FIRE-CLIMATE-VEGETATION INTERACTIONS IN SUBALPINE FORESTS OF
THE SELWAY-BITTERROOT WILDERNESS AREA,
IDAHO AND MONTANA, USA

by

Kurt Foster Kipfmueller

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Signed. [Signature]
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DEDICATION

To my family, especially Kara and Kaylee.

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ABSTRACT

The long term patterns of fire-climate interactions and forest recovery processes in subalpine forests are poorly understood. This study used a suite of dendrochronological techniques to identify tree growth-climate relationships, assess the interactions of fire with interannual climate variability, and reconstruct summer temperature in subalpine forests of the Selway-Bitterroot Wilderness Area on the border of Idaho and Montana, USA.

Comparison of ring-width chronologies from whitebark pine (*Pinus albicaulis* Engelm.) and subalpine larch (*Larix lyallii* Parl.) with modern climate data indicated that summer temperatures were most limiting to growth in these conifers. Warm summers were generally conducive to radial growth. However, the temporal stability of the climate-tree growth relationship weakens from the early to later periods of the record. Alterations to growing season length, possibly modified by snow pack, may be related to the reduction in climate-growth relationships.

A 748-year reconstruction of average summer temperature was developed that explains ≈36% of the variance of the instrumental record. Positive values of the coefficient of efficiency and reduction of error verification statistics indicated that the reconstruction was of good quality. Warm and cool periods in the reconstruction include a warm decade around the 1650s and prolonged cooling around 1700. Peaks in variance in reconstructed average summer temperature occurred at 87, 15, and 2 years.

More than 2000 fire scar and age structure samples were used to evaluate fire-climate relationships. Comparison of widespread fire events to climate variables
indicated dry conditions both during the fire year and one year before a fire. Multiple spatial patterns of drought and El Niño were related to widespread fire occurrence.

Forest recovery following fires generally proceeds from lodgepole pine (*Pinus contorta* var. latifolia Dougl.) toward spruce-fir forests (*Picea engelmannii* Parry-Abies *lasiocarpa* (Hook) Nutt.). Two successional pathways occur, one beginning with an initial lodgepole pine stage, the other a spruce-fir stage. Initial composition was related to the presence of overstory lodgepole pine at the time of fire occurrence as well as the intervals between successive fires. Collectively, these results suggest a strong multi-year drought linkage between climate and fire, and dependence on fire intervals for structuring forest communities.
CHAPTER ONE
INTRODUCTION TO THE DISSERTATION

“As accumulation of knowledge continues, we eventually find facts that will not fit properly into any established pigeon-hole.”

H.A. Gleason (1926)

Explanation of the Problem

Fire has long been recognized as a keystone ecological process in forests of the Rocky Mountains (Mutch 1970; Habeck & Mutch 1973; Arno 1980; Romme 1982). Fire’s legacy on the landscape yields a complex mosaic of forest patches in various stages of recovery. Fire occurrence and its associated impacts vary widely, influenced by forest composition, fuel structure, human impacts, and climate variability. Moreover, these factors can influence one another as well. For example, climate patterns exert important controls over vegetation types, that in turn influence fuel amount and arrangement, and hence the characteristics of fire.

Fire is so pervasive in some forest landscapes that some tree species have developed specialized adaptations to cope with its effects (Habeck & Mutch 1973). Ponderosa pine (Pinus ponderosa Laws), a common montane conifer found across a broad region of the west, possesses thick bark to insulate it from the passing flames of frequent low intensity surface fires common in its environment. Lodgepole pine (Pinus contorta Dougl.) and jack pine (Pinus banksiana Lamb.), common subalpine and boreal trees, have evolved serotinous cones that hold their seed until the passage of high intensity fires initiate their opening. This ensures rapid regeneration following the high

The spatiotemporal occurrence of fire is strongly related to variations in climate characteristics, vegetation composition, and the impact of human land uses (Swetnam 1993; Baisan & Swetnam 1997; Veblen et al. 1999). At interannual scales, climatic events such as seasonal drought alter fuel moisture characteristics that promote conditions conducive to fire occurrence and spread (Renkin & Despain 1992; Barrett et al. 1997). Long-term fluctuations in climate such as shifts in temperature or precipitation patterns can alter fire occurrence by changing the type, condition, and abundance of fuels through modifications in vegetation patterns (Clark 1988; Swetnam 1993).

Humans have also left an indelible imprint on fire regimes in some places and at some times through the intentional ignition of fires by Native Americans to improve wildlife forage, clear trails for ease of movement; and for the purpose of waging war (Barrett 1980; Lewis 1985; Denevan 1992; Vale 1998; Kaib 1998). But the effect of humans on fire regimes is arguably most evident over the last ≈150 years. Humans have altered fuel conditions thereby, altering fire regimes, primarily through activities such as livestock grazing (Savage & Swetnam 1990) and the suppression of fires (e.g., Dodge 1972; Arno & Brown 1991; Stephenson et al. 1991; Covington & Moore 1994). Investigations in many parts of the western United States also suggest that fire regimes of the 20th century are different from those prior to EuroAmerican settlement. Prior to settlement, a higher proportion of fires were of low or mixed severity. At present, however, a greater proportion of stand-replacing fires occur where mixed regimes
dominated in the past (Barrett & Arno 1991; Brown et al. 1994). These changes are most often attributed to unnatural fuel accumulations due to reduced Native American burning practices and improved fire suppression efficiency over the 20th century. In contrast, reductions in area burned over the last century in the Canadian Rockies and the boreal forests of northern Canada have been attributed to synoptic scale shifts in air masses resulting in a reduced drought frequency (e.g., Johnson et al. 1990; Johnson & Larsen 1991; Bergeron & Archambault 1993).

