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Treeline Responses to Climate Change in High-Elevation Landscapes of Western Montana, U.S.A.

David F. Mann
University of Tennessee - Knoxville

To the Graduate Council:

I am submitting herewith a dissertation written by David F. Mann entitled "Treeline Responses to Climate Change in High-Elevation Landscapes of Western Montana, U.S.A.." I have examined the final electronic copy of this dissertation for form and content and recommend that it be accepted in partial fulfillment of the requirements for the degree of Doctor of Philosophy, with a major in Geography.

Henri D. Grissino-Mayer, Major Professor

We have read this dissertation and recommend its acceptance:

Kenneth H. Orvis, Sally P. Horn, Timothy G. Rials

Accepted for the Council:
Dixie L. Thompson

Vice Provost and Dean of the Graduate School

(Original signatures are on file with official student records.)

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Carolyn R. Hodges
Vice Provost and Dean of the
Graduate School

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TREELINE RESPONSES TO CLIMATE CHANGE IN HIGH-ELEVATION LANDSCAPES OF WESTERN MONTANA, U.S.A

A Dissertation
Presented for the
Doctor of Philosophy
Degree
The University of Tennessee, Knoxville

David F. Mann
May 2008

Dedication

This dissertation is dedicated to my wife Caroline, and children Truda, Robert, and Gavin, for all of their support, inspiration, and love.

Through our loyalty to each we become stronger, We Stand Together

"Per nostrum fidelitas ut sulum nos invalesco, Nos Sto una"

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Most importantly, I want to thank my wife, Caroline, and my children, Truda, Robert, and Gavin, for their encouragement and patience during my many years in graduate school. I hope this work inspires and helps direct my children’s lives no matter what they pursue. Finally, I thank my parents, Claus and Helen Mann, for teaching, inspiring, and for their love. They taught me that, no matter who you are or where you come from, you can improve your life and mind through hard work and dedication.

Abstract

The purpose of this research was to use whitebark pine trees at three major sites in western Montana to: (1) determine tree species response to climate, (2) reconstruct past climate conditions, (3) determine the effects of climate shifts on treeline, and (4) reconstruct fire history from fire-scar data. I collected samples from whitebark pine and subalpine fir and from remnant whitebark pine in the western Beaverhead-Deerlodge and Lolo National Forests.

In the climate response analysis, the Palmer Drought Severity Index (PDSI) had the highest correlations with whitebark pine growth. The strongest relationship occurred in the previous year's June and July. Precipitation in the previous year's May and June was also positively related to growth.

I reconstructed the previous year's June and July PDSI at all sites using a transfer function with tree-ring indices as the independent variable. The most intense drought year since 728 occurred in 1468. The reconstruction showed no evidence of a shift in the intensity and duration of wet and drought periods between the Medieval Warm Period and Little Ice Age.

To evaluate possible effects of increased global temperatures since the beginning of the 19th century on tree growth at high-elevation sites in western Montana, I established plots at 10 sites. I examined the establishment dates of all stems in each plot. Additionally, I examined the spatial relationships with establishment dates using a nearest neighbor statistic. Movement of treeline upslope was seen at the lower and upper elevation plots, while the mid-elevation plots remained stable. The greatest degree of movement (150 m) at treeline occurred in the 1980s.

Fire frequency and fire seasonality varied over time, reflecting the influence of climatic conditions. I collected 26 fire-scarred samples from a single site in the Gravely Range. The reconstruction of fire history revealed that fire was most frequent at the Gravely Range site during the Medieval Warm Period, but became less frequent during the Little Ice Age. Fire could possible play a role in the stand dynamics of the whitebark pine/subalpine fir ecosystem and limit the recruitment of whitebark pine.

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CHAPTER ONE

INTRODUCTION

1.1 Purpose

This doctoral dissertation focuses on assessing effects of climate variability on specific environmental processes in the subalpine ecosystem of western Montana. The past and present position of treeline is largely controlled by longer-term climatic conditions, such that the treeline position can act as a recorder of past climate change (Kullman and Kjallgren 2000). Treeline shifts may serve as proxy indicators of climate dynamics throughout the Holocene. During the last ca. 2000 years, the Medieval Warm Period (MWP, ca. AD 900–1400) and the Little Ice Age (LIA, ca. AD 1400–1850) followed the Roman Warm Period (5th century B.C. to the 1st century A.D.) and the Dark Ages Cold Period (AD 100–700) (McDermott *et al.* 2001). These shifts are considered characteristic examples of the warm and cold phases of a millennial-scale climatic fluctuation that has recurred throughout glacial and interglacial periods (McManus *et al.* 1999). The LIA was particularly influential in determining current vegetation patterns (Biondi *et al*. 2001). Studies have suggested that decreased natural regeneration and premature death of full-grown forest stands occurred during the climax phase of the LIA (Kullman 1987). Changes that have occurred since the LIA, particularly during the 19th century, may provide insight into the ecological dynamics of the subalpine ecotone.

Global warming during the 20th century could further complicate our understanding of vegetative responses to climate change. In fact, this warming may not be extensive enough to compensate for the thermal decline associated with the LIA (Payette *et al*. 1989). Furthermore, large explosive volcanic events have produced different extents of cooling in the Northern Hemisphere in the last 600 years (Briffa *et al*. 1998). The eruption of Huaynaputina, Peru, in 1600 produced the most severe short-term cooling event during the last 500 years, beginning in 1601 (Briffa *et al*. 1988). Tissue damage (i.e., frost rings) in subalpine bristlecone pines (*Pinus longaeva* D.K. Bailey) of the western United States (LaMarche and Hirschboek 1984) and light rings in conifers of Fennoscandia (Briffa *et al*. 1990) have been directly linked to volcanic events.

1.2 Proxy Indicators of Past Environments

Interest is steadily mounting for more accurate predictions of the effects of changing climate variability on vegetation dynamics at regional and continental scales (Peterson and Darling 1985, Mote and Mantua 2002). The distribution of ocean areas, land surfaces, and mountain ranges are major controls of broad-scale atmospheric and oceanic circulation patterns over the Earth's surface (Fritts 1971). Changes in one or more of these features have caused major fluctuations in climate patterns and subsequent changes in responses by ecosystem processes (Tessier *et al*. 1997). The most pronounced effects of these linkages can often be found in trees growing in stands that are near their ecological limits, *i.e.* at the upper treeline (Jacoby and D'Arrigo 1997). This dissertation therefore focuses on changes in high elevation ecosystem processes generated by variability in global climate.

The last ice age ended between 20 ka BP and 10 ka BP. This time span was marked by a series of abrupt climate shifts that are documented in high-resolution proxy data (Hannon *et al*. 2000). Ice cores from Greenland and Antarctica revealed that, during the last deglaciation, gradual warming in the Southern Hemisphere preceded an abrupt temperature increase in the northern high latitudes at the onset of the Bolling/Allerod 14,600 years ago, which in turn was followed by a cold reversal in Antarctica (Alley 2000). Consequently, climate conditions in the Northern Hemisphere dropped back to glacial conditions during the cold phase called the Younger Dryas, which ended about 11,500 cal. yr BP (DeMenocal *et al.* 2000). These interhemispheric climate changes have been linked to shifts in the Atlantic thermohaline circulation (THC) and associated variations in northward oceanic heat transport (Crowley 2000). The neoglaciation of the late Holocene culminated in many places during the Little Ice Age (LIA) of the 14th to 19th centuries (Bradley and Jones 1993), but the synchroneity and intensity of the cooling period varied from location to location (Grove 1988).

D'Arrigo *et al*. (1992) suggested that the northern treeline experienced a period of warming following the close of the LIA (ca. 1880). The cause of this warming could be related to solar variations (Friis-Christensen and Lassen 1991), decreased volcanic aerosols (Bryson and Goodman 1980), and natural climatic variations (Wigley and Raper 1990). Between 1950 and 1980, a significant drop occurred in average temperatures similar to that of the LIA period (Bradley and Jones 1993). The following reconstructions of past climate clearly show shifts between cold and warm conditions globally.

Levy *et al*. (2004) focused on Holocene climatic changes and glacial fluctuations in an alpine setting of southwestern Alaska. The study reconstructed the history of neoglacial activity of a basin using both the geomorphic (moraines) and paleolimnological (lake sediment) records. This study was the first in Alaska of its kind and was unique in the fact that it combined the morainal evidence of up-valley glacier fluctuations with the continuous sediment record of a down-valley lake to determine the timing and extent of the glacial fluctuations. The authors found that glaciers larger than any that formed during the late Holocene persisted near the lake until 9100 cal. yr BP. Calkin *et al*. (2001) found that mountain glaciers did not persist long after the Younger Dryas interval (12.8–11.5 cal. yr BP). Evidence from other studies in Alaska (Anderson *et al.* 2003) found an increase in effective moisture after 9100 cal. yr BP. The reduction in glacier extent after 9000 cal. yr BP suggested a shift to precipitation associated with summer (June–July) periods.

Temperatures at high latitudes peaked during the first half of the present interglaciation period with warming occurring at different times and to varying degrees in different locations (Kaplan *et al.* 1998). Cape (2001) found a strong spatial variability in the response to insolation forcing during the early Holocene in the Arctic indicating unequal heating across the earth. Kaufman *et al*. (2004) developed a comprehensive examination of the spatial pattern of the Holocene thermal maximum – the interval of warmth associated with the peak in Holocene temperature. Rather than modeling the spatial pattern the authors relied on the spatio-temporal pattern of time-transgressive intervals when temperatures reached their local Holocene thermal maximum. Using unpublished and published records of Holocene paleoenviromental change (lake sediments, marine sediments, ice cores, and peat deposits), they found that the Holocene thermal maximum in the Arctic was forced by insolation changes associated with orbital variations whose forcing effects differ with latitude. The forcing of the event was symmetrical but occurred earlier in Alaska (11 ka) than in the Hudson Bay region (7 ka) due to the delayed melting of the Laurentide Ice Sheet, with a regional influence of the North Atlantic and Arctic Oceans occurring ca. 9 ka.

1.3. Paleoclimatological Studies in the Northern Rocky Mountains

Mountain environments contain a variety of excellent natural records that document past ecological and climatic changes with differing temporal and parameter resolution (Beniston 2000). Numerous studies have been conducted within mountain ecosystems on glacier variation and past glacial extent (Karlen 1988). The lower limits of ice cover can be reconstructed and dated using C^{14} for the last full glacial period and by examination of lichens for the past few centuries (Beniston 2000).

The potential for the extraction of climatic and additional atmospheric signals from ice at high-elevation sites was illustrated by Thompson *et al*. (1984). Ice preserves records of atmospheric gas concentrations, pollen, and volcanic and aeolian dust, as well as the isotopic composition of snow (Wagenbach 1989). For example, ice-core data showed that greenhouse gas concentrations varied over the glacial-interglacial cycle in a way that can explain a significant fraction of the global temperature change (Barnola *et al*. 1987).

Pollen studies of Tertiary and Pleistocene deposits in the northern Rocky Mountains record the expansion of conifer ecosystems in the Miocene and the response by these communities to glacial/interglacial climate oscillations in the Pleistocene (Barnosky 1987). In addition, Holocene pollen records have been developed in Yellowstone National Park (Waddington and Wright 1974), the Caribou Range

(Beiswenger 1991), and the Bighorn Mountains (Mehringer *et al*. 1977), and these data offer a representation of regional postglacial vegetational change (Whitlock 1993).

Within the northern Rocky Mountains, limited data are available to assess whether the modern subalpine forest has a long history in the postglacial or if the present forest is simply a recent configuration of vegetation representing continuous change (Delcourt and Delcourt 1983). McDonald (1987) investigated the subalpine-boreal transition zone in western Canada using sediment cores from three lakes. The period of greatest change occurred during the early to middle-Holocene, with conifer trees species that established in deglaciated areas between 10000 and 8000 BP, and significant vegetation change continuing until 5000 BP. Cattail (*Typha latifolia* L.), a common emergent aquatic plant, has been interpreted in early Holocene pollen records as an indicator of the initiation of warm, post-glacial climatic conditions in both the Rocky Mountains and in northwestern Canada (Ritchie *et al*. 1983). McDonald's research indicated the presence of cattail at the bases of cores collected from the study sites. This early vegetation was then displaced by expanding *Populus* and *Picea* populations as the climate grew warmer.

During the full glacial and the mid- to late Holocene, vegetation distribution and composition were moderately stable (Crowley 2000). Climatic shifts caused plant distributions to shift northward, but taxa also moved east or west. For example, Thompson and Wallace (2001) found that areas of high concentrations of spruce, pine, and other cold-tolerant species expanded from eastern North America into central and western Canada. Modern associations (such as beech-hemlock and spruce-alder-birch) date to the early Holocene, whereas other associations common to the late glacial no longer exist (Hannon *et al.* 2000).

Kaufman *et al*. (2004) developed a 33,000 year record of paleoenvironmental change from Rolik Lake, Alaska. Their objective was to obtain records of paleoclimate change that extend beyond the last deglaciation (14 ka) and develop a record that would capture the inception and full-glacial phases of the LGM. The investigation used a collection of proxy climatic indicators to reconstruct millennial-scale paleoenvironmental changes. They analyzed the magnetic susceptibility, grain sizes, organic matter and inorganic carbon content, $^{13}C/^{12}C$ ratios, biogenic silica, and radiocarbon (^{14}C) dates of sediment cores. They showed that cores that date to 33 ka show that an ice-marginal lake formed within the lake basin during the LGM. Examination of the pollen collected from the sediments showed that lake level fluctuated synchronously with global climatic change patterns. The lake also responded quickly and sensitively to climate changes during the Pleistocene-Holocene transition.

Doerner and Carrara (2001) investigated the vegetation and climate history of the Long Valley area in west-central Idaho during the late Quaternary. Vegetation and climate of the western U.S. underwent substantial changes during the transition from glacial to interglacial conditions at the end of the late Pleistocene (Thompson *et al.* 1993). They investigated pollen and sediment recovered from the McCall fen in Long Valley, Idaho. An AMS radiocarbon date of 16,330 ± 60 ^{14}C yr BP on the lower portion of the core indicated a period of deglaciation prior to ca. 16,500 ^{14}C yr BP. Following this period, a cold, dry climate lasted until 12,200 ^{14}C yr BP, when *Artemisia* was the dominant pollen. After this, a cool, moist climate prevailed with a spruce-pine forest dominating vegetation around the fen. A dry and cold interval interrupted this period and indicated the onset of the Younger Dryas climate oscillation. From 9750 to 3200 ^{14}C yr BP, the climate was warmer and drier than present with an open pine forest surrounding the fen. The establishment of the current forest was marked by the return of cool, moist conditions ca. 2000 ^{14}C yr BP. Analyzing the effects of glacial and interglacial episodes supports the understanding of the consequences of climate change associated with these events and their effects on vegetation shifts.

Lakes located at the ecological limit of a species such as alpine lakes may be more sensitive to minor changes in climate. Carrera *et al*. (2002) found that high elevation alpine lakes (1) are situated at all latitudes from the equator to the poles, (2) are sensitive to global impacts, (3) have simple ecosystems, and (4) are highly variable in terms of size, depth, UV radiation, ice cover, and nutrient status. Seppa and Birks (2002) reconstructed Holocene climate fluctuations from a Fennoscandian treeline area using a pollen-climate reconstruction model and a new pollen stratigraphy to derive quantitative estimates of annual precipitation and July mean temperature. They discovered that the pollen-based climate reconstruction showed a gradual early Holocene rise in summer temperatures and a gradual mid- and late Holocene cooling with no explicit indications of abrupt climate shifts. This finding is in agreement with the frequently presented view that the Holocene has been characterized by a relatively stable climate with fluctuations of relatively low amplitude (Fronval and Jansen 1997).

Litter mass loss of peat can be used to determine past climatic events because peat that lies above water will have a higher rate of litter mass loss than peat that has been submerged (Frolking *et al.* 1998). Roos-Barraclough *et al*. (2002) used an ombrotrophic bog located in Switzerland to develop a late glacial and Holocene record of climate change. Ombrotrophic bogs, which are raised above the groundwater level, receive moisture only from the atmosphere and are therefore particularly sensitive to climate change. The authors demonstrated that high levels of precipitation occurred during the Younger Dryas cold period with high proxies of precipitation throughout the peat samples. Several periods of low precipitation also occurred during the record with temperature throughout the late glacial period and the late-glacial-Holocene transition aiding in the decomposition of peat.

The quantification of microscopic charcoal particles preserved in lake sediments and mires is commonly used by palaeoecologists to study the occurrence of past fires and their effects on ecosystems (Tinner and Hu 2003). Long and Whitlock (2003) developed a fire and vegetation history for the coastal rain forest of western Oregon using: (1) charcoal analysis to reconstruct the fire history; (2) measurements of sediment magnetic-susceptibility to provide information on sedimentation history; and (3) pollen analysis to reconstruct vegetation history. They showed that fire is an important agent of disturbance in the coastal forest and that fire occurrence was synchronous with climate change over the last 4600 yrs. Frequent fires in the early portion of the record helped to maintain an open forest while a decrease in fire occurrence over the last 2700 yr led to a closing of the forest canopy. The authors felt that this was directly related to a protracted summer fog event and shorter duration of summer drought. Climatic variability is a dominant factor that affects fire frequency in the western United States, an observation supported by paleoecological data from charcoal in lake sediments and reconstructions from fire-scarred trees (Tinner and Hu 2003).

For the purpose of detecting the effects of human activities on climate change, it is important to document natural change in past climate (Thompson and Wallace 2001). Climate shifts such as the Medieval Warm Period (AD 900–1400) and the Little Ice Age (AD 1400–1850) are examples of such change within the last millennium. Pederson *et al*. (2005) investigated the roles of the Medieval Warm Period, Little Ice Age, and European impact on the environment by examining pollen and charcoal from the last 1350 years from the Hudson Valley, New York. The pollen records from two cores revealed the pre-European forest composition (AD 653–1697) as *Pinus-Quercus* with minor elements of *Tsuga*. An increase in pine and a decrease in oak and hemlock during the Medieval Warm Period implied drought and a possible increase in fire as indicated by high inorganic content (erosion) and the presence of the largest amount of charcoal in the sediment record. This period also showed low levels of pollen influx indicating a drought period. The Little Ice Age was indicated by strong peaks in spruce within the pollen record and cooler temperatures. Low proportions of inorganic compounds were found during this same period which indicated minimal erosion rates. Upland disturbances were revealed in the record by the presence of weedy vegetation that correlated to European impact in the late 18th century. European impacts also corresponded with resurgence in the inorganic fraction, possibly caused by changes in land use.

1.4. Dendroclimatological Research

1.4.1 Introduction

The annual rings of trees provide a global, natural archive for studying past environments (Fritts 1976). Trees are sensitive to the environmental conditions that occur during the year, especially during the growing season. Tree-ring series collected from sites over an extended geographical area can assist in the understanding of how climate shifts affect ecosystem patterns and response (D'Arrigo and Jacoby 1991; Grissino-Mayer 1996). Reconstructions developed from high-elevation tree-ring series can provide information concerning fluctuations of Holocene temperature. These reconstructions can also increase our knowledge of how climatic shifts in the past affected global, regional, and micro-site ecosystem processes, such as wildfires.

1.4.2 Dendroclimatic Research in the Northern Rocky Mountains

Tree rings have characteristics that make them an exceptionally valuable source for reconstructing climate information (Fritts 1976). Although instrumental climate records are limited in length, proxy climate data, such as from tree rings, have been useful in extending precipitation and temperature records (Woodhouse 2003). At high elevation sites, growth processes are strongly influenced by temperature (Briffa *et al.* 1990, D'Arrigo *et al.* 1992) with seasonal temperature having a strong effect on tree-ring formation and tracheid production (Vaganov *et al.* 1994). Interaction between ring width and summer temperatures is often found in tree species at their high-elevation or high-latitudinal limits (Fritts 1976).

Graumlich and Brubaker (1986) reconstructed annual temperature using mountain hemlock (*Tsuga mertensiana* (Bong.) Carr.) and subalpine larch (*Larix lyallii* Parl.) in Washington state. The investigation showed a positive relationship between ring growth and summer (July to September) temperature and a negative relationship with spring (March) snow depths. The temperature reconstructions illustrated that between 1590 and

1900 temperatures were generally 1 ºC lower than those of the 20th century, with a short interval between 1650 and 1690 approaching the 20th century values.

Luckman and Wilson (2005) used tree-ring data from Engelmann spruce to examine summer temperatures in the Canadian Rockies for the last millennium. The reconstruction revealed warm intervals, comparable to 20th century warming, during the first half of the eleventh century, the late 1300s, and the early 1400s. A portion of the reconstruction, however, showed below-normal temperatures for the period 1901–1980, with prolonged cool periods from 1200–1350 and from 1450 to the late 1800s. The most extreme cool period was observed in the late 1690s. The reconstructed cool periods agreed with regional records of glacial advances between 1150 and the 1300s, and in the early 1500s, early 1700s, and 1800s.

Villalba *et al*. (1994) examined variations in tree growth in response to climate variations in the subalpine zone of the Colorado Front Range using 25 tree-ring chronologies. At the driest sites, growth of spruce and fir tracked climatic variations similarly. Steep environmental gradients in the subalpine zone accounted for a majority of the observed differences in growth response to climate change. The research demonstrated that inter-site differences in tree growth response to climate can be used as indicators of environmental differences among subalpine habitats.

Both the spatial and temporal variance of climate over the conterminous United States can be partially attributed to the Pacific Decadal Oscillation (PDO), a low frequency (20–30 years) recurring climate pattern somewhat analogous to the El Niño Southern Oscillation (ENSO) (Collins and Sinha 2003). Spatially, PDO effects can be seen in the northwestern United States (Gershunov and Barnett 1998). The PDO modulates the pattern of jet stream activity, which affects the tracking of mid-latitude storm systems, resulting in changes in precipitation and temperature patterns (Kaplan *et al.* 1998). When PDO is in a positive phase, waters in the north central Pacific Ocean tend to be cool, and waters along the west coast of North America tend to be warm (Delworth and Mann 2000). The opposite is true when the PDO index is negative. PDO is generally associated with lower than average rainfall and higher than average air temperatures in the Pacific Northwest during a positive (warm) phase (McCabe *et al.* 2004), while the negative (cool) phase PDO is associated with relatively high precipitation rates and low air temperatures (Kaplan *et al.* 1998).

The Atlantic Multidecadal Oscillation (AMO) is a low-frequency (60–100 years) 10-yr running mean of detrended SST anomalies averaged over the North Atlantic from 0–70ºN and having an amplitude of 0.4 ºC. The AMO acts to modulate El Niño-Southern Oscillation teleconnections over extensive portions of the northern hemisphere (Gray *et al.* 2003), and regulates northern hemisphere temperatures (McCabe *et al.* 2004). In both tree-ring and instrumental records, the wettest decadal-scale events across the U.S. tend to be associated with a negative AMO, the driest with a positive AMO. However, in the American Southwest, positive and negative phase PDO tend to entrain the Southwest in circulation patterns of wetness and dryness, respectively (Sutton and Hodson 2005). Megadroughts (such as the one that occurred in the late 1500s) have been associated with a positive phase AMO and a negative PDO (Gray *et al.* 2003).

Advances in our understanding of climate change over the past 1000 years have resulted largely from reconstructions based on tree-ring records. D'Arrigo *et al*. (2001) reconstructed PDO variability in coastal Alaska and the Pacific Northwest using tree-ring chronologies developed by Wiles *et al*. (1996, 1998). Using both ring-width and density data, they found that PDO shifts such as those that have occurred in recent decades also occurred in prior centuries. A more dominant decadal mode was observed prior to the mid-1880s with a possible shift to more El-Niño/Southern Oscillation (ENSO) forcing after 1890. The authors identified the beginning of a positive phase PDO (1976–1977) as an indicator of a possible return to a more pronounced decadal variability in the Pacific.

Pohl *et al*. (2002) used a 545-year ponderosa pine tree-ring chronology to examine the drought history of central Oregon and determine the relationship between drought, ENSO, and PDO, and to compare the climatic sensitivity of ponderosa pine and western juniper to determine their suitability as interchangeable climate proxies. Severe drought periods occurred during the 1480s, 1630s, 1700s, and 1930s. The most sustained drought period occurred during the 1930s with the most severe single year drought occurring in 1489. The authors found a weak relationship between the ponderosa pine chronology and an ENSO/PDO signal. Western juniper demonstrated more sensitivity to regional and seasonal climate variations whereas ponderosa pine was more responsive to temperature change.

Gray *et al*. (2003) used wavelet analyses to investigate multi-decadal oscillations in precipitation using tree-ring records from Douglas-fir, ponderosa pine, limber pine (*Pinus flexilis* James), and pinyon pine (*Pinus edulis* Engelm.) back to A.D. 1400 from the central and southern Rocky Mountains. Strong multi-decadal phasing of precipitation was present in all regions during the 16th century megadrought, with synchronous dry cycles occurring until the 1590s, resulting in a significant reorganization of climate in the Rocky Mountains. Pederson *et al*. (2004) used tree-ring reconstructions of North Pacific surface temperature anomalies and summer drought as proxies for winter glacial accumulation and summer ablation over the last three centuries. These records showed that the 1850s glacial maximum was likely produced by about 70 years of cool/wet summers coupled with high snowpack values. Glacial retreat after 1850 coincided with an extended period (> 50 years) of summer drought and low snowpack.

1.4.3 Dendroclimatic Studies in the Subalpine Ecosystem

High elevation ecosystems are areas that could be unusually susceptible to changes in global climate (Luckman 1990, Beniston *et al*. 1997, Diaz and Graham 1996, Battarbee *et al*. 2002), with relatively small changes in temperature and precipitation having significant implications for natural ecosystems and disturbance regimes (Pisaric *et al*. 2003). Mountain ecosystems provide direct life support for close to 10% of the world's population, and indirectly to over half (Beniston *et al*. 1997). Our understanding of how these environments may have been influenced by changing climate conditions is therefore critical. Paleoecological studies that focus on climate and vegetation dynamics at extended temporal scales may provide insight into the natural variability of a system, providing a more precise assessment of vegetation dynamics in response to changing climate (Pisaric *et al*. 2003).

Ecological processes in mountain environments are mainly constrained by physical components of the environment. Some high-elevation species can survive extremes in both temperature and precipitation (Korner 2003) and as a result have developed a high level of specialization. High elevation sites are often highly fragmented and also contain higher plant diversities than those in the surrounding lowlands (Barthlott *et al.* 1999). The ability to cope with specific environmental demands at high-elevation can be achieved in three ways: (1) by evolutionary (phylogenetic) adaptation, (2) by ontogenetic modifications (the origin and development of an individual organism from embryo to adult), and (3) by acclimation (Korner 2003). In the western United States, subalpine and treeline pines include two distinctive phylogenetic groups, with Rocky Mountain bristlecone pine (*Pinus aristata* Engelm.), foxtail pine (*Pinus balfouriana* Grev. & Balf.) of the section *Parrya* subsection *Balfouriana*, and whitebark pine (*Pinus albicaulis* Engelm.) and limber pine of the section *Strobus* subsection *Cembrae* (Richardson and Rundel 1998). The upper limit of forest growth occurs at approximately 1400 m on favorable sites in the Rocky Mountains with the treeline progressively rising southward at a rate of 100 m per degree of latitude to an elevation of over 3600 m in northern New Mexico (Peet 2000).

Graumlich (1991) examined growth trends in subalpine trees as related to increases in CO_2 and changes in climate. Using tree-ring series developed from foxtail pine, lodgepole pine (*Pinus murrayana* Balf. ex A. Murray), and western juniper (*Juniperus occidentalis* Hook.) collected in the Sierra Nevada, California, she found that current growth levels were equal to or exceeded pre-industrial periods but that the results did not indicate that CO_2-induced growth enhancement was occurring among subalpine conifers. Tree growth responses were shown to correlate with precipitation during the previous winter and temperature during the current summer.

Graumlich (1993) reconstructed temperature and precipitation back to A.D. 800 for the Sierra Nevada using foxtail pine and western juniper. The temperature reconstruction showed fluctuations on centennial and longer time scales including a period from AD 1100–1375 (corresponding to the Medieval Warm Period) with temperatures that exceeded late 20th century values, and a period of cold temperature from AD 1450–1850 that corresponded to the Little Ice Age. An examination of 20 and 50 year intervals indicated that precipitation levels during the 1000 year record equaled or exceeded 20th century levels. Also, Graumlich *et al*. (2003) used precipitation-sensitive Douglas-fir tree-ring records to reconstruct the annual flow of the Yellowstone River back to AD 1706. The reconstruction showed that 20th century streamflow was not representative of conditions in the past. With the exception of the 1930s, streamflow during the 20th century exceeded average flows during the previous 200 years. The drought of the 1930s resulted in the lowest flows during the last three centuries.

1.4.4 Justification

Tree-ring data can be used as a tool for reconstructing climate in the alpine and boreal zones (Briffa *et al.* 1988, Jacoby and D'Arrigo 1989, Briffa *et al.* 1992). Recent investigations have focused on the effects of changing climate variability on vegetation dynamics at regional and continental scales. Because of differences in the density of tree-ring sites versus that of meteorological stations, a clearer understanding of the regional effects of climate change in the western United States is needed.

Gaps currently exist regarding the spatial and temporal record of climatic fluctuations for high elevation sites in much of the northern Rocky Mountains, including western Montana. This study seeks to address these gaps by conducting a comparative analysis of growth response to climate changes in tree rings in western Montana, and using this information to reconstruct the climate record to analyze both short-term (< 100 yrs) and long-term (> 100 yrs) trends in past climate.

Examination of the variation in past climate is a significant component of the effort to predict the behavior of the climate system in the future. The results of my dendroclimatic and ecological research will result in an increased knowledge of the effects of temperature and precipitation alterations on the whitebark pine ecosystem, the elucidation of the subalpine ecological response to climate variability over millennial scales, and increased comprehension of the effect of the MWP and subsequent LIA on ecosystem processes.

The results of this study will help fill gaps in our understanding of the role of climate as a primary influence on ecosystem dynamics at high-elevation sites in the northern Rocky Mountains. In addition, species composition and ecosystem functions are likely to respond in the future, possibly in unexpected ways, to increases in sea surface temperature and air temperature and to any changes in the thermohaline circulation. The upper treeline in the western United States should be targeted for a more systematic sampling of climate-sensitive specimens to increase our knowledge of past climatic shifts, further quantify the effects of climate shifts on high elevation ecosystems, and to help predict future climate variations and how they may affect changes in treeline and other aspects of high-elevation ecology.

1.4.5 Objectives

The specific objectives of this research are to:

- Develop a multimillennial tree-ring chronology from long-lived trees and subfossil wood;

- Analyze the relationship between climate and tree growth;
- Use this relationship to develop a reconstruction of temperature for the northern Rocky Mountains; and
- Reconstruct the record of short-term, decadal scale (<100 years) and long-term, centennial scale (>100 years) changes in climate.

1.5 Research in Subalpine Environments and Treeline Elevational Migration

1.5.1 Studies in Subalpine Ecology

Latitude, elevation, physiographic position, and parent material all interact to influence environments and the distribution of species, communities, and the nature of disturbance regimes in the Rocky Mountains (Long and Whitlock 2002). Rocky Mountain forest ecosystems reflect typically steep environmental gradients. Topography strongly influences temperature, water availability, and the distribution of species (Peet 2000). Many ranges in the Rocky Mountains have well defined upper and lower treelines. Between these limits, elevation-climatic gradients are often associated with pronounced mountain zonation that reflects not only the upper and lower limits of a species' distribution but also its successional status (Long and Whitlock 2002). Tree species native to the Rocky Mountain ecosystem differ extensively in their adaptive responses to climatic variation (Rehfeldt 1994). Some species, such as the western white pine (*Pinus monticola* Dougl. ex D. Don), may be considered broadly adapted generalists, whereas the adaptive responses of other species, such as Douglas-fir, more closely follow environmental gradients (Rehfeldt 1994). Compared with forests in other regions of

North America, the prevalence of wide-ranging conifer species suggests comparatively high genetic diversity in the forest of the Rocky Mountains (Hamrick and Godt 1989).

The alpine treeline is possibly one of the most acknowledged and studied of all distributional boundaries of trees, but still awaits a conclusive functional explanation that withstands testing at both the macro- and micro-ecological levels (Korner 1998). Lavoie and Payette (1992) investigated the effects of a changing winter environment at treeline in Canada. Spruce growth on slopes and valleys revealed periods of low growth and reduced symmetrical growth between AD 1601 and 1904. Krummholz spruce that grew during the LIA were replaced by symmetrical trees during the warming in the 20th century. The large number of spruce seedlings successfully initiated between 1710 and 1810 and between 1880 and 1990 appeared to be related to milder temperatures and heavier snow fall. Spruce that established during the Little Ice Age tended to assume a krummholz growth form with matted lower branches. These were replaced during the 20th century by symmetrical trees indicating a reduction in snowfall and an increase in winter temperatures.

Lloyd and Graumlich (1997) reconstructed a 3500 year history of fluctuations in treeline elevation and tree abundance in the southern Sierra Nevada. Declines in the abundance of live trees and treeline elevation occurred during two periods: from AD 1000 to 1400 and from AD 1500 to 1900. The first decline was correlated to a period of warm temperatures and multidecadal droughts. The second decline was related to a period of low temperatures that caused a sustained failure of regeneration and an increased rate of adult mortality. The effects of limited precipitation (extend snowpack and drought) suggested that climatic controls over treeline location could be more related to micro-site characteristics than to regional climate.

Luckman and Kavanagh (1998) found that on north-facing slopes in the Canadian Rockies the treeline ecotone, dominated by subalpine fir (*Abies lasiocarpa* (Hook) Nutt.), had changed little in the last 400 years. However, on south-facing slopes treeline showed a catastrophic dieback during the 1600s. Extensive upslope migration of seedling establishment occurred during the 20th century. Lower-elevation trees that established during the 18th and 19th centuries exhibited no expansion in population. This suggests that the response of treeline species to climate change varies with local site conditions.

Camarero *et al*. (2000) focused on the detection of tree regeneration at treeline using tree density, size-structure, growth-form, and estimated age. Using point pattern analysis they mapped each *X–Y* coordinate of all stems within fixed plots at two sites. Examining the distribution of larger and multi-stemmed trees across the treeline ecotone revealed a distinctive boundary of size-structure and growth form. They found that tree populations and growth forms varied between areas within the plots. Those areas in the plots that were affected by high winds showed krummholz growth forms with a rapid increase in the spatial extent of seedlings with warming conditions. Areas not affect by wind showed more vertical growth forms with a more gradual increase in seedlings. The results indicate that seedling survival and any movement of treeline is dependent on micro-site conditions.

Numerous studies of terrestrial ecological response have focused on average 20th century patterns or on specific periods within the 20th century (Easterling *et al*. 2000). In alpine locations, generalized warming trends have been linked to forest expansion

(Payette and Delwaide 1994, Klasner and Farge 2002) and changes in tree-ring width (D’Arrigo *et al*. 2000). Boyce and Saunders (2000) suggested that summer climate strongly affects winter water relations in high-elevation engelmann spruce and subalpine fir habitats, and that change in both the summer and winter climate would alter species’ ranges.

An increased understanding of the past relationship between disturbance and climate variability clarifies the implications of climate change on subalpine ecosystems. Sibold and Veblen (2006) investigated the relationship of subalpine forest fires in the Colorado Front Range with interannual and multidecadal scale climate variations and found that extensive subalpine zone fires were significantly associated with the La Nina phase of ENSO, the negative phase PDO, and the positive phase AMO. Zeglen (2002) surveyed whitebark pine and white pine blister rust (*Cronartium ribicola* J.C. Fisch.) in British Columbia, Canada for mortality, incidence of blister rust, or other damage. The study showed that 50% of whitebark pine were dead or had active blister rust infections. Whitebark pine seedlings revealed lower rates of mortality and infection than mature trees. The combination of mature tree mortality, lack of suitable regeneration, and the incursion of alternative climax species (subalpine fir) presaged a continued decline in whitebark pine populations.

Perkins and Roberts (2003) used predictive models of whitebark pine mortality from mountain pine beetles (*Dendroctonus ponderosae* Hopkins) to estimate the probability of attack. Logistic regression models were calibrated from reconstructed pre-epidemic stand conditions and post-epidemic mortality levels that were caused by an outbreak that occurred from 1909 to 1940. Stands that had a basal area > 10 m^2 had a

100% probability of being attacked. They illustrated that a tree level model may be used to estimate anticipated cumulative mortality in currently or potentially infested areas and that stand susceptibility to infestation may be identified from density thresholds.

1.5.2 Treeline Elevational Migration

An advancing treeline would have important implications for the global carbon cycle and for biodiversity of the ecotone. A number of northern hemisphere treeline migration studies have demonstrated northward shifts during warming in the early part of the 20th century (Hofgaard 1997). Migration was less pronounced during warming trends of the latter part of the 20th century (Kullman 1986, Lescop-Sinclair and Payette 1995, Briffia *et al*. 1998). This response by treeline species could lead to unique ecosystems in these areas over the next century. For example, Chapin and Starfield (1997) simulated the advance of Alaskan treeline and estimated a 150–250 year time lag in the forestation of Alaskan tundra, and suggested that rapid warming under dry conditions could emerge. Evidence of this scenario comes from the lack of response of trees to warming in Alaska (Barber *et al*. 2000).

Laroque *et al*. (2001) described the nature of treeline dynamics and upper-treeline establishment patterns in British Columbia, Canada through the examination of tree growth, climate, and seedling establishment using dendroecological approaches. Their results demonstrated that the area had experienced species-specific pulses of tree establishment. Stem data demonstrated that seedling establishment within the last three centuries had been episodic and linked to historical climatic patterns. The establishment of hemlock was associated with cool summers and shallow snowpack while the establishment of subalpine fir was restricted to intervals of cool growing season temperatures and deep seasonal snowpack. Episodic seedling establishment in the 20th century resulted in a gradual infilling of the local treeline and the development of a more structured parkland.

Wunder *et al*. (2006) used the forest succession model *ForClim* to simulate Holocene treeline dynamics in the Central European Alps to explore the extent and cause of changes in treeline altitude and composition. The modeled results were compared with Holocene pollen and macrofossil records collected from the current treeline and confirmed paleoecological studies of treeline fluctuations. The simulated results also demonstrated changes in species composition and treeline position between 11000 and 4500 B.P. The results indicated that changes in temperature alone can account for alterations in treeline elevation during the first half of the Holocene.

1.5.3 Justification

Past studies have investigated the impact of climate change on the advance and retreat of the treeline ecotone (Lloyd and Gramlich 1997, Kullman and Kjallgren 2000). However, field studies and remote sensing analyses that focused on recent treeline shifts during the last century have yielded ambiguous results, ranging from rapid dynamics to complete inertia (Kullman 2002). Past research has shown that climatic changes and treeline ecotone dynamics were not synchronous and also occurred at different regional and local rates (Holtmeier 2003). The amplitude and magnitude of recent warming may not be large enough to compensate for the residual negative effects of the LIA, which emphasizes the importance of site history to the present treeline dynamics (Payette *et al.*

1989, Lescop-Sinclair and Payette 1995). During the last two decades, modeling relationships between tree growth and site conditions, especially temperature, under changed climate conditions has increased in ecological studies (Kauppi and Posch 1988, Rupp *et al.* 2001). Models, however, must be as simple as possible in order to work, and thus they cannot model the complex reality, and the great regional and local variety, of the treeline ecosystem (Holtmeier 2003).

This study will help to clarify the role of the current global warming period and past climatic shifts on changes in the treeline ecosystem. The methodology includes an extensive local and regional field study focusing on a complex landscape-ecological approach, and also considers multi-century paleoclimatic shifts. Regional climate scenarios and data are currently inadequate in treeline areas. Past studies have relied on climate data collected far from these micro-site treeline environments. This research addresses the spatiotemporal patterns of tree establishment at treeline since the end of the LIA by examining the species, height, age, and mode of establishment of living trees and the locations of remnant wood. This research also examines the seedling and forest composition at treeline and addresses the issue of patterns of seedling establishment as related to climate fluctuations (warm summers versus cool summers). Few studies have reconstructed the details of population dynamics in the treeline ecotone, yet these types of investigations are important because they identify the controls on vegetational processes and could show whether vegetation response at treeline is homogeneous over broad scales.

1.5.4 Objectives

The specific objectives of investigating the response of altitudinal treeline in western Montana to environmental change were to:

- Collect well-preserved remnant wood to assist in the development of a multimillennial tree-ring chronology that will encompass multiple climate shifts;
- Use remnant samples located above the current treeline to provide information concerning the advance of treeline during warm phase climate shifts such as the MWP and subsequent treeline retreat such as during the LIA; and,
- Provide a clearer understanding of the asynchronous impact of climate on high-elevation ecosystems through the examination of the lag time between climate shifts and subsequent regeneration or mortality.

1.6 Temporal Variations in the Fire Regime of a High-Elevation Forest

1.6.1 Introduction

Fire is believed to have been historically the most important form of natural disturbance in forest ecosystems (Romme and Knight 1981, Peet 1988). The interaction of fire with the biophysical environment and vegetation has shaped landscapes and dictated species composition and structure (Heinselman 1981). Although wind and insect outbreaks clearly alter high-elevation forest ecosystems, fire is believed to also have played a dominant role in facilitating changes (Veblen *et al*. 1994). Subalpine forests of the northern Rocky Mountains are prone to large, sporadic stand-replacing fires (Kipfmueller and Baker 2000). Some subalpine forests also experience nonlethal surface fires (Habeck and Mutch 1973, Arno 1980, Huckaby and Moir 1995). The interaction of fire and forest stand development has been studied over various parts of a large elevational gradient, from the montane zone to subalpine forests (Whipple and Dix 1979, Peet 1981, Veblen 1986, Shankman and Daly 1988).

The annual area burned by wildfires has increased since the beginning of the 1970s across North America and in the Rocky Mountains region in particular, possibly in response to recent climate change and the gradual accumulation of fuels during decades of effective fire suppression (Grissino-Mayer and Swetnam 2000, Westerling *et al*. 2006). Wildfires play an essential role in forest ecosystems by removing accumulated ground fuels, maintaining heterogeneity in both age and stand structure, and preserving open, park-like conditions (Dieterich 1980; Wright and Bailey 1982). Since the early 1900s, extensive fire suppression has all but eliminated low-severity, stand maintenance fires. This has resulted in extraordinarily high fuel loads in a majority of forested areas in the Rocky Mountains (Cooper 1960, Swetnam 1990, Mutch *et al*. 1993, Covington and Moore 1992, Fulé *et al*. 1997, Veblen *et al*. 2000), and has dramatically increased the likelihood of high-severity, stand-replacing fires.

To better understand forest dynamics, areas that have been minimally disturbed by humans should be targeted for research because they contain vital information about past ecological processes. Such minimally disturbed areas exist primarily in isolated parts of the Pacific Northwest, in the American Southwest, and at high-elevation sites in the Northern Rocky Mountains.

