified key sources of error in these analyses to be 1) decadal scale precipitation associated with the PDO, and 2) population growth predictions and related heating and cooling energy demand.

3.7.2.3 Addressing uncertainty in water resources impact assessment

In this unpredictable water future, water resource management may become increasingly complex as new objectives and multi-facility coordination requirements are added to reservoir rule curves and water infrastructure. This complexity will be driven by a number of key uncertainties, including the loads and market price of power, hydrologic variability, water demands, environmental and biological requirements, aging infrastructure, and science and policy changes. These uncertainties pose important challenges to the management of hydrosystems.

The uncertainty in climate and hydrology is important for a number of management decisions around water infrastructure, including Biological Opinions, ESA and NEPA, flood risk management, FERC relicensing, evaluating resource adequacy and rates, infrastructure studies and policy (e.g. 2014/2024 Columbia River Treaty) reviews. While decision making under uncertainty is not new (Loucks et al., 1981), the non-linearity of changes associated with climate change presents some new challenges in a) using, interpreting, and communicating data based on climate projects, b) grasping the scope of policy/economic changes, c) resolving inconsistency in timesteps between social change and environmental change, d) meeting new and varied demands on the water system, and d) identifying control points in social-economic-ecological systems. Using dam operations as a specific example, uncertainties in reservoir operation models include fisheries objectives (flow augmentation and spill), forecast errors, hydropower loads, and runoff distributions. These challenges are further complicated by inconsistencies between the temporal and spatial scales of GCM and water resource models. Some (Barsugli et al., 2009) argue that GCMs need better model agreement on key parameters, a narrowing of the range of model output, and climate model output that matches the spatial and temporal resolution needed for water utility models, improved short-term climate model projections that fit water utility planning horizons.

Model uncertainties emerge from a variety of sources, including both the projections of future climate and the hydrologic response to changing climate. Based upon work in the United Kingdom, Kay et al. (2006) suggest that the largest source of uncertainty in estimated future runoff is introduced by the choice of GCM, followed by the chosen emission scenario, and then by the hydrological model which produces estimates of future runoff. This ranking of uncertainty has been supported by others (Praskievicz and Chang, 2009b; Bloeschl and Montanari, 2010) who have suggested that the greatest uncertainty emerges from the fact that the GCM projections cannot be evaluated against measured data for the future. Hydrological models, on the other hand, can be evaluated against historical conditions, and calibrated to ensure that the model captures measured dynamics, albeit historical, of the hydrological system.

The assessment of uncertainty in climate and hydrological predictions is becoming an integral component of climate forecasting (e.g. Solomon et al. 2007). The most common implementation of uncertainty analysis in climate modeling focuses on the utilization of an ensemble of climate results, including both multiple emissions scenarios and the results from multiple GCMs, as input into a single hydrological model. A variety of studies have utilized this model for impact assessment in California and Washington watersheds (Vano et al., 2009a and 2009b; Miller et al. , 2003; Hayhoe et al., 2004; Leung et al., 2004; Dettinger, 2004; Maurer and Duffy, 2005; Maurer, 2007). This procedure makes direct use of the uncertainty methods that are commonly employed by the atmospheric modeling community and climate change impacts research. In addition, Bloeschl and Montanari (2010) make the case that uncertainty introduced into water resource assessment through hydrological modeling should also be acknowledged and quantified. Techniques exist for evaluating hydrological model uncertainty (e.g. Beven and Freer, 2001; Vrugt et al., 2003), which can support climate change planning and designs.

3.7.3 Evaluating vulnerability of infrastructure to climate change

Measuring and communicating the impacts of climate change on water infrastructure is critical as water resource managers reflect on future infrastructure needs and risk. Several metrics may be useful in evaluating the vulnerability of water infrastructure across basins, political governances, and management approaches. Indirectly, measures of a landscape's hydrologic sensitivity to climate change may include percent of watershed with transient snow (resulting in a two peaked hydrograph due to rain and snow) (Swanson et al. 1992), and Hydrologic Landscape Regions (Wiggington et al. in review), which are based on measures of seasonality, climate, aquifer permeability, terrain, and soil permeability.

More directly, minimum reservoir storage, as the ratio of annual runoff to total storage, is another measure of system vulnerability or stress. Studies by Vano et al. (2009a, 2009b) indicate that active storage capacity drops by 50, 25, or 10% in October, when reservoir storage is typically lowest, under the A1B and B1 scenarios as compared to historic conditions. This could result in lower reservoir storage available late spring through early fall in the future. Similarly, Mote et al. (2003) found that basins with higher storage-to-flow ratios may be less vulnerable to stress than those with minimal storage. Another direct measure of storage vulnerability is the projected timing of reservoir filling. Medellín-Azuara et al. (2007) predict that dry-warm climate scenarios, given projected water demands and land use for 2050, will increase the seasonal storage range, with peak storage occurring around one month earlier.