Fire-climate relationships have been well studied in some areas of North America such as the Sierra Nevada of California (Swetnam 1993), the Southwestern United States (Baisan & Swetnam 1990; Grissino-Mayer & Swetnam 2000), the Pacific Northwest (Heyerdahl et al. 2002), the Front Range of Colorado (Veblen et al. 2000; Donnegan et al. 2001), and the boreal forests of Canada (Bergeron & Archambault 1993). In addition, fire-climate interactions have been examined in other regions of the world such as Patagonia, Argentina (Veblen et al. 1999) and Siberia (Swetnam 1996).

Although a number of fire history investigations have been conducted in the northern Rocky Mountains region (e.g., Habeck 1972; Romme 1982; Arno & Petersen 1983; Arno & Gruell 1983; Arno 1985; Arno & Gruell 1986; Barrett 1988; Barrett et al. 1997; Barrett 2000), few have explicitly examined the role of climate in driving fire regimes at time scales longer than the fire season (but see Larsen & Delavan 1922). Seasonal drought is clearly an important factor for the occurrence of large fires in the Northern Rockies (Barrett et al. 1997), but the long-term influence of climate on fire regimes has not been well established in these regions. For the most part, investigations
of fire regimes in the northern Rocky Mountains have either not specifically addressed these relationships, or data sets have lacked the precision necessary to examine long-term trends.

Investigations of fire-climate interactions in subalpine forests are also limited. Determining linkages between fire and climate in subalpine forests is hampered by a fire regime consisting of long return intervals (i.e., few events) as well as high severity fires. Severe fires lead to abundant regeneration of lodgepole pine in the Rockies and jack pine in many boreal regions, but often fire scars recording the event are few. Determination of the precise calendar year of a fire event based solely on age-structure evidence is difficult due to the vagaries of regeneration. Regeneration of a burned forest can lag behind fires by years or even decades (e.g., Leiberg 1899; Larsen 1925; Larsen 1929; Stahelin 1943; Habeck 1972; Turner et al. 1999). This greatly limits the temporal precision of fire dates and hinders comparisons with interannual climate (Kipfmuebler & Swetnam 2001). Further, it is often difficult to determine the exact date of tree establishment using increment borer methods due to the difficulty of intersecting the root-shoot interface and problems associated with missing the exact center of a tree (Villalba & Veblen 1997; Gutsell & Johnson 2002). The failure to intersect a tree’s exact center typically requires the use of imperfect methods to determine the number of rings missed. These methods often rely on matching estimations using concentric circles with constant growth rates that may not be a true reflection of the complexity of tree growth.

This dissertation research investigates the interactions of fire and climate in subalpine forests of the Selway-Bitterroot Wilderness Area located in Idaho and Montana
(Figure 1.1). Multiple lines of tree-ring evidence are used to 1) develop crossdated fire histories in four subalpine watersheds; 2) develop chronologies of tree growth at upper elevation sites suitable for reconstructing summer temperature patterns for comparison with fire events; and 3) examine changes in forest composition and structure with increasing time since fire occurrence.

The use of dendrochronological techniques to reconstruct multiple lines of evidence to develop a holistic understanding of subalpine forest ecology is the foundation of this research. Dendrochronology is the science of assigning exact calendar dates to sequences of annually formed growth rings in woody plant species (Douglass 1941; Stokes & Smiley 1968; Fritts 1976). Crossdating, the technique (and principle) of matching patterns of wide and narrow growth rings within and among trees in a given region ensures that precise calendar dates are assigned to individual growth rings by accounting for false or missing rings within a given tree. More importantly, crossdating allows for the incorporation of remnant wood material into the development of tree-growth chronologies through the overlapping of wide and narrow growth patterns from remnant material to the living material used to anchor a chronology in time (Fritts & Swetnam 1989; Kipfmueller & Swetnam 2001). The inclusion of remnant materials in this study is paramount since many fire-scarred trees have been killed by previous fires, mountain pine beetles (*Dendroctonus ponderosae* Hopk.), or blister rust (*Cronartium ribicola* J.C. Fisch.).
Figure 1.1. The location of the Selway-Bitterroot Wilderness Area and the study sites examined in the dissertation.
Research Objectives

The primary objective of this research is to develop a quantitative comparison of the relationship between fire events and interannual climate variability in subalpine forests. While it is intuitive that large fires occur during dry years, little attention has been given to longer term patterns of fire and climate in subalpine forests in these regions. Specifically, this research examines interannual relationships between various climatic parameters and fire events. In addition, I examine the relationships between El Niño-La Niña events, with an expectation that large fires are related to El Niño because these events have been linked to dry conditions in the Pacific Northwest. An understanding of the interannual relationships between fire and climate can be used by land managers in developing and refining fire management plans that seek to maintain fire as a critical component of subalpine ecosystems.

A second objective of this research is to document the structural and compositional patterns of forest recovery in subalpine forests. Successional processes are slow in these forests making the identification of these patterns difficult. Lodgepole pine is thought to dominate most stands during the early phases of succession, but the degree of variation remains somewhat unknown. I examine the patterns of vegetation recovery in stands at various stages of recovery following fire to investigate the variability in forest recovery patterns. In addition, I investigate the complex environmental controls that lead to differences in the overstory composition of old-growth forests in the subalpine zone.

Finally, the comparison of fire to climate variability necessitates the reconstruction of climate at long time scales. Few reconstructions of climate exist in this
region and those that are available cover a much broader region (Briffa et al. 1992; Briffa & Jones 1994). However, valuable reconstructions of summer Palmer Drought Severity Index (PDSI), an important measure of drought in this region and likely linked to fire occurrence, have been completed (Cook et al. 1999). Comparisons between PDSI and fire events will lend insight into drought-fire relationships in the region.