During the 10,000 years since development of forests after the last glacial retreat, fires have had a major influence on the structure and composition of forests in the northern Rocky Mountains (Arno 1980). These forests typify ecosystems that experience infrequent, high-severity crown fires (Peet 2000, Veblen 2000). In the northern Rocky Mountains of western North America, whitebark pine historically dominated many upper subalpine forests (Tomback *et al*. 1990). In the absence of major disturbance, whitebark pine is eventually replaced by more shade-tolerant subalpine fir and Engelmann spruce. Whitebark pine often survives low-intensity surface fires, which more readily kill associated conifers (Morgan *et al*. 1992). Stand-replacing fires benefit whitebark pine as it regenerates more successfully than many associated tree species (Tomback *et al*. 1990).

A recent decline in whitebark pine abundance has been associated with less frequent fire activity (Keane *et al*. 2001). Fires in whitebark pine forests occurred at mean intervals of 30 to 300 years based on fire history information derived from fire scars and stand ages (Arno 1980, Arno 1986, Morgan and Bunting 1990, Romme 1982). Morgan and Bunting (1990) documented very frequent low-intensity fires on a relatively dry site supporting seral whitebark pine in open, park-like stands within a 100-ha area in northwestern Wyoming. Prior to 1867, the mean interval between fires was 33 years. Larsen (2005) reported a mean fire interval in western Montana between 7–25 years, with an average interval of 15.25 years. Little change between fire frequency during the pre-settlement and post-settlement periods was found as the largest fires occurred from 1719 to 1850.

Arno (1986) calculated a mean fire-free interval of 30 to 41 years with a range from 4 to 78 years for fires occurring in large stands (100 to 300 ha) in upper subalpine habitat types where whitebark pine is a seral species and replaced by climax subalpine fir. For small areas less than 1 ha, Arno and Peterson (1983) found mean fire free intervals to vary between 72 and 94 years. Romme (1982) estimated a fire-free interval of 300 years for much of the lodgepole pine forest of Yellowstone National Park, and this interval may reflect conditions of the adjacent whitebark pine forests (Tomback *et al*. 1990).

Numerous studies have been conducted that focused on the effects of climate on past fire regimes (Larsen and Delaven 1922, Grissino-Mayer 1995, Nash and Johnson 1996, Veblen *et al*. 2000, Kitzberger *et al*. 2001). Examination of the past climate shifts (MWP and LIA) and its effects on high-elevation fire regimes is a significant component of understanding the ranges and behaviors of ecosystems in the future. The proposed dendroclimatic and fire ecology research will result in increased knowledge of the effects of past fire occurrence on the whitebark pine ecosystem, the development of the subalpine ecological response to fire variability over millennial scale changes, and an increased comprehension of the effect of the MWP and subsequent LIA on fire regimes.

1.6.2 Justification

Effects of fire on ecosystem processes can be both intermediate and long-term (Reich *et al*. 1990). Oxidation of organic matter by fire reduces forest floor carbon pools and disrupts carbon and nutrient cycling (Raison 1979, Gosz 1981, DeBano 1991). Vegetation mortality from fire frequently results in a decrease of leaf area, which may cause a reduction in photosynthesis and transpiration, alternating stand microclimate, water use, and decomposition (Rundel 1982, Mushinsky and Gibson 1991). Landscape patterns, structure, and composition that result from fire directly affect species composition by altering propagule dissemination and species migration among most high-elevation tree species (Turner and Romme 1994).

Fire is an important factor in whitebark regeneration and survival on sites where it is seral (Tomback *et al*. 1990). Whitebark pine often survives low-severity fires, which more easily kill associated conifers (Morgan and Bunting 1990). Stand-replacing fires also benefit whitebark pine, for although all trees are damaged or killed, whitebark pine regenerates on burned sites more successfully than many associated tree species (Tomback *et al*. 1990). Two strategies allow whitebark pine to survive in fire-prone ecosystems: survival of large and refugia trees, and post-fire seedling establishment facilitated by Clark's nutcracker (Tomback *et al*. 1990). The Clark's nutcracker (*Nucifraga columbiana* Wilson) commonly transports seeds several kilometers (Hutchins and Lanner 1982). These birds prefer open, burned areas for caching seeds (Tomback *et al*. 1990). Although large fires are infrequent, they are ecologically important in maintaining extensive whitebark pine forests on the landscape.

The results of this study will serve to fill gaps in our understanding of the role of climate as a primary influence on fire dynamics at high-elevation sites in the northern Rocky Mountains. The upper treeline in the western United States should be targeted for a more systematic sampling of climate-sensitive specimens in the future to increase our knowledge of past climatic shifts and fire occurrence, and to further quantify the effects of climate shifts on high elevation ecosystems.

1.6.3 Objectives

The specific objectives of the southwestern Montana fire reconstruction research conducted for this study were to:

- Develop a reconstruction of fire for a high-elevation site southwestern Montana dominated by whitebark pines;

- Investigate the short-term, decadal scale (<100 years) and long-term, centennial scale (>100 years) changes in the mean fire interval;
- Examine the occurrence of fire scars in high-elevation whitebark pine in relationship to reconstructed temperature, and determine the response of fire scar formation to climate shifts that might result from oceanic-atmospheric teleconnections.
- Examine the relationship between changes in fire and changes in climate, over both short-term and long-term time scales.

1.7 Organization of the Dissertation

This dissertation consists of seven chapters. The research is introduced in Chapter One by discussing specific questions to be addressed that will increase knowledge concerning the effects of climate on high-elevation ecosystems. In Chapter Two, I discuss the environmental setting by providing information on the geology, vegetation, and climate of the Beaverhead-Deerlodge and Lolo National Forests. Chapter Three details the development of a continuous, high-resolution record of past climate at treeline in western Montana. It specifically focuses on the development of the complex patterns of negative and positive shifts in the Pacific Decadal Oscillation and the Atlantic sea surface temperatures over the last two centuries.

Chapter Four addresses the spatiotemporal patterns of tree establishment at treeline since the end of the Little Ice Age by examining the species, age, and year of establishment of living trees and the locations of remnant wood. Chapter Five examines the linkages between climatic shifts and wildfire occurrences at a high-elevation site in southwestern Montana and seeks to clarify the role that fire may have played in the paleoenvironment. Chapter Six summarizes the major conclusions from this research. The Appendices contain (1) information on the samples used to develop the tree-ring chronology, (2) the final tree-ring chronologies, and (3) the final reconstructed values of annual drought.

CHAPTER TWO

GENERAL SETTING OF THE BEAVERHEAD-DEERLODGE AND LOLO NATIONAL FORESTS

2.1 Introduction

Vegetation composition and structure vary along gradients of temperature and precipitation whether environmental conditions change with altitude (Weaver 2001) or space (Whittaker 1975). Simple parameters such as average annual temperature and total annual precipitation are predictors of the vegetation that will tend to grow in particular climates. The productivity of vegetation is largely determined by climate and is therefore expected to be highest where temperature, moisture, and nutrients are simultaneously favorable (Weaver 1992).

In the high mountain ranges of North America, relatively favorable climatic conditions during the previous century allowed conifers to advance on to previously treeless alpine sites. In the northern Cascade Mountains, invasion of subalpine meadows by conifers began in the early 1920s during a time of above-average temperatures (Lowery 1972). On Mt. Rainier, invasion of subalpine meadows began around 1910 with some especially rapid advances in the 1970s (Franklin *et al.* 1974). Establishment patterns vary and are site-specific. At Mt. Rainier, recruitment was found to be continuous on the west side since 1930 in contrast to short periods of recruitment occurring on the east side. Warm, dry summers facilitated tree establishment on the west side where snowpack is heavy, whereas cool, wet summers enhanced establishment of trees on the east side where snowpack is lower (Rochefort and Peterson 1996).

The alpine treeline ecotone comprises one of the smallest in area of the major North American ecosystem complexes, occupying high mountain summits, slopes, and ridges above the continuous forest boundary (Marr 1977). Rather than forming a distinct boundary on the landscape, the alpine treeline ecotone occurs as zones within which there is a gradual change from complete cover of one vegetation type to another (Cairns and Waldron 2003). Holtmeier (2003) defined four types of treeline boundaries, which may occur as distinctive lines, as wide ecotone boundaries, as patches in close proximity, and as a gradual transition on a single mountainside (Figure 2.1).

As the limit of a tree species' survival is reached, the forest canopy becomes discontinuous and finally is characterized by patches of trees (Marr 1977, Elliott-Fisk 1983, Malanson 1997). The trend toward discontinuous forest is often accompanied by a physiognomic shift toward low-stature tree forms and the development of krummholz growth forms (Marr 1977). Temperature (Bryson 1966, Marr 1977, Kullman 1998), moisture (Lloyd and Graumlich 1997, Cairns and Malanson 1998), snowpack (Scott *et al*. 1988), and edaphic and substrate characteristics (Timoney 1995, Kupfer and Cairns 1996) have all been shown to play important roles in determining the locations and patterns of alpine treelines. Dry ridges and exposed southern slopes within the subalpine zone in the Rocky Mountains do not support dense forest structure, but instead support forests of shorter, round-crowned, more widely spaced trees of the white pine group (*Pinus* subgenus *Haploxylon*) (Peet 1988). These sites characteristically have undeveloped skeletal soils (Arno and Hoff 1989). On the less extreme of these sites, environmental conditions promote the establishment of *Picea* and *Abies* in the understory

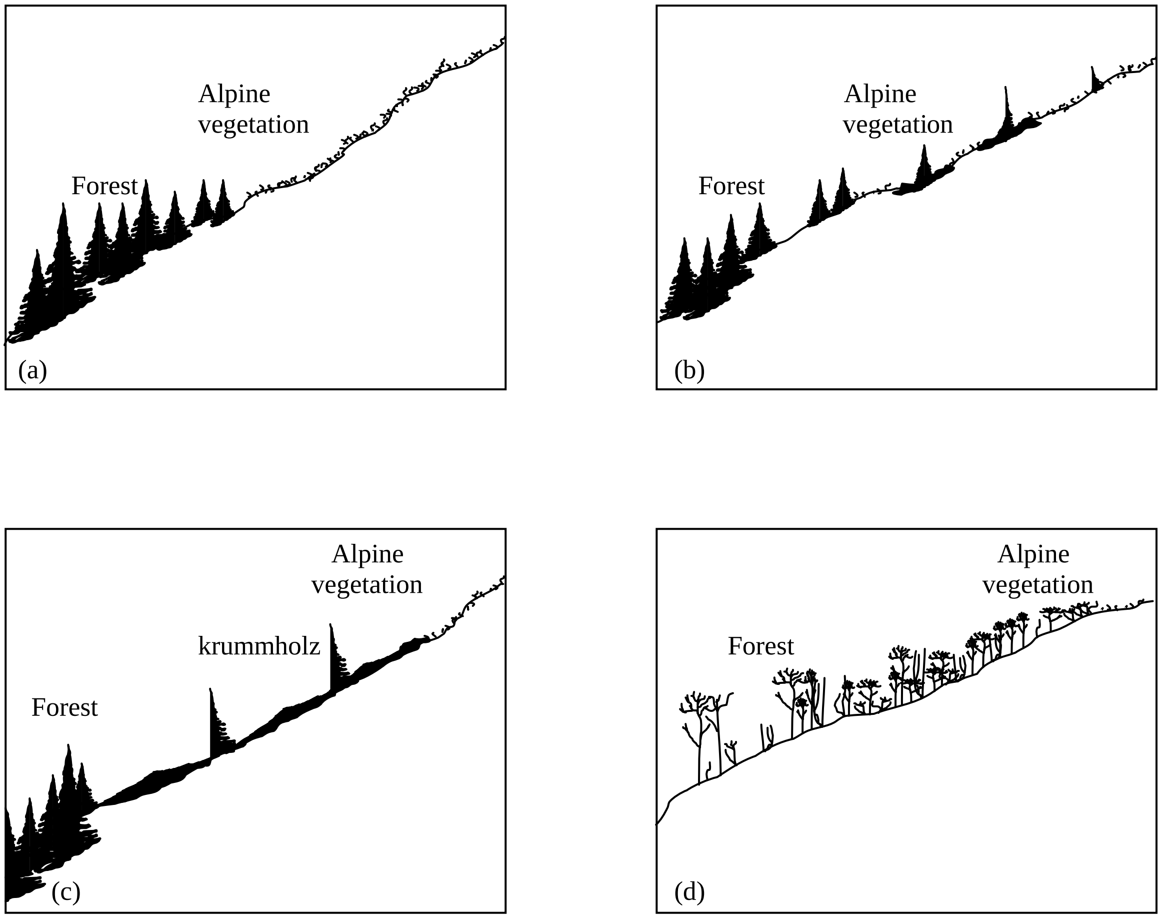


Figure 2.1. Main types of treeline: (a) abrupt forest limit bordering alpine vegetation, (b) transition zone, (c) krummholz belt above the upright growth forest, (d) gradual transition from upright stemmed forest to stunted trees of the same species bordering alpine vegetation (adapted from Norton and Schonenberger 1984).

resulting in competition with the white pines (Beasley and Klemmedson 1980, Peet 1981, Veblen 1986).

2.1.1 Topography

Although the altitudinal position of climatic treeline depends on regional climates, topography and (especially) micro-topography are key factors that control the local site conditions and patchiness of vegetation, and consequently the spatial structure and physiognomy of the treeline (Holtmeier 2003). The effects of micro-topography on radiation load and wind often override the influence of elevation locally (Holtmeier 2003). Exposure to wind and solar radiation influences temperature and moisture in the form of precipitation, relocation of snow, and evaporation, and, as a result, cause different microclimates that modify other site factors and ecological processes (Holtmeier 2003).

On steep slopes, downslope mass movements by debris slides, debris avalanches, and snow avalanches, as well as geologically defined surface structure, are important factors that affect site conditions in the treeline ecotone (Holtmeier 2003). Forest strips or fingers are often restricted to structurally sound benches and bordered by stable talus outcrops that together provide protection from avalanches and other mass movements.

The altitudinal position of treeline and the distributions of tree species are strongly influenced by exposure to solar radiation and wind (Holtmeier 2003). The contrast between southern and northern exposures and leeward and windward slopes increases or reduces the effects of wind and heating. In temperate mountain environments, treeline is found at higher altitudinal positions on western and southern exposures than on more shaded northern and eastern slopes. In the mountain ranges of Oregon and Washington, the altitudinal position of treeline on southern and northern slopes differs by 150 m (Arno 1986). Billings (1979) examined treeline in the Brooks Range of Alaska, and found that treeline on the southern slope was located at 700 m to 800 m with krummholz growth forms occurring at 950 m. On the northern exposure, the same elevations were covered with treeless tundra. Southern slopes are not only warmer but also drier than northern and eastern slopes. In the continental mountain ranges of the western United States, whitebark pine (*Pinus albicaulis* Engelm.) stands with sparse undergrowth occur on southern exposures while subalpine larch (*Larix lyallii* Parl.) with heath vegetation cover northern exposures (Franklin and Dyrness 1973, Arno 1980, Arno and Hammerly 1984, Peterson 1998).

Another additional factor that controls the altitudinal limit of treeline are the landforms found in mountain regions, such as rocky trough walls, talus cones, slope debris, and avalanche chutes. Moreover, small ridges, grooves, rills, and gullies caused by postglacial slope erosion create an irregular micro-site pattern of wide forest strips that alternate with treeless terrain oriented perpendicular to the slope gradient (Holtmeier 2003). Forest and tree fingers usually reach their highest position in small valleys that provide improved moisture conditions and shelter from strong winds (Holtmeier and Broll 1992).

2.1.2 Soil Conditions

Soil development in the upper subalpine and alpine areas of the Rocky Mountains has been studied intensively in the Colorado Front Range (Burns 1980, Holtmeier and

Broll 1992). Johnson and Billings (1962) assigned the alpine soils found under different vegetation types to four categories that are based on a classification system developed by Retzer (1956): alpine turf soils, alpine meadow soils, alpine bog soils, and lithosols. Under the current soil taxonomic definitions (Soil Survey Staff 1999), Johnson and Billings (1962) defined alpine turf as well as alpine meadow soils as Inceptisols.

Due to the complexity and diversity of alpine and subalpine landscapes, climate, and topographical factors, the development of alpine soils is not as complete as that of other lower-elevation areas (Soil Survey 1994). The grain size distribution found in high-elevation soil typically reflects the parent material. Siliceous rock, particularly granite and gneiss as found in the northern Rocky Mountains, produces coarser grain structures than calcareous rock, and this has important consequences for water-holding capacity (Korner 1999). Also, grain size tends to increase with altitude due to less weathering of parent material, resulting in pedalogically younger soils with silt and clay contents highest at the surface (Munn and Spackman, 1990).

Frost action has been proposed as a means of silt relocation. The process of gelifraction, including frost-wedging, has been identified as a means by which sedimentary and igneous bedrock detritus is reduced in size near the surface (Munn and Spackman, 1990) and also results in needle ice formation near the surface. Deeper freezing may also occur (Korner 1999) that can result in vegetation heaving and even the loss of substrate (Johnson and Billings 1962).

2.1.3 Wind

At high-elevation mountain ranges at mid-latitudes that are influenced year-round by the westerly winds, wind speeds are highest during the winter months (November–February) (Holtmeier 2003). At seven valley sites in Montana, wind speeds average 17.0 km/hr annually with the highest wind speeds occurring in January (NOAA 2005).

In mountainous terrain, wind speeds and directions are controlled by the local topography and as a result are different from air currents above the ranges (Holtmeier 2003). Ridges and gullies can alter wind speed when the wind is perpendicular to the ridges (Barry 1992). This results in a variety of wind-exposed and wind-protected sites that strongly influence the ecological conditions at treeline (Holtmeier 2003) and comprise one of the most important agents affecting site conditions in the treeline ecotone. Wind affects tree growth physiologically and mechanically (abrasion, desiccation) and by influencing other site conditions (air and soil temperature, soil moisture, height and duration of snow cover) (Holtmeier 2003). These effects are most distinct in high mountains that typically have high mean wind speeds close to the surface, as in the Rocky Mountains, where evaporation is also amplified by advection of relatively dry air from western directions (LeDrew 1975, Barry 1992).

2.1.4 Snowpack

Snowpack is an important limiting factor to tree growth in subalpine forests with respect to length of growing season (Graumlich 1991, Peterson 1998). Snowpack acts to insulate the ground effectively from the cold. Surface snow is coldest throughout the winter while near the soil it is rarely below freezing (Holtmeier 2003). This was demonstrated by Seppala (1994) in northern Finland who found the upper surface, where the snow is under the direct influence of wind and re-radiation, to reach a minimum temperature of –23.4 ºC, while the bottom edge of the snowpack stayed a constant 0.1 ºC over a six month period.

Snowpack also has significant ecological impacts both through its influence on water supply and its role as a physical force important in controlling the distribution of vegetation (Arno and Hoff 1989). Duration of snowpack also limits seedling establishment in subalpine meadows (Fritts 1976) and after disturbance by fire. In the *Pinus albicaulis-Abies lasiocarpa* habitat, snowpack extends from October to the following April, reaching a mean depth between 460 cm to 1270 cm (Arno and Hoff 1989).

2.1.5 Temperature

Temperature affects tree growth, regeneration, and survival directly through low mean temperatures during the growing season, shortness of the growing season, and by random unusual weather events such as heavy frost during the growing season (Holtmeier 2003). Temperature at treeline can also be influenced by other site factors such as wind and duration of snow cover and its effects on soil temperature. Long-lasting snowpack and high soil moisture due to melt water can keep soil temperatures low until early summer or longer (Tranquillini 1979). On wind-exposed sites, air temperature, needle temperature, and soil temperature are usually low when compared to sites with topographic protection, and this may hamper regeneration and growth (Tranquillini 1979).

Globally, treeline has been estimated to approximate the altitudinal position of the isotherm representing the mean air temperature of the warmest month of 10 ºC (Korner 1998). Comparisons made between mean annual air temperatures at treeline in the subtropics and at high latitudes show significant differences. However, several authors have found that mean air temperature at treeline in the tropics ranges from 5–6 ºC throughout the year (Smith 1980). Similarly, mean air temperature at treeline in the subarctic was found to range between 6–7 ºC during the warmest month (Tuhkanen 1993). In the continental mountains of the northern hemisphere, the mean air temperature for the warmest month instead ranged from 10–13 ºC (Tuhkanen 1993). Korner (1998) suggested that the mean air temperature of the warmest month could be systematically overestimating the actual temperatures during the growing season and that most growth in subalpine forests occurs early in the season when temperatures will vary depending on the specific site locations or micro-site characteristics.

2.2 High-Elevation Tree Species

2.2.1 Whitebark Pine

Whitebark pine is classified as a stone pine of section Strobus that characteristically has five-needled fascicles, wingless seeds, and cones that remain closed at maturity (Little and Critchfield 1969) (Figure 2.2). Whitebark pine is a common constituent of subalpine vegetation in the xeric areas of the northern Rocky Mountains

Figure 2.2. Multi-stemmed whitebark pine in the Beaverhead-Deerlodge National Forest. Inset shows the five-needle fascicles of whitebark pine.

and occurs as a minor element into the alpine zone where it forms dense stands of krummholz (Arno and Hammerly 1984). Whitebark pine is a main forest constituent on warm, dry exposures (Tomback and Linhart 1990). The eastern range of whitebark pine is primarily in the Northern Rocky Mountains of western Wyoming, Montana, and central Idaho with a secondary population along the coastal mountains of British Columbia and the Sierra Nevada (Critchfield and Little 1966) (Figure 2.3). In the continental climate of central Montana, whitebark pine is found as a codominant with subalpine fir, spruce, and alpine larch (Baig 1972) (Figure 2.4). Its seeds are a significant food source for grizzly bears (*Ursus arctos horribilis*) and other wildlife of the high mountain areas (Tomback and Linhart 1990).

The presence and dominance of whitebark pine depend on its environmental tolerances and on its competitive abilities (Arno and Hoff 1989). Whitebark pine is excluded from low-altitude grasslands and Douglas-fir (*Pseudotsuga menziesii* (Mirb.) Franco) forests by low annual precipitation and extended periods of drought and competition for water resources (Weaver 1989). In the alpine zone, whitebark pine often occurs in pure or mixed groves or as tree islands and is considered part of the climax community (Pfister *et al*. 1977, Arno and Hammerly 1984). In the subalpine zone, whitebark pine can be found on steep, rocky ridges from which cold air drains (resulting in warm night conditions) and on which excessive drainage occurs (causing drought conditions better tolerated by whitebark pine than subalpine fir) (Weaver 2001). Its comparatively low capacity to compete restricts it to these harsh sites (talus slopes and bedrock outcrops) where growth of more competitive species is hindered by physical factors or by disturbance (Arno and Weaver 1989).

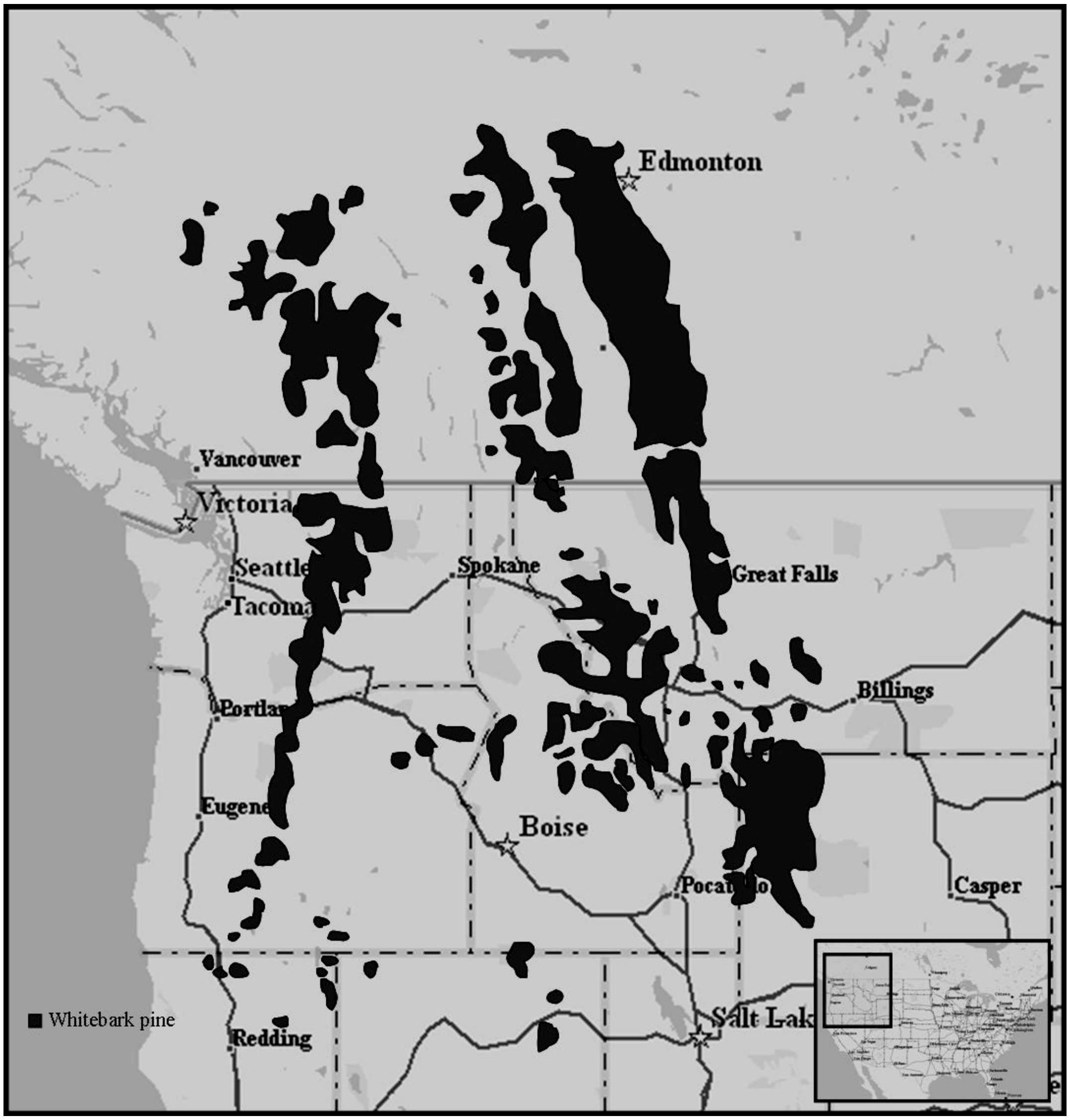


Figure 2.3. Distribution of whitebark pine in the northern Rocky Mountains (adapted from Arno and Hoff 1989).

Figure 2.4. Dense stands of whitebark pine and subalpine fir at treeline (2850 m) in the Beaverhead-Deerlodge National Forest.

Its early success is possible due to the hardiness of whitebark pine in the severe microclimate of a disturbed site and its introduction by the seed-caching Clark's nutcracker (Tomback 1989). Large crops of seed are produced at uneven intervals, with smaller crops and failures occurring in between (Tomback 1978). Dense cones, usually 5 to 8 cm long, contain the large seeds that are heavy and wingless (McCaughey and Tomback 2001). Clark's nutcrackers play a vital role in planting whitebark pine seeds (Tomback 1978, Lanner 1980, Lanner and Vander Wall 1980, Hutchins and Lanner 1982). Nutcrackers cache from one to several whitebark pine seeds in soil suitable for germination (Tomback 1982, Tomback 1986). Whitebark pine seeds sustain these birds but a fair amount of the seed caches are unrecovered (Lanner 1980). Various mammals, such as red squirrels (*Tamiasciurus hudsonicus* Erxleben), also transfer and cache whitebark pine seeds (Tomback 1978, Hutchins and Lanner 1982). Red squirrels gather large quantities of whitebark pine cones and accumulate them in rotten logs and on the ground (Mattson and Reinhart 1994). Black (*Ursus americanus* Pallas) and grizzly bears raid many of these cone caches, dispersing additional seeds.

Short, cool summers occur in the whitebark pine habitat with mean July temperatures ranging from 13–15 °C and from 10–12 °C in the upslope treeline zone (Weaver 2001). A cool growing season lasts between 90 to 110 days (Arno and Hoff 1989) and is defined by mean temperatures higher than 5.5 °C (Baker 1944). Mean annual precipitation for whitebark pine stands ranges between 600 mm and 1600 mm (Weaver 2001). Conversely, whitebark pine is also a long-persisting seral associate on xeric sites in the upper subalpine forest in subalpine fir habitat types in Montana, central Idaho, and western Wyoming (Weaver and Dale 1974, Arno and Hoff 1989).

In Montana, the whitebark pine stands at treeline are identified as the *Pinus albicaulis-Abies lasiocarpa* habitat type (Daubenmire and Daubenmire 1968, Pfister *et al*. 1977). Prior to the early 1900s, whitebark pine was more productive in the subalpine forest as a consequence of natural fires that favored its regeneration over competing fir (Arno 1986). Natural disturbance plays a vital role in shaping the structure of whitebark pine communities and its perpetuation. In large portions of the western United States, the area covered by seral whitebark pine communities has diminished in recent decades (Arno 1986). At treeline, climatic effects from wind, ice, snowload, summer frost events, and winter desiccation prevent stand closure and thereby allow whitebark pine to coexist with tolerant competitors such as subalpine fir (Franklin and Dryness 1973, Arno and Hammerly 1984). Due to fire suppression and epidemics of mountain pine beetle and white pine blister rust beginning in the early 20th century, the natural fire cycle in whitebark pine communities has been disturbed, causing this species to be replaced by competitors such as subalpine fir (Arno 1986). Also, during the period of extensive grazing activity in these high-elevation areas, which lasted from the mid-1800s to the early 1900s, many of these areas were severely impacted by eliminating the fine fuels necessary for a fire to occur.

Although whitebark pine is known to be intolerant of competition or shade (Baker 1949), recent observations (McCaughey and Schmidt 2001) reveal that it is actually intermediate to intolerant. On dry, wind-exposed sites, whitebark pine is a climax species that persists as an associate with subalpine fir and other species because it is longer-lived, drought tolerant, and hardier (Pfister *et al*. 1977). On these harsh sites, whitebark pine establishes and is later locally replaced by tolerant fir (Franklin and Dyrness 1973).

Most whitebark pine stands grow on weakly developed soils. Many of the sites were covered by extensive alpine glaciers during the Pleistocene and have been released from glacial ice for less than 12,000 years (Mehringer *et al*. 1977). Widespread disparity occurs in regional climates and amount of soil development in whitebark pine habitats (Thompson and Kuijt 1976). The majority of soils under whitebark pine stands are characterized as Inceptisols (USDA 1975). The best-developed soils found in spruce-fir/whitebark pine stands are Typic Cryandepts (Nimlos 1963). Open, park-like stands are often found on coarse talus, exposed bedrock, or lava flows that have minimal soil development (Ogilvie 1989). These soils are categorized as fragmental and loamy skeletal families within the order Entisols (Cryorthents in granitic substrates) (USDA 1975).

Primary undergrowth species in the Rocky Mountains include grouse whortleberry (*Vaccinium scoparium* Leib. ex Coville), mountain arnica (*Arnica latifolia* Bong. var. *gracilis* (Rydb.) Cronq.), red mountain heath (*Phyllodoce empetriformis* (Sm.) D. Don), rustyleaf menziesia (*Menziesia ferruginea* Sm.), smooth woodrush (*Luzula hitchcockii* (Hoppe ex Rostk.) Desv. var. *hitchcockii* (Hämet-Ahti) Dorn), beargrass (*Xerophyllum tenax* (Pursh) Nutt.), elk sedge (*Carex geyeri* Boott), Parry rush (*Juncus parryi* Engelm.), Ross sedge (*Carex rossii* Boott), and Idaho fescue (*Festuca idahoensis* Elmer ssp. *idahoensis*) (Peet 2000).

2.2.2 Subalpine fir

In the Rocky Mountains, subalpine fir establishes in combination with Engelmann spruce and creates an Engelmann spruce-subalpine fir forest cover type (Alexander 1980)

(Figure 2.5). Major treeline associates are mountain hemlock and whitebark pine (Alexander *et al*. 1984). At lower limits in the Rocky Mountains of Montana and Idaho, minor associates include western white pine (*Pinus monticola* Douglas ex D. Don), interior Douglas-fir (*Pseudotsuga menziesii* (Mirb.) Franco var. *glauca* (Beissn.) Franco), western hemlock (*Tsuga heterophylla* (Raf.) Sarg.), western larch (*Larix occidentalis* Nutt.), grand fir (*Abies grandis* (Dougl. ex D. Don) Lindl.), and western red cedar (*Thuja plicata* Donn ex D. Don). Also, at higher elevations, minor associates include lodgepole pine (*Pinus contorta* Dougl. ex Loud.), alpine larch, mountain hemlock, and whitebark pine (Franklin and Dyrness 1973).

In the Rocky Mountains, subalpine fir extends from its northern limits in central British Columbia, south to southern New Mexico and Arizona and into western Wyoming (Grant and Mitton 1977) (Figure 2.6). Subalpine fir prefers the coolest and wettest sites in the continental areas of the western United States (Thornthwaite 1948). Subalpine fir grows where mean annual temperatures vary between –3.9 °C to 4.4 °C (Baker 1944, Marr *et al*. 1968, Haeffner 1971).

In the Rocky Mountains of western Montana and Idaho, subalpine fir is found between 610 and 3353 m (Daubenmire and Daubenmire 1968). Subalpine fir begins cone production when trees are 1.2 to 1.5 m tall. Subalpine fir that are 150 to 200 years old produce the highest levels of viable seeds (USDA 1974). High-quality subalpine fir seed crops occur every 3 years with light crops or failures in between (LeBarron and Jemison 1953, Franklin *et al*. 1974). Cones disintegrate when they are ripe. Scales fall away with the large, winged seeds, leaving only a central, spikelike axis (Franklin *et al*. 1974). Dissemination begins in September and is completed by the end of October in the Rocky

Figure 2.5. Subalpine fir at treeline (2850 m) in the Beaverhead-Deerlodge National Forest.

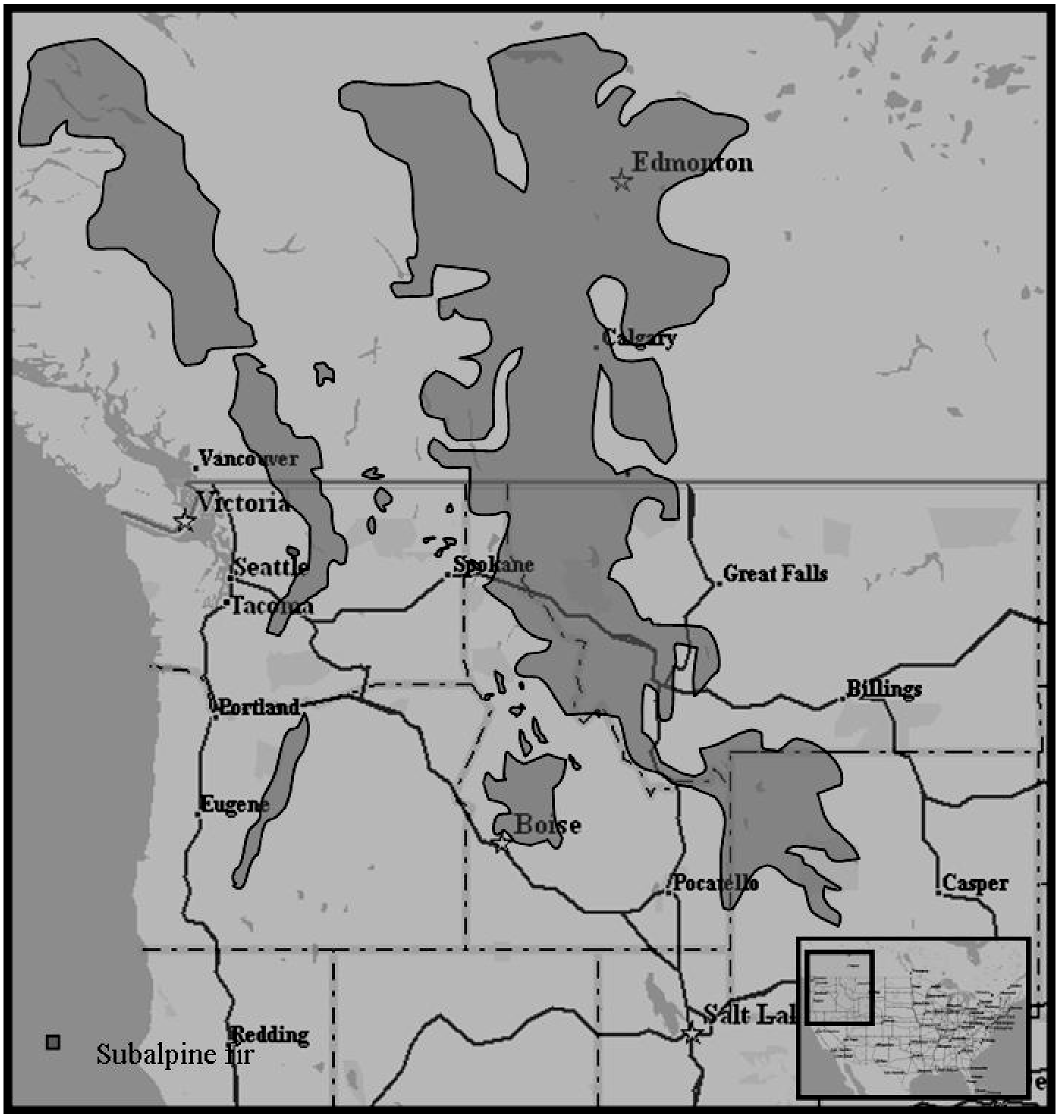


Figure 2.6. Distribution of subalpine fir in the northern Rocky Mountains (adapted from Alexander *et al*. 1984).

Mountains (USDA 1974).

The majority of seed dispersal is by wind. Subalpine fir is constrained to cold habitats due to its limited tolerance to high temperatures (Shearer 1980). Numerous seedlings are killed and injured by spring frosts, frost heaving, birds, and rodents (Alexander 1987). Subalpine fir commonly reproduces vegetatively on severe sites by layering (Franklin and Dyrness 1973). On sites exposed to wind, subalpine fir is frequently reduced to a krummholz form, under closed-forest conditions it reaches diameters between 30 cm and 61 cm and heights between 14 m and 30 m (Marr *et al*. 1968).

Subalpine fir is commonly found on soils that are too wet or too dry for other species to establish (Hanley *et al*. 1975). Increased growth is found on lower-elevation slopes, alluvial floodplains, and glacial moraines, and on well drained soils (Alexander *et al*. 1984). Growth is limited on shallow soils and on saturated soils, but subalpine fir establishes on severe sites, such as avalanche zones, prior to any of its common associates (Kirkwood 1922). Under these conditions, it may pioneer the site earlier than other species resulting in the exclusion of other species (Franklin and Mitchell 1967).

Understory plant growth commonly found in the Rocky Mountains includes mountain bluebells (*Mertensia ciliata* (James ex Torr.) G. Don) and heartleaf bittercress (*Cardamine cordifolia* Gray var. *cordifolia*) (cool, moist sites); thimbleberry (*Rubus parviflorus* Nutt.) (warm, moist sites); red buffaloberry (*Shepherdia Canadensis* (L.) Nutt.), mountain snowberry (*Symphoricarpos oreophilus* Gray var. *oreophilus*), Rocky Mountain whortleberry (*Vaccinium myrtillus* L.), grouse whortleberry (*Vaccinium scoparium* Leib. ex Coville), fireweed (*Chamerion angustifolium* (L.) Holub ssp.

angustifolium), heartleaf arnica (*Arnica cordifolia* Hook.), polemonium (*Polemonium pulcherrimum* Hook. ssp. *delicatum* (Rydb.) Brand), daisy fleabane (*Erigeron eximius* Greene), elksedge (*Carex geyeri* Boott), boxleaf myrtle (*Pachystima myrsinites* (Pursh) Raf.), prickly currant (*Ribes lacustre* (Pers.) Poir.), sidebells pyrola (*Orthilia secunda* (L.) House), and mosses (Alexander 1988).

2.3 The Beaverhead-Deerlodge National Forest

The Beaverhead-Deerlodge National Forest (BDNF), located in southwestern Montana, is the largest national forest in Montana, covering 1,343,000 hectares and divided into nine separate sections (USDA 1997) (Figure 2.7). The BDNF consists of deep valleys separated by forested mountain ranges and bounded by the Beaverhead Mountains to the west, the Anaconda Range to the north, the Pioneer Mountains to the east, and the Big Hole Divide to the south (USDA 1997). The BDNF comprises two major drainage basins, the Beaverhead River basin and the Big Hole River basin (USFS 1998). This network of upper watershed drainage basins makes up the headwaters of the Missouri River (USDA 1997).

2.3.1 Geology, Soils, and Topography

The BDNF is located at the junction of two thrust plates, the Medicine Lodge Thrust Plate and the Grasshopper Thrust Plate (Lopez 1981). During the middle Proterozoic, the Yellow Jacket Formation was deposited on the Medicine Lodge Plate and is made up of thin to medium bedded fine-grained feldspathic quartzite interbedded

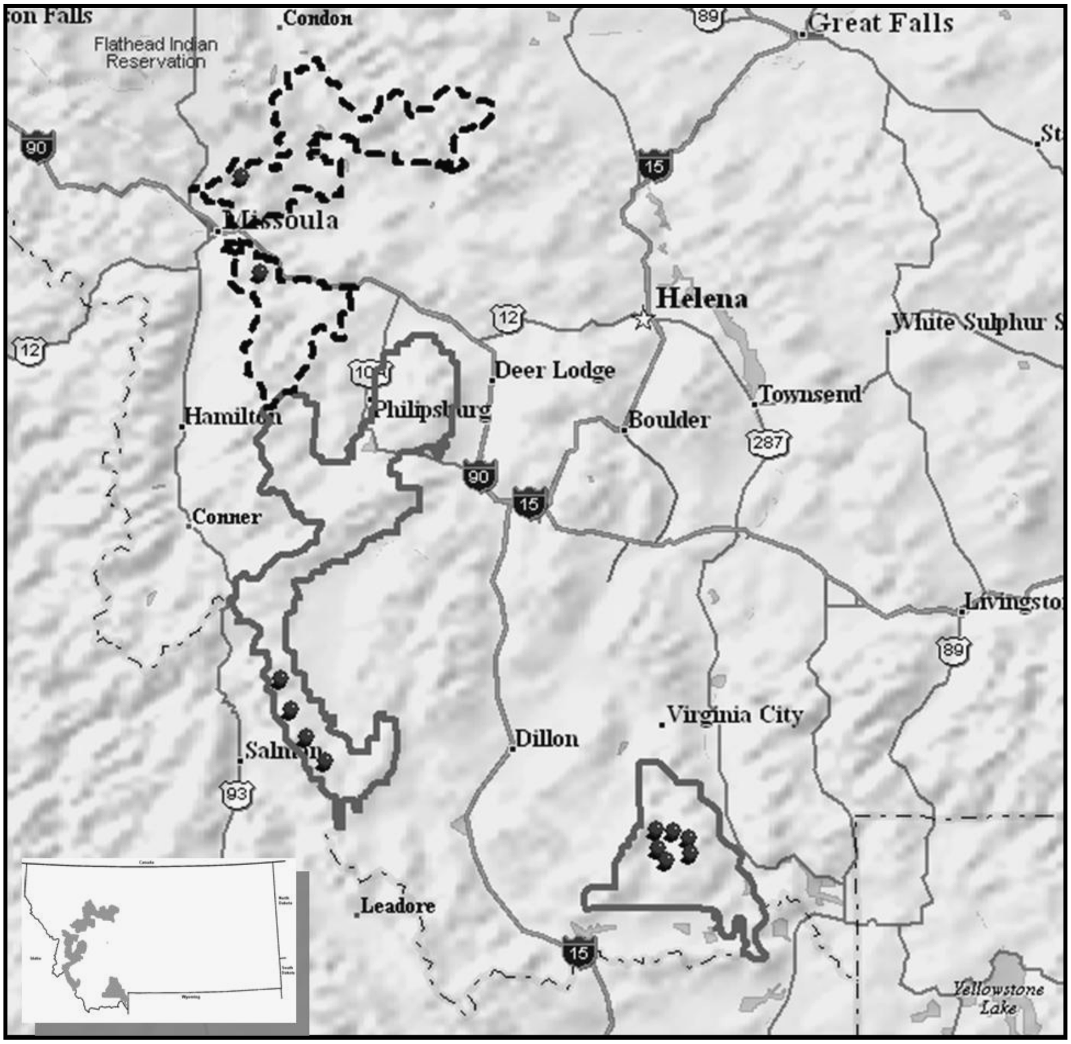


Figure 2.7. The Beaverhead-Deerlodge National Forest is outlined by a heavy solid line and the Lolo National Forest is outlined by a heavy dashed line. Study sites are indicated by dots (map created using Delorme mapping software).

with siltite and argillite (Lopez 1981). The Grasshopper Thrust Plate has an overlying Missoula Group/Mount Shields Formation deposited during the middle Proterozoic and consists of fine- to coarse-grained feldspathic quartzite interlayered with quartz-pebble conglomerate (Dyman and Nichols 1988). The valleys that surround the BDNF consist of glacial till composed of an unsorted mixture of boulders, cobbles, pebbles, and sand. The till was deposited by three separate episodes of glaciation (Dyman and Nichols 1988). The East Pioneer, Tobacco Root, Pintlar, Beaverhead, Gravelly, Snowcrest, Tendoy and Centinnial Ranges all experienced Pleistocene alpine glaciations (Lopez 1981). Boulder fields in the southern portion of the study area are characterized by frost riving and frost-heaved boulders in boggy and terraced terrain that formed by creep and solifluction on flanking hills (Lopez 1981). These boulder fields formed as a result of weathering in severe cold climate adjacent to Pleistocene glaciers (Dyman and Nichols 1988).

