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

Philip W. Mote¹, Daniel Gavin², and Adriana Huyer³

Contributing authors: Jack A. Barth³, Dudley B. Chelton³, Michael Fortune⁴, Burke Hales³, P. Michael Kosro³, Stephen D. Pierce³, Roger Samelson³, Kipp Shearman³, Robert L. Smith³, P. Ted Strub³

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

¹ Oregon Climate Change Research Institute, College of Oceanic and Atmospheric Sciences, Oregon State University
² Department of Geography, University of Oregon
³ College of Oceanic and Atmospheric Sciences, Oregon State University
⁴ Climate science writer, Corvallis, Oregon
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 δ18O 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 CO2 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), CO2 and CH4 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 CO2 and CH4 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 CH4 record shows that the abrupt temperature fluctuations during the glacial period had global influence (Flückiger et al. 2004; EPICA Community Members 2006). High resolution CO2 since the LGM is from Monnin et al. (2004).
atmospheric CO\textsubscript{2}; thus, increased weathering from the uplift of the Himalaya and Tibet Plateau, which began 55 million years ago, could increase movement of CO\textsubscript{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\textsubscript{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\textsubscript{2} during glacial periods include moving atmospheric CO\textsubscript{2} into the ocean, by changes in solubility of CO\textsubscript{2} in the ocean, increased biological productivity in the ocean surface waters, and changes in ocean circulation. In contrast, methane (CH\textsubscript{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 CO\textsubscript{2} 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.
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 CO₂ 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

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 (CO₂) 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 (CO₂), methane (CH₄), 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.
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 observations, 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 physically based models 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.
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...
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

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.
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\textsubscript{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 influenced 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

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).

---

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.
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 1.7 (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).](image1)

![Figure 1.7 (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.](image2)

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
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
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 1.9 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.
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 1.10** 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).
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 45ºN and by about 50% at 41ºN.

Third, monthly values of the Coastal Upwelling Index at 45ºN, 125ºW and at 42ºN, 125ºW 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 m², 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 $0.63 \pm 0.27 \mu\text{mol/kg/yr}$ (Figure 1.18). Inshore, the NH-line shelf has summer oxygen decreases of 1.8-2.0 $\mu\text{mol/kg/yr}$ 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 125.1°W 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 $\mu\text{M}$, 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.
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.

**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.
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 1.16. 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.](image)
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,
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.
References Cited


Bakun, A. (1990), Global climate change and intensification of coastal ocean upwelling, Science 247, 198201.


Barth, J. A., et al. (2010), Spatial and Temporal Variability in Near-Bottom Hypoxia over the Pacific Northwest Continental Shelf, in Ocean Sciences Meeting, edited, Portland, OR.


Erofeev, A., J. A. Barth, R. K. Shearman, L. Rubiano-Gomez, and J. Brodersen (2010), Seasonal and interannual variability of hydrographic and bio-optical fields off central Oregon from glider observations, in Ocean Sciences Meeting, edited, , Portland, OR.


Garzione, C. N. (2008), Research Focus Surface Uplift of Tibet and Cenozoic Global Cooling, Geology 36, 1003-1004.


Lisiecki, L. E., and M. E. Raymo (2005), A Pliocene-Pleistocene stack of 57 globally distributed benthic d18O records, Paleoceanography, 20, PA1003


Minckley, T. A., C. Whitlock, and P. J. Bartlein (2007), Vegetation, fire, and climate history of the northwestern Great Basin during the last 14,000 years, Quaternary Science Reviews 26, 2167-2184.


Pagani, M., J. Zachos, K. H. Freeman, B. J. Tipple, and S. Bohaty (2005), Marked change in atmospheric carbon dioxide concentrations during the Oligocene, Science 309, 600–603


Siegenthaler, U., et al. (2005), Stable Carbon Cycle-Climate Relationship During the Late Pleistocene, Science, 310 1313-1317.


Whitlock, C. (1992), Vegetational and climatic history of the Pacific Northwest during the last 20,000 years: implications for understanding present day biodiversity., Northwest Environmental Journal 8, 5-28.