In addition, the summer temperature reconstruction developed here will add to the existing tree-ring database for the region and extend our understanding of paleoclimate further into the past. The dendroclimatic relationships of the tree species used for reconstructing climate, whitebark pine (*Pinus albicaulis* Engelm.) and subalpine larch (*Larix lyallii* Parl.), have not been widely used in other reconstructions. I present a detailed investigation of the temporal patterns of tree-growth response to temperature and precipitation variability in these long-lived species.

**Organization of the Dissertation**

This research comprises five separate studies. Four of these are written as separate articles to be submitted for publication. Similar datasets are used in several of the studies resulting in some repetition, particularly in the brief study area descriptions provided. Chapter Three entitled “Research Setting”, is meant to serve as an introduction to the physical and administrative setting of the Selway-Bitterroot Wilderness Area. This chapter will not be published elsewhere. The geographic context of the Selway is too detailed to include in manuscripts for publication. It is nonetheless important to highlight some of the unique features of this remarkable landscape.
Summaries of Individual Studies

Appendix A--Climate-tree growth relationships in subalpine conifers in the Northern Rocky Mountains, USA.

The first article describes the response of whitebark pine and subalpine larch to temperature and precipitation variability. This article was prepared for submission to the journal *Arctic, Antarctic, and Alpine Research*. This investigation provides the basis for the development of the climate reconstruction presented in Appendix B. Six tree-growth chronologies from three sites were used to identify the response of these conifers to climate. All the chronologies were at least 748 years long and one whitebark pine chronology extended to AD 721. The oldest known living whitebark pine, a 1,278 year old strip bark tree, spanned the entire chronology developed at Baker Lake, Montana. The Baker Lake subalpine larch chronology, jointly developed with Peter M. Brown and Connie A. Woodhouse, extended to 12 AD.

Both standard and residual chronologies from the six chronologies (three from each species) were reduced to four chronologies using principal components analysis (PCA). In other words, two chronologies were developed for each species, a PC chronology that used the standard tree-growth chronology and a PC chronology that was based on the residual chronology. The first principal component (PC) was retained for each and compared to historical records of climate in order to identify the "climatic window" for the two species studied (Fritts 1976).

I identify the strongest tree-growth response to climate and also examine the temporal stability of the climate signal. In addition, I examine potential mechanisms that
result in changes to tree-growth response from the early half of the 20th century to the latter half. The temporal changes I identify could result from either alterations of the climatic environment or changes to the ecosystem unrelated to climate variability. The mechanisms for altering tree-growth response to climate include the potential impacts of increasing summer temperatures, representing the potential of a threshold being met whereby temperatures are no longer limiting to growth. In addition, I investigate the potential role of a decadal scale oscillation in Pacific Ocean-atmosphere teleconnections that may alter climate-growth responses. Changes to the ecosystem in which these species persist may also play a role in altering growth-climate relationships. Mountain pine beetle (*Dendroctonus ponderosae* Hopk.) outbreaks earlier in the century have left a legacy of ghost forests of whitebark pine snags over a broad region. Mountain pine beetles may have killed only those trees most sensitive to climate, resulting in a possible bias introduced by sampling trees killed by the beetles. Those trees used to span the remainder of the 20th century may be less strongly limited by climate resulting in a weakened response during the later 20th century.

Appendix B—A 748-year reconstruction of summer temperature in the northern Rocky Mountains, USA.

Appendix B presents a ≈750-year reconstruction of average summer temperature using the six tree-growth chronologies from Appendix A. This article was prepared for submission to the journal *Holocene*. It uses a stepwise regression procedure to screen potential predictor variables from the pool of potential chronologies and results in the
inclusion of three chronologies: two subalpine larch chronologies, and one whitebark pine chronology lagged forward one year. This step is crucial given the reductions in sensitivity identified in Appendix A. In addition, calibration and verification is conducted using two methods to ensure the regression model is reliable for the period of record. The first method uses the traditional split sample approach common to most dendroclimatological studies while the second uses a "leave-one-out or jack knife procedure". Periods of relatively warm or cool summers in the reconstruction are identified and their temporal persistence is examined. Potential climate forcing mechanisms such as explosive volcanism are discussed in addition to the identification of the spectral response of the reconstruction. The summer temperature reconstruction is also compared to other published reconstructions from the western United States and Canada to differentiate regionally synchronous warm and cool periods from those expressed locally.

Appendix C—Fire-climate interactions in subalpine forests of the Selway-Bitterroot Wilderness Area, Idaho-Montana, USA

This article, to be coauthored with Dr. Thomas W. Swetnam, focuses on the reconstruction of past fire events in four subalpine watersheds. Dr. Swetnam provided assistance with study and field sampling design, sample collections, interpretation of the fire-climate relationships, suggestions for alternative analysis techniques, and manuscript preparation assistance. In addition, Dr. Swetnam secured funding through the National Science Foundation for this research. This article was prepared for submission to the
A combined fire-scar analysis and stand-origin approach is used to identify fire events and to differentiate between widespread and smaller, localized fire events. Age-structure information from more than 2,000 trees is used to help identify widespread stand-replacing fires. In addition, the ten largest fire years in the SBW that occurred during the modern period were examined to identify differences between past fire-climate interactions and those of the 20th century. Fire events are compared with summer temperature, summer Palmer Drought Severity Index (PDSI), and an index of the El Niño-Southern Oscillation using superposed epoch analysis (SEA). We also examine the spatial patterns of drought associated with widespread fires using a gridded network of PDSI for the conterminous United States (Cook et al. 1999).