High-elevation soils in the BDNF are unstable due to their parent material including calcareous and non-calcareous sedimentary as well as crystalline materials (Lopez 1981). The combined exposure of soil, gravel, and rock usually exceeds 50% and runs as high as 90% with leaf litter the only other important component (Alden 1953).

The BDNF ranges in elevation from 1400 m at the Beaverhead River to 4000 m at the summit of Tweedy Mountain in the East Pioneer Range (Klepper 1950). The diverse topography of the BDNF, high relief, and climatic variability result in an abundance of unique environments (DeBlander 2001). Elevations within the Beaverhead and Gravely Mountains range from 1768–3415 m (DeBlander 2001).

2.3.2 Climate

The climate of the BDNF is typical of higher elevations in southwestern Montana with very cold, dry winters and mild summers. Precipitation and temperature from five surrounding climate stations for the 1955–2005 period of record indicate an average annual precipitation of 68 cm and an average temperature of 3.7 ºC (NOAA 2005). The average annual temperature for the same period at Wisdom in the central BDNF is 6.8 ºC, which is the lowest of all recording stations in the study area (NOAA 2005). The average last occurrence of 0 ºC is late May, and the average first occurrence is early September (NOAA 2005). The BDNF lies in the rainshadow of Oregon's Blue Mountains and central Idaho's high mountain mass, resulting in lower amounts of precipitation from Pacific storm systems when compared to areas in northern Montana (Ross and Hunter 1976).

2.3.3 Vegetation

Of the 480 plant associations identified for Montana, over half (57%) occur within the BDNF (Pfister *et al*. 1977). Given that the BDNF comprises less than one tenth of Montana's total land area, the region exhibits the greatest ecological diversity in the state (Pfister *et al*. 1977). Glaciation and periglacial climates have resulted in environmental gradients that produce communities with high diversity (Peet 1988).

Vegetation within the BDNF ranges from Douglas-fir and limber pine forests in the montane zone to western spruce and fir in the montane to subalpine zone to alpine meadows and barrens above 2900 m (Peet 2000). Alpine turf and cushion plant communities are limited and often found on the highest ridgelines (USDA 1997).

Additional vegetation characteristics include mountain grasslands dominated by Idaho fescue (*Festuca idahoensis* Elmer) and shrub lands dominated by mountain big sagebrush (*Artemisia tridentata* Nutt. ssp. *vaseyana* (Rydb.) Beetle) (Peet 2000).

2.3.4 Land Use History

2.3.4.1 Pre Euro-American settlement

Southwestern Montana, including the BDNF, has been occupied by human groups for at least the last 12,000 years (Sanders 1998). The oldest documented human cultural groups in the BDNF are from the Paleo-Indian Period (12000–7500 yrs BP) (Soffer and Praslov 1993). Evidence of occupation for the Archaic Period (7500–1500 yrs BP) and the Late Prehistoric Period (1500–250 yrs BP) has also been collected in the BDNF (Sanders 1998).

The nomadic Indian tribes of Lemhi, Flathead, Shoshoni, Bannock, Crow, Blackfeet, and Nez Perce used the BDNF as common hunting grounds prior to the arrival of settlers from the east (Brown 1975). Of these tribes, the Lemhi and Nez Perce stayed within the BDNF year round (Brown 1975). The Lemhi presence in the BDNF was due to raiding by the Blackfeet which pushed them into the southwestern corner of Montana (Sanders 1998). This forced the Lemhi to move away from a dependence on buffalo to mountain sheep and deer found in higher-elevations (Brown 1975). The Lemhi were the first Indian tribe Lewis and Clark encountered after leaving the Mandan Indian villages on the plains (Brown 1975). The Nez Perce, located in Oregon, maintained buffalo hunting grounds in the BDNF (Brown 1975). In 1877, Colonel Joseph Gibbon attacked the Nez Perce encampment in the Big Hole Valley, driving the tribe across the Continental Divide and into north central Montana (Sheridan 1882).

2.3.4.2 Post-settlement

The first record of European movement into the BDNF occurred in August 1805 when the Lewis and Clark expedition traveled through the Big Hole Valley and made contact with the Lemhi Indian tribe (Appleman 1993). Organized fur trade in the BDNF began with Manuel Lisa's voyage up the Missouri River in 1808 (Oglesby 1963). Beginning in 1810, the Missouri Fur Company sent men into the area and in 1830 the American Fur Company began operations in the area (Chittenden 1935). As the fur trade declined, trappers began to harvest buffalo hides in the BDNF (Brown 1975). In 1830, the Blackfeet buffalo trade alone was estimated at 10,000 to 20,000 robes (Chittenden 1935).

Gold was discovered in the BDNF in 1863 and within a year 10,000 miners had begun working these deposits (Barsness 1962). Hydraulic mining operations began in the 1870s and their effect on the landscape can still be seen today in the form of exposed talus slopes (Burlingame 1957). Beaverhead County was first established in the Territory of Idaho in January 1864 and became part of Montana when the Montana Territory was admitted into the Union in May 1864 (Brown 1975).

2.4 The Lolo National Forest

The Lolo National Forest (LNF), located north of Missoula, Montana, covers an area of 810,000 hectares (USFS 1986). Due to its location west of the continental divide, the LNF is influenced by both continental and maritime climates (USFS 1986). These climates provide a wide range of environmental gradients that produce a forest of high species richness. Elevation ranges from less than 732 m at the Clark Fork River to 2743 m at Lolo Peak (USFS 1998).

2.4.1 Geology, Soils, and Topography

The LNF is located on the western end of the Garnet Range and at the northern edge in the Sapphire tectonic block at the junction of the Clark Fork fault and the Blackfoot thrust plate (Obradovich 1993). The Garnet Range formation consists of medium brown, poorly sorted, very fine-grained sandstone and siltstone (Cobban 1945). The Sapphire tectonic block moved east into Montana between 75 and 70 million years ago (Cobban 1945). The formation varies in thickness due to pre-Middle Cambrian erosion and ranges from less than 1 m to 490 m in thickness (Cobban 1945). The Clark Fork fault and Blackfoot thrust plate are overlain by the Belt formation which consists mostly of Precambrian sedimentary rock formations made up of the Snowslip Formation and the Mount Shields Formation (Obradovich 1993). The Snowslip Formation consists of argillite and siltite with thin beds of fine-grained quartzite (Obradovich 1993). Thin beds of limestone and flat pebble conglomerate also occur. The age of the Snowslip Formation is uncertain. Obradovich (1993) dated the southern portion of the formation to 1,170 ± 20 Ma.

2.4.2 Climate

The climate of the LNF is representative of higher elevations in western Montana, with very cold, dry winters and mild summers. Average annual precipitation is 480 mm with the majority falling in the winter and early spring (NCDC 2005). Temperatures range from –2 ºC in the winter to 13 ºC during the summer (NCDC 2005). The western section of the LNF lies in the rainshadow of the Bitterroot Mountains, a position that results in lower amounts of precipitation from Pacific storm systems when compared to areas in the eastern section of the national forest (Ross and Hunter 1976).

2.4.3 Vegetation

The LNF is covered 95% by forested area made up of 14 forest types comprising five hardwood species and 17 conifer species (DeBlander 2001). Plant communities, which together encompass over 1,500 plant species, include non-forest vegetation, dry-warm and dry-cool Douglas-fir types, alpine fir types, and spruce types (Peet 2000). Whitebark pine, subalpine fir, Engelmann spruce, and alpine larch dominate high elevations and create a park-like landscape. Grouse whortleberry (*Vaccinium scoparium* Leib. ex Coville), red mountain-heath (*Phyllodoce empetriformis* (Sm.) D. Don), woodrush (*Carex luzulina* Olney), and bear grass (*Nolina* Michx.) dominate the understory (Peet 1988). Areas above treeline are dominated by shrub and herbaceous plant communities made up of elk sedge (*Carex garberi* Fern.), pinegrass (*Calamagrostis rubescens* Buckl.), twin flower (*Linnaea borealis* L.), shooting star (*Dodecatheon* L.), yellow avalanche-lily (*Erthronium grandiflorum* Pursh), mountain arnica (*Arnica* *montana* L.), arrowleaf ragwort (*Senecio triangularis* Hook.), and dwarf blueberry (*Vaccinium caespitosum* Michx.) (Peet 1988).

2.4.4 Land Use history

2.4.4.1 Pre-Euroamerican settlement

The LNF was used by the Blackfeet, Kootenai, and the Salish Indian tribes before settlers arrived in the 1800s (Chittenden 1935). The Blackfeet comprised three major tribes, the Piegan, the Blood, and the Blackfeet (Brown 1975). Estimates of their population in 1780 were as high as 15,000. Early trading with the English allowed them to acquire guns and horses (Chittenden 1935). Due to their high population numbers, the Blackfeet raided competing tribes and pushed out of the area (Brown 1975). In 1805, the Lewis and Clark Expedition passed through what is now the LNF using the Lolo Trail (Appleman 1993). On the expedition's return trip through the area, the Blackfeet attacked and killed a number of party members (Appleman 1993). The Blackfeet's strength was greatly reduced by disease, especially smallpox, and by 1830 their population had been reduced to an estimated 10,000 (Brown 1975).

2.4.4.2 Post-settlement

The LNF has historically had abundant mineral resources and since 1860 continuous mining operations have been carried out (Safford 2004). Prospecting occurred throughout the region. Following the initial increase in mining activity for precious metals, mining interests broadened to include numerous resources including antimony, barite, copper, sapphire, gold, and silver. In the mid-1980s, large-scale mining operations were underway with 107 hectares being mined for gravel and sand. Oil and gas have not yet been exploited within the forest, but large-scale exploration began in the mid-1980s, with over 360,000 hectares under lease (USFS 1986).

Euro-American settlers began moving into the LNF near the mouth of Rattlesnake Creek in 1858 (USFS 1986). Large-scale logging was restricted by terrain and did not begin until the region was opened to rail at the turn of the last century (DeBlander 2001). Since the 1920s, logging operations have focused on clear-cutting operations at low and mid-elevation forests of ponderosa pine, Douglas-fir, western larch, and lodgepole pine (USFS 1998). Due to inaccessibility, many high-elevation forests in the LNF have never been logged (DeBlander 2001). Over 410,000 hectares have been considered appropriate for harvesting and timber production (USFS 1986).

CHAPTER THREE

DENDROCLIMATOLOGICAL ANALYSES AND CLIMATE RECONSTRUCTION IN THE NORTHERN ROCKY MOUNTAINS

3.1 Site Selection

To acquire the longest and clearest dendroclimatic record, it is essential to sample sites where climatic conditions are most likely to restrict physical processes that affect tree growth (Fritts 1976). The justification for selecting the sites described in Chapter 2 was predicated on two of the principal objectives of this study, which were to develop a multi-century tree-ring chronology from high-elevation sites, and to reconstruct climate based on these chronologies. I selected high-elevation, climate-sensitive sites with steep exposed slopes, open-grown stands, coarse well-drained soils, and southerly aspects. These areas were located in the western section of the Beaverhead-Deerlodge National Forest (BDNF), the Gravely Range in the eastern BDNF, and the Lolo National Forest (LNF) (Figure 3.1). These areas are dominated by whitebark pine and subalpine fir. The sites are characterized by open forest of multi-stemmed whitebark pines that displayed classic signs of exceptional longevity, such as flat-topped crowns, a common trait of open forest whitebark pine communities (Tomback *et al*. 2001). Site selection was based on (1) southern slope aspect to increase the possibility of extracting a temperature signal (Fritts 1976), (2) elevations at or near the ecological limit of whitebark pine and subalpine fir, and (3) presence of remnant wood in the form of both standing dead trees and downed woody debris.

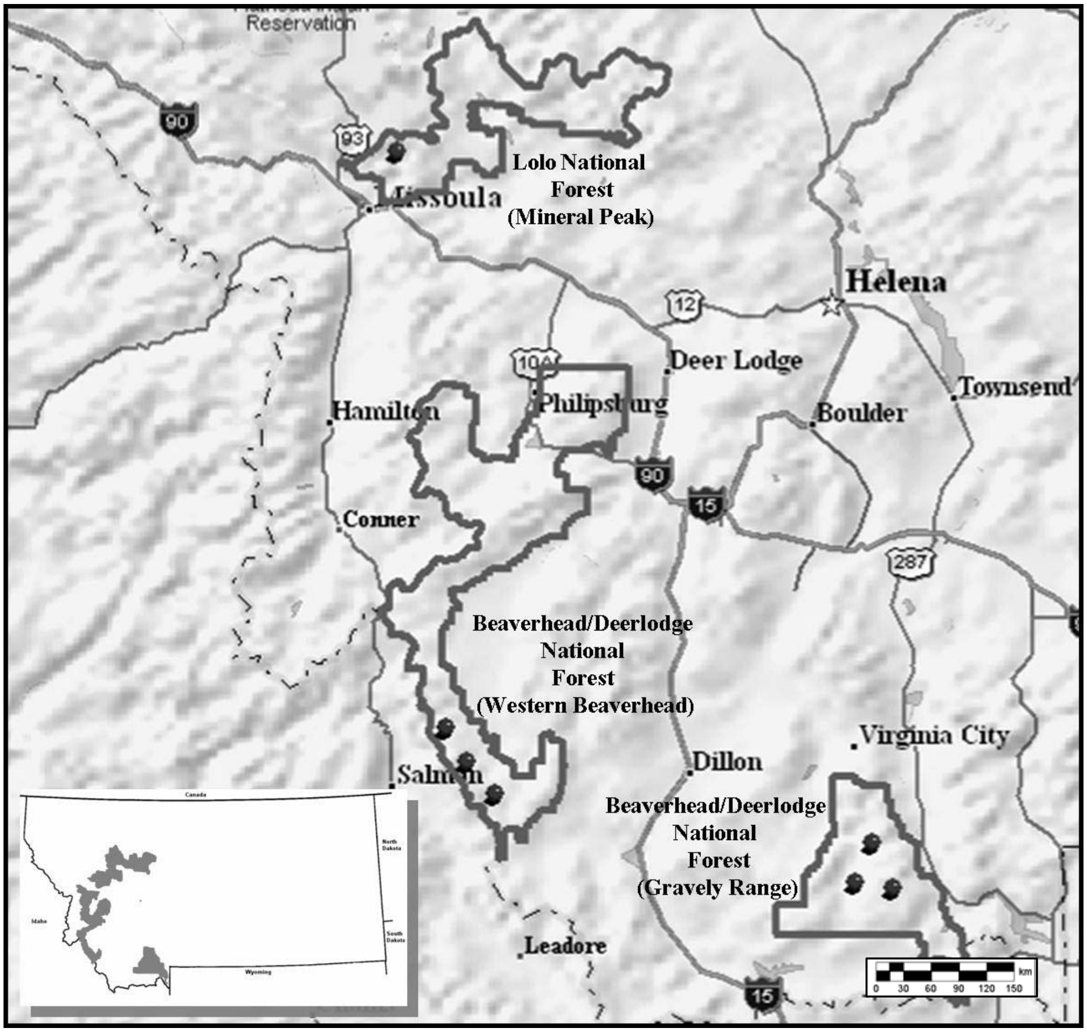


Figure 3.1. Map of study area showing the Western Beaverhead (WB = N45.25, W113.42), Gravely Range (GR = N45.00, W111.51), and Mineral Peak (MPK= N47.00, W113.48) (map created using Delorme mapping software).

3.2 Methods

3.2.1 Field Methods

Increment cores and cross sections were collected from selected whitebark pine and subalpine fir trees growing below and at the present treeline on southerly exposures. Sampling focused on subjectively located stands of older trees with collection sites widely dispersed to fully sample the elevational range of whitebark pine and subalpine fir. I collected two radii from all trees with a diameter at breast height (dbh) of 5 cm or greater. Although the smaller diameter trees are young, these were necessary to ensure accurate crossdating in the last two centuries as older whitebark pine trees can have problematic ring sequences in the most recent areas of growth. Trees were sampled with an increment borer at a height of 30 cm (Figure 3.2). Cores were then placed in paper straws for protection, and all relevant information on each tree sampled was recorded on standard forms. Cross sections were collected using a chain saw from standing dead and downed whitebark pine (Figure 3.3). Sections were not collected from living whitebark pine trees, however, because the species is currently being decimated in many areas of its native range from infection by the near-ubiquitous white pine blister rust.

3.2.2 Laboratory Methods

3.2.2.1 Sample Preparation and Measuring

In the laboratory, cores were air-dried, glued to wooden core mounts, and allowed to dry overnight. Cores and cross sections were sanded using a progression from ANSI 40-grit (500–595 µm) to 400-grit (20.6–23.6 µm) to ensure the cellular features of the individual tree rings were visible under standard 7–10X magnification (Orvis and

Figure 3.2. Collecting cores with an increment borer from whitebark pine at the Western Beaverhead site.

Figure 3.3. Collecting cross-sections using a chain saw at the Western Beaverhead site.

Grissino-Mayer 2002). This process allowed the cell wall structure and the earlywood and latewood boundaries of each annual ring to be examined and identified under 10X magnification. Steel wool was also used to remove excess resin to help differentiate the tree-ring boundaries.

To more closely examine tight clusters of rings, rotten wood, or poorly sanded areas, a razor blade was used to surface these areas. Initial analyses used all cores regardless of tree age, but eventually only cores with > 250 annual rings were used to preserve as much low frequency variability in the final tree-ring chronology as possible (Cook *et al*. 1995).

3.2.2.2 Crossdating and Master Chronology Development

The fundamental technique of dendrochronology is crossdating (Douglass 1934, Douglass 1935, Stokes and Smiley 1968), the evaluation of ring patterns within a single tree, between trees, between species, within age classes of single species, and between sites and regions (Stahle 1978). Tree rings from samples collected in the BDNF and LNF were crossdated using established visual match- mismatch methods (Phipps 1985), and labeled (Stokes and Smiley 1968) by noting the occurrence of extremely wide or narrow marker rings. Total ring widths were measured using a Velmex movable stage micrometer and recorded with Measure J2X software to the nearest 0.001 mm.

Measurements were then processed through the computer program COFECHA to check the visual crossdating with correlation analysis using overlapping 50-year segments of all measured cores lagged 25 years (Holmes 1983, Grissino-Mayer 2001). COFECHA identifies tree-ring segments that should be re-evaluated for crossdating errors and performs quality checks (Holmes 1983). In cases for which COFECHA indicated ring segments were possibly misdated, I visually re-examined the sample and corrected the date as required. In the case of missing or false rings, parts 5 and 6 of the COFECHA output were used. Part 5 of the COFECHA output shows the correlation of each tree-ring series broken down by 50-year segments with the master chronology, while Part 6 provides critical diagnostic information for identifying segments with potential dating problems (Grissino-Mayer 2001). Flagged segments were also examined to ensure no measurement errors occurred. To remove low-frequency variance that may hamper crossdating, COFECHA fits a cubic smoothing spline to each series (Cook and Peters 1981) as well as performing a log transfer and autoregressive modeling. A master dating series was then developed by taking the mean of all transformed series by year after indices were developed for each year by dividing the actual value by the predicted value obtained from the detrending process. The resulting series is then tested against the master dating series using correlation analysis conducted on each 50 yr segment. The COFECHA program checks for a positive and significant correlation for each segment. It also checks whether the correlation is higher when segments are shifted forward or backward from that point. Single measurements that were statistical outliers after filtering and transformation were also noted (Holmes 1983).

To develop a high-resolution tree-ring chronology, physiological and competing environmental factors (age, slope, competition) must be removed from the raw ring-width measurements. This process is achieved by standardizing the ring measurements and developing a site chronology. Standardizing is necessary because individual trees grow at different rates and so faster-growing individuals with higher absolute ring widths would dominate a simple average of ring width (Fritts 1976, Cook 1985). Also, most individuals put on narrower rings as age increases (age-related growth trends) and a simple average of tree growth would be skewed by the long-term ring width decline due to increasing stem circumference instead of reflecting the desired signal, such as temperature, precipitation, or drought. The standardization process is used to remove these growth factors and to clarify the desired climatic signal. The chronologies produced in the standardization process are an index of ring-width values with a mean of one and relatively homogeneous variance through time (Fritts 1976).

Tree-ring measurement series were standardized by entering the measurement data into the program ARSTAN (Cook 1985). ARSTAN computes ring-width indices by dividing the ring-width value by the value computed by a cubic smoothing spline fit to each series with a 50% cutoff wavelength (Cook and Peters 1981, Cook and Holmes 1986). The smoothing spline is a centrally-weighted moving average on the data, and splines work better than polynomials at approximating functions that are episodic, such as tree-ring series (Cook and Peters 1981).

Master chronologies were developed for each site in the BDNF (the western Beaverhead and the Gravely Range) and in the LNF (Mineral Peak). I developed a combined master chronology for the larger study area (BDNF and LNF) by combining series from each site into a composite chronology. The chronologies for three individual study areas were further combined to develop a regional chronology for western Montana.

3.2.2.3 Instrumental Climate Data

Assessment of the climate-tree growth relationship was then carried out utilizing divisional climate data from the National Climatic Data Center (NCDC 2007). Regional climate data from 1895–2003 rather than single station data were used because regional data provide calibration and verification statistics better than do single stations (Blasing *et al*. 1981). This allowed for less total variance to estimate and less potential variance in regression models (Cook *et al*. 1996). For the BDNF chronologies, I used climate data for NOAA Climate Division MT-02-Southwestern 2402. For the LNF chronologies, I used climate data for NOAA Climate Division MT-01-Western 2401.

The climate variables used in the climate-response analysis included monthly average temperature, monthly total precipitation, Palmer Drought Severity Index (PDSI), and the Palmer Hydrological Drought Index (PHDI). PDSI and PHDI are used by the National Weather Service to monitor drought and wetness conditions in the United States and are a measure of the moisture conditions during the growing season (Stahle *et al*. 1998). PDSI has been used in dendroclimatic studies and is often significantly correlated with tree-ring indices in North America (Grissino-Mayer and Butler 1993, Watson and Luckman 2001, Woodhouse and Brown 2000). PDSI describes the severity of both wet and dry periods and incorporates temperature, precipitation, and evapotranspiration as an estimate of soil moisture availability as a monthly index (Palmer 1965). The index is a weighted average of soil moisture conditions for the current and preceding months resulting in a strong month-to-month autocorrelation that represents how soil moisture conditions change over time (Stahle and Cleveland 1988). The index ranges from –6 (very dry) to +6 (extremely wet), with zero values illustrating normal conditions (Palmer

1965, NCDC 2004). Values greater than +4 or less than –4 are considered extreme climatic conditions (Meldahl *et al*. 1999).

PHDI is a hydrological variation of the PDSI that is used to evaluate long-term moisture supply (NCDC 2007). PHDI estimates subsurface hydrological characteristics and reacts more gradually to changes in weather conditions than does the PDSI (Grissino-Mayer and Butler 1993). The values for PHDI are similar in range and sign to PDSI values, with negative values indicating drought.

I also examined possible relationships between high-elevation tree growth and oceanic-atmospheric teleconnections associated with the Pacific Decadal Oscillation (PDO), Atlantic Multidecadal Oscillation (AMO), Northern Oscillation (NOI), and Arctic Oscillation (AO).

The PDO is the dominant decadal mode of North Pacific sea surface temperature (Mantua *et al*. 1997). The PDO varies between warm (positive) and cool (negative) phases every 20 to 30 years (Mantua *et al*. 1997). Warm phases prevailed from 1925 to 1946 and from 1977 through the mid-1990s, while cool phases were dominant from 1890 to 1924 and from 1947 to 1976 (Mantua and Hare 2002). A number of scientists believe PDO modulates ENSO in the U.S. (Kerr 1999, Biondi *et al*. 2001). Moreover, climate anomalies associated with PDO are similar but less extreme than those linked to ENSO variations, as the warm-phase PDO produces anomaly patterns similar to El Niño (Mantua and Hare 2002). The warm-phase ENSO events are typically stronger when they are in phase with a PDO event, so ENSO-related rainfall anomalies are often associated with PDO phases (Enfield *et al*. 2001). For example, in the PDO warm phase, El Niño events exhibit a generally stronger pattern of wet winters in the southern tier of the U.S. (Gershunov and Barnett 1998).

The AMO is a multi-decadal pattern of surface temperature variability centered on the North Atlantic Ocean (0^o – 70^o N) (Kerr 2000, Knight *et al.* 2005). The AMO index is a 10-yr running mean of detrended sea surface temperature (SST) anomalies averaged over the North Atlantic (Enfield *et al.* 2001). The oscillatory nature of the phenomenon, which occurs on approximately 50 to 70-year cycles, is related to the variability in oceanic thermohaline circulation and the associated meridional heat transport (McCabe *et al*. 2004, Knight *et al.* 2005).

In the 20th century, the AMO had an important role in modulating boreal summer (June, July, and August) climate in the U.S. on multidecadal time scales (Sutton and Hodson 2005). In the warm (positive) phase of the AMO, summer precipitation in many regions of the U.S. is lower and temperatures are warmer than in the cool phase (McCabe *et al*. 2004, Sutton and Hodson 2005). Warm phases of the AMO occurred in the late 19th century and from 1931 to 1960. Midwest droughts in the 1930s and 1950s have been linked to AMO (Enfield *et al.* 2001). Cool phases of the AMO occurred from 1905 to 1925 and from 1965 to 1990 (Sutton and Hodson 2005).

The NOI is an index of climate variability based on the difference in sea level pressure anomalies at the North Pacific High in the northeastern Pacific and near Darwin, Australia (Schwing *et al*. 2002). The NOI contains alternating decadal scale periods dominated by positive and negative values, suggesting significant climate shifts on a roughly 14-year cycle (Schwing *et al*. 2002). Positive shifts occurred during 1970–1976 and 1984–1991 (Schwing *et al*. 2002). Negative values predominated in 1965–1970, 1977–1983, and 1991–1998 (Schwing *et al*. 2002).

The dominant pattern of atmospheric variability during winter over the extratropical Northern Hemisphere is referred to as the Arctic Oscillation (AO) (Thompson and Wallace 2001) and is characterized by a redistribution of atmospheric mass between the polar latitudes and midlatitudes (Deser 2000). A positive phase of the AO corresponds to reduced sea level pressure over the Arctic and increased westerly winds at midlatitudes (Thompson and Wallace 2001). The largest changes in midlatitudes associated with the AO occur over the Atlantic sector in Northern Hemisphere winter and are known there as the North Atlantic Oscillation (NAO) (Delworth and Dixon 2000).

3.2.2.4 Statistical Analysis of Climate Response

Correlation analysis is a practical first step in establishing the association between tree growth and climatic variables (Gholz 1982) and can provide information similar to analysis of variance (Fritts 1976). Pearson correlation coefficients were calculated between growth indices and climate variables (temperature, precipitation, PDSI, and PHDI) for a 24-month period (previous January–current December). Monthly and seasonal climate variables were used in the calculations. I included the prior year's climate because carbon is fixed during the previous growing season and the dormant season (Fritts 1976, Kozlowski 1979, Grissino-Mayer and Butler 1993, Foster and Brooks 2001).

3.2.2.5 Developing the Climate Reconstruction

I selected the climate variable with the highest correlation coefficient with the tree-ring indices to perform the reconstruction. To develop the strongest association with climate for calibration, I divided the full period of instrumental climate record (1895–2005) into subsets (1900–2005, 1910–2005, 1920–2005, and 1930–2005) (Stahle and Cleaveland 1992). The subsets were necessary because climate data in earlier periods may be sparse and/or suspect. I developed calibration equations that calculated the selected climate variable using tree-ring indices as predictors by means of ordinary least-squares regression analysis. These equations are commonly known as transfer functions (Fritts 1976). Outliers that exceed tolerances after inspection of the Studentized residuals (< –2 or > +2) and Cook's d (approx. 0.1) statistics were evaluated for possible removal from the model (Grissino-Mayer 1995). I withheld outlier observations from the final model according to the applicable tolerance criteria (Grissino-Mayer 1995).

I verified the calibration equations using the subset withheld from the calibration computations to determine how well the tree-ring reconstruction replicated the instrumental record of climate (Fritts 1976). To test the statistical integrity of the reconstructions, I used the Pearson correlation and the t-test over the verification period (Stahle and Cleaveland 1992, Grissino-Mayer 1996, Cook *et al*. 1996). These tests compare the predicted values of climate to the actual values of the instrumental record. If the statistical tests proved statistically significant, I used the entire period of the instrumental record to compute a new transfer function for the final reconstructions. The final transfer function is more statistically rigorous than in the test sequence because the number of observations used in the model is increased (Grissino-Mayer 1996).

I also examined annual, decadal, and multi-decadal trends to determine the coherence of climate conditions at high-elevation sites in western Montana. I first examined extreme individual years of drought and wetness and compared the composite chronology to determine if a regional pattern of extreme conditions existed during those years. I also identified periods of above and below average moisture conditions by smoothing the climate reconstructions with a centrally weighted moving average. I analyzed decadal trends using a 10-year moving average and multidecadal trends using a 50-year moving average. To enhance the calculated degree of past climatic fluctuations, I converted the reconstructed values to standard deviation units (z-scores) by subtracting the series mean and dividing by the standard deviation of the series (Grissino-Mayer 1995). As a guideline, I arbitrarily defined the initiation of an extreme period when the standard deviation unit first fell below a value of –1.0 sd or above +1.0 sd. The termination of an extreme period was considered complete when the yearly value exceeded –1.0 sd or dropped below the +1.0 sd level. I also compared the persistence of dry or moist conditions at the individual sites at the interannual, decadal, and multidecadal level to determine climate variability through time.

3.3 Results

3.3.1 Tree-ring Chronologies

The three sites together totaled 332 measured series, while the composite chronology for western Montana totaled 213 measured series (Table 3.1). Series with weak or nonexistent crossdating were not used. This included series that crossdated poorly and would have amplified nonclimatic noise in the chronologies and therefore

Table 3.1. Summary of tree-ring data for the western Montana master chronologies. WB = Western Beaverhead-Deerlodge, GR = Gravely Range, MPK = Mineral Peak.

	Period of record	Number of samples	Interseries Correlation	Mean sensitivity
WB	AD 1116–2003	140	0.54	0.22
GR	AD 728–2005	52	0.41	0.20
MPK	AD 981–2003	34	0.52	0.23

weakened the desired climate signal. The number of series that were not included from WB, GR, and MPK were 32, 24, and 63 series, respectively. The number of MPK series that did not crossdate was high in large part due to the effects of wind and possible differences in tree growth on the windward and leeward sides of the tree stem. The MPK site is also more open with fewer trees being grouped together than at the WB and GR sites.

Two descriptive statistics, interseries correlation and mean sensitivity, were used to describe the statistical quality of the site and composite chronologies. Interseries correlation represents the Pearson correlation coefficient among individual series at each site and is considered a measure of the signal-to-noise ratio in the developed chronology (Foster and Brooks 2001). The interseries correlations of the WB, GR, and MPK chronologies were 0.54, 0.41, and 0.52, with the composite chronology exhibiting an overall interseries correlation of 0.47. These values are similar to the Larson (2005) study in which the interseries correlations between sites in western Montana ranged from 0.28 to 0.48.

Mean sensitivity is a measure of the comparative differences in ring width from one ring to the next and is calculated by averaging the percent change from ring to ring (Fritts 1976) and is an acceptable indicator of climate responsiveness. Mean sensitivity for the individual sites and composite chronologies ranged from 0.20 (GR) to 0.23 (MPK), with an overall composite value of 0.21. The values are comparable to the Biondi et al. (1999) study in which the mean sensitivity for whitebark pine ranged from 0.20 to 0.22 in central Idaho.

The lengths of the GR and composite chronologies are equal in length (1277 years) to the longest whitebark pine chronology for the western United States developed by Kipfmueller (2003) from the Selway-Bitterroot region of Idaho. The MPK site chronology is the seventh longest whitebark pine chronology and the WB the ninth longest, comparing with data in the International Tree-Ring Data Bank (ITRDB). The most long-lived tree in all chronologies came from the WB site and contained 607 rings (sample AJ002).

I used the STANDARD rather than the RESIDUAL chronologies produced by ARSTAN because I found better correlations between STANDARD chronologies and climate variables than with the RESIDUAL chronology. Also, the RESIDUAL chronology represents a white noise series with all low-frequency variation removed, and this would not have proved useful for analyzing long-term (>100 years) trends in climate (Grissino-Mayer 1995).

A comparison of the site and composite chronologies with other tree-ring chronologies from the region illustrates the strength of association and the commonality of the climate signal. I chose three regional chronologies for comparison with the site chronologies (Table 3.2). I computed Pearson correlation coefficients between the developed site chronologies and the region chronologies. In all cases, the correlation coefficients were statistically significant ($p < 0.001$). Due to the strength of the relationships, the sites appear to indicate a common climate signal.

Table 3.2. Correlation coefficients between study sites and adjacent sites.

Site	Species	Correlation Coefficient
WB		
Helena, Montana (MT100)*	Douglas-fir	0.41
GR		
Helena, Montana (MT100)*	Douglas-fir	0.44
Butte, Montana (MT103)*	Douglas-fir	0.38
MPK		
Mineral Peak (Larson 2005)	Whitebark pine	0.48
Selway-Bitterroot (Kipfmueller 2003)	Whitebark pine	0.42

* from the International Tree-Ring Data Bank

3.3.2 Climate Response

3.3.2.1 Precipitation

The effects of preconditioning by rainfall in the previous and current year were significant. The highest correlations between tree growth and precipitation occurred in the previous year's May and June and current year's May for the composite chronology (Figure 3.4). The positive correlation indicates that an increase in precipitation in the previous year's May and June and current May result in increased tree growth. This could possibly indicate that, during the growing season (June to September), the most influential months for precipitation are the previous year's May and June. In western Montana, maximum snow pack occurs in May and decreases exponentially into June, resulting in an increase of available moisture during this period. Rainfall in western Montana reaches its highest levels in June (Figure 3.5). However, yearly snowpack levels far exceed levels of rainfall. The pattern of snowpack melt may reflect periods of seasonal growth activity. For whitebark pine, shoot growth and bud break occur in late May (Weaver 2001). This would indicate that snowpack melt in previous year's May and June could possibly be the most important factor for growth and sustainability of whitebark pine.

3.3.2.2 Temperature

The relationship between temperature and whitebark pine growth was somewhat stronger than that of precipitation. The most comprehensive trend concerning temperature was the negative influence of previous year's temperatures on current year's cambial growth in the composite chronology. The most common physiological explanation for

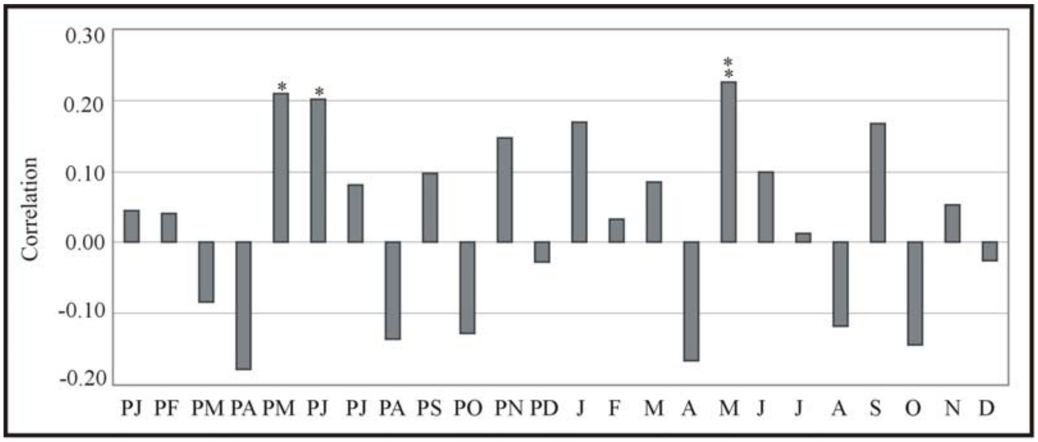


Figure 3.4. Correlation coefficients between the western Montana composite chronology and Montana climate region 2 precipitation. Asterisks indicate significance (* $p < 0.05$, ** $p < 0.01$).

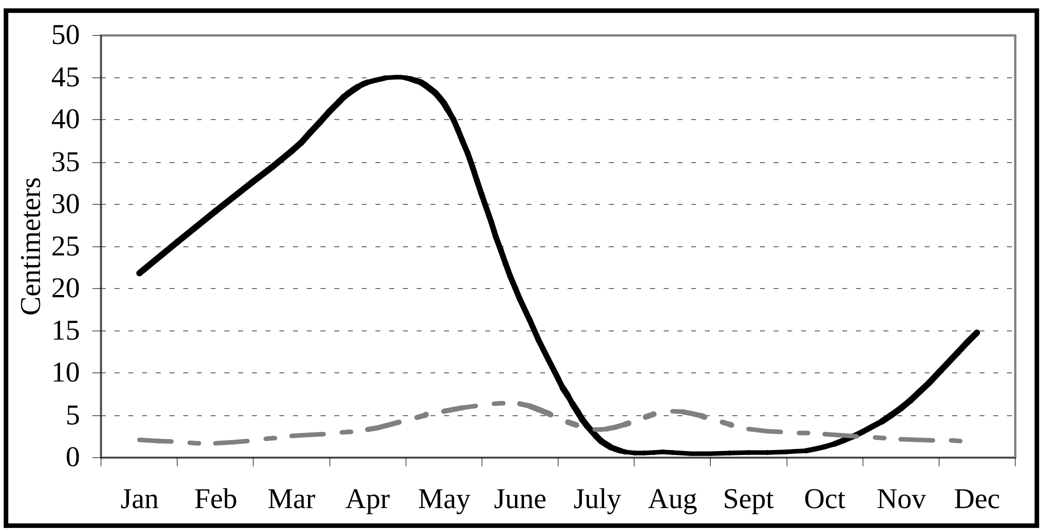


Figure 3.5. Yearly snowpack levels (black line) and rainfall (gray line) in centimeters for western Montana. Levels of snowpack are significantly higher than rainfall levels with snowpack moisture being released in early May and June.

this relationship is the carryover effect of food storage from one year to the next (Fritts 1976). Perkins and Swetnam (1996) found significant negative relationships between whitebark pine growth and spring temperatures in central Idaho. Positive associations with summer temperature have also been documented in the whitebark pine ecosystem (Luckman *et al*. 1997, Biondi *et al*. 1999). The composite chronology illustrated a negative relationship to previous and current year's July temperatures (Figure 3.6), suggesting that growth in whitebark pine has an inverse relationship to growing season temperatures. The negative correlation implies that warm temperatures in the previous and current year's July would result in reduced ring growth while cooler temperatures would result in increased growth. This could indicate that, during the growing season (June to September), July temperatures have the greatest degree of influence on ring growth. In western Montana, snowpack is at its lowest levels from July to October (< 3 cm) and temperatures are at their highest levels (> 15 °C) (Figure 3.7). If snowpack levels in the previous year are low, this could possible induce moisture stress during periods of higher temperatures in the current year. The significant correlation between both previous year's and current year's temperature and growth may not actually be a causative relationship because temperatures are frequently associated with precipitation and cloud cover. At high-elevation sites, anomalous weather patterns can quickly reduce temperatures for extended periods.

3.3.3 Palmer Drought Severity Index and Climate Teleconnections

Correlation coefficients between PDSI variables and whitebark pine growth were significant throughout most months during the previous and current years. However,

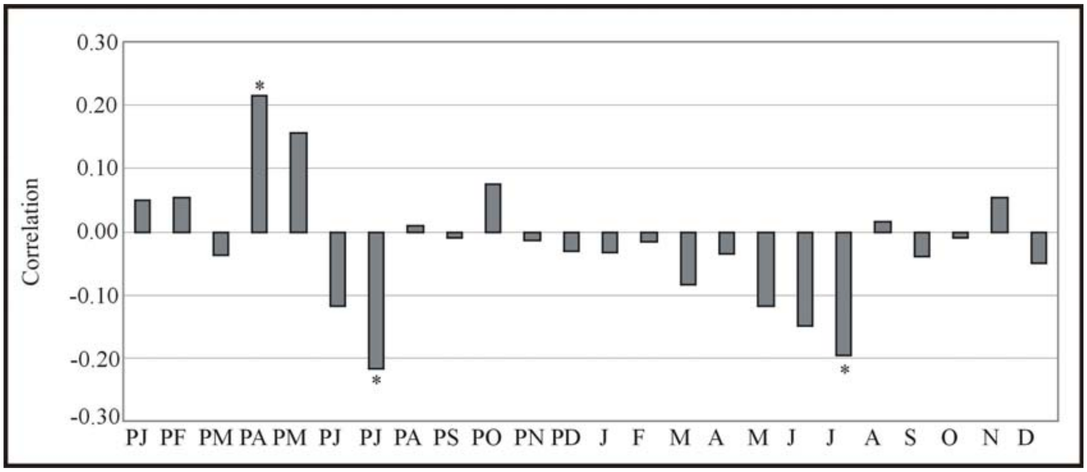


Figure 3.6. Correlation coefficients between western Montana composite chronology and Montana climate region 2 temperature (* $p < 0.05$).