3.7.4 Conclusions

As Oregon's population and economic activity increase over time, and as changes in climate increasingly impact the hydrological and ecological systems of Oregon, management of water resources will become increasingly complex. Difficult decisions regarding tradeoffs, modifications to current water infrastructure, and coordinated, thoroughly-analyzed operations of hydrosystems are going to be necessary.

3.8 Conclusions and Recommendations

As illustrated by several case studies, climate change will affect various sectors of water resources in Oregon in the 21st century. First, the amount and seasonality of water supply is expected to shift as seasonal distribution of precipitation changes and temperatures rise. While annual total precipitation may not change or even increase under some climate change scenarios, hotter summers accompanied by reduced precipitation will decrease stream flow. Although there are no anticipated spatial patterns of precipitation and temperature change across the state for the 21st century (temperature is projected to increase uniformly across the region), significant regional variations do exist. Models suggest that spring and summer streamflow in transient rain-snow basins, such as those in the Western Cascade basins, will be sensitive to these changes in precipitation and temperature; analyses of existing long-term streamflow data in western Cascade basins reveal declining spring streamflow, but no changes in summer or winter streamflows, since the 1950s. The High Cascade basins that are primarily fed by deep groundwater systems are expected to sustain low flow during summer months despite declining snowpacks, although the absolute amount of summer flow will decline. Basins in the east of the Cascades underlain by the Columbia River basalts are expected to have low summer flow in a distant future as groundwater recharge declines over time. April 1 snow water equivalent (SWE) will decline and the center timing of runoff will become earlier in transient rain-snow basins as snowpack is projected to decline consistently in the 21st century.

Second, water temperature is projected to rise as air temperature increases in the 21st century, particularly in urban streams where natural riparian vegetation is typically lacking. A decline in summer streamflow is expected to exacerbate water temperature increases, because the low volume of water will absorb solar radiation or blackbody radiation from the streambed and banks more quickly than during times with larger instream flows. Changes in water temperature can have significant implications for stream ecology and salmonid habitat in many Oregon streams. Lower order streams in transient rain-snow basins and in semi-arid eastern Oregon will be the most vulnerable to rising summer air temperature and diminished low flow. A new dam or reservoir might be required to maintain environmental flow in summer.

Third, as shown in the Portland water use study, when other demand factors are constant, increases in temperature alone would result in higher demands for peak season water. While demand during winter months is expected to remain constant, urban water demand is positively correlated with air temperature, particularly among single family residential (SFR) households. These impacts are also evident at multiple scales, including the household, neighborhood, and region. At the regional scale, urban land uses have different water demands, and will have varying impacts on water availability. Overall, people living in single-family residential areas are the largest consumers of water. At the neighborhood scale, the density of development helps to predict future water use, where higher density residential developments have lower water demand. Finally, at the household scale, the results of empirical research in the Portland region suggests a coupling of structural attributes (e.g. building and lot area) and temperatures that affect water demand. For the competing demands on regional water resources, if this development had contained smaller homes, and higher densities, other land uses would likely have more total water available.

Uncertainty is still high in projecting future changes in runoff, urban water demand, and water quality in Oregon. While the main source of uncertainty stems from the choice of global circulation models, additional sources of uncertainty include GHG emission scenarios, downscaling methods, hydrologic model structure and parameterization, and impact assessment methods. Multi-ensemble models that take into account all sources of uncertainty with different weights might provide a means of quantifying different sources of uncertainties. Communicating uncertainty to water resource decision makers is another challenge for adaptive water resource management in a changing climate. While a more sophisticated hydrologic impact assessment model yet to be developed, climate adaptation strategies can be implemented at multiple spatial scales.

Since one objective of land use planning is to coordinate regional activities, planning is one tool that may be helpful in meeting the future water needs of the State. Currently, land use and water resource management agencies have limited coordination in their responsibilities. The analyses provided here suggest that two characteristics of land-use plans, namely zoning and public involvement, can be instrumental to improving the coordination between land and water management agencies. Zoning can be used to link types of future development (e.g., for 2030 and 2040) to include a combination of infill, expansion, connecting existing developments, with explicit identification of water demands on different land uses in the region. To date, few plans have explicitly included dimensions of water management. Outreach and education campaigns can help inform the public about the relationship between water demand and supply, but can also assist in adapting to a future with increasingly limited resources. The details of those plans and the precise nature of the outreach and education campaigns will require further investigation, and will likely be part of the second assessment of Oregon's water resources.

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