Appendix D—Structural and compositional characteristics in relation to time-since-fire in subalpine forests of the Selway-Bitterroot Wilderness Area, USA

This article examines the evolution of structural and compositional traits of forest stands that have experienced different periods of time since the last fire. This paper is to be coauthored with Dr. John A. Kupfer and was prepared for submission to The Professional Geographer. Dr. Kupfer provided assistance with developing and refining the approach used for ordination as well as assistance with data interpretation and manuscript revision. Age, forest composition, and vertical structure is examined for 23 sampled stands. In all, more than 1100 trees were successfully dated and used in the age structure analysis. Summaries of species composition and forest structure (basal area,
stem density, and species importance values) are presented using information collected in intensive age structure transects. Both overstory and understory characteristics are examined using nonmetric dimensional scaling to examine groupings of stands with similar characteristics. These groupings are compared with time since fire to determine how forest compositions and structures change through time.
CHAPTER TWO
PRESENT STUDY

The methods, results, and conclusions of this study are presented in the articles appended to this dissertation. The following summarizes the most important findings of those articles. Collectively, these studies add to our understanding of subalpine forest dynamics and climate variability at long time scales in the northern Rocky Mountains. The long term record provided by this dendrochronological examination is crucial in these forests because disturbances occur after relatively long periods and successional processes are slow (Habeck & Mutch 1973).

Dendroclimatological Findings

Tree growth response to climate variability

Whitebark pine and subalpine larch have been little used in dendroclimatological reconstructions (but see Colenutt & Luckman 1995; Luckman et al. 1997). These species are extraordinarily long lived and exist at the upper limits of forests in the northern Rocky Mountains making them ideal candidates for dendroclimatological investigations. Six chronologies were developed for use in reconstructing climate and all extend at least 748 years (Figure 2.1).

Tree-growth in these species responds favorably to warm summer temperatures with whitebark pine most responsive to July temperature and subalpine larch to June
Figure 2.1. Tree-growth chronologies used for reconstructing climate in the Selway-Bitterroot Wilderness Area. Chronologies were detrended using a 250 year cubic spline. Heavy black line is a 50 year cubic spline that preserves 50% of the variance at wavelengths of 50 years. Note that the Baker Lake subalpine larch chronology was truncated at AD 700.
temperature (Figures 2.2 & 2.3). However, this research identified important shifts in the response to climate of these conifers from the early half of the 20th century to the later half (Figures 2.2 & 2.3). This reduction was more pronounced in whitebark pine than was generally evident in chronologies developed from subalpine larch. In essence, whitebark pine appeared to have a strong positive relationship with summer temperature during the early part of the 20th century but the strength of the association diminished during the later half. Instead, a strong negative response to spring temperatures occurred. Although a similar change in response also occurred in subalpine larch it was not so dramatic.

Several potential causes for this shifting response were examined, but none proved entirely satisfactory. It was first reasoned that mortality in whitebark pine caused by either mountain pine beetle (*Dendroctonus ponderosae* Hopk.) or white pine blister rust (*Cronartium ribicola* J.C. Fisch.) during the early 20th century may have biased the climate response. For example, whitebark pine killed by the beetles may have been more sensitive to variations in climate and these trees were already limited. The additional stress of beetle attacks may have overwhelmed the trees resulting in their death. Those whitebark pines that survived outbreaks may have been growing in better conditions allowing them to beetle resist beetle attacks. This was supported by the fact that the subalpine larch did not show such a strong reduction in sensitivity to summer temperatures. However, an examination of the response patterns using only living trees was similar, suggesting that the shift was climate related. The second, and most plausible explanation for this reduction relates to the delivery and subsequent melt of snow in the
Figure 2.2. Correlation coefficients for standard and residual PIALPC chronologies. (A) Full time period (1901-1998). (B) Split time periods. Dark circles represent the early period (1901-1949), white circles the late period (1950-1998). Dotted line is the approximate 95% confidence limit.
Figure 2.3. Correlation coefficients for standard and residual LALYPC chronologies. (A) Full time period (1901-1998). (B) Split time periods. Dark circles represent the early period (1901-1949), white circles the late period (1950-1998). Dotted line is the approximate 95% confidence limit.
region. While further testing is required, including detailed phenological investigations, it seemed plausible that snow altered the effective growing season with a greater impact on whitebark pine than on larch. The Pacific Decadal Oscillation modifies snow delivery patterns in the area and a coincident shift in the PDO occurred around 1947 (Mantua & Hare 2002). It is speculated that PDO, while having no direct relationship to tree-growth in these chronologies, may modulate climatic response. An examination of tree-growth climate response during different phases of the PDO seems to support this hypothesis to some degree (Figure 2.4). When PDO is in its positive (warm) phase, the strength of the association between growth and climate is typically stronger during the summer months and resembles the response identified during the earlier period. During the negative (cool) phase of the PDO, however, these relationships weaken considerably (Figure 2.4).

Reconstruction of summer temperature

Due to the diminished response of whitebark pine to tree-growth it is perhaps not surprising that screening potential predictors for the climate reconstruction using stepwise multiple regression selected two larch chronologies and a whitebark pine chronology lagged forward one year. Calibration and verification statistics were generally of excellent quality and variance explained in the 748 year reconstruction was ≈36%. The reconstruction suggests decadal-scale oscillations between warm and dry conditions with no obvious centennial scale departures (Figure 2.5).