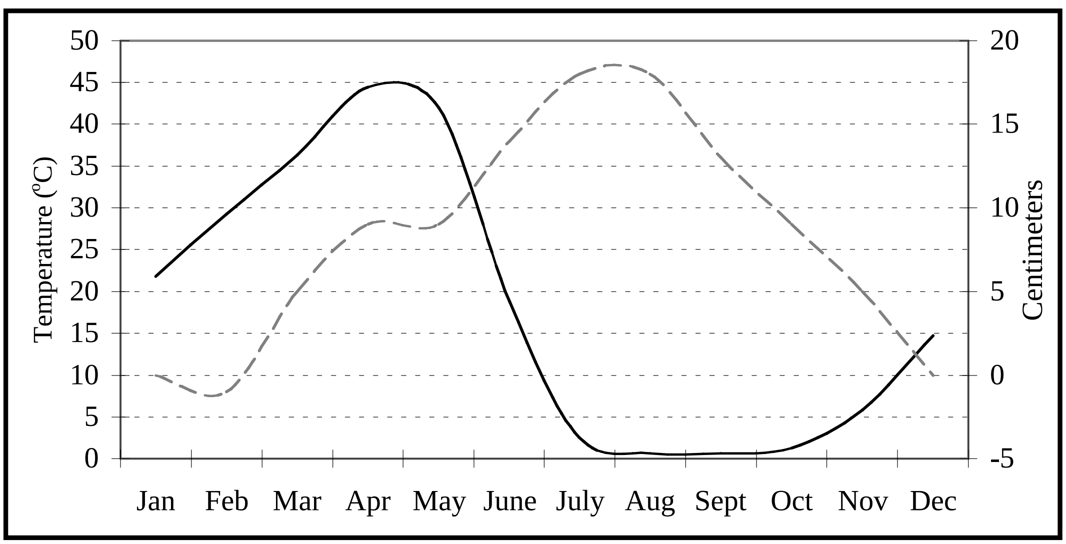


Figure 3.7. Yearly snowpack levels (black line) and temperature (dashed gray line) for western Montana. Levels of available moisture are lowest during the growing season (June to September) while temperatures are at their highest levels.

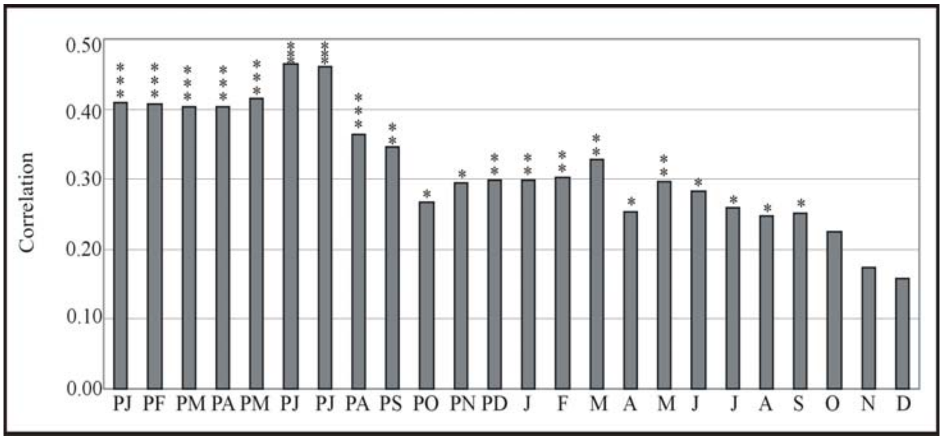
trends differed between the WB/GR sites and the MPK site. At the WB site, all months in the previous and current years, with the exception of current year's October, November, and December, were significant and positively correlated to PDSI (Figure 3.8a). At the GR site, all months during the previous and current years were positive and significant with the exception of previous year's April (Figure 3.8b). At the MPK site, only the previous year's January to December illustrated a positive and significant relationship with PDSI (Figure 3.8c). The composite chronology showed a strong positive relationship in nearly all months, with the highest correlation occurring in previous year's June and July (Figure 3.9). Correlation analysis between PDO, AO, NAO, and NOI and the composite chronology showed weak relationships (Figure 3.10).

3.3.4 Climate reconstructions

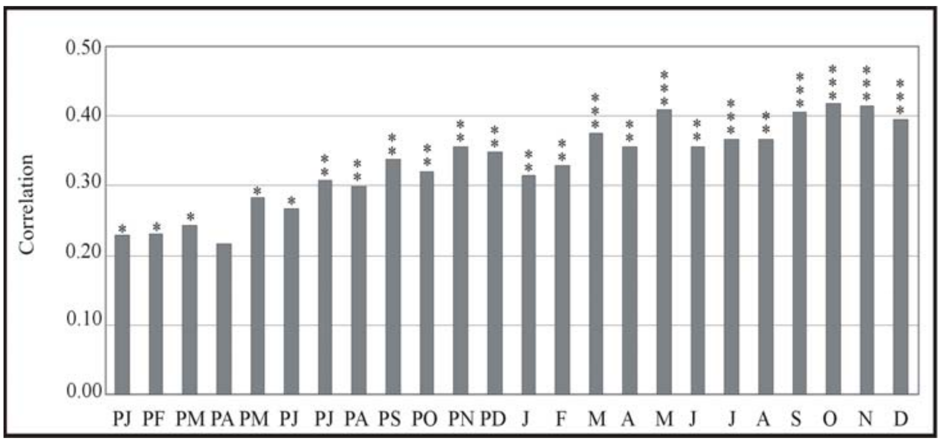
3.3.4.1 Model results

I reconstructed previous year's June and July average PDSI because of the high level of variance explained in the regression model developed for all sites back to 1930. Because of the resolution built into the Palmer indices, previous year's June and July provided a high-quality representation of the common moisture conditions throughout the summer growing season in high elevation whitebark pine ecosystems.

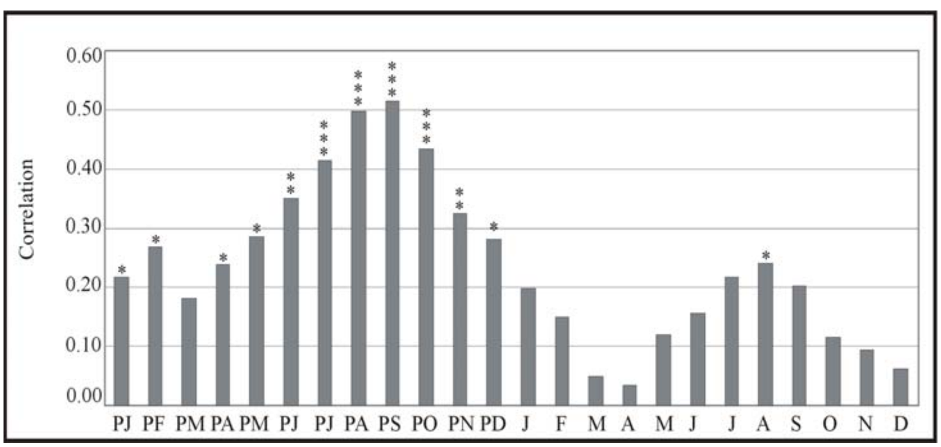
The result of the calibration and verification procedure showed that the percent of variance explained by the regression model ranged from 18% to 32% (Table 3.3). Outlier observations distinguished by high values of Studentized residuals were identified and withdrawn from the calibration model. All models were verified with a statistically



a. Western Beaverhead



b. Gravely Range



c. Mineral Peak

Figure 3.8. Correlation coefficients between Western Montana site chronologies and PDSI (* $p < 0.05$, ** $p < 0.01$, *** $p < 0.001$).

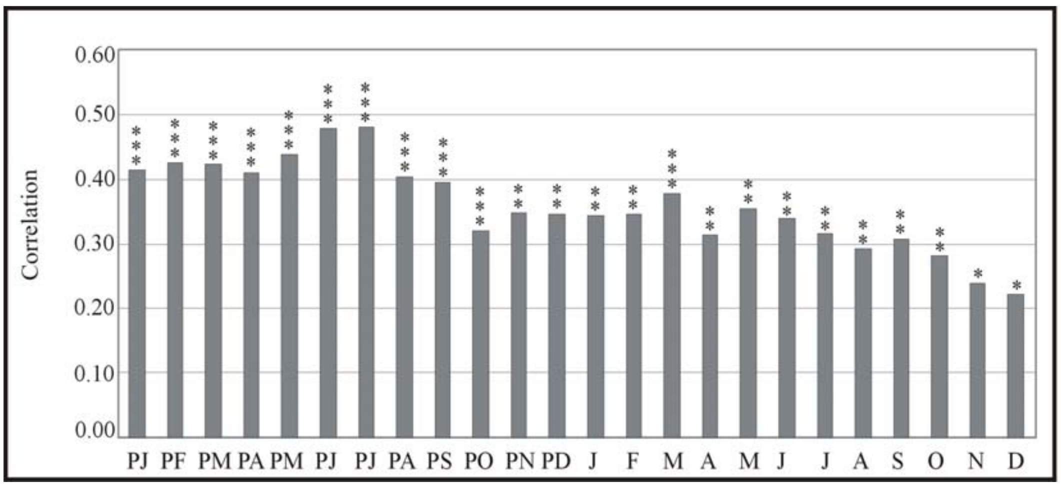
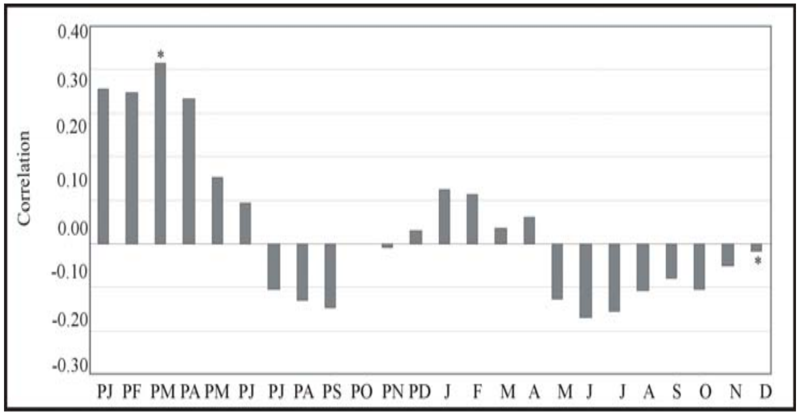
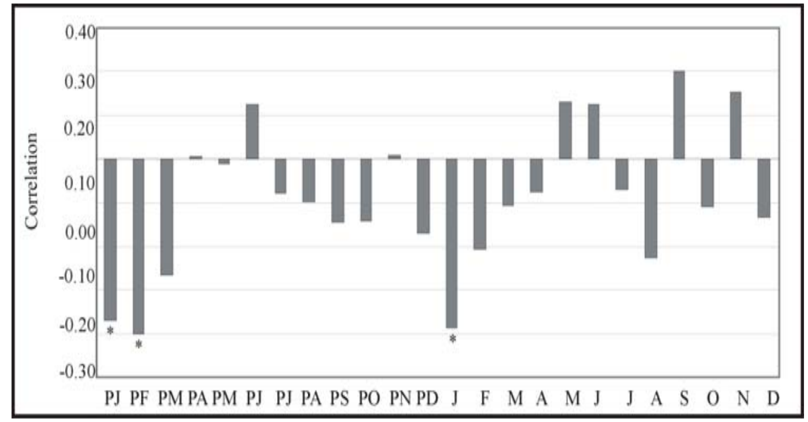


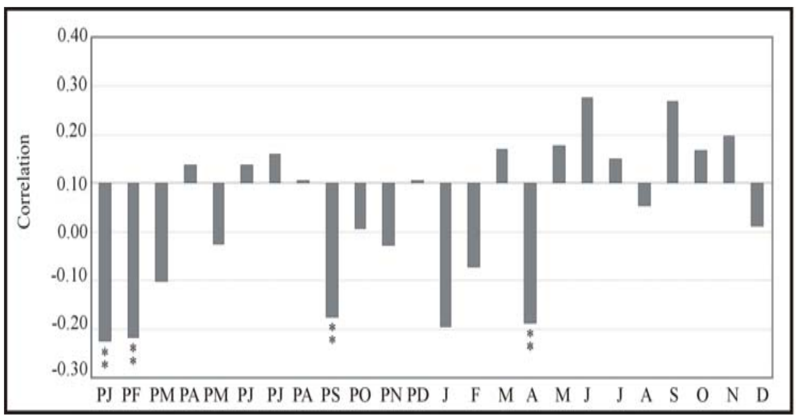
Figure 3.9. Correlation coefficients between western Montana composite chronology and PDSI ($* p < 0.05$, $** p < 0.01$, $*** p < 0.001$).



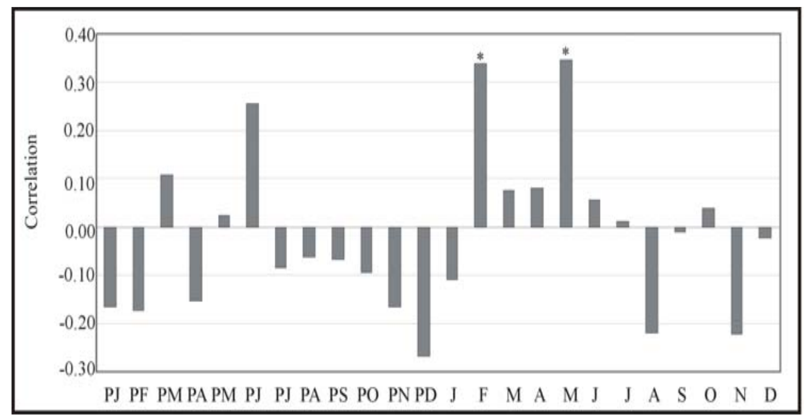
a. Pacific Decadal Oscillation



b. Arctic Oscillation



c. North Atlantic Oscillation



d. Northern Oscillation Index

Figure 3.10. Correlation coefficients between the western Montana composite chronology and PDO, AO, NAO, and NOI (* $p < 0.05$, ** $p < 0.01$).

Table 3.3. Results of the verification between actual and predicted previous year's average June-July PDSI.

Calibration Period	Verification Period	Correlation Coefficient	t-test Means
1930–1966	1967–2005	0.47*	1.99*

*$p < 0.05$ level

significant Pearson correlation coefficient ($p < 0.05$), and a relatively close fit between actual and estimated PDSI values were apparent for the model (Figure 3.11). I used the regression equation developed from the calibration period to reconstruct previous year's June and July average PDSI for the entire period of the composite chronology (Figure 3.12):

$$\textit{Reconstructed (PJ-PJ PDSI)}_t = -14.60\ (TR)_t + 14.63$$

where t is the year, *PJ-PJ PDSI* is previous year's June and July, and *TR* is the tree-ring index for year t.

3.4 Discussion

3.4.1 Climate-Tree Growth Response

3.4.1.1 Composite Variables

The degree of cambial cell division (i.e., ring growth) depends on the rate of precipitation and its distribution during specific seasons. A typical characteristic of whitebark pine is a direct cambial growth response to precipitation during the 61-day growing season (late June to early September) (Weaver 2001). The amount and distribution of precipitation between complete snowmelt (middle to late June) and fall freeze-up (early to mid-September) may be the most critical factor in whitebark pine survival and growth (Farnes 1990).

Decreased runoff from precipitation that falls during the snow-free period occurs, as most of the growing season's precipitation enters the shallow soil profile and is subsequently used by the vegetation or evaporates. Lack of significant moisture over long periods might easily stress whitebark pine because the water holding capacity of the soils

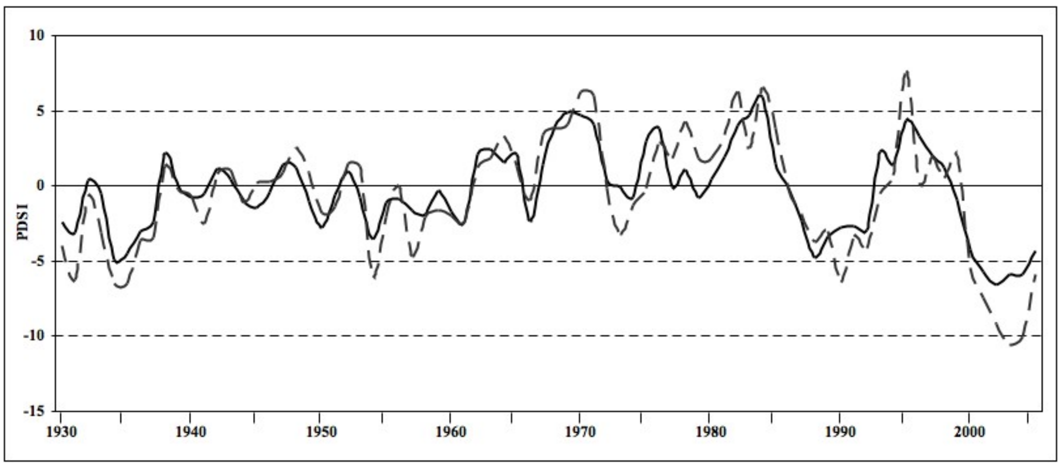


Figure 3.11. Reconstructed (dashed line) and actual (solid line) PDSI values AD 1930–2005.

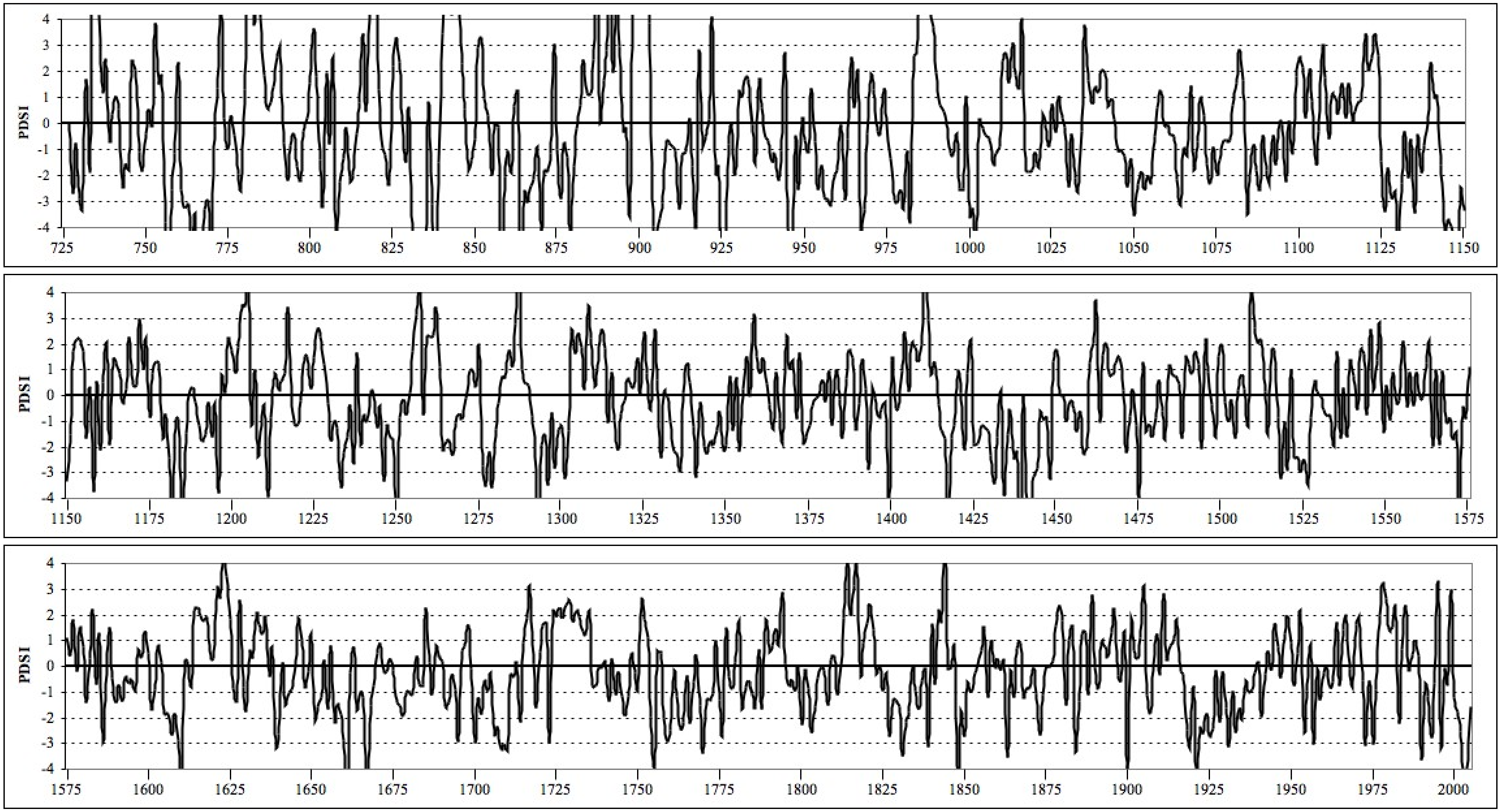


Figure 3.12. Reconstructed June-July average PDSI for western Montana for the period AD 730–2005.

is low. From June to September, cell division in whitebark pine is occurring in the cambium, and expansion of the growing xylem and phloem results in a net increase of these tissues, *i.e.* ring growth. The rate of change in size of tissues in the stem can be shown to be a function of both the difference between water uptake and water loss by the plant, and the ability of water to move, to be absorbed, and to be held by the tissues of the plant (Fritts 1976). Drought conditions decrease cambial growth significantly, and growing season shortages when evapotranspiration demands are high have a large impact on internal water relations (Cook *et al*. 1988).

The growth response to PDSI was much more pronounced than with precipitation and temperature. The fact that the site and composite chronologies illustrated limited sensitivity to both precipitation and temperature suggests that it may be possible to reconstruct climatic indices that incorporate both varables (*i.e.* PDSI). A large degree of autocorrelation is present in the PDSI values, so the high correlations with whitebark pine growth over consecutive months of the year can be attributed, in part, to this persistence (Grissino-Mayer and Butler 1993). Because the contribution of each successive month is given an increasing weight, up to the month for which the drought index is computed, the individual contribution of each month in the growing season may not be accurately reflected (Blasing *et al*. 1988).

However, the autocorrelation does not account for the more statistically robust association in general. Because the PDSI incorporates temperature, precipitation, and the available water content of soil, the composite of the variables more closely replicates the environmental conditions necessary for tree growth than do precipitation and temperature only. Warm summer temperatures concurrent with reduced moisture availability could induce a drought effect and reduced cambial activity, with negative effects on growth processes that span more than one growing season. Interactions can occur among prior growth and the concurrent climatic factors if drought is followed by favorable conditions in late summer. The growth processes that usually consume carbohydrate reserves are checked by the drought, and all photosynthates go into such reserves, which are then available for the next season's growth (Fritts 1976).

The most significant relationship between whitebark pine growth and the PDSI found in this study occurred in the previous year's June and July. The PDSI indices were strongly correlated to growth from June to August with July having the highest correlation. This indicates that drought (precipitation and temperature) conditions are much more critical later in the growing season than earlier. Perhaps moisture conditions late in the previous year (August-September) affect this year's bud break and the initiation of growth more than climate during the current year's growing season. The net effect of drought may be a reduction in current growth, but an increase in future growth (Fritts 1976).

3.4.1.2 Factors that Affect Correlation Results

The model of tree growth acquired from the composite chronology explains less than 40% of the variance in PDSI. Additional factors related to tree growth at high-elevation could possibly explain the collateral variance, and these include type of precipitation, temperature, phenotypic characteristics, tree age, and climatic factors not included in PDSI. The majority of precipitation at high-elevation sites in Montana falls as snow (October to June). Deep snowpack and an increase in the length of time a particular site is covered by snow may prolong the length of the photosynthetically inactive period and delay the beginning of growth (Fritts 1976).

At the WB and MPK sites, sampled early in the 2004 growing season (June), snowpack was still visible. At the GR site, sampled in the middle of the 2006 growing season (July), snowpack was also visible. In addition, precipitation falling as snow may be subject to redistribution due to wind, transporting it away from certain exposures and enhancing snowpack elsewhere (Fritts 1976). Variations in wind direction or intensity from one year to the next can obscure growth-precipitation correlations as they affect snow redistribution (Fritts 1976).

At high-elevation sites, where the potential growing season is short and plant temperatures can be low while the cambium is still active, growth may be reduced by cold temperatures during any time of the growing season (Mikola 1962). Trees on south-facing slopes may receive direct solar radiation during the daytime, causing their tissue temperatures to be higher than those trees on adjacent slopes. On south-facing exposures, temperature is most likely to become a limiting factor to growth during cloudy weather, at night, or when high winds dissipate the energy and reduces cambial temperatures (Fritts 1976). High winds can also reduce photosynthesis by 20–40% (Welkie and Caldwell 1970). Given whitebark pine’s short growing season, any extended period (> 24 hours) of cloud cover or high winds may reduce temperatures, resulting in lower than average growth at the micro-site level which could create noise in the climate signal.

The seasonal development of needles and cones, through the hormonal control of cambial activity, is vital to the distribution of seasonal growth (Kramer and Kozlowski 1979). Because whitebark pine retains its needles from 5–8 years, needles can affect cambial activity for several years (Weaver 2001). Whitebark pine reproduces by masting when they reach 60 to 80 years of age, and the amount of cones produced depends on internal factors (Day 1967, Krugman and Jenkinson 1974). Years of high cone productivity are usually preceded by a year of low productivity (Weaver and Forcella 1985). Intervals between large cone crops for whitebark pines range from three to five years (McCaughey and Tomback 2001). Whitebark pine cone production is also influenced by tree canopy size due to a larger number of fertile shoots (Spector 1999). In years with heavy cone production, more carbohydrates may be allocated to cone production than in years of low productivity, and the effects of reproduction could also interfere with the climate signal.

3.4.2 Climate Reconstruction

The high variability in the early portion of the reconstruction is characteristic of low sample depth in tree-ring chronologies. The higher level of variance in the early portion of the reconstruction does not necessarily reflect a climate period with high variability and exceptional events shifting to a more moderate climate regime over time. Accordingly, any conclusions derived from the data where the sample depth is low are made with the recognition of limitations of the information and the reliability of the reconstruction.

3.4.2.1 Analysis of Model Outliers and Stand Disturbance

In the period of the instrumental record used in the reconstruction (1930–2005), the outliers removed from the reconstruction model appear to be related to low moisture levels, and possibly low snowfall totals. I removed six outlier observations in the regression analysis that occurred in the years 1961, 1975, 1976, 1983, 1984, and 1996. Two of the outliers removed (1961 and 1996) occurred when previous-year (1960 and 1995) precipitation was higher than normal and tree growth during 1961 and 1996 was below normal. One possible explanation for this inverse relationship could be persistent snowpack at high elevations. Periods of persistent snowpack that extend from the previous year into the current year may lengthen the photosynthetically inactive period and delay the beginning of growth leading to an inverse relationship between precipitation and tree growth (Fritts 1976).

Stand or regional scale disturbances also may have caused anomalous growth patterns. Positive relationships between moisture balance and insect epidemics have been identified in a number of studies (Swetnam and Lynch 1993, Logan *et al*. 1998). In the composite chronology, some of the identified outliers appear to have been related to several pine beetle outbreaks in the western United States during the period 1930 to 2005.

Between 1974 and 1977, an extensive pine beetle outbreak occurred in the western United States and Canada (Safranyik and Linton 1998). During this period, the number of pine beetle infestations in the western United States and Canada increased from 250 to 500 (Carroll *et al*. 2004). The composite chronology shows that whitebark pines experienced below-average growth in 1975 and 1976 while precipitation was high in 1974 and 1975. From 1980 to 1984, a peak in pine beetle outbreaks occurred in the northern Rocky Mountains (Safranyik and Linton 1998). During this period, pine beetle infestations reached a total of 1400 infestations (Carroll *et al*. 2004). The composite chronology shows that 1983 and 1984 experienced below average growth while precipitation was high during 1982 and 1983.

This examination of outliers and possible disturbances associated with them points to one of the major weaknesses of dendroclimatic reconstructions based on tree species prone to disturbance processes. Prior to the 20th century, the effects of disturbances, especially those that occurred several hundred years in the past, are difficult to differentiate from the climate signal. As a result, values derived from a reconstruction can only be interpreted to represent climatic conditions. This issue may be resolved partially when spline curve-fitting techniques are employed to resolve noise linked with non-climatic disturbances; nevertheless, disturbances common to the entire site will remain to some degree in the final chronology (Cook 1985).

3.4.2.2 Trend Analysis

The reconstructed June-July PDSI (Figure 3.13), 10-year average (Figure 3.14) and 50-year average (Figure 3.15) revealed both interannual and decadal trends for the period 735–2005. Reconstructed PDSI values that extended above +1 or below –1 were considered to be representative of wet or dry periods, respectively. I used the 10-year average to determine the occurrence of interannual drought periods between 730 and 2000. To determine decadal-scale drought occurrence between 750 and 1980, I used the 50-year average. A small number of short-term periods of drought (4) and wetness (2) are embedded in long-term periods. These were not listed in order to avoid overlapping of these cycles between interannual and decadal time scales.

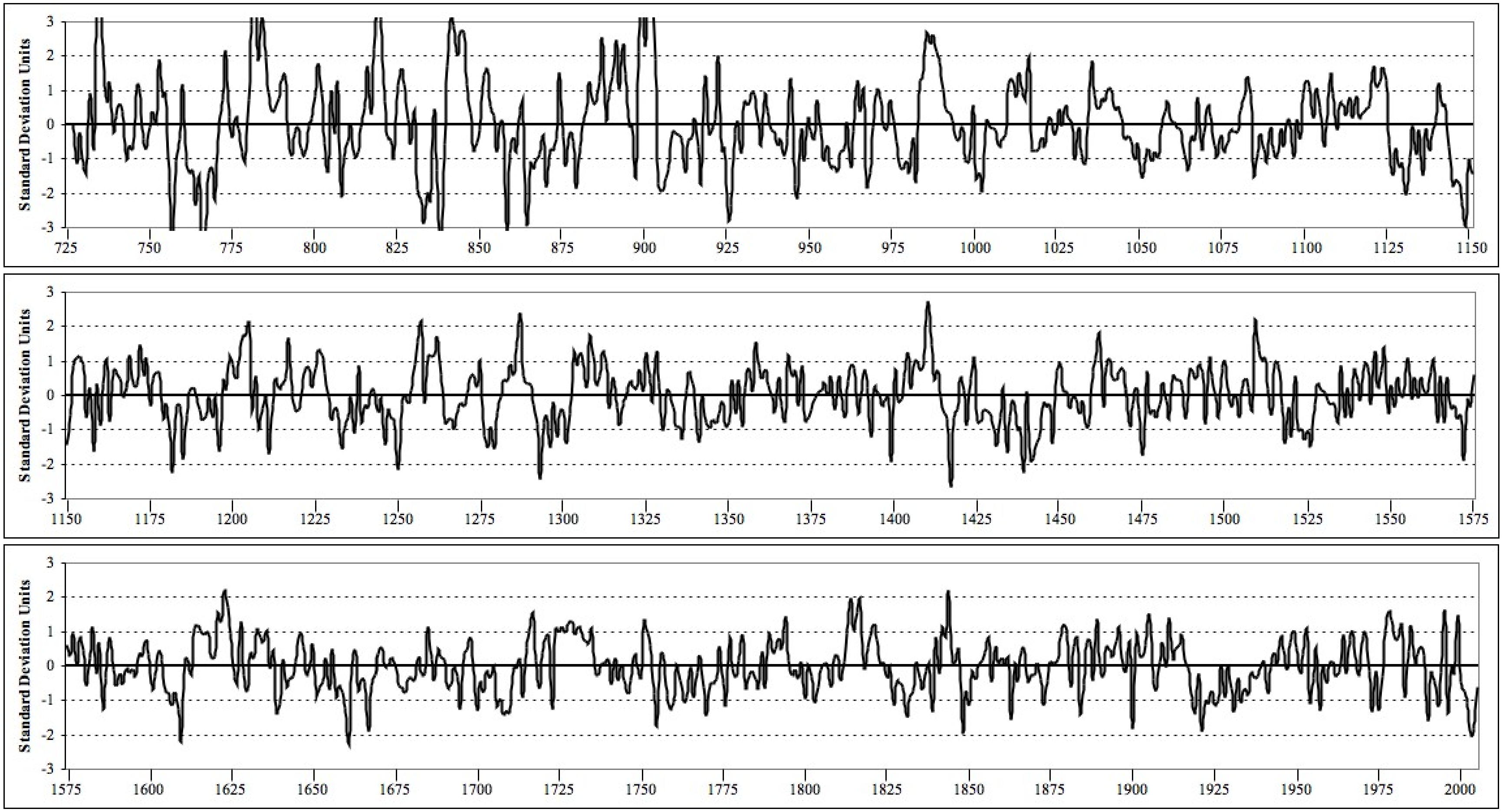


Figure 3.13. Reconstructed June-July average PDSI for western Montana AD 730–2005 converted to standard deviation units.

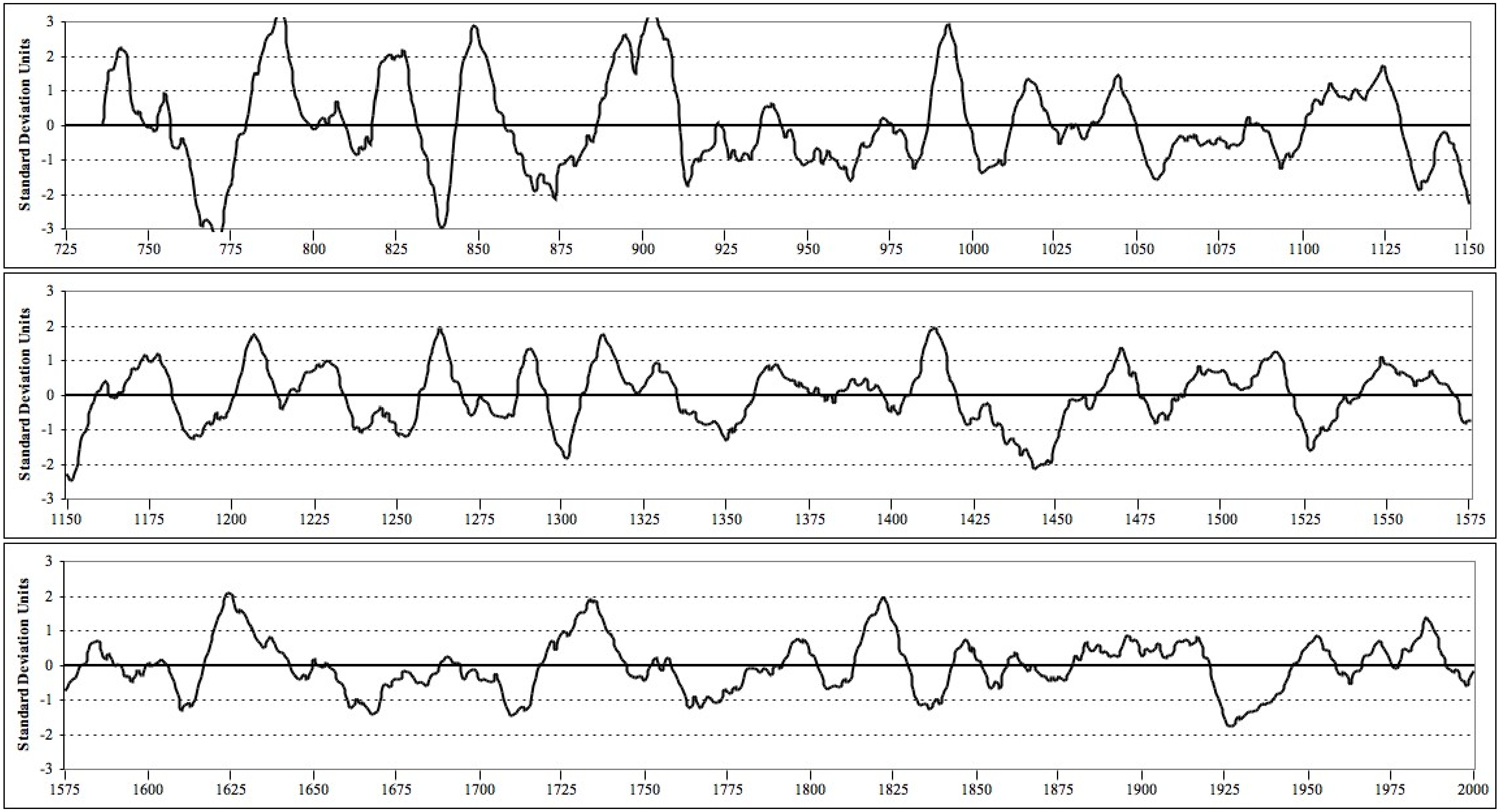


Figure 3.14. Reconstructed June-July average PDSI for western Montana for the period AD 730–2000 represented as a 10-year moving average and converted to standard deviation units.

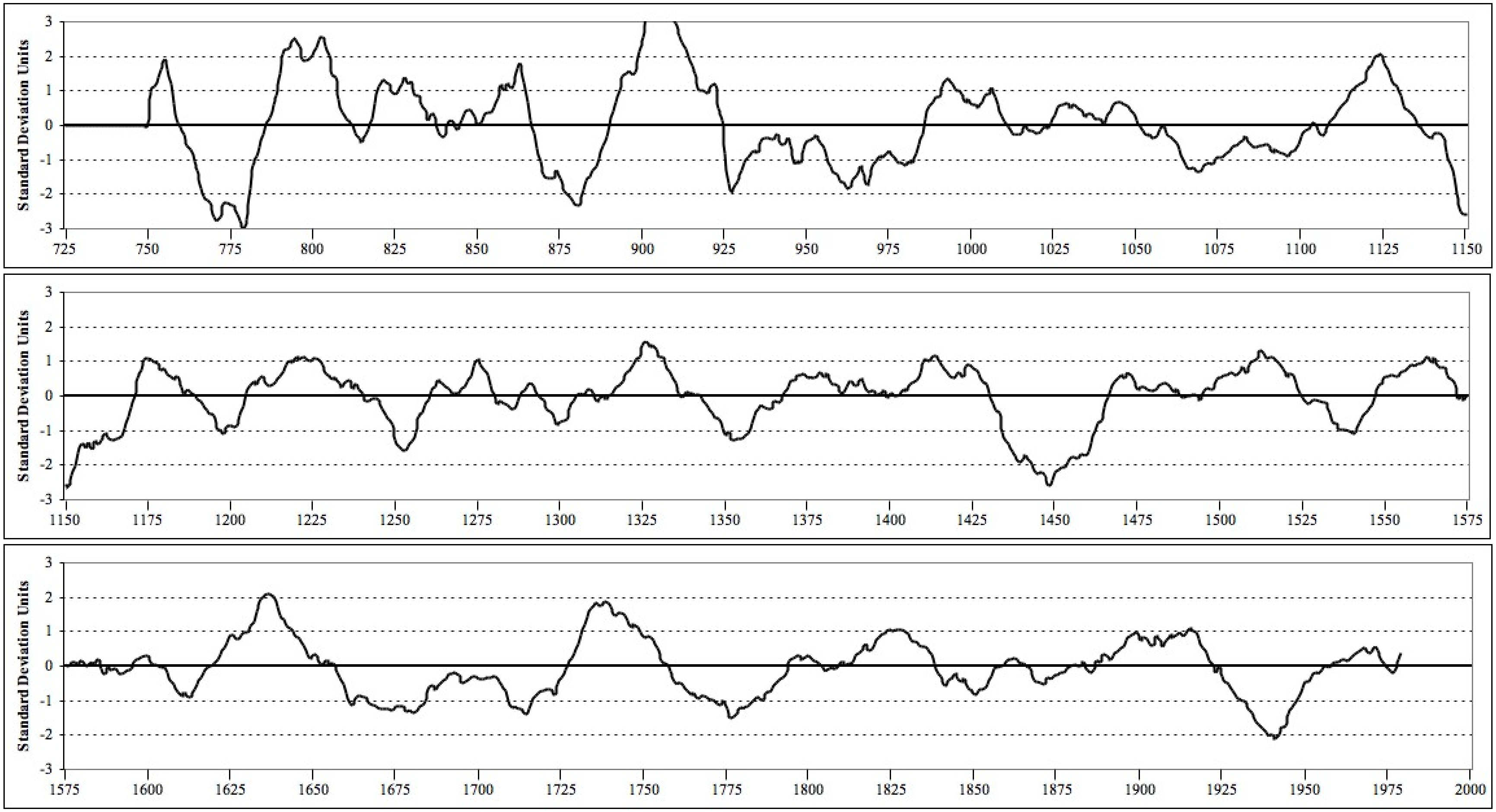


Figure 3.15. Reconstructed June-July average PDSI for western Montana for the period AD 750–1980 represented as a 50-year moving average and converted to standard deviation units.

The climate regime fluctuated between periods of above average dry and wet periods, separated by short transition periods. For example, an intense dry period occurred between 869 and 886 (Figure 3.15), but this was preceded by the most severe of the short-term period droughts between 835 and 841 (Table 3.4), which had an average PDSI value of –2.25 for this 7-year period. During the last 1100 years, two other severe short-term droughts occurred between 1133–1139 (average PDSI of –1.55) and 1297–1303 (average PDSI of –1.45). The wettest short-term period occurred between A.D. 988 and 996 (Table 3.5) which had an average PDSI of 2.18 over the 9-year period. The second wettest climate episode occurred between A.D. 845 and 854 with an average PDSI of 2.14.

The reconstruction also revealed major decadal trends in past climate. Since A.D. 750, eight periods of protracted extreme drought (Table 3.6) and five periods of extreme wetness (Table 3.7) have occurred. The most severe extended drought occurred between 1434 and 1462 which had an average PDSI of –2.21 for the 29-year period. Two additional periods of extended drought, 1145–1167 (average PDSI of –1.64) and 764–782 (average PDSI of –2.21), lasted 23 and 19 years, respectively. In terms of wetness, the most extreme period and longest in duration occurred between 894 and 918, lasting 25 years, which had an average PDSI of 2.49. Two additional periods of extended wetness, 790–807 (average PDSI of 2.07) and 1732–1748 (average PDSI of 1.52), lasted 18 and 17 years, respectively.

Comparisons between 20th century droughts and those reconstructed for the preceding 1100 years point to fundamental differences between dry events during the instrumental period and those in past centuries. Decadal scale dry events in the past

Table 3.4. The most severe inter-annual droughts in western Montana since AD 900 by magnitude.

Period (AD)	PDSI	Duration (yrs)	Rank
835–841	–2.25	7	1
1133–1139	–1.55	7	2
1297–1303	–1.45	7	3
1348–1350	–1.45	3	4
1610–1614	–1.44	5	5
912–915	–1.43	4	6
1053–1058	–1.37	6	7
1525–1529	–1.33	5	8
1708–1715	–1.30	8	9
1002–1009	–1.30	8	10
1763–1764	–1.19	2	11
1661–1662	–1.18	2	12
1186–1191	–1.16	6	13
1766–1768	–1.13	3	14
982–983	–1.13	2	15
1250–1254	–1.12	5	16
1832–1837	–1.11	6	17

Table 3.5. The most severe inter-annual wet periods in western Montana since AD 900 by magnitude.