The reconstruction has many notable cool periods that were common to other reconstructions of temperature in the western United States and Canada (Figure 2.6).
Figure 2.4. Climate-tree growth relationships during different phases of the Pacific Decadal Oscillation. Chronologies are standard PC chronologies.
Figure 2.5. Reconstructed summer temperature near the Selway-Bitterroot Wilderness Area, 1250-1997. Data are shown as anomalies from the 1900-1997 mean. The smooth black line is the reconstruction smoothed using a 50-year cubic smoothing spline.
Figure 2.6. Comparison of low-frequency patterns of reconstructed growing season temperature anomalies from other reconstructions. Values are smoothed using a 20-year cubic smoothing spline fit to anomalies based on each reconstruction’s calibration period. Dashed lines near zero are used to highlight deviations from the calibration period mean from each chronology. Note that the central Idaho reconstruction is displayed as standard deviation units rather than anomalies.
Prominent cool periods occurred in the late 1300 and 1400s as well as around 1600. The longest cold departure occurred in the early 1700s. Conditions were near the longterm average throughout much of the 1800s until around 1880 and remained cool through the early 1900s. Warm periods were in place in the early 1400s, the mid 1600s and near the end of the 1700s. There was no evidence of a Medieval Warm Period signal at the beginning of the record and limited evidence of prolonged cooling related to the Little Ice Age. Spectral analysis suggested peaks in variance in reconstructed summer temperature around 87, 15, and 2 years.

The use of these conifers for reconstructing climate warrants further research into the causes of reduced sensitivity. Limited phenological data exists for these conifers but will prove important in further development of climate reconstructions that utilize these species. Elucidating the causal mechanisms of reduced sensitivity will require focused research and hypothesis testing to refine climate reconstructions and take advantage of the longevity of these species.

Fire-Climate Findings

*Fire history characteristics of subalpine forests*

A total of 96 fire-scar samples were dated in four watersheds, yielding 45 fire dates (Figure 2.7). Widespread fires within watersheds were identified in 17 years. Estimated germination dates from nearly 2,000 trees aided in the differentiation of probable widespread fires versus smaller, or localized fire events.
The average interval between all fires varied from about 20 to 170 years. The average interval between widespread fires was longer and varied between about 40-300 years. However, within a given stand in these subalpine forests, fire intervals more closely resemble the prevailing pattern of 100-300 year intervals between fires with MFI for small stands ranging from 139-341 years.

The effects of fire suppression on subalpine forests are not well understood. Successional processes are slow so the impacts of fire suppression over the past 50 years on forest structure and composition are probably slight. However, there did appear to be a reduction in fire occurrence during the 20th century that suggests suppression may have limited the spreading of fires in subalpine forests (Figure 2.7).

Figure 2.7. Selway-Bitterroot four site composite fire chronology. Solid lines at the bottom represent probable widespread fires and dashed lines depict local fires occurring in any of the four sites.
Fire-climate interactions

Superposed epoch analysis (SEA) was used to compare climate variables with fire events during the fire year and the five years prior to fire occurrence. SEA is a compositing technique that compares average conditions during fire event years and the years prior to a Monte Carlo simulation to produce confidence intervals (Swetnam & Betancourt 1992). Tree-ring reconstructed climate variables were used to compare fire events that occurred prior to the instrumental period of record. Only widespread and modern fires exhibited significant relationships with the climate variables (Figure 2.8). Widespread fires were significantly related to dry conditions (as measured by PDSI) both during the fire year and one year prior to fire occurrence. While dry conditions during the fire season are well recognized as important drivers of large fire occurrence, the importance of lagging relationships had not been previously identified in these forests. Modern fires were significantly related to warm summer temperatures and dry conditions during the fire year but did not exhibit lagging relationships. There was no direct relationship between ENSO and fire occurrence identified using SEA. However, some large fires might be related to El Niño events. Large fires in 1889 and 1919 both coincided with strong El Niño events. Analysis of the available climate data suggested El Niño conditions can lead to summer drought in the vicinity of the SBW.

Examination of the spatial patterns of drought related to large fire occurrence suggested multiple patterns of regional drought were related to widespread fire occurrence (Figure 2.9). A principal components analysis performed on reconstructed PDSI over the coterminous United States suggested that during most widespread fire
Figure 2.8. Comparisons of lagged responses of fire events to climatic variables using superposed epoch analysis. Gray bar is the departure of the actual measured variable minus the simulated value estimated from 1000 Monte Carlo simulations. Solid horizontal lines are the 95% confidence limits and dashed lines are the 99% confidence limits.
Figure 2.9. Rotated factor loadings between widespread fire years and PDSI in the SBW. Variance explained is shown in parentheses. Heavy dark lines differentiate positive (wet conditions) and negative loadings (dry conditions), indicated by "+" and "-". Note that factor 1 has an inverse relationship to fire years suggesting conditions are dry during the fire years that load most strongly on this factor.
years in the SBW region climate conditions were dry over a broad region of the western United States. Additionally, one spatial pattern resembled the spatial pattern of droughts that commonly occur El Niño events in the Pacific Northwest and northern Rockies.

**Successional Dynamics Findings**

The generalized pattern of forest succession proceeds through early stages with canopy composition largely dominated by lodgepole pine (Figure 2.10). Subalpine fir became more prominent in the stands as successions proceeded. The oldest forests sampled had canopies dominated by subalpine fir, Engelmann spruce, or whitebark pine.

Ordination using nonmetric dimensional scaling grouped stands with similar fire-free periods together with few exceptions (Figure 2.11). Additionally, when NMDS was applied treating overstories and understories as separate observations the understories of relatively young stands (≈100-150 years) appeared to be more similar to their overstories than older forests (Figures 2.11). The largest change between overstory/understory similarity occurred in the two oldest lodgepole pine stands (≈250 & ≈350 years respectively) indicating that canopy breakup and species replacement occurs after more than 200 years.

The most important finding of this research documented the existence of two successional pathways (Figure 2.12). The most common pathway involved lodgepole pine initially colonizing a site but field evidence suggested subalpine larch or Engelmann spruce could also colonize a site if lodgepole pine was either absent from the prior stand or if it was younger than required to reach seed bearing age. In the case of subalpine
Figure 2.10. Stem density and basal area relationships with time-since-fire. The number of stands in each TSF category is shown in parentheses. Error bars are 1 standard deviation from the mean.
Figure 2.11. Two dimensional NMDS ordination of overstory and understory importance values (Approach 2, see Appendix D for details). Circles indicate the axis scores for the overstory and squares the understory. Shading indicates TSF categories for individual stands (see legend). The line connects the overstory axes scores for a stand to its understory scores. Each stand is labeled at its overstory symbol. Solid lines enclose overstories dominated by the individual species based on importance values and dashed lines delimit understories (indicated by the first two letters of the genus and species).
Figure 2.12. Conceptual framework of stand development based on the age structures and disturbance history of sampled stands and the life history characteristics presented by Noble and Slatyer (1980) and Cattelino et al. (1979). The upper figure (A) shows the relevant life history traits for each of the four common species. A lower case p indicates propagules are present, and m indicates the point at which a species has reached seed bearing age, and l indicates the maximum longevity of the individual species persistence within a stand (either via propagules or living individuals). The lower figure (B) is a flow chart showing stand development with increasing time-since-fire. Black lines in the flow chart indicate progression with time while gray lines denote resetting a stand due to fire occurrence. Examples of each stage are indicated by the stand numbers. In parentheses next to each stand are listed the dominant overstory and understory species in each stand.
Lodgepole Pine (PICO) CI:
Lodgepole Pine (PICO) DI:
Whitebark Pine (PIAL) DT:
Subalpine Fir (ABLA) DT:
Engelmann Spruce (PIEN) DT:

Stand Age

Stand Initiation
Post-fire colonization dependent on seed dispersal from PICO, PIAL, ABLA, & PIEN

Spruce-Fir Pathway

Serotinous lodgepole present to colonize post-fire stand plus dispersal of other species

Lodgepole Pine Pathway

TSF Categories 1 & 2

TSF Categories 3 & 4
forests, historical legacies appeared to have an important impact on the composition of developing forests.

The occurrence of large areas of lodgepole pine suggested that fire intervals were typically shorter than the lifespan of lodgepole pine, allowing its persistence due to the abundant seed rain supplied by serotinous trees. If fire intervals were longer than lodgepole pine longevity it would be expected that more areas of early recovery would be dominated by subalpine fir or spruce. This suggests that the final stages of succession ("climax") are rare on the landscape and may be relegated to sites with particular environmental characteristics. Moreover, some forest stands could possess "old-growth" characteristics even if they are relatively young. This has important implications for the development of management plans based on the concept of potential vegetation (McCune & Allen 1985; O'Hara et al. 1996). For example, the use of fire as a management tool can be varied to achieve different goals by altering fire intervals.

**Forest Management Implications of this Research**

The multiple facets of this research will provide those involved in the management of the Selway-Bitterroot Wilderness Area important information concerning the linkages between fire and climate, forest development, and climate variability at long time scales. This research provides important baseline information that can be incorporated into management plans focused on sustaining subalpine ecosystems in the Selway and will aid in the understanding of forest dynamics.
From the perspective of fire-climate interactions, the identification of persistent dry conditions as factors in promoting widespread fire occurrence will aid the decision making process during suppression activities. Currently, fire management activities involve the use of “prescribed natural fires”, or those fires that ignite naturally and are allowed to continue burning under a predetermined set of conditions. The identification of two successive dry years in widespread fire development will enable managers to consider different management scenarios based not only on current conditions but also those of the prior year. Moreover, the potential linkage between widespread fire occurrence and El Niño events might provide fire managers advance notice of potentially severe fire seasons allowing for modifications to man power distributions and awareness of the potential for fires to grow.

There is also a need to better understand the development of subalpine forests following fire events at both local and landscape scales. Fire intervals within small areas, and presumably at larger spatial scales, have a strong impact on forest composition. Alteration to these intervals could lead to shifts in forest composition that could have important feedbacks on future fuel or fire severity conditions. Moreover, an awareness of the multiple initial conditions related to forest development in this area is crucial to developing an understanding of the long term effects of fire suppression. Forests that are dominated at their earliest stages by lodgepole pine progress toward spruce-fir forests over much longer periods of time and fuel buildups are also likely slower. However, initial dominance by spruce or fir might lead to a more rapid succession toward forests
with highly connected overstories and understories that could lead to large or severe fires at shorter intervals.

Identification of climate variability at long time scales provides a context to interpret present climate scenarios. This research provides a nearly 750-year perspective on summer temperature variability that can be used as baseline information within which to judge current temperature regimes. Steadily rising temperatures have important implications for a number of processes including the frequency of disturbances such as fire and mountain pine beetle epidemics.
CHAPTER THREE

Research Setting: The Selway-Bitterroot Wilderness Area

“From this elevated spot we have a commanding view of the surrounding mountains, which so completely enclose us that though we have once passed them, we almost despair of ever escaping them without the assistance of the Indians.”

--Captain Meriwether Lewis, 1806, describing the view from somewhere in the Bitterroot Mountains.

Introduction

The Selway-Bitterroot Wilderness Area is a remarkable landscape. Traveling west in search of a Northwest Passage, Captains Meriwether Lewis and William Clark were greeted by a massive wall of mountains as they left the Montana Plains. The Bitterroot Mountains, situated along the eastern boundary of the SBW rise from the \( \approx 1000 \) m Bitterroot Valley to the east to more than \( 3000 \) meters. A series of steep glacial canyons trending east-west punctuate the Bitterroot front’s eastern side, masking the complex ridges of the SBW’s interior. The dramatic landscape comprising the SBW region was a formidable test for Capt. Meriwether Lewis’ Corps of Discovery and today remains one of the largest wilderness landscapes in the continental United States.