Period (AD)	PDSI	Duration (yrs)	Rank
988–996	2.18	9	1
845–854	2.14	10	2
737–744	1.76	8	3
1409–1416	1.61	8	4
1817–1825	1.56	9	5
1260–1265	1.55	6	6
1204–1210	1.40	7	7
1310–1316	1.40	7	8
1016–1019	1.25	4	9
1468–1470	1.24	3	10
1042–1045	1.23	4	11
1289–1292	1.22	4	12
1985–1988	1.22	4	13
1177–1178	1.15	2	14
1108–1109	1.14	2	15
1512–1517	1.12	6	16
1173–1174	1.12	2	17

Table 3.6. The most severe decadal droughts in western Montana since AD 900 by magnitude.

Period (AD)	PDSI	Duration (yrs)	Rank
764–782	–2.21	19	1
1434–1462	–1.84	29	2
869–886	–1.68	18	3
1145–1167	–1.64	23	4
1932–1947	–1.61	16	5
957–971	–1.46	15	6
1772–1783	–1.21	12	7
1667–1684	–1.20	18	8

Table 3.7. The most severe decadal wet periods in western Montana since AD 900 by magnitude.

Period (AD)	PDSI	Duration (yrs)	Rank
894–918	2.49	25	1
790–807	2.07	18	2
1631–1644	1.57	14	3
1732–1748	1.52	17	4
1115–1129	1.47	15	5

tended to persist longer than those experienced in the 20th century. A single period of extreme drought in the 20th century (1932–1947) lasted 16 years, while drought events before 1900 lasted on average approximately 20 years in length. Moreover, two significant dry periods (1145–1167 and 1434–1462) before 1900 lasted for an average of 26 years.

The reconstruction also suggests that droughts on the scale of the 20th century drought occurred on average once every 208 years prior to the instrumental period. Before 1900, the severity and extent of decadal scale droughts in western Montana appears more extreme than those in the instrumental period. Also, the most extreme and longest drought period that occurred between 1434 and 1462 shows a notable relationship with the North American 15th century mega-drought between 1444 and 1481 (Stahle *et al*. 2007).

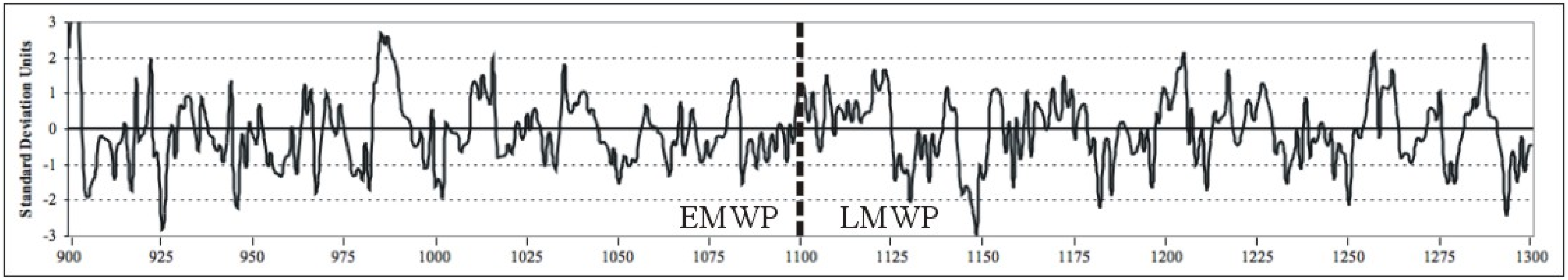
3.5 Climate During the Medieval Warm Period and the Little Ice Age

The conventional perception of climate development during the last millennium had been a simple sequence of a warm "Medieval Warm Period" (MWP) and a cool "Little Ice Age" (LIA) followed by global warming since the latter half of the 19th century and especially into the 20th century. Numerous researchers have presented convincing evidence regarding the existence of the MWP and LIA and how these episodes varied both temporally and spatially (Hughes and Diaz 1994). One region may have experienced persistent drought, while another may have experienced a significant wet period. The reconstruction of any climatic variable depends on the accuracy and sensitivity to shifts in climate by the proxy record used to infer these long-term episodes (Bradley and Jones 1993).

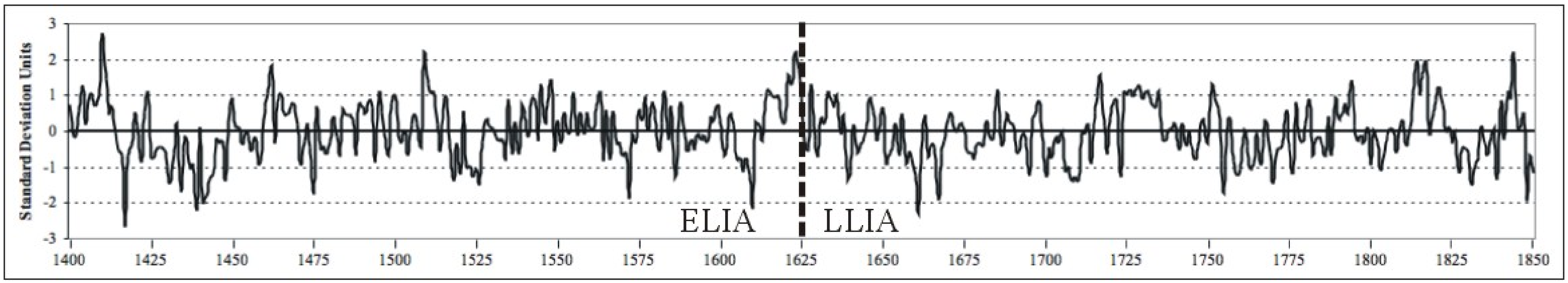
To analyze climate during the MWP and the LIA, I subdivided each period into four separate periods: (1) early MWP (900–1100), (2) late MWP (1101–1300), (3) early LIA (1400–1625), and (4) late LIA (1626–1850) (Figure 3.16). The MWP displayed higher variability than the LIA (Table 3.8). Of the four subdivided periods, the late MWP displayed the highest variability in the record.

Comparisons of climate during both the MWP and the LIA show that the climatic episodes during the MWP, a period marked by above average temperatures, had a higher level of severity and duration than during the LIA, a period of lower than average temperatures (Lamb 1977). However, researchers have acknowledged that both phenomena were not globally synchronous and possibly operated on different spatial scales (Hughes and Diaz 1994, Bradley *et al*. 2003, Jones and Mann 2004)

The reconstruction offers evidence that long-term climate episodes that coincide with the MWP and the LIA did not leave evidence in the moisture-availability record at high-elevation sites in western Montana. The number of wet and drought years during the MWP and LIA illustrate only marginal differences. During the MWP and LIA the number of drought years is relatively equal, 223 and 234, respectively. The number of wet years showed a negligible difference with 178 wet years occurring during the MWP and 217 during the LIA. The severity of drought during the MWP was more intense then during the LIA with the average duration of drought between the two periods relatively equal (Table 3.9). Wetness during the MWP was also more intense than that of the LIA



(a)



(b)

Figure 3.16. Illustration of persistent drought periods indicated by negative values and periods of wetness represented by positive values for the periods (a) Medieval Warm Period, and (b) Little Ice Age converted to standard deviation units.

Table 3.8. Periods of climatic variability during the Medieval Warm Period and the Little Ice Age.

Period	Coefficient of Variation
Medieval Warm Period	0.41
Early Medieval Warm Period	0.30
Late Medieval Warm Period	0.73
Little Ice Age	0.23
Early Little Ice Age	0.22
Late Little Ice Age	0.24

Table 3.9. The average PDSI value and duration of drought occurrence in western Montana during the MWP and LIA.

Period	PDSI	Duration (yrs)
MWP	–1.34	10
LIA	–1.26	14

with the average length of wetness between the two periods also comparatively equal (Table 3.10).

Reconstructed PDSI for the MWP and LIA shows that shifts between dryness and wetness occurred much more rapidly during the MWP when compared to the LIA. During the MWP, shifts between dry and wet periods averaged 13 years in duration, while during the LIA shifts between dry and wet periods averaged 28 years.

Persistent droughts during the MWP and LIA were common in the reconstruction. The instrumental period, however, is unusual because the strongest dry event during this interval is slightly shorter (16 years) compared to the duration of droughts during the MWP and LIA (20 years). Many persistent droughts prior to the instrumental period resulted in major precipitation deficits over decadal periods. During the MWP, two intense decadal droughts occurred from 957 to 971, and from 1145 to 1167. Three periods of extensive drought occurred during the LIA: 1434–1462, 1667–1684, and 1772–1783.

Numerous hypotheses have been proposed to explain climatic fluctuations during the MWP and LIA, including solar variability (Eddy 1977), volcanism (Grove 1988), and changes in global oceanic and atmospheric circulation (Lamb 1977, Broecker *et al*. 1985). While secondary factors, such as volcanism, could elicit such fluctuations, the most important internal mechanism is the teleconnection between atmosphere and ocean circulations (Gray *et al*. 2003).

Whitebark pine ecosystems in western Montana do not show conclusive evidence of climatic shifts that can be equated with the MWP and LIA. No prolonged

Table 3.10. The average PDSI value and duration of wetness in western Montana during the MWP and LIA

Period	PDSI	Duration (yrs)
MWP	1.35	7
LIA	1.22	9

environmental deviations that correspond to these climatic shifts can be identified within the reconstruction.

3.6 Discussion

Tree growth in western Montana more closely reflects changes in drought occurrences than any other climatic variable. The rate of cambial growth is clearly responsive to the intensity of drought and its seasonal distribution. PDSI from June–July was one of the primary factors that controlled whitebark pine growth in this study. However, other climatic factors act to exacerbate drought occurrence in the region such as snowpack. Snowpack variability is a central force that limits tree growth at high-elevation sites (Peterson 1998). In the northern Rocky Mountains, changes in the timing and amount of snowpack could reflect climatic shifts (Stewart *et al*. 2004). In the western United States, snowpack amounts have declined since the 1950s, and snowmelt occurs 20 days earlier than in the late 1940s (Stewart *et al*. 2004).

The PDSI reconstruction presented could possibly be a record of changing snowpack levels at high elevation. Due to the limited soil profile at these sites, available moisture may be stored in the snowpack until summer melt occurs. Any decrease in snowpack level due to climatic change would result in a reduction of moisture available for tree growth. Interactions between snowpack and summer drought over decadal to multidecadal timescales would also have a strong impact on disturbances (pine beetle outbreaks and wildfire occurrences) in western Montana. For example, fires that occurred in Glacier National Park in 1910 resulted from extreme summer drought conditions and below average snowpack during the year of the fire event (Peterson 1998). While I did not attempt to reconstruct snowpack levels, the tree-ring chronology that has been developed presents a viable resource for such an investigation.

The influence of climate change on the forest systems of western North America is a pervasive topic in modern forest management, and is particularly relevant to the management of whitebark pine (Larson 2005). Applications of historical data are critical for survival of whitebark pine. Interpretation of the influence of climate change on high-elevation vegetation, as revealed in the PDSI reconstruction, can improve our application of historical variability to current contexts. Climate change is likely to impact whitebark pine significantly over the next century. Long life span and late maturity of whitebark pine limits its ability to adjust rapidly to change. Current projections of a doubled CO_2 level could reduce whitebark pine to less than 10% of its current range, where it is a major component of alpine and subalpine ecosystems (Mattson and Reinhart 1994). A warmer climate will favor less-hardy species that have been restricted to lower elevations by temperature, but will likely find more opportunities to compete with whitebark pine at higher, more rugged locations. Whitebark pine will be less successful at regeneration, and future stands are likely to be more mixed.

The results of this study substantiate the importance of drought for high-elevation ecosystems. The evidence presented here suggests that climate change may increase the threat to whitebark pine survival and is hypothesized to affect whitebark pine communities through three mechanisms: (1) causing a shift in pathogens, which may lead to new regions of hospitable climate for white pine blister rust and increase the potential for pine beetle infestation; (2) increasing drought, which may lead to decreases in range availability for whitebark pine, due to exclusion by more drought-tolerant species; and

(3) changes in the frequency of severe fires, which may lead to overall decreases in whitebark pine numbers.

CHAPTER FOUR

SENSITIVITY AND RESPONSE OF TREELINE IN WESTERN MONTANA TO ENVIRONMENTAL CHANGE AT THE LANDSCAPE LEVEL

4.1 Introduction

High mountain forests often display a characteristic discontinuity in their distribution (*i.e.* treeline). This boundary represents the upper limit of forest canopies and is associated with temperature decrease along elevational gradients (Lloyd and Graumlich 1997). The patterns of vegetation at treeline may indicate the process that produced them, and may help in developing a prediction of future ecotone dynamics (Malanson 1997). The possible loss of high-elevation tree species under a global warming regime is an important concern (Luckman and Kavanagh 1998). Considerable variation in global climate has occurred on varied time scales during the last 3000 years (Briffa *et al*. 1990, Hughes and Diaz 1994, Hughes and Graumlich 1996), but biological responses to climate change at sub-millennial frequencies are not yet well understood (Lloyd and Graumlich 1997). The rate of vegetation response to these long-term climatic fluctuations is pertinent to understanding the possible ecological consequences of anthropogenically induced climate change.

In alpine locations, widespread warming trends have been associated with forest expansion (Payette and Delwaide 1994, Klasner and Fagre 2002). Biogeographic models that predict ecological consequences of future global change often assume that species will make directional shifts (Sykes *et al*. 1996). Upper-elevation forest ecosystems offer unique opportunities to assess decadal-scale climate effects directly. Seasonal precipitation, temperature, wind, and snow regulate species composition, density, distribution, tree growth, crown architecture, age-class structure, and reproductive capacity (Tranquillini 1979, Korner 1998). In western Montana, trees in subalpine communities are known to be sensitive indicators of climate change and have persisted for many centuries (Kipfmueller 2003).

In the western United States, the Little Ice Age marked the most extensive cold period since the late Pleistocene (Clark and Gillespie 1996). Paleoecological changes, such as lowered treeline (Lloyd and Graumlich 1997), decreased forest productivity (Graumlich 1993), and changes in species distributions and fire regimes (Swetnam 1993), mark a transition from the previous warm and dry medieval centuries (Graumlich 1993) into the generally cold and dry Little Ice Age. Subsequent warming of the late 19th and early 20th centuries was accompanied by changes in precipitation as well as temperature which may have similarly resulted in ecological boundary changes (Graumlich 1993).

This chapter seeks to determine recent dynamics of lower, middle, and upper elevation species in western Montana. I examine the establishment rates of the two dominate species at treeline: whitebark pine and subalpine fir. I investigate the spatial patterns of high-elevation trees over time. Also, I discuss the results in terms of alterations in climate on recent treeline dynamics.

4.2 Site Selection

I selected sites located in the western Beaverhead-Deerlodge (BDNF), Gravely Range, and Lolo National Forests (LNF), Montana (Figure 4.1). The sites are south-

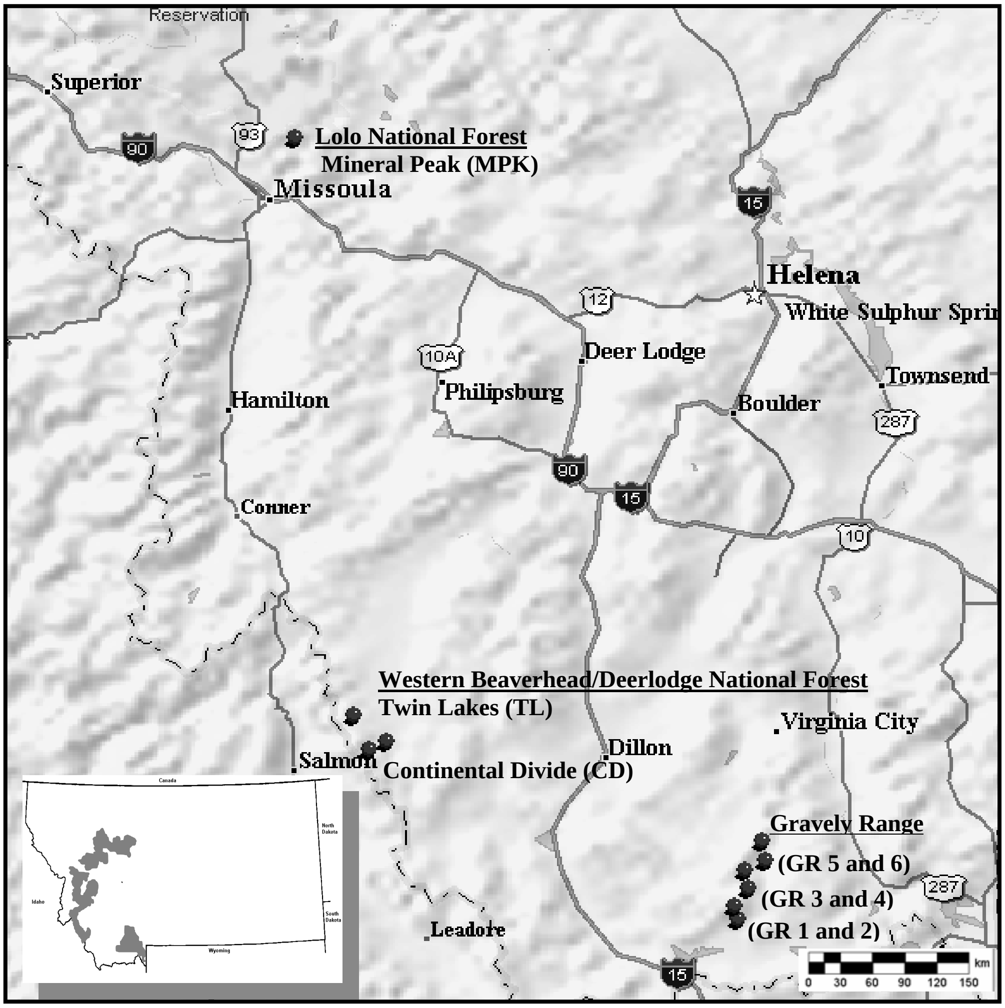


Figure 4.1. Map of study sites showing lower-elevation (Mineral peak/Twin Lakes), mid-elevation (Continental and Gravely Range), and upper-elevation (Gravely Range) (map created using Delorme mapping software).

facing, encompassing the subalpine forest and alpine-tundra transition zone between 2100 and 2850 m. Treeline patterns vary from tree islands at upper-elevation sites to treeline fingers extending from the lower-elevation continuous forest, to individual seedlings, saplings, and small trees in areas between treeline fingers and near tree islands. These areas are dominated by seral whitebark pine with a sub-canopy of subalpine fir.

In the LNF, Mineral Peak represents the lowest elevation site in the study. Tree islands dominate the site with areas of talus exposed between islands. The mid- and upper-elevation sites are located in the western BDNF and Gravely Range. These sites are composed of tree islands and treeline fingers made up of whitebark pine and subalpine fir. Alpine turf, mountain grassland, and shrub lands dominate the areas between islands and fingers. Soils are thin with exposed rock and gravel between vegetated areas.

4.3 Methods

4.3.1 Field Methods

To analyze the effects of climate on treeline dynamics in western Montana, I established 40 plots (10 m x 10 m) from lower to upper elevations in the LNF, BDNF, and Gravely Range, and further divided these plots into 600 subplots (1 m x 1 m) (Figure 4.2). The plots were subjectively located to represent the “average” condition within the stands. Plots were further grouped into low, middle, and upper elevations, resulting in a diverse altitudinal range for different plots (Table 4.1). Lower elevation plots were located in the LNF (Mineral Peak) and the Western BDNF (Twin Lakes). Middle elevation plots were located in the western BDNF (Continental Divide) and the

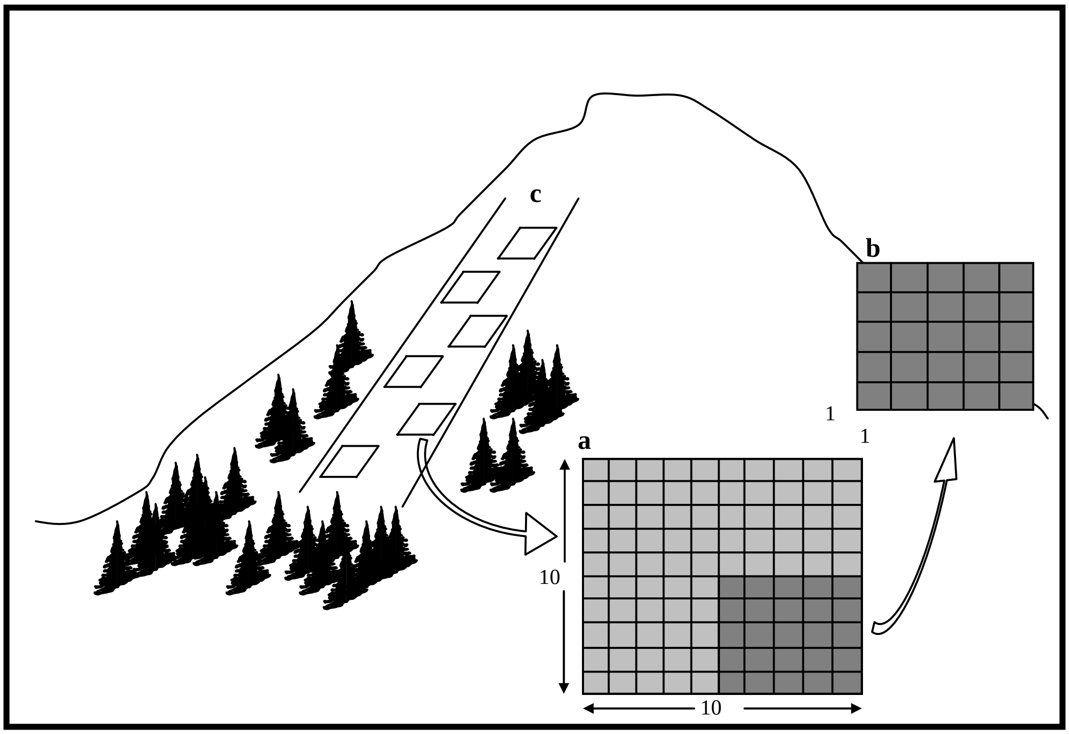


Figure 4.2. Illustration of study plot position, size, and orientation. 10 m x 10 m plots (a) were established and all stems > 5cm dbh documented. The plots were further subdivided into 1m x 1m sub-plots (b). Plots at the Continental Divide site (*n* = 2) and the Gravely Range sites (*n* = 6) were shifted sideways as plots were established upslope (c) (adapted from Hofgaard 1997).

Table 4.1. Characteristics of sites where treeline dynamics were studied in Western Montana.

Site	Site Location	Elevation (m)	Lat./Long.	Aspect	Number Samples	Establishment Date Ranges
Lower plot	BDNF/LNF	2100–2250	N47.00/W113.48	South	72	1825–1976
Middle plot	BDNF/ Gravely Range	2400–2550	N45.21/W113.42	South	114	1724–1977
Upper plot	BDNF/ Gravely Range	2700–2850	N45.09/W111.51	South	43	1887–1990

Gravely Range (1 and 2). Upper elevation plots were located in the Gravely Range (3 and 4, 5 and 6).

To calculate spatial patterns within each plot, all stems, saplings, and seedlings were mapped, identified by species, and assigned a grid location (latitude and longitude). To calculate establishment dates (pith dates), 229 increment cores and cross sections were collected from whitebark pine and subalpine fir trees growing below and above the present treeline. I collected two radii from all stems within each plot using an increment borer at breast height. Cores were then placed in paper straws for protection, and all relevant information on each tree sampled was recorded on standard forms. I also inventoried all seedlings (individuals less than 40 cm in height, but less than 3 cm dbh) and saplings (individuals greater than 40 cm in height, but less than 3 cm dbh) in each subplot. To analyze the affects of climate on establishment rates, I used the PDSI reconstruction developed in chapter 3.

Samples from whitebark pine and subalpine fir seedlings were not collected due to the threatened status of whitebark pine. As an alternative, I counted all whitebark pine and subalpine fir seedlings within each plot.

4.3.2 Laboratory Methods

In the laboratory, cores were air-dried, glued to wooden core mounts, and allowed to dry overnight. Cores and cross sections were sanded using a progression from ANSI 40-grit (500–595 µm) to 400-grit (20.6–23.6 µm) to ensure the cellular features of the individual tree rings were visible under standard 7–10X magnification (Orvis and Grissino-Mayer 2002). This process allowed the cell wall structure and the earlywood and latewood boundaries of each annual ring to be examined under 10X magnification. To examine tight clusters of rings, rotten wood, or poorly sanded areas, I used a razor blade to surface these areas.

4.3.3 Crossdating and Master Chronology Development

Total ring widths were measured using a Velmex movable stage micrometer and recorded with Measure J2X software to the nearest 0.001 mm. Measurements were then processed through the computer program COFECHA to check the visual crossdating with correlation analysis using overlapping 50-year segments of all measured cores lagged 25 years due to the length of the tree-ring series (Holmes 1983). If COFECHA indicated rings that were incorrectly dated, I visually examined the sample and corrected the date as necessary. In the case of missing or false rings, I used parts 5 and 6 of the COFECHA output (Grissino-Mayer 2001). Flagged segments were also checked to ensure no measurement errors occurred. To remove low-frequency variance, each measured series was then filtered with a cubic spline (Cook and Peters 1981). To enhance the effect of variability among smaller rings, a log transfer was also performed.

A master dating series was then developed using COFECHA by taking the mean of all filtered and transformed series by year after indices were developed for each year. The resulting filtered series was then tested against the master dating series, and correlations between each series compared between segments of time series.

Also, most individuals put on smaller rings as age increases. A simple average of tree growth would be skewed by the long-term ring width decline due to increasing stem circumference instead of reflecting the desired signal such as climatic variations. The standardization process is used to remove these growth factors and to clarify the desired climatic signal. The chronologies produced in the standardization process are an index of ring-width values with a mean of one and relatively homogeneous variance through time (Fritts 2001). To standardize the tree-ring measurement series, I entered the measurement data into the program ARSTAN (Cook 1985). A master chronology was then developed for the high-elevation sites in the BDNF. All dated cores were entered into Excel for graphical and statistical analysis. Histograms were used to visually analyze species density, composition, age structure, and size classes by the elevation of each study plot.

4.3.4 Spatial Analysis

To assess the univariate spatial pattern of trees within plots, I chose a nearest neighbor analysis for its relatively straightforward interpretation and its ability to reveal the scale of spatial pattern (Getis and Franklin 1987). To analyze the pattern of dispersion of whitebark pine and subalpine fir individuals, I used the x-y coordinates of the 242 mapped individuals to calculate the distance to the nearest neighbor for each tree using the program Biotas 1.03 (Ecological Software Solutions, Urnasch, Switzerland).

Nearest neighbor analysis is a test for complete spatial randomness. The aggregation index by Clark and Evans (1954) describes the ratio between the observed average distance (*XY*) of a tree to its nearest neighbor and the expected average distance for random tree distribution. Aggregation values < 1 indicate a tendency towards clustering, values of 1 indicate random distribution, and those > 1 show a tendency towards regular distribution. Distributional values are derived by calculating the distances

Y and *X* to their nearest neighbors for each of *N* trees, using these distances to obtain the average distance to the nearest neighbor. The actual, observed distance to the nearest neighbor is related to the expected average distance for random tree distribution. The aggregation index thus measures the extent the distribution pattern diverges from a Poisson distribution.

Adjacent ecotones can influence spatial patterns within the specific ecotone being studied, known as the edge effect. Because such an area contains habitats common to both communities as well as others unique to the transition zone itself, the edge is typically characterized by greater species diversity and population density than occur in either of the individual communities. Edge effects were eliminated from the study plots using an edge correction formula for unit areas in Biotas 1.03.

4.4 Results

4.4.1 Establishment Rates

Whitebark pine establishment at the lower-elevation plot began in the 1820s (Figure 4.3A). Between 1840 and 1910, whitebark pine establishment maintained moderate levels, ranging from 5% to 9% of all present trees. Whitebark pine establishment peaked in the 1920s (13%) and 1930s (13%). Establishment began to decline in the 1940s and ceased by the 1980s. Between 1840 and 1890, subalpine fir maintained an establishment rate of 5%. Subalpine fir establishment increased during the 1900s (17%) and peaked in the 1910s at 23%. Rates of establishment began to decline in the 1920s and terminated after the 1960s.

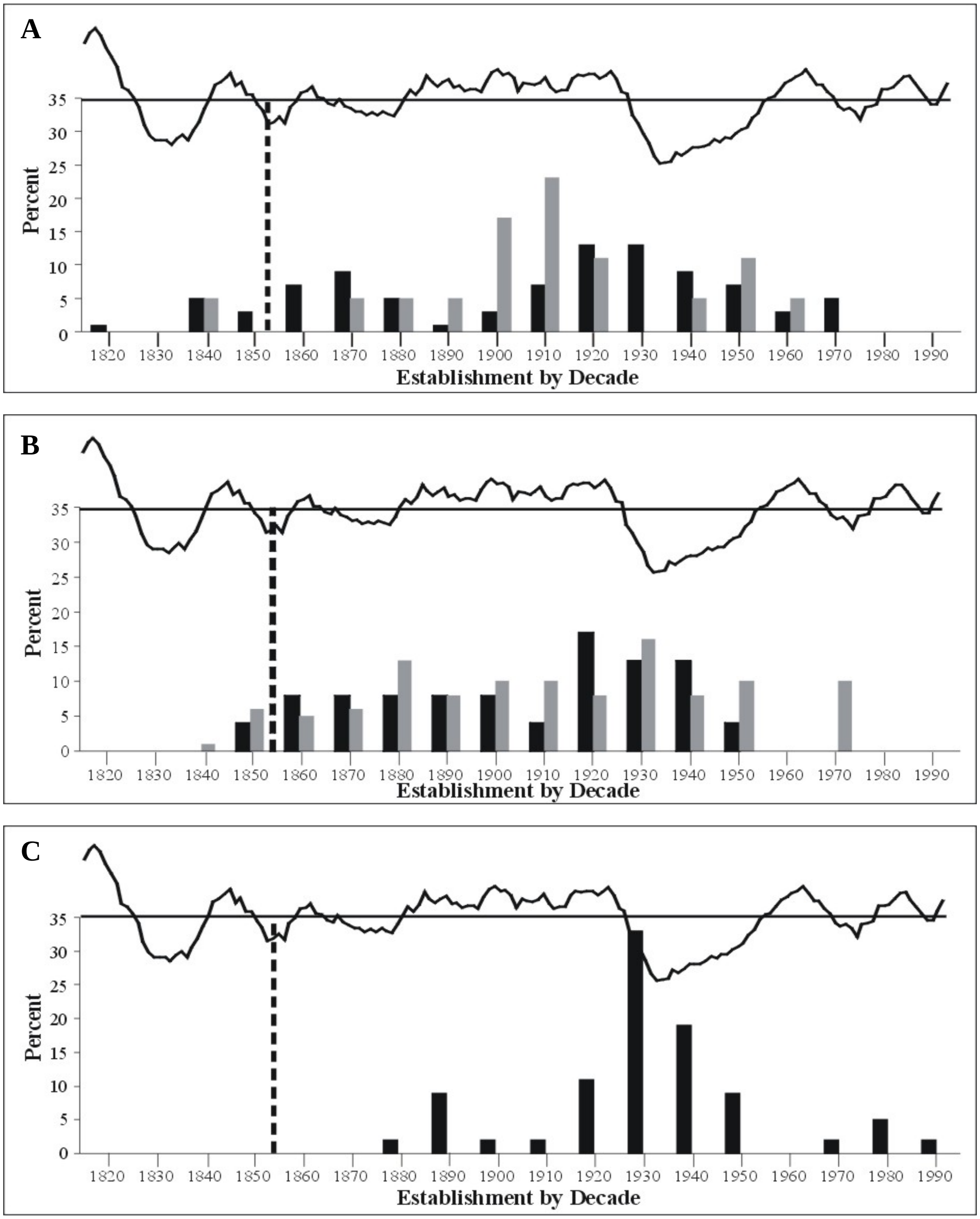


Figure 4.3. Establishment of whitebark pine (black bars) and subalpine fir (light gray bars) represented in percentages by decade at the lower (A), middle (B), and upper (C) elevation plots in relation to reconstructed previous year's June–July PDSI (black line) for western Montana.

At the mid-elevation plot, establishment of whitebark pine began in the 1850s (Figure 4.3B). Between the 1860s and 1910s, establishment percentages remained constant, ranging from 3% to 8%, and peaked between the 1920s and 1940s (17%, 13%, and 13%, respectively). Establishment rates declined rapidly in the 1950s (4%) and ceased after this decade. Subalpine fir showed two distinctive peaks of establishment. Establishment began in the 1840s and increased progressively, peaking in the 1880s at 13%. Establishment maintained constant levels between 1890 and 1920, ranging from 8% to 10%. A second peak in establishment began in the 1930s (16%) with rates remaining constant between the 1940s and 1970s (8%, 10%, 0%, and 10%, respectively). Establishment of subalpine fir ceased by the 1980s.

At the upper-elevation plot, whitebark pine establishment began in the 1880s (Figure 4.3C). Between the 1890s and 1920s, establishment rates ranged from 2% to 11%. Between the 1930s and 1940s, whitebark pine establishment reached its highest levels (33% and 19%, respectively). Beginning in the 1950s, establishment rates declined to 9% and maintain a moderate level between the 1970s and 1990s (2%, 5%, and 2%, respectively). At the upper-elevation site, subalpine fir seedlings, saplings, and stems were not present.

4.4.2 Spatial Analysis

The stem diagrams created by Biotas revealed differences in movement of whitebark pine and subalpine fir at the lower-, middle-, and upper-elevation sites. At the lower-elevation plot, the examination of stems by decade for the 160 year record suggested that two periods of upslope movement occurred. During the 1860s, whitebark pine established 50 meters above the previous decade's upper limits (Figure 4.4). In the 1920s, subalpine fir established 25 meters above the previous decade's upper limit.

At the middle-elevation plot, the stem diagram showed no movement over the 140 year record (Figure 4.5). Stability was maintained by comparative levels of establishment for both species within the plot. Whitebark pine and subalpine fir had an average establishment rate of 8% by decade, illustrating the stability within the plot. The greatest degree of movement, however, was found at the upper-elevation plot. During the 1980s, whitebark pine established 150 meters above the previous limit (Figure 4.6), but this movement is represented by just two stems.

4.5 Discussion

Examination of establishment rates in the whitebark pine/subalpine fir ecosystem presented here excludes the effects of wildfire. The interaction of climate and wildfire on the establishment of these species and treeline movement is paramount. As a result, the interpretations presented here of climatic effects on establishment are preliminary.

Patterns in establishment rates and spatial analysis among plots were evident. At lower elevations, establishment began in the 1820s (Figure 4.3A). Establishment at the middle-elevation plot began 22 years later (1840s), while establishment began 70 years later (1880s) at the upper-elevation plot (Figure 4.3 B and C). Establishment rate increases were seen beginning in the 1860s at the lower- and middle-elevation plots. These increased rates of establishment may be due to the termination of the unfavorable climatic conditions during the Little Ice Age (LIA) (1400–1850) at lower and middle

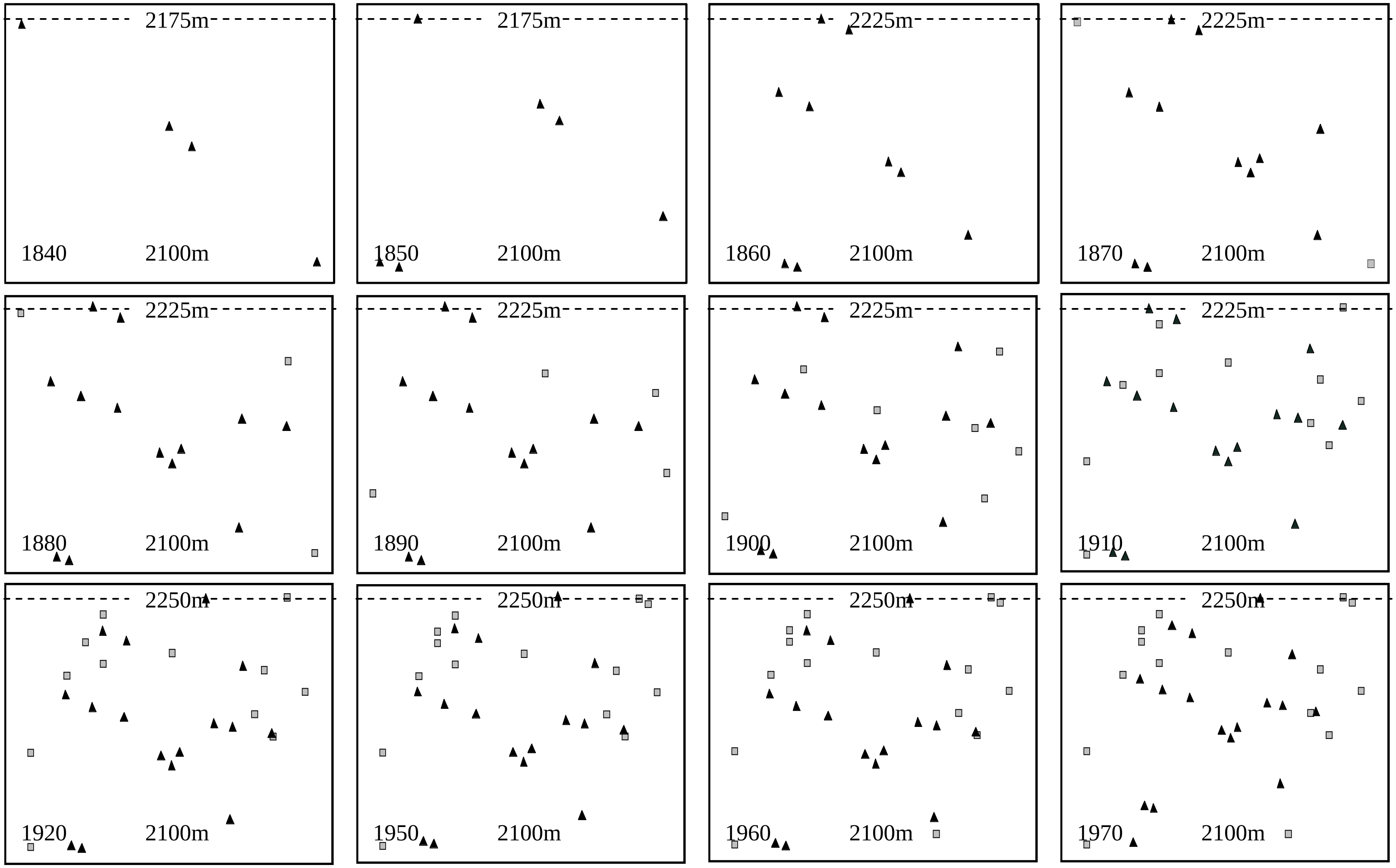


Figure 4.4. Identification of spatial patterns by decade of whitebark pine (black triangle) and subalpine fir (gray square) using Biotas 1.03 at the lower-elevation plot. Dashed line represents the upper extent of tree establishment.

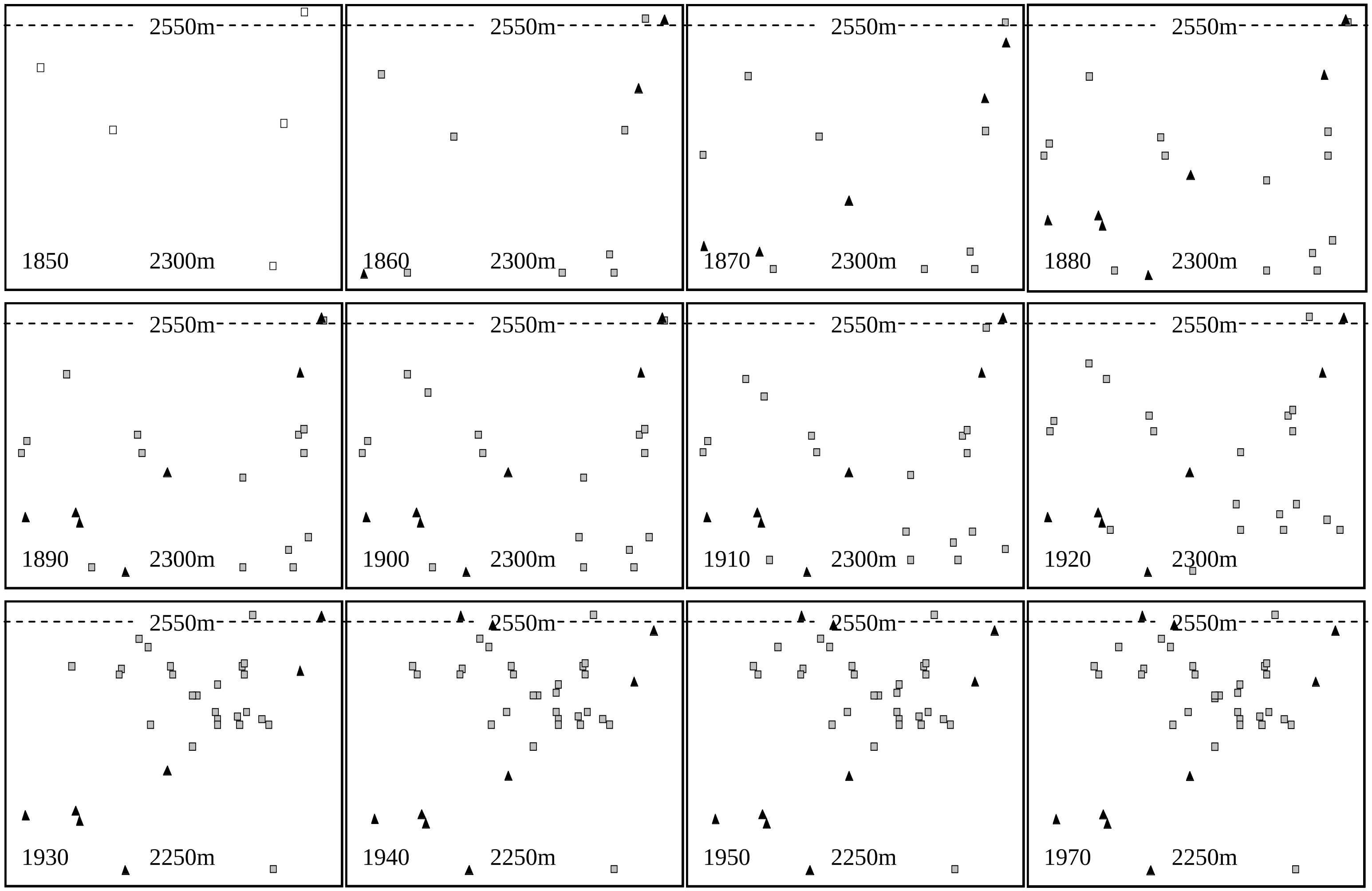


Figure 4.5. Identification of spatial patterns by decade of whitebark pine (black triangle) and subalpine fir (gray square) using Biotas 1.03 at the middle-elevation plot. Dashed line represents the upper extent of tree establishment.