The SBW has an important place in the formulation of fire management policy as well as the drafting of important federal policies such as the Wilderness Act of 1964 and Federal Lands Policy and Management Act of 1976 (Bolle 1997). Further, the
mountainous region is an important irrigation source for the rapidly expanding Bitterroot Valley by capturing and storing winter precipitation for incorporation into reservoirs. Perhaps more importantly, the SBW, together with its neighbor the Frank Church River-of-No-Return Wilderness Area represents the largest contiguous tract of wilderness in the lower 48 states.

This chapter details the physical and administrative setting of the Selway-Bitterroot Wilderness Area. The intent of this chapter is to summarize some important details of the physical setting that are too numerous and detailed to include in manuscripts centered on the fundamental questions of this research including fire-climate-vegetation interactions. However, the context of the SBW is important to understanding the SBW in a holistic sense.

Geographic and Administrative Setting

The Selway-Bitterroot Wilderness Area (SBW) is located on the border of Idaho and Montana in the Northern Rocky Mountain Physiographic Province (Figure 3.1). The SBW has as its origin the Bitterroot Preserve, a 1.6 million ha set aside in 1897 (Moore 1996). The Selway-Bitterroot primitive area was established in 1936 (Habeck 1972). Designated as a wilderness in 1964 with the passage of the Wilderness Act, the SBW today is composed of 546,627 ha, representing the third largest wilderness area in the continental United States. The Bitterroot Mountains lie along the eastern edge of the wilderness area and extend ≈100 km from north to south. The complex terrain of the interior SBW includes the Clearwater Mountains in the central and southwest portions of
Figure 3.1. The location of the Selway-Bitterroot Wilderness Area along the border of Idaho and Montana. Lighter shading depicts areas of higher elevation and darker shading lower elevation. Major Rivers in the region are labeled.
the wilderness area. Elevations along the Selway River in the western portion of the SBW are around 450 m. At its widest point, the SBW is approximately 100 km wide and 130 km long. The SBW is administered jointly by four National Forests of the Northern Region (Region 1) of the United States Forest Service: the Bitterroot, Clearwater, Lolo, and Nez Perce. Adjacent to the SBW, separated by only a one lane dirt road, is the largest wilderness area in the continental United States, the Frank Church River-of-No-Return Wilderness (958,177 ha).

The SBW is topographically diverse, physically divided into eastern and western portions by the north-south trending Bitterroot Mountains. The western interior portion forms the largest proportion of the wilderness area and contains the Clearwater Mountains. The western portion is typified by more moderate slopes, lower peaks and more contiguous forest than the region immediately adjacent to the Bitterroot mountains.

**Geology and Soils**

The SBW region is underlain by granitic bedrock derived from the Idaho Batholith (Greenwood & Morrison 1973; Foster et al. 2001; Foster & Raza 2002). The Idaho Batholith was intruded around the Cretaceous-Tertiary boundary (Foster et al. 2001). Granite bedrock from this period dominate the majority of the region from the Bitterroot Mountains westward while granite derived mylonite form much of the main chain of the Bitterroot Mountains (Foster et al. 2001). The bedrock of the region was exposed following broad scale warping and crustal extension during the Eocene (Foster & Raza 2002).
The SBW region was glaciated extensively during the Pleistocene and evidence of past alpine and valley glaciers in the form of cirque lakes and glacially carved valleys is abundant in the region, particularly along the Bitterroot Front where a series of east-west trending U-shaped valleys are present (Habeck 1972). In addition, the terminus of the Cordilleran ice sheet was located in the northwestern part of Montana and that of the Laurentide ice sheet in northeastern Montana (Mickelson et al. 1983; Waitt & Thorson 1983). A large proglacial lake, Lake Missoula, occupied much of the Bitterroot Valley at elevations below ≈1250 m (Baker & Bunker 1985; Locke 1990). Glacier development in the region was apparently constrained by regional airflow patterns related to precipitation delivery to the region that was modified by the presence of continental ice sheets and Lake Missoula (Locke 1990; Hostetler & Clark 1997). There are no active glaciers within the SBW at present although alpine glaciers are present a short distance to the north in Glacier National Park, Montana. These glaciers have been receding at a rapid rate throughout the 20th century (Butler & DeChano 2001).

Limited soil data exists in the SBW region. Natural Resources Conservation Service Soil surveys, available for most of the United States, have not been completed in this region due to its limited value for agriculture as well as access difficulties. For the most part soils are young and derived primarily from granitic bedrock. Soils are generally thin in most places within the SBW. Soils developed from granite bedrock are nutrient poor but volcanic ash and loess deposits likely contribute to nutrient availability in some areas (Clayton 1974; Mehringer et al. 1977; Clayton et al. 1979).
Forest Composition and Fire Regimes

The vegetation composition and associated fire regimes reflect the topographic and climatic complexity of the SBW. Ponderosa pine-Douglas-fir (*Pseudotsuga menziesii* Mirb.) habitats characterize lower forest zones (between about 500-1500 m), merging to lodgepole pine and Engelmann spruce-subalpine fir (*Picea engelmannii* Parry-*Abies lasiocarpa* (Hook.) Nutt.) as elevation increases (Crane & Fischer 1986). Lower elevation zones, particularly along riparian corridors, contain mixed forests of western red cedar and western hemlock. The SBW is near the easternmost extension of these species' ranges. Long-lived whitebark pine (*Pinus albicaulis* Engelm.) and subalpine larch (*Larix lyallii* Parl.) are found at the highest elevations and on more extreme sites (Arno & Habeck 1972).