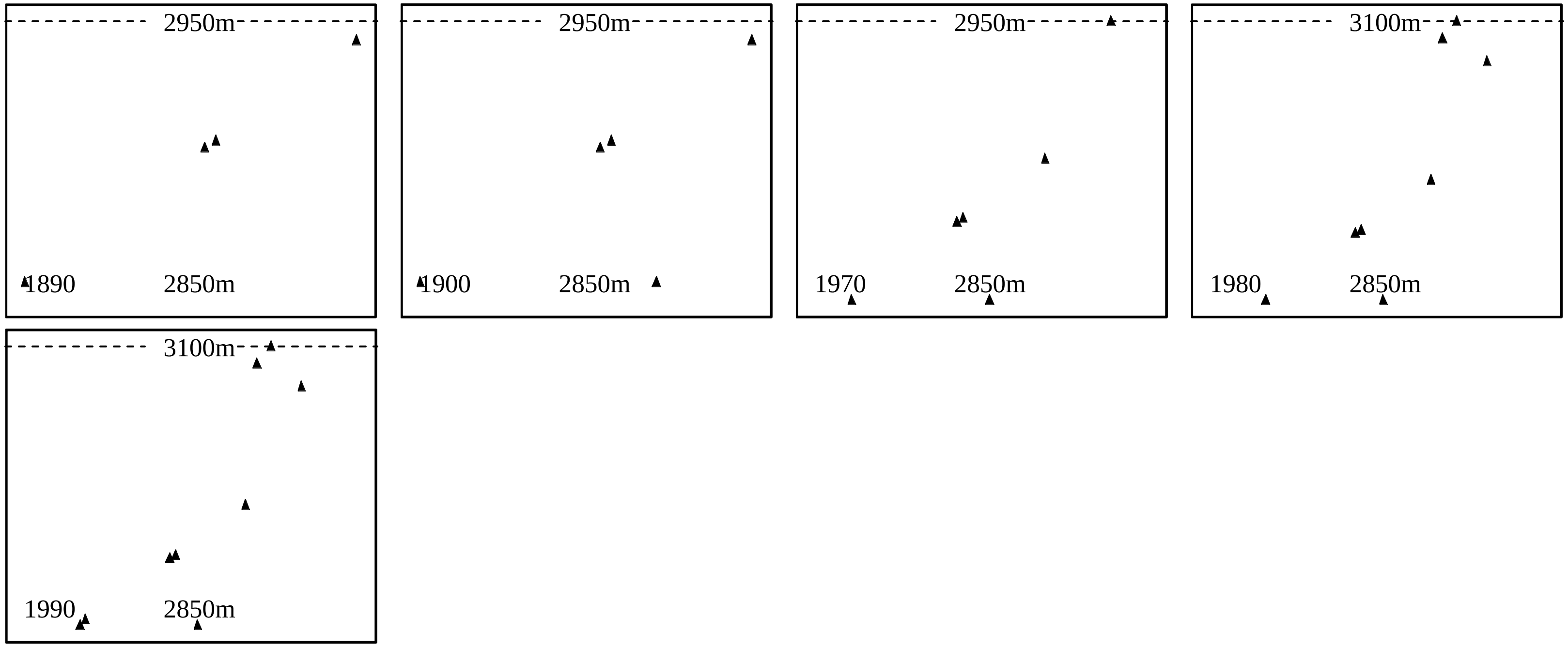


Figure 4.6. Identification of spatial patterns by decade of whitebark pine (black triangle) using Biotas 1.03 at the upper-elevation plot. Dashed line represents the upper extent of tree establishment.

elevations. In addition, the greatest movement in the lower-elevation plot (1860) coincided with the end of the LIA, which would indicate the development of more favorable conditions for establishment upslope. These climatic conditions were maintained and resulted in a stable treeline at the lower- and middle-elevations during the 1860s and 1990s.

The highest percentage of subalpine fir establishment at the lower-elevation plot occurred during the 1900s and 1910s. Subalpine fir prefers to establish during wet conditions and reconstructed previous year June–July PDSI for western Montana (developed in Chapter Three) showed wet conditions occurred during this time. In contrast, establishment of whitebark pine at the lower-elevation plot reached its highest levels during the 1920s and 1930s. Whitebark pine prefers to establish on warm, dry slopes and the reconstructed PDSI shows this period of increased establishment was the driest period during the 20th century.

Movement seen at the lower elevation plot in the 1860s indicates the development of favorable climate conditions for upslope establishment at the end of the LIA. While movement was not indicated at the middle-elevation plot, increased establishment in the 1860s verifies that climatic conditions had been more favorable for whitebark pine and subalpine fir. The movement seen in the lower elevation plot in the 1920s does not appear to be related to favorable climate conditions but rather could be a response to disturbance, possibly mountain pine beetle.

At the middle-elevation plot, subalpine fir establishment reached its peak in the 1930s. While the PDSI reconstruction shows an extensive drought in this decade, this peak in subalpine fir establishment could possibly be related to a decrease in whitebark pine overstory due to extensive mountain pine beetle outbreaks during the beginning of the 20th century (Carroll *et al*. 2004). Extensive whitebark pine establishment occurred during the 1920s, 1930s, and 1940s with the initiation of an extensive drought period in western Montana.

The upslope increase of stems during the 1860s and 1920s suggests that site conditions favored establishment at upper limits. During the 1860s, the plot was characterized by open, park-like conditions with individual whitebark pines randomly distributed, favoring the establishment. During the 1920s, major mountain pine beetle outbreaks occurred in the northern Rocky Mountains (Carroll *et al*. 2004). The upslope movement of subalpine fir during this decade could indicate the effects of whitebark pine mortality on the plot and an increase in favorable conditions for the expansion of subalpine fir. An opening of the overstory canopy due to mountain pine beetle would increase favorable conditions for subalpine fir to establish within the plot.

Establishment of whitebark pine at the upper-elevation plot reached its highest percentage in the 1930s and continued into the 1940s and 1950s. The PDSI reconstruction shows a significant drought during this period that would favor whitebark pine establishment. High percentages of whitebark pine establishment are also evident during this period at the lower- and middle-elevation plots. The upslope movement at the upper-elevation plot during the 1980s occurred after a moderate period of dryness and could be related to decreased levels of extended snowpack which would favor caching of seeds by the Clark’s nutcracker (discussed in Chapter Two).

The degree of movement displayed during the 1980s could be related to limited snowpack and an increase in available sites for the caching of seeds by the Clark’s nutcracker. During the 120 year record of establishment, the plot has been characterized by open-park like conditions, favorable for whitebark pine establishment.

4.6 Conclusion

This study provides a summary of the comparison of historical and current forest establishment rates in the lower, middle, and upper elevations of western Montana. The data indicate that a shift of species establishment occurred at a variety of elevations during the period 1820 to 1990. In general, dense, whitebark pine-dominated stands have been largely impacted by a lower canopy of subalpine fir at lower and middle-elevations beginning in the late 1800s. The upper-elevation plot did not exhibit incursions of subalpine fir establishment. The effects of wildfire could possibly explain the difference between elevational gradients and subalpine fir lower canopy dominance. The lack of wildfire could have allowed subalpine fir to establish in the subcanopy and could result in whitebark pine being out-competed for resources, consequently lowering the recruitment rates at the lower and mid-elevation plots.

Spatial movement varied across elevations. The findings indicate that during the period 1820–1990, only minimal advancement of treeline occurred at the lower-elevation plot with the earliest movement occurring in the 1860s, and a second advancement occurring in the 1920s. The early movement at the lower-elevation plot occurred at the termination of the unfavorable climatic conditions associated with LIA. The middle-elevation plot remained stable since establishment began in the 1840s. The upper-elevation plot showed the most significant movement of treeline occurring in the 1980s.

I found a more significant correlation between establishment rates and climate (PDSI) than between treeline shifts and climate. Periods of whitebark pine establishment appear to be related to dry conditions, while subalpine fir establishment favored wet conditions. Initiation of establishment at the lower- and middle-elevation plot was synchronous with the termination of the LIA (1850). At the upper-elevation plot, establishment was delayed 50 years after the termination of the LIA.

The results suggest that the period between the end of the LIA and the current period is one marked by climatic fluctuations in western Montana that affected the establishment rates of high-elevation species rather than facilitating a rapid movement of treeline. The lack of natural disturbance (*i.e.* wildfire) in high-elevation ecotones since the beginning of the 20^{th} century may also be affecting the advancement of treeline.

CHAPTER FIVE

FIRE HISTORY OF A WHITEBARK PINE (*PINUS ALBICAULIS* ENGELM.) STAND IN THE GRAVELY RANGE, BEAVERHEAD-DEERLODGE NATIONAL FOREST, MONTANA

5.1. Introduction

At high-elevation sites in the Northern Rocky Mountains where human effects have been minimal, climate variability has been the dominant influence on interannual-scale fire dynamics. Since the beginning of the 20th century, however, effects of fire suppression in some areas have allowed young, fire-intolerant trees to establish during lengthened fire intervals. An increase in the understanding of the interannual relationships between climate and wildfire could provide information for helping plan and employment of fire management programs. Linkages between wildfire and climate have been established in certain areas of the western U.S., in particular in the American Southwest (Swetnam 1990, Grissino-Mayer 1995, Grissino-Mayer and Swetnam 2000, Allen *et al*. 2002, Lewis 2003) and in the Northern Rocky Mountains (Arno 2000, Kipfmueller and Baker 2000, Buechling and Baker 2004, Schoennagel *et al*. 2004). However, a gap currently exists in our knowledge concerning high-elevation (> 2700 m) fire in the Northern Rockies in ecosystems dominated by whitebark pine (*Pinus albicaulis* Engelm.).

Examination of high-elevation fire regimes prior to the initiation of fire suppression in the 20th century can enhance our understanding of how natural climate shifts affected fire regimes. Fires that occurred during longer-term climatic episodes (such as the Medieval Warm Period (A.D. 900–1300) and Little Ice Age (A.D. 1400–

1850)) may have had a major influence on the structure and composition of forests in the northern Rocky Mountains (Arno 1980).

Increasingly, ocean-atmospheric teleconnections that function on different time scales and originate from different regions of the world have been shown to affect regional climate and fire activity in the high-elevation forests of North America (Gray *et al.* 2003, Schoennagel *et al.* 2004, Sibold and Veblen 2006). At the annual scale, widespread fires in the Rocky Mountains have been associated with individual years of considerable drought (Kipfmueller and Swetnam 2000, Sherriff *et al.* 2001). Precipitation also plays a significant role in the fire regimes of dry forests throughout western North America (Baisan and Swetnam 1990, Grau and Veblen 2000, Veblen and Kitzberger 2002), with years of above-average precipitation increasing the growth of fine fuels, and consequently increasing the likelihood of fires in the years that follow.

In the absence of wildfires, high-elevation whitebark pines appear to be replaced by more fire-intolerant subalpine fir (*Abies lasiocarpa* (Hook) Nutt.) and Engelmann spruce (*Picea engelmannii* Parry ex Engelm.) (Morgan and Bunting 1990). Due to specific fire adaptations, whitebark pine can survive low-intensity surface fires, which kill associated conifers (Morgan and Bunting 1990). Low to moderate intensity fires benefit whitebark pine regeneration by eliminating the sub-canopy associates, such as subalpine fir (Tomback *et al*. 1990), and creating open conditions that enable the Clark's nutcracker to cache seeds (Morgan and Bunting 1990). In the absence of fire, cone production diminishes as more shade-tolerant and less fire-resistant trees increase in abundance, thereby competing with or replacing whitebark pine (Morgan and Bunting 1990).

This chapter develops a reconstruction of high-elevation fire regimes in the whitebark ecosystem of the Gravely Range. I investigate the linkages between climate of the Medieval Warm Period and the Little Ice Age and how these climate shifts affected fire regimes in the whitebark pine ecosystem. Also, I examine the relationship between long-term and short-term climate and the mean fire interval.

5.2. Site Selection

The study area is located in the Gravely Range in the eastern part of the Beaverhead-Deerlodge National Forest, in southwestern Montana (Figure 5.1). Sampling was restricted to a high-elevation, subalpine forest dominated by whitebark pine and subalpine fir. The area is characterized by an open forest of multi-stemmed whitebark pine in the overstory, and dense subalpine fir in the understory. Topography is characterized by steep slopes generally oriented in a north to south or northwest to southeast direction.

I chose this site for several reasons. First, the site has a southern aspect, which increases the possibility of examining the relationship between reconstructed PDSI and fire occurrence because these trees should be sensitive to climatic factors. Second, the site is dominated by long-lived whitebark pine, increasing the ability to investigate the decadal scale (< 50 years) and longer-term, centennial scale (> 100 years) changes in the fire regime. Specifically the longevity of whitebark pines makes it possible to examine the occurrence of fire activity in high-elevation ecosystems during the Medieval Warm Period and the Little Ice Age. Third, numerous fires-scarred trees, both living and dead, were found in this area. Visual signs of wildfire, such as charred bark and fire-scarred

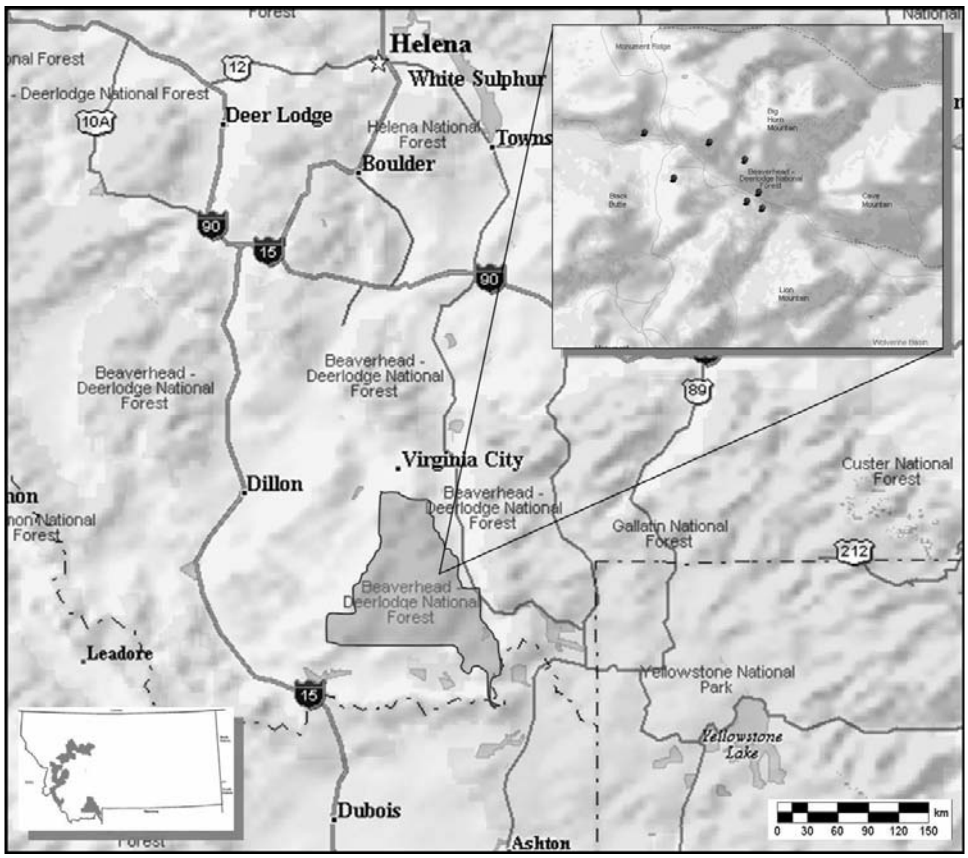


Figure 5.1. Map of the study area showing the Gravely Range (N45.00, W111.51) (map created using Delorme mapping software).

trees, were evident throughout the site.

5.3. Methods

5.3.1. Field Methods

I collected 26 fire-scarred whitebark pine cross-sections during the summer of 2006 using a chain saw. Because whitebark pine is a threatened species, cross-sections were taken from non-living fire-scarred logs, stumps, and standing snags. Samples were collected in a relatively small geographic area that was adjacent to the study plots established for the treeline sensitivity and response study presented in Chapter 4. All samples were wrapped in plastic wrap to protect them during transportation. All cross-sections were numbered and drawn on a sample form to facilitate reassembly, if needed.

5.3.2. Laboratory Methods

All samples were frozen at –40 °C before analysis in the laboratory to kill any insects contained in the wood. If necessary, fire-scarred sections were reassembled and mounted on plyboard for stability. All samples were sanded using ANSI 40-grit (500–595 μm) to ANSI 400-grit (20.6–23.6 μm) sandpaper (Orvis and Grissino-Mayer 2002) to ensure the cellular features of the individual tree rings were visible under standard 7–10X magnification.

5.3.3. Quality Control

I crossdated all fire-scarred samples with the Gravely Range chronology (developed in Chapter 3) to attain dates for the fire scars contained in the ring series.

Quantitative crossdating was performed by measuring the ring widths on each sample with a Velmex moveable stage micrometer and recorded with Measure J2X software to the nearest 0.001 mm. Measurements were then processed through the computer program COFECHA to check the visual crossdating with correlation analysis using overlapping 50-year segments lagged 25 years (Holmes 1983, Grissino-Mayer 2001). COFECHA identifies tree-ring data that should be re-evaluated for crossdating errors (Holmes 1983). The COFECHA program checks for a positive and significant correlation for each segment. It also determines if the correlation is higher when segments are shifted forward or backward from that point (Holmes 1983).

If COFECHA indicated segments that were possibly misdated, I visually re-examined the sample and corrected the date as required. If missing or false rings were indicated, I used parts 5 and 6 of the COFECHA output to help diagnose the problem area. Part 5 of the COFECHA output shows the correlation of all segments for each tree-ring series with the same segments from the master chronology, while Part 6 provides critical information for identifying segments with potential dating problems (Grissino-Mayer 2001). Flagged segments were also examined to ensure no measurement errors occurred. Finally, single measurements that were statistical outliers after filtering and transformation were also noted and evaluated (Holmes 1983).

Once all tree-ring series were successfully dated, I assigned dates to all fire scars and to the season of occurrence (seasonality). To determine the seasonality of the fire event, the intra-annual position of the fire scar in the ring was noted (Dieterich and Swetnam 1984, Baisan and Swetnam 1990). The seasonality of a fire event falls into five categories (Grissino-Mayer 1995):

- Dormant season fire (D): The fire scar is situated between the latewood of the preceding growth ring and the earlywood of the following growth ring. The fire may have occurred anywhere between the end of the previous year's growing season and the initiation of the following year's growing season.
- Early season fire (E): the fire scar is located in the first one-third of the earlywood.
- Middle season fire (M): the fire scar is located in the middle one-third of the earlywood.
- Late season fire (L): The fire scar is located in the latter one-third of the earlywood.
- End of growing season fire (A): the fire scar is located in the latewood.

After determining the date and seasonality of every fire scar contained in each sample, I recorded all data on fire-scar data sheets, which included the fire-scar dates, inner- and outer-ring dates and types (i.e., pith, non-pith, bark, or non-bark), descriptions of the injuries, site description, and recorder and non-recorder years. A tree is considered fire-susceptible if previously scarred by fire (Romme 1982), consequently becoming a recorder tree (Grissino-Mayer 1995). A tree becomes a recorder when it is previously scarred by wildfire and the ring pattern is very clear, thus indicating that the series contains all rings. A tree is a "non-recorder" before it has been scarred for the first time. On occasion, a tree may revert to non-recorder status after it has been scarred. This occurs when sections of the tree-ring surface are burned off, decayed, or obscured by beetle galleries. Each fire-scar date was confirmed by an independent observer to ensure accurate dating was achieved.

5.3.4. Statistical Analysis

I entered the fire-scar data in the FHX2 software program for statistical analysis. FHX2 is a software program that supports the compilation, organization, and analysis of tree-ring data for the fire history of forest ecosystems (Grissino-Mayer 1995, 2001). This software is specifically designed for (1) entering and archiving of fire history data, (2) creating graphs that visually depict both the temporal and spatial characteristics of the site of fire history study, and (3) conducting statistical analysis on fire-free intervals and seasonal changes (Grissino-Mayer 2001).

The FHX2 software allows the user to examine the percentage of trees scarred in any given year, and facilitates analysis of the temporal and spatial patterns of past wildfires. The percentage-scarred class is a tool for investigating the scope of past wildfires (Swetnam 1990), based on the assumption that a greater percentage of scarred trees at a given site represents a more widespread fire occurrence. The number of "recorder" trees scarred is used to calculate the percentage scarred. In FHX2, non-recorder years are referred to as "null years" and are not included in the percentage-scarred analysis. It is possible for a tree to contain fire scars in these null years, but it is not possible to be certain that all fires are recorded. Due to the limited number of samples in this study and the spatial extent of sampling, I used the all-scarred class to analyze fire events for the Gravely Range sample site. I began by examining the entire range (A.D. 728–1928). I further divided the entire range into separate periods that coincide with climatic periods: Medieval Warm Period as a whole (MWP: 900–1300), Early Medieval Warm Period (EMWP: 900–1100), Late Medieval Warm Period (LMWP: 1101–1300), Little Ice Age as a whole (LIA: 1400–1850), Early Little Ice Age (ELIA: 1400–1625), and the Late Little Ice Age (LLIA: 1626–1850).

The period of reliability is the period considered to be suitable for conducting statistical analysis (Grissino-Mayer 1995), and is considered by fire ecologists as the period deemed suitable for statistical analyses of fire regimes at a particular site (Wong *et al*. 2004, Arabas *et al*. 2006). Due to the limited number of fire-scarred samples, the period of reliability for the Gravely Range site was established using a minimum number of one fire-scarred sample and a minimum of one sample overall for that period. The period of reliability therefore brackets the beginning fire year and last fire year.

5.3.5. Measures of Central Tendency

I analyzed fire frequency using the Weibull distribution, which is a flexible statistical distribution that can be fit to a wide variety of fire interval data (Grissino-Mayer 1995, 1999, 2001). The Weibull Median Interval (MEI) and the Weibull Modal Interval (MOI) are the principal measures of central tendency used in this study. The MEI is the fire interval that corresponds to the 50th percentile of the modeled distribution (Grissino-Mayer 2001). Because the MEI represents the midpoint in the theoretical distribution fit to the actual data, it provides a measure of central tendency not easily influenced by larger values in the actual data set. The MOI is the interval that corresponds to the greatest amount of area under the probability density function

(Grissino-Mayer 2001). Lastly, the Mean Fire Interval (MFI) represents the average of all fire intervals.

5.3.6. Measures of Dispersion

The standard deviation (SD) was calculated to illustrate variability about the measures of central tendency. Sites with elevated values of MFI, however, typically display increased variability about the mean. The coefficient of variation (CV) is a standardized statistic that combines both the standard deviation and the MFI, so that intra-site variability in fire intervals can be evaluated (Grissino-Mayer 1995). Higher values for the coefficient of variation signify more variability in the length of fire intervals.

5.3.7. Measures of Range

The minimum (MIN) fire interval and maximum (MAX) fire interval are the shortest and longest actual fire-free intervals. The Lower Exceedence Interval (LEI) and the Upper Exceedence Interval (UEI) define the theoretical fire-free intervals equivalent to the 12.5 and 87.5 percentiles in the modeled Weibull distribution (Grissino-Mayer 2001). These intervals emphasize extremes in the data set, with the LEI delineating the statistically significant shorter intervals and the UEI delineating the statistically significant longer intervals. The final measure of range I used in this study is the Maximum Hazard Interval (MHI), the longest theoretical fire-free interval that an ecosystem can sustain before it should burn (Grissino-Mayer 1995, 1999). FHX2 calculates the MHI using the 50% threshold level because once the interval passes the 50% threshold level, that particular ecosystem may burn.

5.3.8. Measures of Shape

For the Gravely Range study site, the Weibull probability density function (pdf) was graphed to visually depict the shape of each fire interval distribution (Grissino-Mayer 1999). The pdf shows the probability that a fire will occur for a given interval. Skewness and kurtosis were calculated to provide additional information about the shape of the distributions.

The shape of the pdf is sensitive to extremely long fire intervals. For example, a site where fire was frequent but also had a few longer fire-free intervals will be positively skewed. If the number of unusually long intervals increases, the shape of the distribution will tend to flatten (i.e. have lower values of kurtosis) and will appear more symmetric. This also affects the MHI, as the pdf is a major component of the MHI calculation (Grissino-Mayer 1999). Therefore, it is possible that both the pdf and the MHI can be adversely affected by unusually long fire intervals.

5.3.9. Seasonal Analysis

The FHX2 statistical package also allows the user to analyze seasonal changes in fire regimes over time by pinpointing the dominant season of fire activity for a selected period. Seasonal analysis on the Gravely Range site was calculated over the same periods previously designated as the Medieval Warm Period and the Little Ice Age. The analysis compared the percentage of early-season fires (D and E events) to the percentage of late-season events (M, L, and A) for the subdivided periods.

5.3.10. Fire/Climate Relationship

To test the relationship between climate and fire on interannual timescales, I used Superposed Epoch Analysis (SEA) (Baisan and Swetnam 1990, Swetnam 1993, Grissino-Mayer 1995). SEA provides graphical and statistical techniques for evaluating moisture conditions leading up to, during, and immediately after a fire occurrence by first stacking the fire event years and then calculating the average climate conditions leading up to and following individual fire years (Grissino-Mayer 1995). To determine statistical significance, confidence intervals were calculated using bootstrapping techniques on randomly selected fire events from the pool of observations (Grissino-Mayer *et al*. 2004).

Finally, I decided to segment the entire record into 200-year periods to more clearly examine the effects of climate on the Gravely Range site. Separation of the SEA helps assess whether changes may have occurred over time, and that may be concealed in the overall climate/wildfire relationship.

5.3.11. Changes in Temporal Patterns of Fire

Observed changes in fire occurrence during the MWP and LIA were tested to examine statistical significance and to either accept or reject the null hypothesis that there is no difference in mean, variance, or shape, or the alternate hypothesis that there is a difference in mean, variance, and shape between the two periods. Three tests were used to analyze the temporal characteristics of fire occurrence between the two periods: (1)

Student's t-test to test differences in mean fire interval (MFI) periods; (2) folded f-test to test whether differences exist in the variability about the MFI from one period to the next; and (3) two-sample Kolmogorov-Smirnov test to test differences in the distributions of fire-free intervals between periods (Grissino-Mayer 1995).

5.3.12. Master Fire Chart

I visually examined past fire events using the graphics module in FHX2 to produce a master fire chart (Grissino-Mayer 2001). On the chart, horizontal lines represent a single tree while tick marks represent dated fire scars recorded by that sample. The dashed sections of the lines represent non-recorder years, while the solid portions of the lines represent recorder years. I examined the Gravely Range master fire chart to intially analyze temporal patterns of wildfires, and to identify any changes that might have occurred over time.

5.4 Results

5.4.1 Fire Chronologies and Sample Information

The number of samples collected at the Gravely Range site was limited, so the data and interpretations of fire history should be considered preliminary. Likewise, even though the periods of reliability signify an extended time scale, some segments of the fire record contain only one fire-scarred sample. However, this research represents an exploratory use of high-elevation whitebark pine for fire-scar history and consequently is an initial evaluation of the effectiveness of these data to present information concerning fire history analysis. Twenty fire-scarred trees showed the characteristic "catface" appearance (Figure 5.2) and resulted in a fire chronology that spans multiple climatic episodes (Figure 5.3). A total of six scarred samples had injuries with healed wounds within the cross-section. Eighteen samples contained multiple fires scars.

5.4.2 Fire Interval Analysis

Between 900 and 1850, the Gravely Range experienced a Weibull median fire return interval (MEI) of 11 years (Table 5.1) for the all-scarred class. The MWP displayed the shortest MEI (8 years), while I found the most extended fire-free interval during the LIA (20 years). During the MWP, the longest fire-free interval (10 years) coincided with the EMWP (900–1100). In the LIA, the longest fire-free interval (22 years) occurred during the LLIA (1626–1850). For the MWP and LIA, the LEI and UEI are 2.0 and 7.0 years, respectively. Values for the LEI and UEI were longest during the LLIA (7.0). The maximum fire-free interval (MAX) for the entire period was 111.0 years. Of the periods examined (MWP and LIA), the EMWP displayed the shortest MAX (35 years), while the LLIA showed the longest MAX (49 years). The short return intervals during the MWP emphasizes the significance of fuel availability during this period (Figure 5.4). If fuels were not consumed in a particular fire year, the following year's production could ensure adequate fine fuels for the site to burn again.

5.4.3 Fire Seasonality

I could determine season of fire occurance for 81% of the samples. The majority of fires during the entire period, the MWP, and the LIA occurred during the early season:

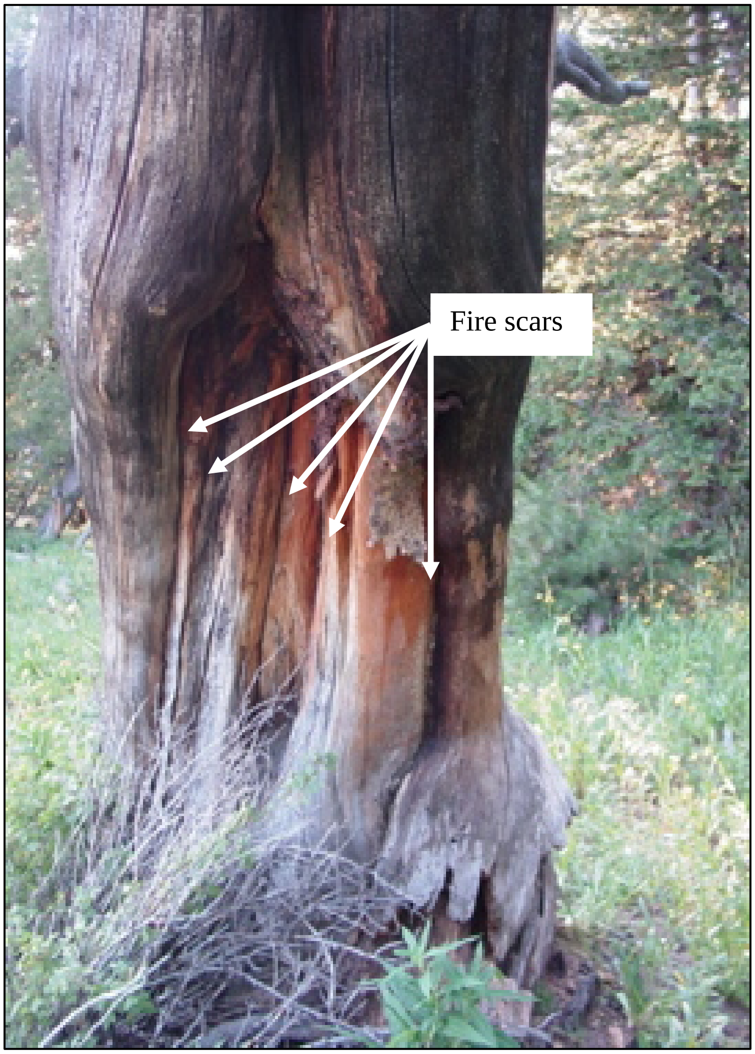


Figure 5.2. Standing dead whitebark pine at the Gravely Range site showing a cat face and multiple fire scars visible around the open wound. Arrows point to fire scars.

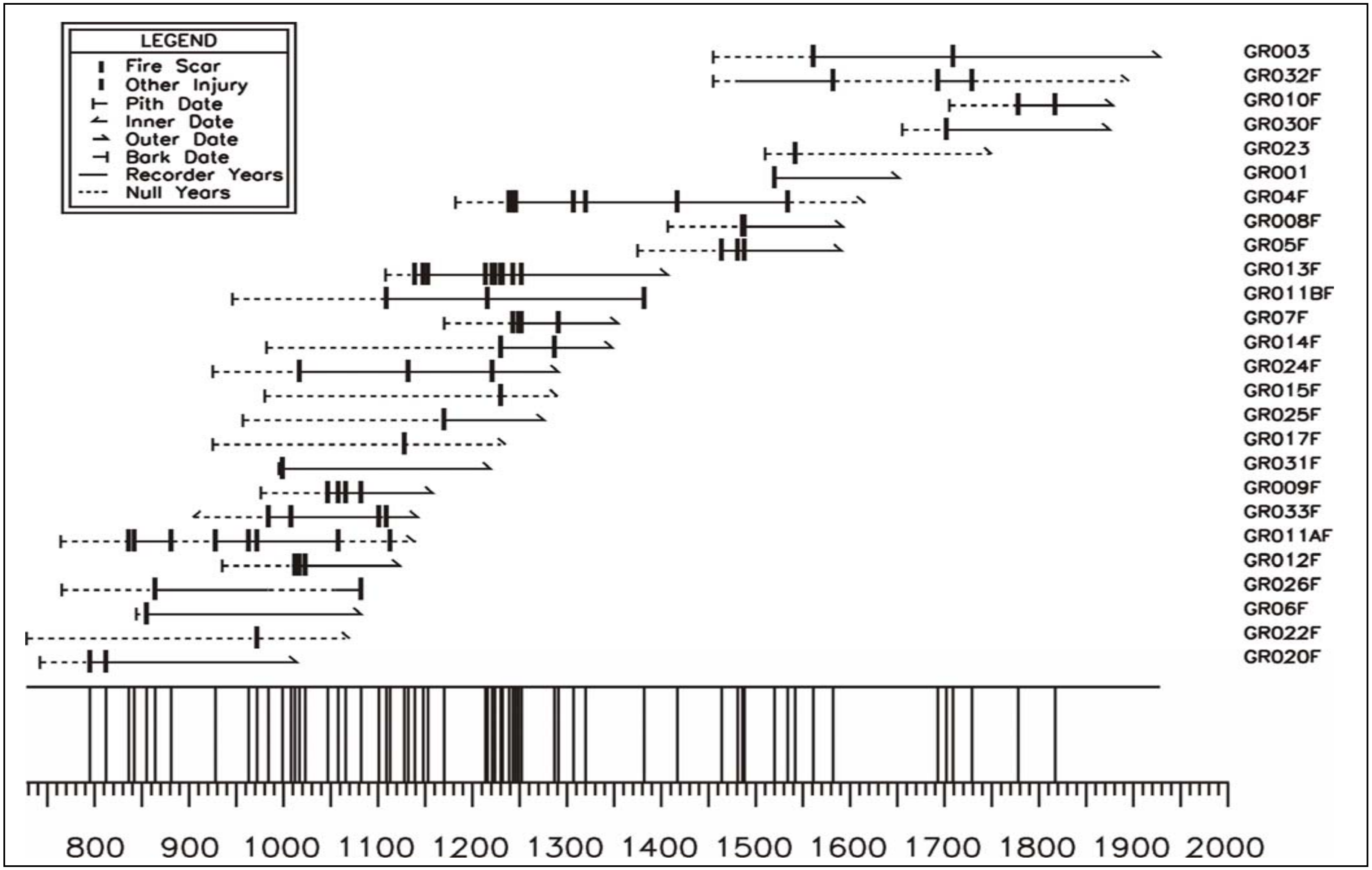


Figure 5.3. The master fire chart for the Gravely Range site. The horizontal lines represent the samples, and the vertical tick marks indicate dated fire scars. Solid lines represent recorder years, while dashed lines represent non-recorder years.

Table 5.1. Descriptive statistics for fire-free interval analysis for the Gravely Range site in the Beaverhead-Deerlodge National Forest.

Statistics†	All	MWP	EMWP	LMWP	LIA	ELIA	LLIA
End year	1850	1300	1100	1300	1850	1625	1850
Total no. intervals	49	33	11	22	16	11	5
MFI (yr)	16	10	14	8	25	16	24
MI (yr)	9	6	11	5	17	15	20
MEI (yr)	11	8	12	6	18	14	21
MOI (yr)	1	2	8	0	2	8	13
MIN (yr)	1	1	4	1	2	2	7
MAX (yr)	111	44	35	44	111	47	49
LEI (yr)	2	2	4	1	4	4	7
UEI (yr)	34	21	24	18	50	30	44
MHI	>1000	>1000	221.42	>1000	>1000	961.69	>1000
Kurtosis	9.48	1.92	0.45	4.15	4.10	1.37	1.85
cv	1.16	0.99	0.70	1.25	1.08	0.75	0.75
Skewness	2.73	1.65	0.91	2.28	2.13	1.35	0.29

† Abbreviations: MFI= Mean Fire Interval; MI= Median Interval; MEI = Weibull Median Interval; MOI = Weibull Modal Interval; MIN = Minimum Interval; MAX = Maximum Interval; LEI = Lower Exceedance Interval; UEI = Upper Exceedance Interval; MHI = Maximum Hazard Interval; cv = coefficient of variation.

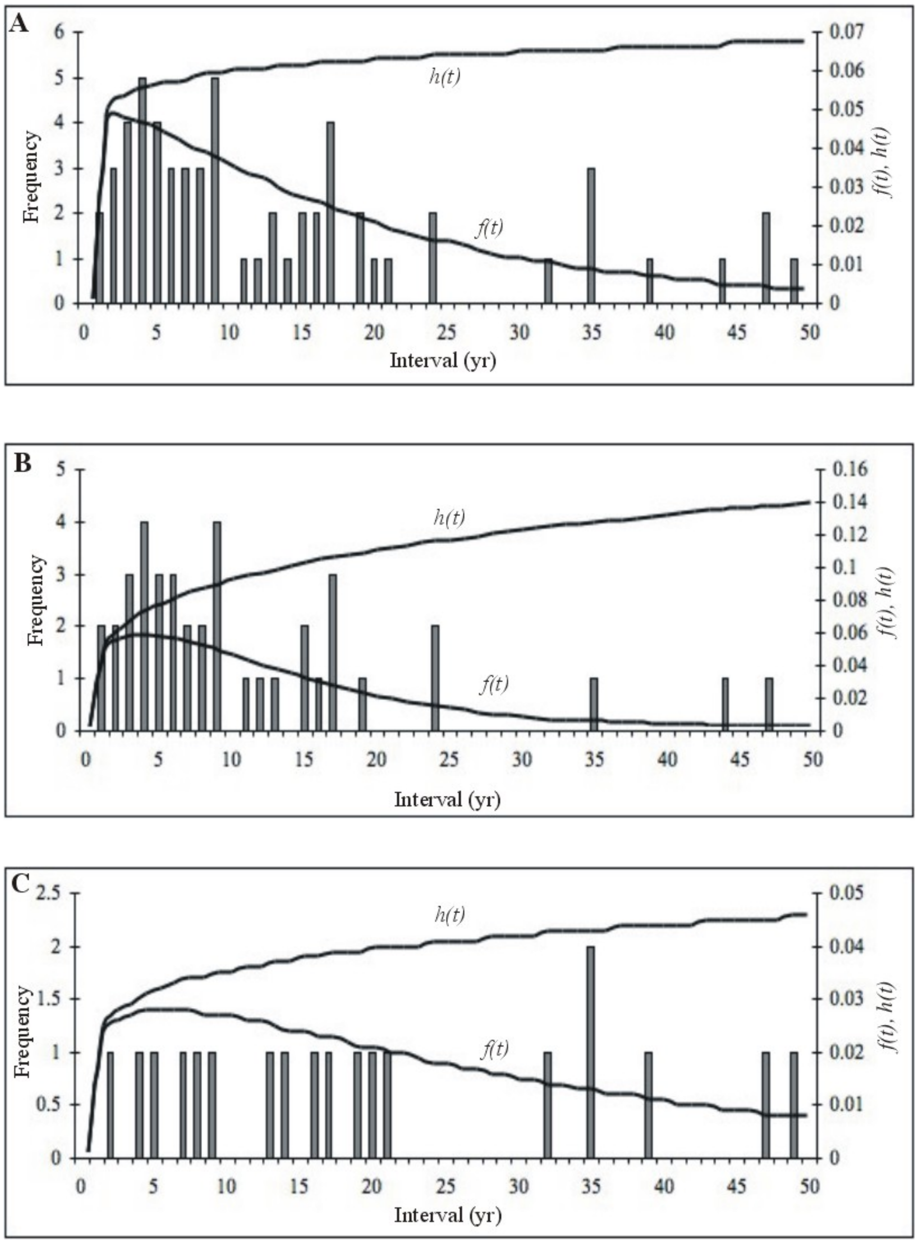


Figure 5.4. Distributions of the fire-free intervals for the Gravely Range site during the period of Reliability; A = combined reference period (900–1850), B = MWP (900–1300), and C = LIA (1400–1850). The probability density function, *f (t)*, models the fire-free interval data based on the Weibull distribution, and can be used to assess the goodness of fit. The hazard rate is represented by *h(t)*. Steep slopes indicate periods of shorter fire-free intervals and greater degree of fire hazard.

46, 33, and 13, respectively (Table 5.2 and Figure 5.5). An analysis of temporal changes in seasonality during the MWP and LIA showed that the highest occurrence of fire activity occurred during the early season of the MWP (68%). Of the subdivided periods during the MWP (EMWP and LMWP), the highest occurrence of fire activity was displayed during the LMWP (22) (Figure 5.6 A and B). During the LIA, the most prominent shift occurred during the ELIA with fires occurring during the early and late season (43% and 37%, respectively) (Figure 5.6 C and D).

5.4.4 Fire/Climate Relationship

The first analysis addressed the relationship between all fire events and climate for the entire record. At the Gravely Range site, no statistical relationship was found between PDSI and fire occurrence (Figure 5.7A). However, a moderate relationship was found two years (t – 2) and six years (t – 4) prior to all fire events. Also, the negative PDSI values for all years from t – 2 and t – 6 prior to the fire year suggests that persistent drought, or possible lower snowpack levels, over several years possibly preconditions fuels that eventually lead to a wildfire at the Gravely Range site.

The second analysis focused on the relationship between changes in the fire/climate relationship between the MWP and LIA. During the MWP, no statistically significant relationship was found between PDSI and fire events, although drought two years (t – 2) prior to fire events showed a near-significant relationship (Figure 5.7B). The fire/climate relationship during the LIA showed a moderate relationship between fire events and PDSI values four years (t – 4) and five years (t – 5) prior to the fire event

Table 5.2. Number of fire occurrences during the dormant (D), Early season (E), Middle (M), Late (L), and End of Season (A) for the Gravely Range site in the Beaverhead-Deerlodge National Forest for the entire period, MWP, EMWP, LMWP, LIA, ELIA, and LLIA.

	Entire Period	MWP	EMWP	LMWP	LIA	ELIA	LLIA
Dormant Season	0	0	0	0	0	0	0
Early Season	46	33	11	22	13	7	6
Mid-Season	9	6	2	4	3	2	1
Late Season	13	7	3	4	6	6	0
End of Season	3	2	1	1	1	1	0

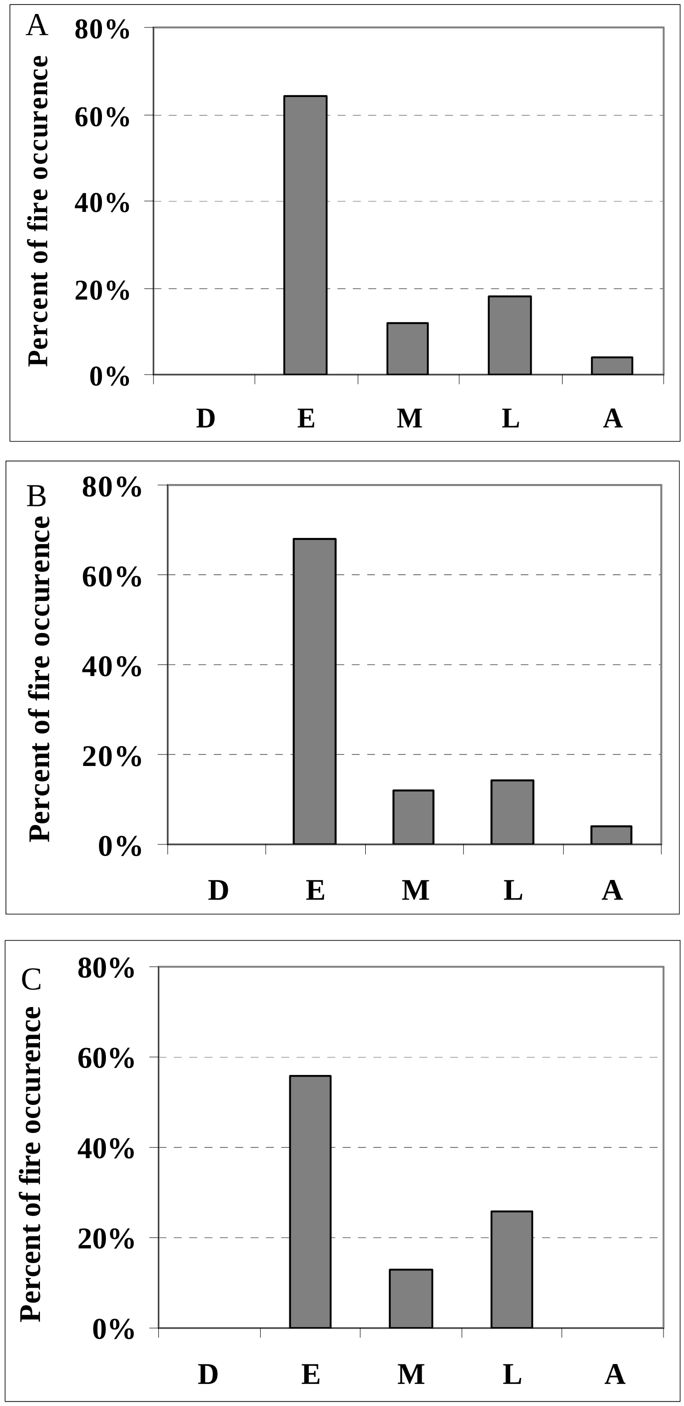


Figure 5.5. Seasonality of fires at the Gravely Range site for the full period of reliability. A= Total period (900–1850), B= Medieval Warm Period (900–1300), C= Little Ice Age (1400–1850).

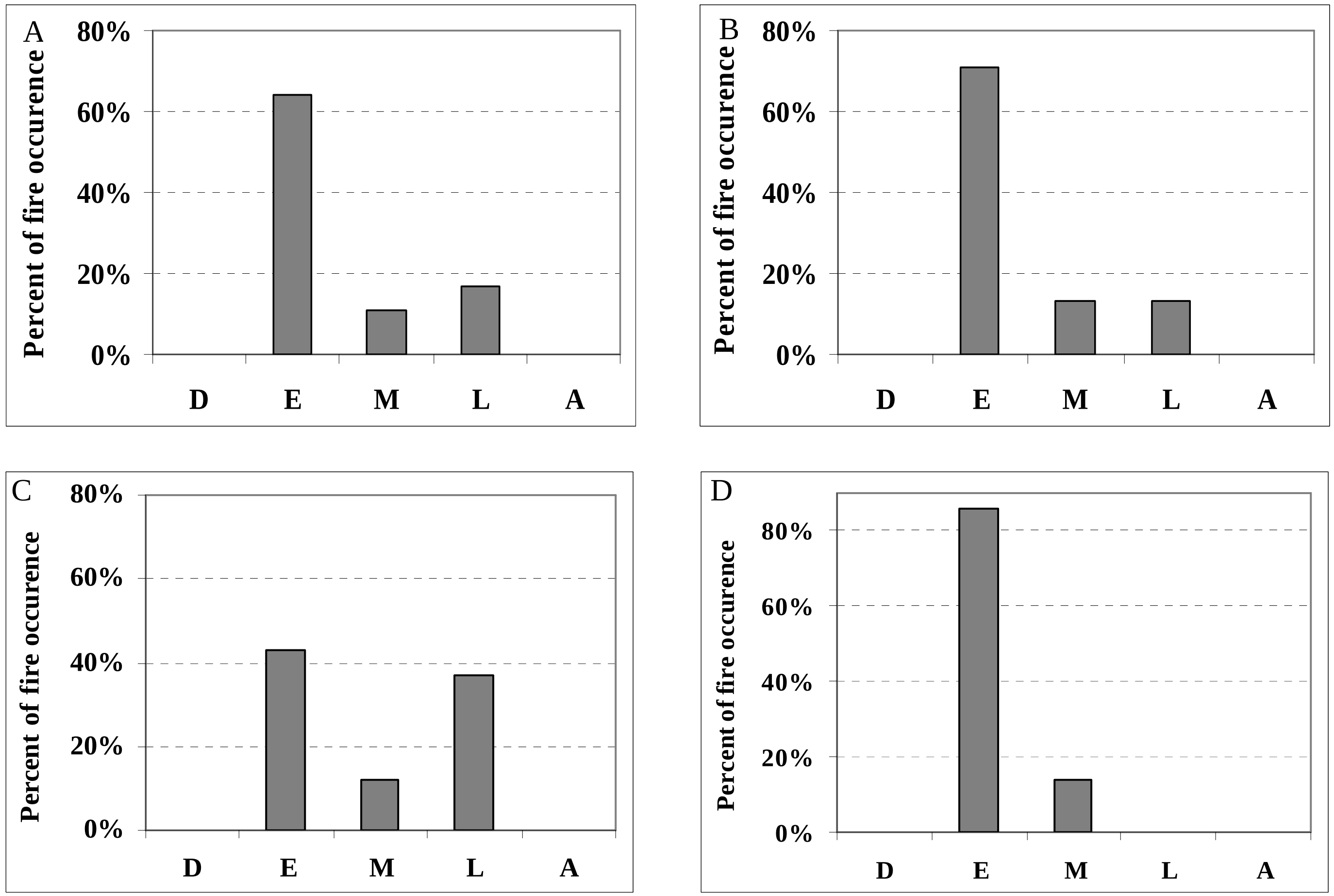


Figure 5.6. Seasonality of fire for the Gravely Range site during the (A) early MWP (900–1100), (B) late MWP (1101–1300), (C) early LIA (1400–1625), and (D) late LIA (1626–1850).

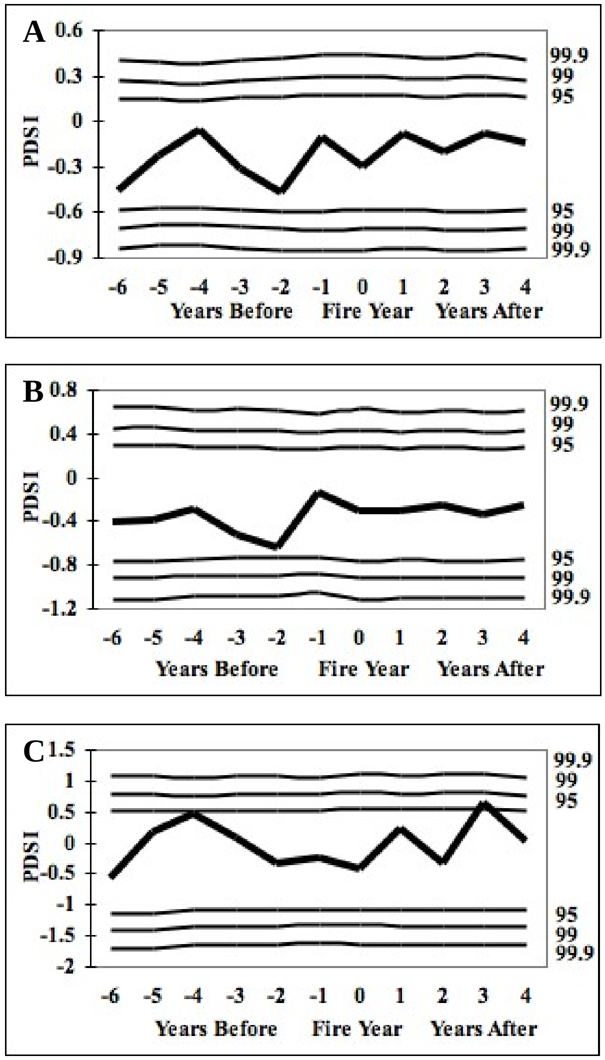


Figure 5.7. Relationships between western Montana reconstructed previous year's June–July PDSI and all fire years from the Gravely Range site; (A) entire period (900–1850), (B) MWP (900–1300), and (C) LIA (1400–1850). Positive PDSI values indicate wet periods and negative PDSI values indicate dry periods.

(Figure 5.7C). This result suggests that climate during the MWP, a period of above average temperatures, was more likely to result in lower snowpack levels and result in preconditioning fuels at the Gravely Range site prior to the fire event. Climate during the LIA, a period of lower than average temperatures (Lamb 1977), would have increased snowpack levels and resulted in the inability for fuels to be preconditioned at the site.

During the MWP, the period 900–1099 showed no statistically significant relationship between drought and fire activity (Figure 5.8A). However, two years (t – 2) prior to the event does show indication of moderate preconditioning during this period. The period 1000–1199 does show a near-significant relationship for a four-year period (t – 5 to t – 2) prior to the fire event (Figure 5.8B). The following period, 1100–1299, shows a moderate relationship 2 years prior (t – 2) between drought and the fire event (Figure 5.8C). The final period of the MWP (1200–1399) showed no significant relationship between drought and fire (Figure 5.8D). Throughout the period when climate shifted from the MWP to the LIA (1300–1499) no statistically significant relationship between drought and fire existed (Figure 5.8E).

During the LIA, the period 1400–1599 showed no statistically significant relationship between drought and fire occurrence (Figure 5.8F). The period 1500–1699 does show a near-significant relationship five years prior (t – 5) to the fire event (Figure 5.8G). The following period, 1600–1799, shows a limited relationship 2 years (t – 2) and four years (t – 4) prior to the fire event (Figure 5.8H). The subsequent period, 1700–1899, showed a near-significant relationship during the fire year (t), one year prior (t – 1) and 4 years prior (t – 4) between drought and the fire event (Figure 5.8I).

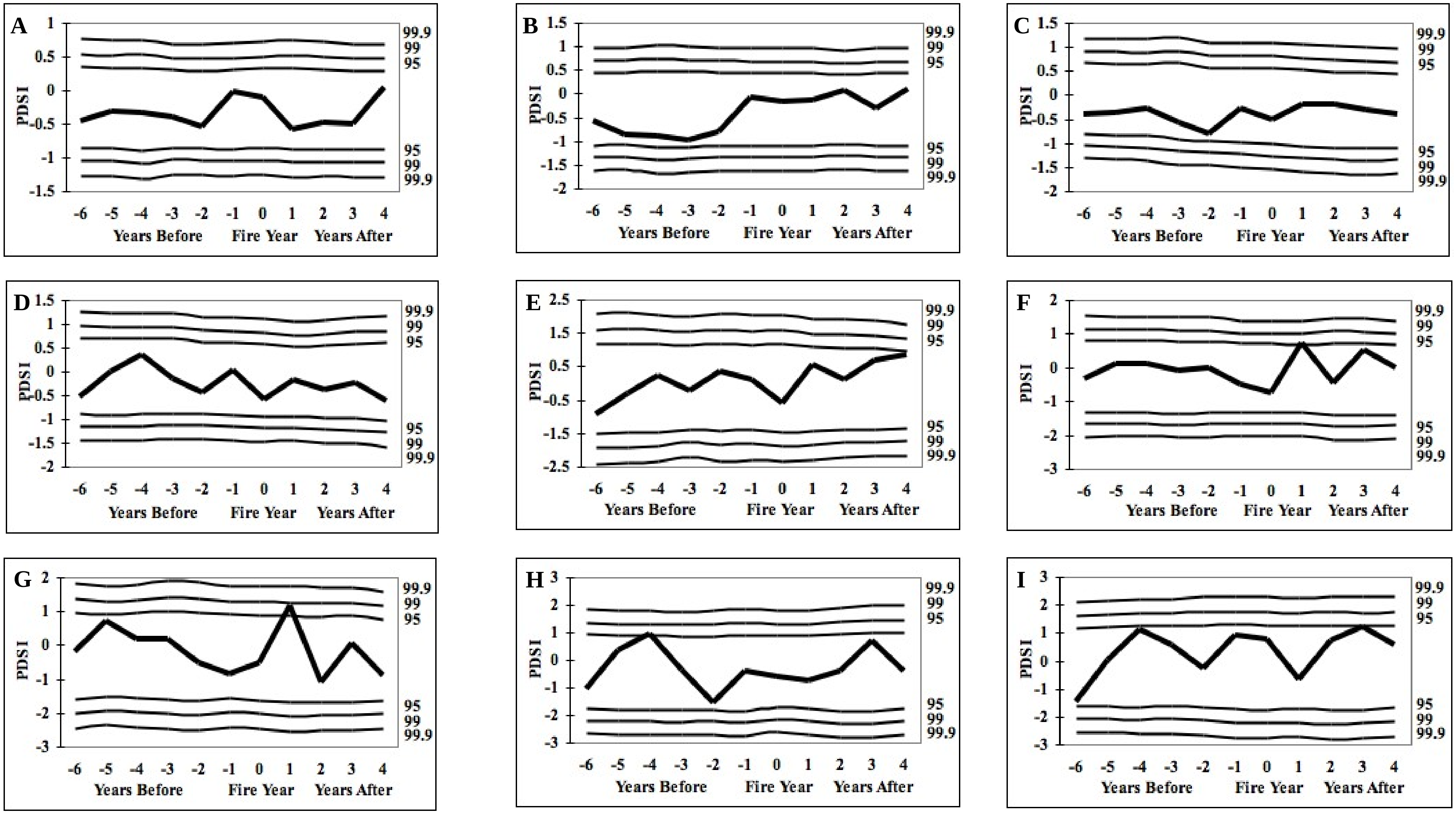


Figure 5.8. Relationship between the western Montana reconstructed previous year's June–July PDSI and all fire events for the Gravely Range site by 200 year segments; (A) 900–1099, (B) 1000–1199, (C) 1100–1299, (D) 1200–1399, (E) 1300–1499, (F) 1400–1599, (G) 1500–1699, (H) 1600–1799, and (I) 1700–1899.

An overall pattern in the fire/climate relationship is evident when comparing the MWP and LIA. The year's necessary for preconditioning during the MWP (t – 2 and t – 3) shifted during the LIA (t – 4 and t – 5) with the greatest degree of alteration in the fire regime occurring in the period 1700–1899. The period 1400–1599 is interesting in the fact that it exhibits the least significant relationship between drought and fire occurrence. This could possibly be an indication of the climate shift between the MWP and the LIA and the alteration of the fire regime between the two periods.

5.4.5 Statistical Analysis of Temporal Changes

The temporal analysis showed statistically significant differences in the MFI between the MWP and LIA ($p < 0.0008$) (Table 5.3). During the MWP, the MFI was short (9 years), but increased significantly during the LIA (25 years). These findings confirm that a significant change occurred in fire regimes between the MWP and LIA. The analysis of variance between the MWP and LIA found no statistical difference between the periods. Differences in the overall distributions between the two periods were significant ($p < 0.01$).

5.5 Discussion

5.5.1 Collection of Samples

The fire statistics reflected in the study are conservative estimates of fire intervals due to sample size, and because trees do not scar with every low- to moderate-severity fire occurrence (Swetnam *et al*. 1999). The small sample size is problematic because it affects the interpretations and lessens the probability of detecting all fires that occurred at

Table 5.3. Testing differences in fire intervals between the MWP (900–1300) and the LIA (1400–1850) at the Gravely Range site. Test statistics are computed on interval data transformed to approximate normality. Asterisks indicate statistically significant (* $p < 0.0008$, and ** $p < 0.01$) test results. Values in parentheses represent sample size.

Test	Value
Mean:	
900–1300	9.31 yrs (33)
1400–1850	25.00 yrs (16)
t -value	3.57*
p > *t*	0.0008
Variance:	
900–1300	0.74
1400–1850	1.02
F-value	1.38
p > F	0.44
Distributions:	
K-S *d*-statistic	0.46**
p > *d*	0.01

the Gravely Range site. This analysis is essentially a site-specific study, and composite data from a larger area would yield more detailed information concerning the fire return interval at long time scales.

From the numerous fire scars I found in these limited number of samples, fire was obviously frequent in the whitebark pine ecosystem at this site in the past. The abundance of whitebark pine trees that exceed several hundred years in age indicates that stand-replacing fires of high severity were uncommon. If most fires were low to moderate severity, whitebark pine trees could easily scar due to its relatively thin bark (Figure 5.9). In the study, few trees had healed themselves by growing completely over the fire scar (Figure 5.10). Normally, trees must be injured first to provide an opening in the bark for fires to penetrate (Wahlenburg 1946). Once the tree has been scarred, it can continue to easily record additional fires until the bark overgrows and seals the injury. In the study, trees appeared to be more susceptible to scarring by fire when they were older due to thinner bark that resulted in less protection.

5.5.2 Fire Frequency in the Whitebark Pine Ecosystem

Although the minimum fire interval at the Gravely Range site was one year during the MWP and two years during the LIA, the fire frequency is variable enough over time to allow for the recruitment of whitebark pine and subalpine fir during post-fire periods with extended fire-free periods. The research indicates that no fires occurred at the Gravely Range site from 1817 to 1928. Increased fire frequency during the MWP would have favored the recruitment of whitebark pine by the reduction of understory species such as subalpine fir.

Figure 5.9. Multiple fire scars in whitebark pine cross section collected from the Gravely Range site (A.D. 1231 and A.D. 1287).

Figure 5.10. Whitebark pine cross section collected from the Gravely Range site showing a healed fire scar wound.

Two strategies allow whitebark pine to survive in fire-prone ecosystems: survival of large and refugia trees, and post-fire seedling establishment facilitated by Clark's nutcracker (Tomback *et al*. 1990). Both light surface fires and stand-replacing fires favor whitebark pine in relation to its shade-tolerant competitors (Arno and Brown 1991). The Clark's nutcracker commonly transports seeds several kilometers (Hutchins and Lanner 1982) and these birds prefer open, burned areas for caching seeds (Tomback *et al*. 1990). After establishment, whitebark pine communities have been perpetuated by low-intensity fires that killed understory fir and spruce (Arno and Hoff 1989). Fire exclusion during the 1900s has postponed the natural fire cycle of seral whitebark pine communities, resulting in a decrease of new establishment and replacement by its competitors (Arno 1986).

5.5.3 Factors that Affect Fire Frequency

Several factors may contribute to alterations in fire frequency. First, temperature amplifies the effects of summer drought conditions and the flammability of fuels. Drought conditions can increase the curing of fine and coarse fuels, thereby facilitating the spread of fire (Kipfmueller 2003). In addition, low humidity at high elevation amplifies the effects of temperature and drought on this process by allowing moisture to evaporate more quickly, resulting in drier fuels. Fuels that have been preconditioned due to drought, low-humidity, and temperature will require less heat for ignition. Fuel temperatures will also affect the rate at which a fire will spread. Warmer fuels will ignite and burn faster because less energy is needed to raise the fuels to their ignition point.

Second, precipitation has a direct effect on fuel moisture. The amount of precipitation between snowmelt and fall freeze-up (growing season) would affect the flammability of litter because most growing season precipitation that enters the shallow soil profile at higher elevations is either used by the vegetation or evaporates. Increased fire frequency between elevations 1680 m and 2590 m has been documented in the western United States to correspond with earlier snowpack melt (Westerling *et al*. 2006). During periods of higher temperatures and increased drought conditions, snowpack melt would occur earlier, resulting in drier fuels. During long periods of drought, moisture that is deeper in surface fuels is able to evaporate. As a result, a larger percentage of fuel is available to burn. If this period extends to two or more years, the likelihood that a fire would occur increases.

5.5.4 Temporal Changes in Fire Frequency

Fire intervals during the MWP were short and increased considerably during the LIA. Some periods of changes in fire were observed at the Gravely Range site. During the MWP, four periods of high fire activity occurred that were preceded by fluctuations between wet and dry periods (Figure 5.11). Between 988 and 996, a period of wetness occurred, followed by a period of drought (1002–1009). Three fires occurred following the dry period (1008, 1012, and 1023). A subsequent series of fire events (1066, 1082, and 1083) followed a second period of fluctuating between wetness (1042–1045) and dryness (1053–1058). This was followed by a third series of fire events (1128, 1132, 1139, 1148, and 1153) that was preceded by wetness (1108–1109) and then drought (1231–1256). The largest number of fire events associated with fluctuations between wet and dry periods occurred in the years 1230, 1232, 1233, 1239, 1243, 1246, 1249, and 1251. These events were preceded by a wet period from 1204 to 1230. This period was

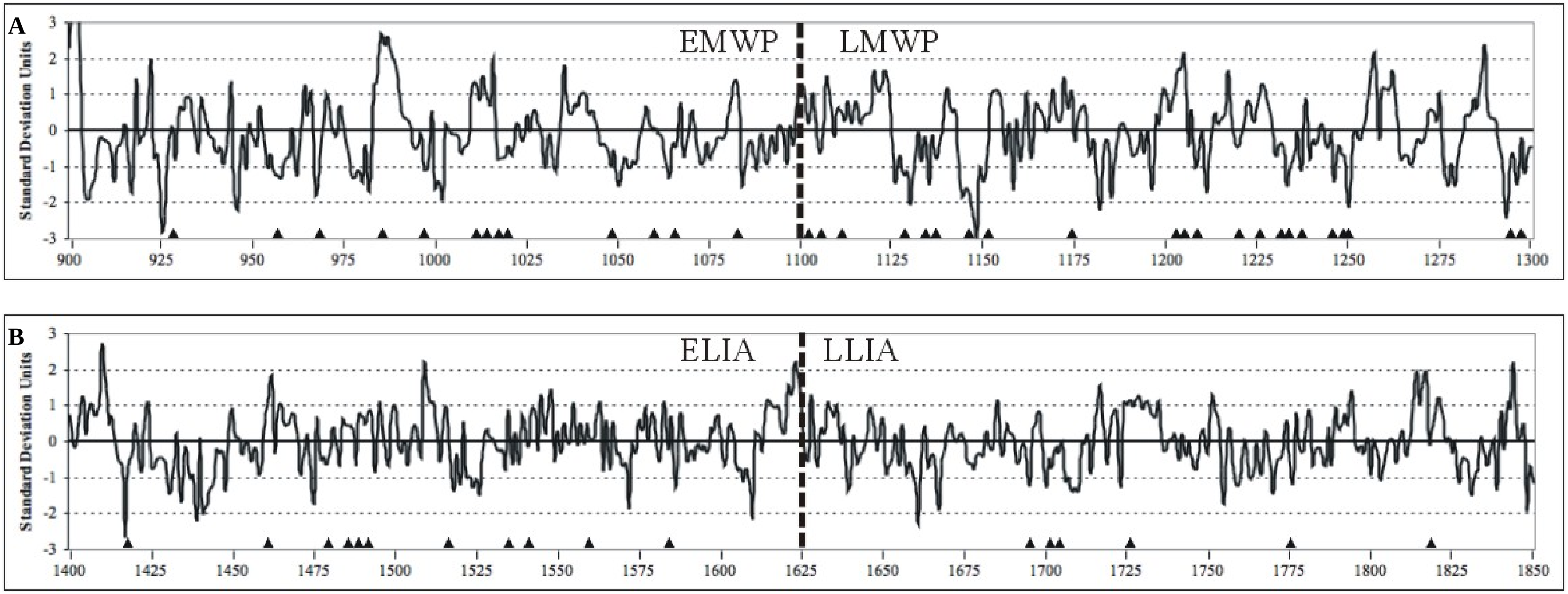


Figure 5.11. Illustration of PDSI and fire occurrence during the (A) Medieval Warm Period and (B) Little Ice Age. Areas above the dark horizontal line represent wet periods and areas below the line represent drought. Fire occurrence is indicated by the black triangles.

followed a drought that extended from 1231 to 1256.

During the LIA, two periods of high fire activity occurred after extended periods of wetness and drought. Between 1468 and 1470, a period of wetness occurred that was followed by drought (1476–1490). Fires associated with these drought conditions occurred in 1481, 1486, and 1488. This was followed by a second series of fire events (1520, 1534, 1542, and 1561) that was preceded by wetness (1512–1517) and drought (1525–1529). During the MWP (900–1300), a total of 33 fires occurred. Throughout the LIA (1400–1850), fire frequency decreased significantly, with 16 fires occurring. During the LIA, more frequent, short-duration periods of less intense drought and wetness occurred than during the MWP.

Higher fire frequencies during the MWP appear to indicate the extensiveness of fires. Fires occurred at the Gravely Range site during dry years that were typically preceded by wet years. The year before a fire event occurred was also typically wetter than the fire year itself. The build-up of fine fuels possibly made the fires more capable of spreading. Additionally, the wet years may have resulted in increased production of fine fuels, making more trees vulnerable to scarring. Because fires during the growing season are hotter than dormant-season fires (Hodgkins 1958), more trees could record the fire because low to moderate intensity fires are more likely to scar trees.

5.5.5 Fire Seasonality

5.5.5.1 Factors that Influence Fire Seasonality

Early-season fires generally occurred over the entire period, but a shift was observed in the seasonality of fires between the MWP and LIA. Fires recorded during the

MWP burned predominantly during the early season. During the LIA, the seasonality of fire alternated between early-season fires and late-season fires. This shift could be indicative of a change in snowpack levels or the initiation of snowpack melt due to climate shifts. Fires that occurred during drought years could indicate earlier snowpack melt and lower fire season precipitation, resulting in drier conditions at the beginning of the growing season. The shift to late-season fire during the early LIA could be indicative of less extensive periods of wet conditions. Due to the drier conditions during this period, it may be possible that lower-elevation fires during this time were able to move upslope. I did not sample lower-elevation trees in the study. However, doing so may lead to a clearer indication of the role lower-elevation fire plays in the fire regimes at upper elevations in the Gravely Range.

5.5.6 Fire/Climate Relationship

Because significant relationships were not evident in the SEA for the entire period, the MWP, or the LIA, interpretations of climatic preconditioning factors may be less reliable. However, the general trends in climate and fire are worthy of examination.

In the American Southwest, fire activity often peaks when several unusually wet years are followed by a drought year (Grissino-Mayer 1995, Grissino-Mayer *et al*. 2004). The wet years allow fine fuels to accumulate and to ignite readily during drought conditions (Grissino-Mayer 1995). In the xeric climate at high-elevation sites in the northern Rockies, dry years (limited snowpack) could exacerbate these conditions and increase the likelihood that a fire would occur. The lack of statistically significant relationships between moisture conditions in the year of fire and fire occurrence at the

Gravely Range site may be related to soil type and depth. Thin, young soils dominate high-elevation sites in the Gravely Range (Figure 5.12). Edaphic conditions at high elevations can temper the effects of climate conditions. At high-elevation sites, even with abundant precipitation, the thin, coarse-textured soil could become extremely dry within a few hours (Beniston *et al*. 1997).

Similarities in fire and climate trends were also evident by visually examining the year of fire occurrence and the long-term climate trends during the MWP and LIA. The highest frequency of fire occurred during the MWP, a period of warmer temperatures and increased duration of wet periods, resulting in higher biomass production. An increased level of biomass production followed by sustained drought would amplify the occurrence of fires at the Gravely Range site. For example, a significant wet period occurred from A.D. 880–924, with no fires occurring during this time. This was followed by an extended period of drought from A.D. 925–980, during which four fires occurred. The LIA was dominated by shorter-duration and less-intense wet periods and periods of drought compared to the MWP. When the climate shifted to the LIA, the shorter-duration and less intense wet periods did not produce enough fuel to sustain the level of fire evident during the MWP.

A statistically significant difference in the MFI was shown between the MWP and LIA. The short mean fire interval of the MWP and the subsequent lengthening during the LIA supports the hypothesis that climate shifts directly influenced the fire regime at high elevations.

Figure 5.12. Photo illustrating the thin soils found in high-elevation meadows (above treeline) in the Gravely Range.

5.6 Conclusions and Future Research

This study documented the fire regime at high-elevation (> 2850 m) sites in the Gravely Range using fire-scar data. Fire scars and other indicators of fire injury are present at the upper treeline of the whitebark pine ecosystem. Snags and stumps of fire-scarred whitebark pine can be seen throughout the stands of the Gravely Range. This study also attempted to disprove the common belief that fire is not a significant factor at the upper treeline of the whitebark pine ecosystem. While fire has been examined at lower and middle elevations (Keane 2001, Larson 2005), investigation of fire at the treeline/alpine tundra border has been limited.

The results of the study show that the whitebark pine ecosystem does experience frequent low- to moderate-severity fires, especially during periods of drought. These low- to moderate-severity fires are important for the survival of whitebark pine and for the regeneration of the species. A lack of fire within the ecosystem would provide an opportunity for competing species to possibly out-compete whitebark pine for resources. Further, fire occurrence at the Gravely Range site has been altered by trends in climate since the beginning of the Medieval Warm Period. Climate during the Little Ice Age shifted the fire regime and resulted in fewer fires. The effects of the LIA on the fire regime may have facilitated an increase in tree density and invasion of sub-canopy tree species.

The investigation of fire frequency at high-elevation sites is imperative to the survival of whitebark pine. While the study presents a limited perspective of fire within the whitebark pine ecosystem, it does, however, provide an initial view of the dynamics of fire over an extended period, and of the effects of shifting climate regimes on fire occurrence. Future research in the Gravely Range should include additional sites at varied elevations, analysis of additional lower-elevation species, and an increased number of fire-scarred whitebark pine samples to amplify the reliability of the statistical significance of fire at high elevations. Future investigation should also include lower-elevation plots to examine the possible connection between fire occurrence at low elevation and its effect on high-elevation fire regimes.

<u>CHAPTER SIX</u>

MAJOR CONCLUSIONS AND FINDINGS

The purpose of this study was to reconstruct climate and assess the effects of climate variability on specific environmental processes in the subalpine ecosystem of western Montana using whitebark pine and subalpine fir. Specifically, this study developed a reconstruction of past climate, reconstructed fire history, and analyized the treeline dynamics of ecosystems dominated by whitebark pine and subalpine fir. Whitebark pine is an essentially unutilized resource for exploring the climate and fire history of high-elevation sites. The research presented in this dissertation provides a basis for future applications of dendroclimatology, spatial analysis, and dendropyrochronology using this tree species. This chapter summarizes the major findings of the research.

6.1. Climate Response and Climate Reconstructions

1. Whitebark pine is an excellent species for dendroclimatological reconstructions because the species is long-lived and subfossil remnants commonly persist on the landscape.

The results of this study demonstrate the feasibility of reconstructing past climate at high-elevation sites in western Montana using whitebark pine. Whitebark pine trees are long-lived (500–1000 years maximum age) and their remnants are well preserved due to both the low humidity at high elevation and the resin content of the pine. Living whitebark pines of considerable age can be found in many areas of the northern Rocky Mountains. Trees in excess of 400 years of age were found at the western Beaverhead, Gravely Range, and Mineral Peak sites. The oldest living tree from all chronologies dated to the year A.D. 728. In addition to the living trees, numerous deadwood remnants of whitebark pine were found at all sites.

2. The whitebark pine samples collected at all three sites produced some of the longest, continuous tree-ring chronologies obtained in the western Unites States.

The final chronologies dated to A.D. 1116 in the western Beaverhead, A.D. 728 in the Gravely Range, and A.D. 981 for Mineral Peak. The chronologies developed for this study filled major gaps in the coverage of whitebark pine chronologies in western Montana and are an excellent complement to the existing whitebark pine chronologies from the region.

3. The correlation between whitebark pine growth and the Palmer Drought Severity Index was the highest among all climate variables analyzed in this study.

The growth response of whitebark pine to drought was much more pronounced than to precipitation and temperature. The fact that the site and composite chronologies illustrated limited sensitivity to both precipitation and temperature suggests that it may be possible to reconstruct climatic indices that incorporate both variables (*i.e.* PDSI). The most significant relationship between whitebark pine growth and the PDSI found in this study occurred in the previous year’s June and July. The PDSI indices were strongly correlated to growth from June to August with July having the highest correlation. This indicates that drought (precipitation and temperature) conditions are much more critical later in the growing season than earlier. This also suggests that drought conditions late in the previous year (August-September) affect bud break and the initiation of growth more than climate during the current year's growing season.

4. Fluctuations in wet and dry periods were evident, but certain periods showed protracted (> 10 years) climate events.

The reconstructed June–July PDSI revealed both interannual and decadal trends for the period A.D. 735–2005. The reconstruction also revealed major decadal trends in past climate. Since A.D. 750, eight periods of protracted extreme drought and five periods of extreme wetness occurred. The most severe extended drought occurred between 1434 and 1462 which had an average PDSI of –2.21 for the 29-year period. Two additional periods of extended drought, A.D. 1145–1167 and A.D. 764–782, lasted 23 and 19 years, respectively. In terms of wetness, the most extreme period and longest in duration occurred between A.D. 894 and 918. Two additional periods of extended wetness, A.D. 790–807 and A.D. 1732–1748, lasted 18 and 17 years, respectively.

5. Comparisons between 20th century droughts and those reconstructed for the preceding 1100 years point to fundamental differences in the duration of drought events in the last 100 years.

Decadal scale dry events in the past tended to persist longer than those experienced in the 20th century. A period of extreme drought in the 20th century (A.D. 1932–1947) lasted 16 years, while protracted drought events before 1900 lasted on average approximately 20 years in length. Moreover, two major dry periods (A.D. 1145–1167 and A.D. 1434–1462) before 1900 lasted for an average of 26 years. The reconstruction also suggests that droughts on the scale of the 20th century drought occurred on average once every 208 years prior to the instrumental period. Before 1900, the severity and extent of decadal scale droughts in western Montana appears more extreme than those in the instrumental period.

6. The reconstruction offers evidence that long-term climate episodes coinciding with the MWP and the LIA did not occur at high-elevation sites in western Montana.

The number of wet and drought years during the MWP and LIA illustrate only marginal differences. During the MWP and LIA the number of drought years is relatively equal, 223 and 234, respectively. The number of wet years showed a negligible difference with 178 wet years during the MWP and 217 during the LIA. The severity of drought during the MWP was more intense than during the LIA with the average duration of drought between the two periods relatively equal. Wetness during the MWP was also more intense than that of the LIA with the average length of wet periods between the two periods also comparatively equal. Whitebark pine ecosystems in western Montana do not show conclusive evidence of climatic shifts that can be equated with the MWP and LIA. No prolonged environmental deviations that correspond to these climatic shifts can be identified within the reconstruction of PDSI, which is basically a moisture availability index.

6.2. Treeline Sensitivity and Response to Environmental Change

1. Species density and composition varies with elevation

Periods of recruitment and treeline movement varied across elevational gradients in western Montana during the period A.D. 1820–2006. In general, dense whitebark-pine-dominated stands have been strongly affected by an understory of subalpine fir at lower and middle-elevations beginning in the late 1800s. High-elevation plots, however, did not exhibit incursions of subalpine fir establishment. This could be an indication of extended snowpack and its effects on seedling survival rates. Also, the effects of wildfire could possibly explain the difference along elevational gradients of subalpine fir understory dominance. The lack of wildfire could possibly allow subalpine fir to establish in the sub-canopy and result in whitebark pine being out-competed for resources, with consequent lowering of recruitment rates at the lower- and mid-elevation plots.

2. The whitebark pine/subalpine fir ecosystem in western Montana has thus far experienced minimal advancement of the upper extent of treeline in response to warming since the LIA.

Vertical movement varied across elevations. The association between climate shifts and the advancement of treeline upslope was negligible. During the period A.D. 1820–2006, only minimal advancement of treeline occurred. Lower-elevation plots showed the earliest movement beginning in the 1850s (25 m), while mid-elevation plots have remained approximately stable since 1840. Upper-elevation plots illustrated the most significant movement of treeline beginning in the 1970s (150 m). Movement of treeline during the mid-1800s at the lower-elevation plots could possibly indicate climatic warming at the end of the LIA affecting growth of competing taxa at lower elevations and, as warming continued into the late 20th century, advanced upslope.

6.3. Historic Fire Regimes

1. Whitebark pine trees produce fire scars that can be used to reconstruct fire history.

This study is the first to use fire scars from whitebark pines collected at upper treeline for dendropyrochronology, and demonstrates that fire history in whitebark pine ecosystems at the treeline-alpine tundra interface can be reconstructed using fire scars. Twenty fire-scarred trees showed the characteristic “catface” appearance and resulted in a fire chronology that spans multiple climatic episodes. Six scarred samples had injuries with healed wounds within the cross-section. Eighteen samples contained multiple fire scars.

2. Even though the number of fire-scarred samples from the Gravely Range was low, the length of the fire chronology was exceptional for fire history in western Montana.

Although the sample size was relatively small, the samples contained well-preserved fire scars. I obtained a total of 26 fire-scarred samples from the Gravely Range. The earliest scars dated to A.D. 795. The length of the fire chronology for the Gravely Range extends from A.D. 795 to 1817. Two existing fire chronologies from western Montana developed by Kipfmueller (2003) and Larson (2005) extend from A.D. 1204–2000 and A.D. 1470–2006, respectively. While the fire chronology from the Gravely Range does not include 20th century fire regimes, additional sampling from the adjacent area should provide increased knowledge of alterations between current fire regimes and those that operated in the past.

3. The statistics of the fire regime at the Graveky Range site indicated a range of historical variation.

Between A.D. 900 and 1850, the Gravely Range experienced a Weibull median fire return interval (MEI) of 11 years. The MWP displayed the shortest MEI (8 years), while the most extended fire-free interval occurred during the LIA (20 years). During the MWP, the longest fire-free interval (10 years) coincided with the EMWP (A.D. 900–1100). In the LIA, the longest fire-free interval (22 years) occurred during the LLIA (A.D. 1626–1850). For the MWP and LIA, the LEI and UEI were 2 and 7 years, respectively. Values for the LEI and UEI were longest during the LLIA (7.0). The maximum fire-free interval (MAX) for the entire period was 111.0 years. Of the periods examined (MWP and LIA), the EMWP displayed the shortest MAX (35 years), while the LLIA showed the longest MAX (49 years). The short return intervals during the MWP emphasizes the significance of fuel availability during this period. If fuels were not consumed in a particular fire year, the following year's production could ensure adequate fine fuels for the site to burn again.

4. Fire frequency was influenced by climate during the Medieval Warm Period and Little Ice Age.

The frequency of fire at the Gravely Range site seemed to be controlled to some degree by moisture conditions related to edaphic factors. Xeric settings appeared to be more conducive to frequent fire because of the accelerated accumulation of fuels that can occur during extended intense wet periods. During periods of drought, these xeric conditions would have been amplified, increasing the probability for fire. Throughout the Medieval Warm Period, extended periods of intense wetness occurred and were followed by intense periods of drought conditions. During this period, fire was more frequent due to the increase in biomass production. During the Little Ice Age, fire frequency was reduced due to limited periods of minimal wetness followed by short-duration droughts of lower intensity. The production of fine fuels during the Medieval Warm Period could have possibly facilitated a higher fire frequency than that of the Little Ice Age.

6.4. Future Studies using Whitebark Pine

While this study demonstrated the usefulness of whitebark pine for research in the climate history, treeline dynamics, and fire history of western Montana, much remains to be answered, and the prospects for future research are encouraging. Four major areas for future study include:

1. extending the length and coverage of existing whitebark pine chronologies,
2. expansion of the spatial area relating climate fluctuations to treeline sensitivity and response,
3. increase in the sample depth of trees at the treeline-alpine tundra interface to provide an increased knowledge of high-elevation fire regimes, and
4. combined research into the effects of climate and fire on the composition of high-elevation tree stands.

While several whitebark pine chronologies are available from the International Tree-Ring Data Bank, these do not cover the full temporal scale (> 1000 years) possible from whitebark pines. Given the presence of well-preserved whitebark pine remnants on the landscape, the potential to create extended chronologies is high. The coverage of current whitebark pine tree-ring chronologies for southwestern Montana is particularly sparse. Further, whitebark pine trees from remote areas of southwestern Montana have yet to be fully exploited. Expansion of the number of chronologies from high-elevation sites in southwestern Montana will increase our knowledge of past climate by improving the spatial resolution of the available data.

Much larger datasets of fire-scarred whitebark pine samples are needed to better characterize the historic frequency and seasonality of fire at extended time scales. Moreover, knowledge concerning fire at extreme elevations in the whitebark pine ecosystem is minimal. Therefore, much work in dendropyrochronology is necessary to increase our understanding of historic fire regimes.

The chronology developed in this study extends into the 8th century, but it is believed that whitebark pine wood that dates to several centuries older may exist in certain high-elevation areas. Remnants from sites located at and above the current treeline could yield samples of substantial age. Due to low humidity and a relatively high level of resin, whitebark pine remnants at these sites should be readily available.

Expansion of the coverage and length of whitebark pine chronologies in southwestern Montana will enable more intensive investigations into past climatic shifts and possibly the effects of ocean-atmospheric teleconnections. Several reconstructions of PDO have been developed in recent years that used whitebark pines, and an increase in the spatial extent of whitebark pine chronologies would enhance our knowledge of the effects of these teleconnections on growth at higher elevations.

Combining research on the whitebark pine ecosystem with the role that climate and fire play in the dynamics of the ecosystem is imperative to helping the survival of this keystone species. The initiation of fire suppression policies has played a role in shaping the whitebark pine ecosystem. The lack of fire over the last 100 years has caused a dramatic shift in the ecosystem by facilitating the encroachment of subalpine fir. Fire is essential for the survival of whitebark pine and increasing our knowledge of the role fire plays in shaping this landscape is paramount.

New advances in technology, such as near infrared spectroscopy, show increased promise in gaining more accurate insight into the growth of trees. Spectroscopy can provide high-resolution information concerning fiber angle, cellular structure, and chemical composition of tree rings. It shows its most promise in dendrochronology by providing chemical information. In the case of fire scars, it may be used to gain information related to fire intensity and alterations in soil composition. By examining the chemical composition of the tree ring prior to a fire occurrence and comparing this to the chemical composition of post-fire tree rings, we may possibly gain insight into fire intensity. In the future, it may also be utilized to measure tree-ring width. However, this is a time-intensive process.

In conclusion, the need to exploit available whitebark pine wood is urgent because of the threats placed upon the species from insects and pathogens. Many sites investigated during this research had already become devastated by mountain pine beetles and blister rust. Additionally, while many remnants are present at high-elevation, these will soon disappear from the landscape, thus eliminating the possibility for extended reconstructions. Therefore, research into climate, stand dynamics, and fire history using whitebark pine is urgent.

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APPENDICES

APPENDIX A1. Statistical descriptions of the series in the Gravely Range total ring-width chronology.

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
1	GR1001	1940 2004	65	.602	.246	.331	–.060	2
2	GR1004	1907 2004	98	.425	.201	.452	.028	3
3	GR1012	1922 1993	72	.357	.278	.540	–.043	1
4	GR1021	1910 2005	96	.412	.280	.498	–.012	1
5	GR1026	1905 2003	99	.347	.268	.359	.003	1
6	GR1029	1934 2001	68	.410	.311	.398	–.105	1
7	GR2010	1937 2003	67	.385	.225	.597	–.027	2
8	GR3002	1936 2002	67	.408	.159	.538	.093	1
9	GR3004	1939 1996	58	.394	.293	.453	–.054	1
10	GR3005	1930 1941	12	.507	.191	.709	–.003	1
11	GR3006	1934 2003	70	.501	.287	.404	–.047	2
12	GR3007	1923 2005	83	.449	.194	.417	–.029	1
13	GR3010	1949 2003	55	.598	.210	.443	–.045	2
14	GR3012	1952 2005	54	.374	.278	.376	–.029	2
15	GR3015	1936 2002	67	.438	.218	.608	–.058	1
16	GR3018	1930 2003	74	.534	.202	.463	–.044	1
17	GR3020A	1923 1942	20	.417	.157	.541	–.068	1
18	GR3020B	1950 2005	56	.458	.239	.539	.051	1
19	GR3021	1930 1998	69	.340	.204	.542	–.031	1
20	GR3026	1928 2001	74	.345	.317	.494	–.017	2
21	GR3027	1944 1995	52	.443	.226	.578	.005	1
22	GR3030	1942 2003	62	.382	.215	.411	–.089	1
23	GR3031	1955 2005	51	.354	.229	.579	–.022	1
24	GR3032	1929 2002	74	.447	.175	.497	–.092	1
25	GR6001	1972 1995	24	.384	.342	.557	–.024	1
26	GR6002	1975 1991	17	.389	.289	.660	.067	1
27	GR6003	1982 1998	17	.643	.310	.676	.150	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A2. Statistical descriptions of the series in the Ajax Peak total ring-width chronology.

Seq	**Series ID**	**Interval**	**No. years**	**Corr. w/ Master**	**Mean sens**	**Std dev**	**Auto corr**	**AR ( )**
1	AJ002	1292 1898	607	.554	.176	.333	.001	1
2	AJ004	1116 1412	297	.379	.177	.406	–.001	1
3	AJ008	1681 1911	231	.518	.215	.401	.002	3
4	AJ012	1765 1987	223	.470	.231	.361	–.079	1
5	AJ013	1758 1986	229	.418	.214	.415	–.021	3
6	AJ018	1574 1801	228	.388	.134	.416	–.020	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A3. Statistical descriptions of the series in the Slag-a-Melt Lake total ring-width chronology.

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
1	SM001	1718 1931	214	.528	.211	.443	–.018	1
2	SM003	1719 1919	201	.541	.199	.339	–.032	1
3	SM011	1690 1926	237	.473	.186	.421	–.028	1
4	SM014	1696 1929	234	.517	.186	.369	–.024	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A4. Statistical descriptions of the series in the Continental Divide total ring-width chronology.

Seq	Series ID	Interval		No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
1	CDB003B	1902	1942	41	.398	.400	.576	–.047	1
2	CDB004A	1931	2003	73	.502	.150	.468	–.077	2
3	CDB006A	1867	2993	137	.601	.175	.324	–.022	1
4	CDB013A	1852	2003	152	.395	.282	.449	–.015	5
5	CDB025A	1977	2003	27	.414	.285	.663	.062	1
6	CDB025B	1977	2003	27	.639	.338	.556	–.014	1
7	CDB027A	1930	1983	54	.512	.211	.445	.037	1
8	CDB028A	1949	2003	55	.549	.229	.397	.033	1
9	CDB028B	1949	2003	55	.454	.245	.637	–.094	1
10	CDB029B	1952	2003	52	.472	.277	.388	–.057	1
11	CDB041A	1937	2003	67	.494	.297	.382	–.095	1
12	CD1002A	1943	2003	61	.521	.224	.421	–.130	2
13	CD1002B	1930	2003	74	.431	.205	.567	–.025	2
14	CD1005A	1944	2003	60	.597	.267	.615	.030	1
15	CD1008A	1850	2003	154	.499	.196	.535	–.044	1
16	CD1008B	1851	2003	151	.544	.182	.349	–.014	1
17	CD1010A	1900	2003	104	.522	.222	.385	.035	1
18	CD1014A	1881	2003	123	.418	.202	.525	.006	1
19	CD1017A	1880	2003	124	.508	.464	.583	.047	1
20	CD1019A	1847	2003	157	.520	.250	.434	.029	1
21	CD1020A	1887	2003	117	.575	.197	.433	–.022	1
22	CD1023A	1885	2003	119	.571	.235	.583	–.023	1
23	CD1028A	1914	2003	90	.477	.163	.434	.089	1
24	CD1036A	1923	2003	81	.575	.191	.482	–.005	2
25	CD1038A	1884	2003	120	.476	.248	.585	–.080	1
26	CD1039A	1877	2003	127	.486	.246	.379	–.089	1
27	CD1040A	1865	2003	139	.438	.291	.471	.025	2
28	CD1040B	1890	2003	86	.446	.234	.478	–.005	1
29	CD1046A	1947	2003	57	.563	.214	.518	.052	1
30	CD1046B	1947	2003	57	.528	.189	.437	–.084	1
31	CD2001A	1777	2003	227	.467	.251	.380	.020	2
32	CD2007A	1883	2003	116	.491	.279	.430	–.016	1
33	CD2008A	1783	2003	221	.493	.244	.426	–.012	1
34	CD2008B	1783	2003	221	.597	.214	.440	–.044	1
35	CD2009B	1759	2003	243	.457	.208	.368	–.024	1
36	CD2013A	1803	2003	201	.541	.179	.374	–.010	1
37	CD2013B	1803	2003	201	.463	.285	.462	–.022	2
38	CD2026A	1949	2003	55	.461	.260	.375	–.012	1
39	CD2026B	1949	2003	55	.555	.249	.446	.066	1
40	CD2031A	1899	2003	105	.426	.225	.505	–.027	1
41	CD2032A	1787	2003	217	.532	.172	.451	–.007	1
42	CD2032A	1769	2003	235	.478	.188	.469	.007	1

APPENDIX A4. *continued*

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
43	CD2034A	1802 2003	202	.623	.200	.396	–.026	1
44	CD2039A	1938 2003	66	.705	.249	.484	–.022	1
45	CD2039B	1938 2003	66	.595	.264	.498	.002	1
46	CD2042A	1917 2003	85	.611	.311	.542	.049	1
47	CD2043A	1819 2003	185	.638	.201	.358	–.005	1
48	CD2046A	1937 2003	67	.659	.300	.559	–.016	1
49	CD2047A	1929 2003	75	.672	.197	.475	–.027	1
50	CD2050B	1945 2003	58	.502	.402	.615	–.118	2
51	CD2056A	1903 2003	100	.641	.241	.523	–.049	1
52	CD2057A	1895 2003	109	.541	.223	.513	–.027	2
53	CD2057B	1895 2003	109	.444	.234	.388	–.039	2
54	CD2058A	1846 2003	158	.382	.217	.420	–.033	2
55	CD2061A	1909 2003	95	.465	.225	.603	.011	1
56	CD2062A	1875 2003	129	.622	.219	.364	.000	1
57	CD2062B	1875 2003	129	.528	.225	.366	–.032	1
58	CD2063A	1845 2003	159	.572	.233	.340	–.021	1
59	CD2066A	1834 2003	170	.491	.279	.412	–.026	1
60	CD2067A	1886 2003	118	.581	.240	.462	–.016	1
61	CD2067B	1886 2003	118	.453	.211	.507	–.025	1
62	CD2068A	1817 2003	187	.584	.260	.449	–.047	1
63	CD2068B	1817 2003	187	.520	.251	.410	–.082	1
64	CD2070A	1891 2003	110	.472	.298	.397	–.023	1
65	CD2071A	1899 2003	105	.478	.279	.430	–.010	2
66	CD2071B	1899 2003	105	.554	.251	.385	–.066	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A5. Statistical descriptions of the series in the Twin Lakes total ring-width chronology.

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
1	TLB001A	1874 2003	130	.492	.163	.401	.030	1
2	TLB002B	1876 2003	128	.458	.192	.422	.049	1
3	TLB003A	1896 2003	108	.730	.233	.435	–.095	1
4	TLB003B	1896 2003	108	.592	.240	.367	–.101	1
5	TLB004A	1876 2003	128	.468	.213	.391	–.051	1
6	TLB004B	1876 2003	128	.462	.252	.259	–.028	1
7	TLB006A	1973 2003	31	.683	.238	.532	–.007	1
8	TLB008A	1958 2003	46	.551	.264	.627	–.088	1
9	TLB008B	1958 2003	46	.650	.246	.689	–.017	1
10	TLB009A	1950 2003	54	.622	.189	.496	–.053	2
11	TLB014A	1943 2003	61	.557	.277	.394	–.099	1
12	TLB017A	1985 2003	19	.334	.330	.607	.018	1
13	TLB017B	1916 2003	88	.646	.270	.419	–.050	1
14	TLB072A	1897 2003	107	.554	.158	.467	–.050	1
15	TLB076A	1917 2003	87	.604	.164	.517	–.017	1
16	TLU004A	1924 2003	80	.355	.221	.488	.025	1
17	TLU004B	1924 2003	80	.494	.190	.475	–.035	1
18	TLU005B	1968 2003	36	.642	.355	.424	–.030	1
19	TLU006B	1958 2003	46	.577	.215	.551	–.025	1
20	TLU007A	1936 2003	68	.514	.175	.617	.053	1
21	TLU007B	1936 2003	68	.386	.184	.526	.101	1
22	TLU008A	1924 2003	80	.518	.161	.555	.045	1
23	TLU008B	1924 2003	80	.523	.198	.417	–.062	1
24	TLU009B	1953 2003	51	.417	.175	.487	–.029	1
25	TLU010A	1930 2003	73	.452	.153	.525	–.058	1
26	TLU010B	1930 2003	74	.533	.155	.592	–.087	2
27	TLU012A	1937 2003	67	.548	.195	.482	–.036	1
28	TLU013A	1957 1984	28	.699	.258	.538	–.119	2
29	TLU018A	1935 2003	69	.703	.256	.500	–.046	1
30	TLU018B	1935 2003	69	.685	.258	.512	.020	1
31	TLU020A	1924 2003	80	.740	.252	.495	–.072	1
32	TLU020B	1924 2003	80	.657	.212	.367	–.076	1
33	TLU021B	1875 2003	129	.556	.221	.334	.023	2
34	TLU022A	1948 2003	56	.555	.195	.384	.004	1
35	TLU022B	1948 2003	56	.679	.236	.498	–.091	1
36	TLU024A	1934 2003	70	.660	.239	.354	–.006	2
37	TLU024B	1933 2003	71	.597	.240	.465	–.046	1
38	TLU025A	1938 2003	66	.497	.317	.455	–.005	1
39	TLU026A	1921 2003	83	.647	.236	.422	–.023	1
40	TLU026B	1927 2003	83	.640	.265	.480	.028	1
41	TLU028A	1940 2003	64	.548	.263	.386	–.060	1
42	TLU029A	1947 2003	57	.649	.258	.423	–.070	1

APPENDIX A5. *Continued*

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
43	TLU031A	1889 2003	115	.633	.216	.427	–.004	1
44	TLU031B	1889 2003	115	.624	.217	.369	–.002	1
45	TLU036A	1943 2003	61	.504	.321	.447	–.015	1
46	TLU036B	1943 2003	61	.568	.287	.710	–.031	1
47	TLU037B	1901 2003	103	.507	.271	.364	–.054	1
48	TLU038A	1932 1995	64	.480	.364	.373	–.082	1
49	TLU040A	1913 2003	91	.553	.203	.511	–.008	1
50	TLU040B	1916 2003	88	.473	.203	.507	–.019	1
51	TLU041B	1908 2002	95	.518	.224	.592	–.008	1
52	TLU043A	1926 2003	78	.557	.322	.478	–.032	1
53	TLU043B	1926 2003	78	.494	.312	.528	–.057	1
54	TLU044A	1952 2003	52	.474	.243	.518	–.034	1
55	TLU044B	1952 2003	52	.552	.268	.428	–.008	1
56	TLU048A	1938 2003	66	.522	.165	.415	–.027	1
57	TLU048B	1938 2003	66	.62	.164	.499	–.037	1
58	TL003A	1751 1990	240	.397	.160	.615	.010	1
59	TL004A	1663 1958	296	.409	.154	.386	–.004	2
60	TL009A	1705 1914	210	.348	.145	.409	–.028	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A6. Statistical descriptions of the series in the Point Six total ring-width chronology.

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
1	PS004	1343 1640	298	.333	.257	.366	–.010	1
2	PS009	1744 1906	163	.360	.226	.289	–.048	1
3	S02016	1864 2003	140	.363	.247	.420	.006	1
4	PS2018	1886 2003	120	.342	.289	.562	–.031	1
5	PS3006	1960 1996	37	.356	.268	.593	–.033	1
6	PS3012	1900 2005	106	.353	.206	.395	–.032	1
7	PS3013	1962 2002	41	.363	.312	.557	–.059	1
8	PS3018	1871 1995	125	.328	.221	.373	–.075	1
9	PS4003	1873 1994	122	.341	.263	.477	–.019	1
10	PS4010	1915 2005	91	.330	.216	.402	.005	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A7. Statistical descriptions of the series in the Mineral Peak total ring-width chronology.

Seq	Series ID	Interval	No. years	Corr. w/ Master	Mean sens	Std dev	Auto corr	AR ( )
1	MPK001A	1300 1562	263	.424	.217	.407	–.086	1
2	MPK007A	1777 1850	74	.350	.124	.457	.085	1
3	MPK008A	981 1289	309	.455	.208	.498	–.069	1
4	MPK009A	948 1105	158	.377	.168	.481	.028	1
5	MPK011BA	1139 1500	362	.373	.192	.380	–.032	3
6	MPK013B	1722 2002	281	.356	.196	.402	–.036	2
7	MPK016A	1329 1768	440	.385	.204	.343	.006	1
8	MPK016B	1206 1609	404	.372	.199	.455	.009	1
9	MPK020A	1333 1749	417	.364	.175	.368	–.020	1
10	MPK1001A	1940 2003	64	.499	.230	.461	–.041	1
11	MPK1001B	1974 2003	30	.675	.251	.635	–.010	1
12	MPK1004B	1878 2002	125	.438	.171	.396	–.022	1
13	MPK1010A	1849 2003	155	.358	.205	.438	.049	1
14	MPK1010B	1849 2003	155	.440	.240	.410	–.034	2
15	MPK1021A	1901 2003	103	.460	.243	.432	.000	4
16	MPK1023A	1953 2003	51	.472	.201	.605	.098	1
17	MPK1076A	1924 2003	80	.419	.294	.517	–.075	2
18	MPK1090B	1945 1989	45	.491	.201	.687	.145	2

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX A8. Statistical descriptions of the series in the Gravely Range Fire total ring-width chronology.

Seq	**Series ID**	**Interval**	**No. years**	**Corr. w/ Master**	**Mean sens**	**Std dev**	**Auto dev**	**AR ( )**
1	GR001F	1326 1652	327	.434	.134	.424	.044	1
2	GR003F	1455 1928	474	.435	.189	.352	–.038	1
3	GR004F	1182 1615	434	.411	.143	.372	–.015	1
4	GR005F	1375 1591	217	.340	.162	.329	–.011	1
5	GR006F	844 1082	239	.426	.194	.388	.007	1
6	GR007F	1170 1354	185	.350	.26	.385	.006	1
7	GR008F	1407 1592	186	.377	.213	.431	.017	2
8	GR009F	976 1158	183	.357	.276	.394	.004	3
9	GR010F	1705 1878	174	.466	.173	.383	–.036	1
10	GR011AF	764 1138	375	.416	.168	.311	–.024	1
11	GR011BF	946 1328	383	.448	.210	.370	–.026	1
12	GR012F	935 1123	189	.348	.187	.383	.020	1
13	GR013F	1108 1407	300	.375	.201	.464	–.025	1
14	GR014F	982 1348	367	.335	.214	.391	–.007	1
15	GR015F	980 1290	311	.472	.232	.393	–.050	1
16	GR017F	382 1235	311	.533	.211	.402	–.004	1
17	GR020F	742 1014	273	.340	.190	.320	–.011	1
18	GR022F	728 1070	343	.369	.216	.329	–.023	1
19	GR023F	1510 1750	241	.400	.185	.391	.022	1
20	GR024F	925 1291	367	.368	.183	.392	–.008	1
21	GR025F	957 1276	320	.482	.157	.370	–.023	2
22	GR026F	765 1054	290	.338	.244	.421	–.035	2
23	GR030F	1655 1875	221	.363	.170	.424	.007	1
24	GR031F	995 1219	225	.395	.201	.450	.003	1
25	GR032F	1455 1895	441	.367	.208	.298	–.001	1
26	GR033F	904 1142	239	.408	.182	.424	–.036	1

Correlation with master (Corr. w/master) is the Pearson correlation coefficient of each individual series to the master chronology creates with the remaining series. Mean sensitivity (Mean sens) is the relative difference in width from one ring to the next and is calculated by averaging the percent change from ring to adjacent ring. Standard deviation (Std dev.) is the variability about the mean which has been standardized to 0.0 for all series. AR ( ) represents the order of the autoregressive model used in the detrending. Higher values are indicative of persistence over several years and the normal range for AR coefficients is between 1 and 4 (Grissino-Mayer 2001).

APPENDIX B. Standard tree-ring chronology for western Montana.

Year	0	1	2	3	4	5	6	7	8	9
720								815	947	801
730	773	1115	875	1459	1611	1283	1080	1167	955	1021
740	1071	1037	834	893	878	1158	1138	923	873	972
750	1037	995	1262	1102	1123	737	379	822	918	1159
760	824	775	786	634	759	424	543	774	796	662
770	990	1090	1305	1013	928	1017	966	864	825	1087
780	1417	1625	1257	1450	1276	1125	1038	1052	1093	1174
790	1196	947	848	941	969	884	852	978	1013	1104
800	1249	1120	902	785	1129	942	1170	681	844	907
810	987	847	863	971	1054	1235	1029	1306	1507	1440
820	1086	931	916	844	1150	1222	1158	1003	897	1038
830	704	664	557	687	648	1051	785	455	897	1235
840	1459	1410	1280	1385	1383	1154	1034	877	902	959
850	1195	1224	1063	1019	862	997	989	520	858	925
860	873	1042	1081	549	821	789	833	851	933	712
870	877	870	830	1208	929	801	948	911	709	886
880	984	1045	1167	1076	1077	1166	1362	1009	1109	1189
890	1347	1133	1329	1200	973	987	759	1297	1641	1318
900	1698	1678	1263	728	696	749	790	952	947	936
910	925	772	915	924	1013	886	736	1191	939	962
920	1007	1279	866	888	569	632	942	998	861	1083
930	1066	1121	1114	995	897	1109	1039	960	952	894
940	918	946	927	1181	726	663	964	823	1007	954
950	925	1089	950	825	890	802	801	781	865	876
960	995	795	957	1171	1041	1130	720	772	971	1126
970	1088	978	967	1089	998	907	796	791	824	777
980	925	737	1174	1179	1374	1331	1369	1297	1252	1132
990	1047	1030	1002	920	912	969	828	826	1068	751
1000	777	702	1004	985	966	960	889	919	937	1163
1010	1181	1103	1210	1123	1095	1270	873	871	876	924
1020	887	1018	995	937	1048	974	1069	1012	988	830
1030	972	856	821	979	1255	1109	1048	1093	1081	1138
1040	1133	1046	1063	1009	922	931	912	832	906	754
1050	836	873	825	862	843	963	1039	1082	995	996
1060	967	962	853	787	932	916	1098	882	951	1066
1070	1018	893	841	937	963	902	954	952	954	1046
1080	1104	1193	1111	768	861	942	875	828	969	876
1090	843	968	889	1006	993	843	1006	921	1120	1172
1100	1144	1012	1140	1055	889	978	1199	1119	957	1072
1110	1067	1024	1102	1012	1104	1109	1030	1056	1060	1234
1120	1143	1171	1232	1176	900	766	883	800	817	678
1130	842	978	856	958	761	967	868	910	999	1160
1140	1062	1073	888	715	747	730	649	552	824	771
1150	896	1114	1146	1145	1116	884	1017	741	1035	854
1160	1032	1132	868	1097	1083	1056	976	1015	1154	1026

Year	0	1	2	3	4	5	6	7	8	9
1170	1027	1202	1069	1146	939	1056	1086	108	888	949
1180	833	652	935	936	705	899	998	1018	980	919
1190	876	894	979	887	974	739	1051	1006	1147	1119
1200	1057	1126	1240	1235	1294	922	1069	840	928	970
1210	727	940	1055	1011	1034	1058	1232	997	922	921
1220	986	1103	1102	1016	1098	1178	1155	1081	979	878
1230	931	911	753	844	886	975	817	1114	869	945
1240	927	1012	1000	912	966	769	913	883	884	664
1250	941	987	946	1004	1117	1213	1292	946	1156	1153
1260	1156	1234	1074	857	877	880	838	925	948	935
1270	992	1064	1067	1022	1130	810	755	849	753	909
1280	978	1050	1043	1117	1074	1188	1336	1066	1031	1025
1290	942	859	623	867	895	758	954	805	909	916
1300	775	904	1168	1109	1166	1126	1015	1235	1149	1026
1310	1067	1176	1122	924	1060	933	853	948	985	1042
1320	1024	1018	1097	1002	1169	1057	962	1173	1041	931
1330	1010	994	929	848	847	800	1003	1086	1031	916
1340	781	981	864	843	877	866	904	957	849	866
1350	1050	904	1047	849	1016	1101	1000	1211	1101	1056
1360	1097	1029	946	911	1060	937	878	1151	1066	1086
1370	935	1111	868	888	919	969	992	1002	1038	980
1380	1065	919	1068	989	884	1043	1118	1098	911	976
1390	1098	1015	802	1015	992	939	960	976	694	1091
1400	962	965	1063	1169	1008	1137	1128	1089	1134	1390
1410	1199	1056	1087	975	905	884	589	885	919	1062
1420	1006	852	1058	1140	871	871	918	913	905	839
1430	763	886	1010	732	894	960	828	850	652	1001
1440	694	715	782	804	952	932	842	776	1088	1120
1450	1002	997	928	984	963	901	958	843	858	1085
1460	1138	1247	930	1104	1137	1053	1052	1086	1098	1013
1470	845	944	1014	918	725	1085	904	921	886	941
1480	1041	958	885	1074	1050	1049	1030	883	1100	1082
1490	1057	1117	1082	856	1139	1060	976	890	1100	1132
1500	980	930	989	884	1012	1054	1015	926	1305	1177
1510	1137	1147	1009	897	1122	1200	939	776	930	802
1520	1068	931	937	793	828	762	940	1033	1007	994
1530	960	943	948	867	1115	885	1019	887	1086	1050
1540	960	1118	1109	945	1175	1016	1091	1185	901	1026
1550	934	1060	974	986	1143	966	1060	970	1061	993
1560	991	1084	1135	862	1069	866	1063	926	939	882
1570	896	704	967	934	1074	1025	1124	983	1103	1049
1580	899	1017	1150	948	1085	801	930	1103	968	900
1590	949	905	971	952	964	936	1041	1021	1092	1031
1600	887	937	1052	1009	869	872	816	872	799	665
1610	1015	1011	948	1146	1155	1133	1119	1130	1015	1020
1620	1209	1176	1311	1235	1122	939	908	1174	942	876
1630	1040	1017	1135	1109	1083	1126	970	1046	875	817

Year	0	1	2	3	4	5	6	7	8	9
1640	1009	898	929	936	998	1122	1068	946	983	1080
1650	857	892	914	1014	884	1053	853	898	872	801
1660	651	980	1051	882	932	916	701	896	945	1032
1670	1059	980	986	1018	986	900	894	867	928	930
1680	921	1024	950	951	1152	1009	921	1051	1024	1012
1690	892	901	977	936	798	1040	1050	1108	989	792
1700	916	871	960	931	979	815	840	776	792	778
1710	977	971	1004	848	1059	1127	1210	1004	888	1038
1720	1107	1115	975	1147	1130	1153	1127	1161	1171	1115
1730	1148	1129	1090	1082	1142	945	945	994	993	1013
1740	905	1018	921	986	919	865	918	1032	982	936
1750	1174	1122	1074	863	732	1036	1034	860	796	827
1760	985	920	827	849	975	845	955	1028	985	767
1770	910	910	925	1069	909	817	1099	971	939	1007
1780	1115	891	944	961	926	890	1044	8885	1118	1040
1790	1053	1099	1058	1194	1004	945	965	932	998	841
1800	983	938	823	892	967	994	988	893	1016	1026
1810	922	1001	1125	1277	1135	1213	1267	977	1060	1112
1820	1164	1094	982	997	933	977	812	912	910	892
1830	760	849	876	1002	931	884	959	986	784	1107
1840	948	1147	1114	13609	993	993	1052	697	878	813
1850	953	927	982	1002	1040	1098	915	1067	996	1017
1860	988	1017	754	985	914	1048	1059	917	944	941
1870	996	926	815	922	1000	994	1010	1106	1160	1118
1880	894	1067	1103	777	904	1084	1088	922	1187	996
1890	939	1065	1051	1023	1087	1146	974	956	1084	713
1900	1119	1039	1020	1073	1210	923	874	1001	1084	978
1910	1193	1018	1054	1070	1118	985	977	828	782	947
1920	701	835	818	873	813	833	987	849	983	888
1930	785	931	822	888	861	955	930	9636	980	1004
1940	862	976	1039	987	1111	1048	966	1086	1133	1059
1950	947	1031	1140	829	933	1055	791	989	909	982
1960	999	939	956	1117	946	1096	1132	990	945	1110
1970	1128	960	788	960	790	935	1132	1218	1179	1094
1980	1101	1156	847	1042	1159	992	1030	1066	1025	748
1990	948	918	907	955	1224	785	990	950	1203	914
2000	878	824	680	714	892	920				

These values are the tree-ring indices for each year in the chronology. The indices are displayed without the decimal points, but the actual value can be obtained by dividing the numbers by 100. The mean value for all indices is 1.0. Each line represents one decade of indices and the decades are shown in the lefthand column. The numbers across the top of the table are the last numbers of the decade year for each decade. This is called the "Tucson format" and is the internationally accepted format of the World Data Center for Paleoclimatology.

APPENDIX C. Previous June-July reconstructed PDSI for western Montana.

Year	0	1	2	3	4	5	6	7	8	9
720									–2.681	–0.746
730	–2.880	–3.292	1.713	–1.974	6.745	8.969	4.168	1.202	2.474	–0.630
740	0.337	1.062	0.574	–2.398	–1.533	–1.760	2.342	2.048	–1.096	–1.832
750	–0.382	0.563	–0.048	3.681	1.520	1.833	–3.820	–9.050	–2.572	–1.170
760	2.357	–2.542	–3.267	–3.095	–5.328	–3.500	–8.398	–6.664	–3.283	–2.961
770	–4.915	–0.123	1.347	4.487	0.223	–1.020	0.282	–0.472	–1.959	–2.534
780	1.305	6.131	9.180	3.789	6.608	4.063	1.853	0.580	0.788	1.385
790	2.573	2.894	–0.746	–2.201	–0.834	–0.423	–1.674	–2.138	–0.287	0.211
800	1.552	3.669	1.779	–1.409	–3.115	1.918	–0.818	2.516	–4.644	–2.256
810	–1.337	–0.164	–2.209	–1.975	–0.395	0.822	3.460	0.447	4.508	7.448
820	6.446	1.286	–0.974	–1.200	–2.252	2.218	3.277	2.345	0.072	–1.481
830	0.579	–4.306	–4.883	–6.453	–4.453	–5.120	0.779	–3.122	–7.946	–1.484
840	3.472	6.751	6.033	4.130	5.569	5.636	2.286	0.530	–1.767	–1.400
850	–0.575	2.876	3.305	0.943	0.311	–1.996	–0.019	–0.139	–6.989	–2.054
860	–1.066	–1.835	0.639	1.218	–6.566	–2.585	–3.060	–2.408	–2.146	–0.952
870	–4.192	–1.776	–1.867	–1.141	3.065	–1.009	–2.880	–0.730	–1.277	–4.230
880	–1.640	–0.020	0.688	2.469	1.147	1.161	2.459	5.323	0.162	1.629
890	2.788	5.112	1.973	4.846	2.962	–0.364	–0.167	–3.493	4.373	9.413
900	4.688	10.237	9.947	3.871	–3.957	–4.413	–3.693	–3.039	–0.667	–0.745
910	–0.904	–1.066	–3.310	–1.219	–1.080	0.216	–1.642	–3.835	2.822	–0.870
920	–0.519	0.135	4.110	–1.933	–1.608	–6.279	–5.350	–0.822	–0.007	–2.000
930	1.249	0.994	1.802	1.692	–0.041	–1.478	1.628	0.592	–0.551	–0.667
940	–1.516	–1.167	–2.223	–1.043	2.680	–3.979	–4.903	–0.496	–2.563	0.128
950	–0.649	–1.065	1.325	–0.696	–2.536	–1.586	–2.863	–2.887	–3.174	–1.949
960	–1.782	–0.048	–2.972	–0.603	2.528	0.624	1.931	–4.069	–3.300	–0.393
970	1.869	1.318	–0.288	–0.455	1.338	0.007	–1.336	–2.961	–3.036	–2.552
980	–3.237	–1.064	–3.823	2.576	2.643	5.501	4.874	5.424	4.380	3.722
990	1.953	0.722	0.470	0.056	–1.146	–1.252	–0.431	–2.490	–2.518	1.028
1000	–3.610	–3.240	–4.333	0.089	–0.191	–0.467	–0.551	–1.598	–1.156	–0.887
1010	2.406	2.680	1.534	3.096	1.830	1.420	3.981	–1.826	–1.863	–1.786
1020	–1.084	–1.627	0.285	–0.048	–0.889	0.735	–0.349	1.037	0.198	–0.146
1030	–2.465	–0.388	–2.078	–2.584	–0.284	3.763	1.629	0.735	1.392	1.217
1040	2.048	1.969	0.709	0.949	0.156	–1.107	–0.983	–1.260	–2.432	–1.345
1050	–3.569	–2.366	–1.827	–2.533	–1.994	–2.275	–0.510	0.611	1.227	–0.038
1060	–0.033	–0.448	–0.524	–2.127	–3.080	–.0960	–1.198	1.467	–1.699	–0.685
1070	0.994	0.296	–1.539	–2.294	–0.892	–1.972	–1.402	–0.638	–0.671	–0.639
1080	0.706	1.556	2.851	1.653	–3.365	–2.012	–0.818	–1.797	–2.493	–0.431
1090	–1.781	–2.289	–0.440	–1.589	0.116	–0.073	–2.271	0.116	–1.131	1.786
1100	2.545	2.136	0.200	2.083	0.836	–1.590	–0.297	2.935	1.771	–0.597
1110	1.076	1.009	0.374	1.526	0.207	1.546	0.156	0.472	0.852	0.902
1120	3.445	2.118	2.523	3.429	2.601	–1.431	–3.399	–1.675	–2.904	–2.644
1130	–4.684	–2.284	–0.291	–2.084	–0.585	–3.474	–0.452	–1.896	–0.851	0.008
1140	2.365	0.941	1.100	–1.165	–4.137	–3.676	–3.927	–5.107	–6.531	–2.544
1150	–3.326	–1.494	1.701	2.169	2.152	1.726	–1.662	0.279	–3.768	0.538
1160	–2.108	0.491	1.960	–1.902	1.445	1.245	0.845	–0.320	0.244	2.286
1170	0.409	0.418	2.986	1.037	2.168	–0.870	0.842	1.291	1.252	–1.609
1180	–0.724	–2.411	–5.063	–0.920	–0.908	–4.280	–1.450	–0.001	0.298	–0.259

Year	0	1	2	3	4	5	6	7	8	9
1190	–1.157	–1.789	–1.519	–0.275	–1.618	–0.355	–3.782	0.779	0.116	2.181
1200	1.771	0.868	1.869	3.539	3.460	4.325	–1.110	1.032	–2.312	–1.020
1210	–0.407	–3.965	–0.846	0.830	0.182	0.532	0.874	3.429	–0.020	–1.118
1220	–1.132	–0.181	1.536	1.151	0.267	1.467	2.629	2.294	1.209	–0.284
1230	–1.754	–0.977	–1.273	–3.583	–2.247	–1.634	–0.333	–2.654	1.692	–1.890
1240	–0.769	–1.033	0.204	0.031	–1.264	–0.472	–3.357	–1.239	–1.683	–1.674
1250	–4.883	–0.835	–0.165	–0.759	0.091	1.741	3.151	4.303	–0.758	2.310
1260	2.266	2.316	3.459	1.114	–2.057	–1.770	–1.721	–2.334	–1.075	–0.733
1270	–0.922	–0.090	0.972	1.013	0.352	1.932	–2.743	–3.554	–2.180	–3.578
1280	–1.302	–0.289	0.788	0.6546	1.746	1.106	2.785	4.951	0.988	0.478
1290	0.400	–0.825	–2.038	–5.493	–1.920	–1.503	–3.510	–0.648	–2.827	–1.306
1300	–1.201	–3.264	–1.380	2.484	1.629	2.450	1.877	0.249	3.461	2.216
1310	0.403	1.010	2.605	1.818	–1.084	0.904	–0.944	–2.127	–0.733	–0.190
1320	0.643	0.384	0.295	1.445	0.055	2.499	0.858	–0.522	2.560	0.630
1330	–2.448	0.173	–0.064	–1.007	–2.196	–2.206	–2.897	0.077	1.280	0.484
1340	–1.203	–3.177	–0.254	–1.964	–2.274	–1.769	–1.927	–1.383	–0.600	–2.184
1350	–1.924	0.757	–1.368	0.719	–2.184	0.255	1.509	0.024	3.119	1.501
1360	0.842	1.449	0.454	–0.329	–1.276	0.902	–0.887	–1.757	2.245	0.993
1370	1.293	–0.919	1.648	–1.898	–1.615	–1.154	–0.428	–0.086	0.059	0.585
1380	–0.262	0.975	–1.154	1.031	–0.136	–1.661	0.652	1.763	1.466	–1.277
1390	–0.326	1.435	0.247	–2.872	0.249	–0.092	–0.867	–0.563	–0.325	–4.450
1400	1.365	–0.521	–0.481	0.952	2.502	0.146	2.033	1.894	1.328	1.982
1410	5.737	2.946	0.852	1.294	–0.338	–1.356	–1.675	–5.979	–1.655	–1.162
1420	0.942	0.122	–2.136	0.880	2.079	–1.861	–1.862	–1.165	–1.241	–1.366
1430	–2.332	–3.436	–1.636	0.168	–3.885	–1.516	–0.562	–2.486	–2.165	–5.057
1440	0.046	–4.448	–4.139	–3.163	–2.843	–0.674	–0.971	–0.825	–3.247	1.321
1450	1.790	0.053	–0.009	–1.018	–0.209	–0.506	–1.421	–0.589	–2.272	–2.044
1460	1.268	2.044	3.646	–0.993	1.558	2.041	0.804	0.794	1.280	1.457
1470	0.213	–2.233	–0.796	0.230	–1.169	–3.996	1.268	–1.375	–1.132	–1.643
1480	–0.841	0.630	–0.591	–1.658	1.106	0.762	0.750	0.465	–1.680	1.495
1490	1.234	0.866	1.742	1.227	–2.079	2.064	0.909	–0.329	–1.583	1.495
1500	1.956	–0.268	–0.993	–0.133	–1.665	0.210	0.823	0.247	–1.058	4.487
1510	2.622	2.033	2.174	0.165	–1.485	1.820	1.779	–0.865	–3.244	–0.993
1520	–2.871	1.028	–2.448	–2.363	–2.997	–2.486	–3.486	–0.846	0.514	0.127
1530	–0.055	–0.551	–0.803	–0.736	–1.192	1.704	–1.649	0.302	–1.627	1.294
1540	0.753	–0.559	1.752	1.619	–0.775	2.591	0.266	1.356	2.740	–1.415
1550	0.419	–0.938	0.901	–0.352	–0.178	2.128	–0.465	0.909	–0.408	0.924
1560	–0.079	–0.102	1.258	2.008	–1.987	1.034	–1.931	0.953	–1.052	–0.870
1570	–1.699	–1.489	–4.306	–0.458	–0.935	1.106	0.397	1.845	–0.225	1.540
1580	0.743	–1.441	0.282	2.216	–0.733	1.278	–2.878	–0.996	1.543	–0.442
1590	–1.435	–0.717	–1.358	–0.393	–0.679	–0.502	–0.903	0.628	0.331	1.378
1600	0.484	–1.631	–0.888	0.789	0.157	–1.884	–1.846	–2.659	–1.846	–2.904
1610	–4.880	0.247	0.189	–0.734	2.159	2.297	1.978	1.768	1.934	0.245
1620	0.326	3.088	2.608	4.574	3.470	1.818	–0.859	–1.318	2.578	–0.816
1630	–1.788	0.608	0.283	2.005	1.619	1.240	1.880	–0.408	0.699	–3.114
1640	–2.651	0.168	–1.467	–1.017	–0.900	–0.000	1.818	1.018	–0.765	–0.215
1650	1.202	2.063	–1.546	–1.229	0.232	–1.674	0.800	–2.123	–1.457	–1.849

Year	0	1	2	3	4	5	6	7	8	9
1660	–2.885	–5.073	–0.260	0.772	–1.691	–0.958	–1.194	–4.349	–1.494	–0.769
1670	0.492	0.896	–0.257	–0.181	0.295	–0.172	–1.438	–1.529	–1.915	–1.028
1680	–0.995	–1.129	0.381	–0.701	–0.685	2.259	0.165	–1.134	0.773	0.374
1690	0.198	–1.555	–1.421	–0.308	–0.906	–2.931	0.620	0.763	1.615	–0.126
1700	–3.008	–1.192	–1.855	–0.557	–0.985	–0.282	–2.672	–2.312	–3.244	–3.013
1710	–3.222	–0.310	–0.396	0.084	–2.192	0.889	1.887	3.097	0.091	–1.602
1720	0.579	1.588	1.711	–2.973	2.180	1.935	2.270	1.880	2.392	2.532
1730	1.719	2.200	1.918	1.350	1.223	2.108	–0.780	–0.771	–0.064	–0.070
1740	0.217	–1.358	0.286	–1.119	–0.177	–1.151	–1.946	–1.170	0.500	–0.235
1750	–0.903	2.569	1.814	1.113	–1.972	–3.892	0.552	0.529	–2.024	–2.950
1760	–2.499	–0.194	–1.146	–2.502	–2.174	–0.339	–2.243	–0.632	0.432	–1.505
1770	–3.374	–1.286	–1.295	–1.072	1.035	–1.304	–2.653	1.473	–0.389	–0.866
1780	0.138	1.714	–1.573	–0.791	–0.546	–1.049	–1.579	0.665	–1.656	1.761
1790	0.617	0.810	1.474	0.877	2.864	0.086	–0.778	–0.487	–0.971	–0.004
1800	–2.293	–0.222	–0.882	–2.562	–1.555	–0.452	–0.061	–0.139	–1.530	0.269
1810	0.413	–1.113	0.037	1.862	4.078	1.998	3.144	3.941	–0.303	0.907
1820	1.667	2.443	1.411	–0.241	–0.019	–0.958	–0.301	–2.726	–1.255	–1.295
1830	–1.557	–3.478	–2.184	–1.785	0.058	–0.985	–1.675	–0.578	–0.169	–3.137
1840	1.602	–0.726	2.174	1.700	4.553	–0.074	–0.069	0.785	–4.398	–1.757
1850	–2.713	–0.660	–1.036	–0.240	0.064	0.615	1.458	–1.208	1.003	–0.029
1860	0.282	–0.149	0.271	–3.564	–0.184	–1.233	0.732	0.896	–1.184	–0.788
1870	–0.835	–0.035	–1.047	–2.670	–1.110	0.031	–0.052	0.179	0.574	2.365
1880	1.755	–1.526	1.010	1.537	–3.229	–1.377	1.252	1.316	–1.105	2.762
1890	–0.036	–0.866	0.986	0.776	0.372	1.300	2.159	–0.349	–0.614	1.258
1900	–4.172	1.763	0.596	0.318	1.091	3.096	–1.090	–1.811	0.048	1.255
1910	–0.297	2.845	0.290	0.825	1.047	1.754	–0.191	–0.308	–2.486	–3.153
1920	–0.745	–4.350	–2.379	–2.638	–1.823	–2.710	–2.413	–0.167	–2.180	–0.218
1930	–1.612	–3.122	–0.980	–2.572	–1.614	–2.004	–0.626	–1.002	–0.913	–0.256
1940	0.090	–1.983	–0.323	0.595	–0.167	1.647	0.735	–0.469	1.290	1.978
1950	0.889	–0.746	0.487	2.074	–2.467	–0.951	0.833	–3.029	–0.136	–1.308
1960	–0.241	0.007	–0.870	–0.614	1.741	–0.755	1.427	1.963	–0.120	–0.777
1970	1.640	1.896	–0.551	–3.074	–0.550	–3.045	–0.920	1.966	3.224	2.651
1980	1.400	1.504	2.310	–2.211	0.636	2.355	–0.085	0.470	0.996	0.390
1990	–3.659	–0.728	–1.166	–2.790	–0.635	3.314	–3.118	–0.115	–0.704	2.996
2000	–1.233	–1.757	–2.553	–4.650	–4.156	–1.555				

These values are the reconstructed previous June-July PDSI for western Montana for each year in the chronology. Negative PDSI values indicate dry conditions, while positive values indicate wet conditions. Each line represents one decade of indices and the decades are shown in the lefthand column. The numbers across the top of the table are the last numbers of the decade year for each decade. The PDSI values are displayed in the internationally accepted format of the World Data Center for Paleoclimatology.

VITA

David F. Mann earned a Bachelor of Science degree from the Department of History at Austin Peay State University with a concentration in Civil War History in 1991. He received the Master of Science degree from the Department of Geography at the University of Tennessee, Knoxville in 2002. While at the University of Tennessee, he served as a Teaching Assistant and taught laboratory classes in introductory physical geography. In 2007 David was commissioned as an officer in the United States Naval Reserve as a Meteorologist and Oceanographer. In 2008, he was awarded the Doctorate of Philosophy degree from the Department of Geography at the University of Tennessee. While working towards the doctorate, he worked as a Graduate Teaching Associate in the Department of Geography and taught lectures in introductory physical geography. After graduation, David continued his work with the Naval Meteorologist and Oceanography Command.

"We are tied to the ocean. And when we go back to the sea, whether it is to sail or to watch - we are going back from whence we came."

John F. Kennedy