The importance of whitebark pine in the SBW is also important to mention. Whitebark pine was one of the most frequently encountered fire-scarred trees in the area and recorded fires as far back as AD1204. Further, the great longevity of whitebark pine (and subalpine larch) coupled with slow rates of decomposition permit the development of millennial length chronologies of tree-growth such as those used here for reconstructing past climate (Appendices A & B).

Whitebark pine is a keystone species of subalpine forest ecosystems, providing watershed protection, aesthetics, habitat, and food for wildlife, such as the Clark’s nutcracker and black and grizzly bears (Arno & Hoff 1990). However, the continued presence of whitebark pine in the subalpine environment is uncertain due to an introduced fungal pathogen, white pine blister rust, changes in fire regimes, and
advancing succession. Whitebark pine populations have declined by as much as 42% in the past 20 years in parts of the northern Rocky Mountains (Keane & Amo 1993) and by more than 45% in the Interior Columbia River Basin (Keane 2001). On sites where it is seral to other conifers, whitebark pine has declined more than 98% (Keane et al. 1996). The decline of whitebark pine is not restricted to the northern Rocky mountains. Campbell and Antos (2000) reported 21% of dead whitebark pines examined in British Columbia, Canada were likely killed by white pine blister rust infections as well as an infection rate of between 21-44% on living trees.

Large areas of whitebark pine killed in the early part of the century remain as “ghost forests” today, persisting as sun bleached snags (Ciesla & Furniss 1975). Many of these trees were likely killed by white pine blister rust (Cronartium ribicola) (Lachmund 1926; Hoff & Hagle 1990) and mountain pine beetle (Bartos & Gibson 1990; Kipfmueller et al. 2002). The blister rust is a Eurasian fungus introduced to the Pacific Northwest in 1910 (Hoff & Hagle 1990). However, the mountain pine beetle is a native forest insect. Although it is known that these two agents were proximate causes of much of the whitebark pine mortality that occurred during the 20th century, drought and fire suppression may have predisposed whitebark pine to mortality.

The loss of whitebark pine from subalpine environments could have serious impacts on the ecology of the region because whitebark pine is considered a “keystone” species. Many species of wildlife rely on its seeds for food, most notably the grizzly bear and Clark’s nutcracker. Currently, intensive research is being conducted to develop rust-
resistant whitebark pines as well as focused research on the effects of fire suppression on whitebark pine communities in an effort to restore and preserve these unique forests.

The importance of fire as an ecosystem disturbance has been long recognized in northern Rocky Mountain forests (Leiberg 1899; Larsen 1929; Habeck & Mutch 1973; Pfister et al. 1977; Arno 1980; Arno 1996). Throughout much of the 20th century, however, fires were viewed as detrimental to forest resources. Leiberg (1899), the first to conduct a thorough examination of the forest conditions in the SBW region, and map burned areas, recognized the ubiquitous nature of fires and suggested there were few if any areas that had not been burned at least once since the early 1700s. Leiberg viewed fire occurrence in these regions as primarily an economic hardship on forest resources. Leiberg is perhaps best remembered for his statement “The after effects of fire in this region are various, but are always evil, without a single redeeming feature…” (Leiberg 1899, p. 388). Fires reduced stocking density in ponderosa pine forests and completely razed lodgepole pine and spruce-fir forests. This validated the need for widespread fire protection throughout the region partly to protect water resources but also to ensure an adequate timber supply for the development of the region.

Perhaps at least part of Leiberg’s distaste for fire was the result of his failure to recognize its natural role in shaping the forest system. The causes of fires by most early surveyors of the forest reserves were attributed primarily to prospectors and Native Americans. While some fires were undoubtedly the result of human-set fires, natural ignitions during dry periods probably were responsible for most of the fire activity in the
area. Leiberg’s (as well as other early forest surveyor’s) failure to recognize lightning as an important ignition source in this region is a remarkable omission (Baker 2003).

Aggressive suppression of fires, though ineffective during dry years, began following the Great Fires of 1910 that left 85 people dead and destroyed the nearby town of Wallace, Idaho. Suppression did not become effective until after the 1940s when aerial suppression of fires through the use of smoke jumpers became established at nearby Missoula, Montana. Smoke jumpers could be rapidly deployed to fires before they grew out of control, but inevitably some still reached large size.

The detrimental effects of fire suppression were recognized within the scientific community perhaps as early as 1940 as recognition of increasing fuels and anomalous fire behavior, particularly in lower elevation forests were observed (e.g., Weaver 1943; Weaver 1961; Dodge 1972). Continued pressure from within the scientific community, as well as from within the US Forest Service, and the passage of the Wilderness Act of 1964 eventually led to an acceptance of fire’s natural role. The Wilderness Act, while seeking to preserve America’s natural heritage and “pristine” landscapes, also recognized the importance of preserving the natural processes important to their development. This necessarily included the incorporation of fire into management plans as an important natural feature of many landscapes (Habeck & Mutch 1973). The Wilderness Act prompted additional research into fire’s role and helped to recognize the effects of fire suppression with respect to changing forest conditions. Wilderness areas were in many ways viewed as natural laboratories, and one of the first prescribed fire programs was
operating over the entire Selway-Bitterroot Wilderness area by 1979 after experiencing success in limited areas of the wilderness beginning in 1970 (Brown et al. 1994).

The close proximity of the SBW to the United States Forest Service Fire Laboratory in Missoula Montana led to a large and diverse number of investigations centering on the effects of fire in this landscape. Most of this research focused on the reduction of fire in the region and the effects on forest communities. Few have explicitly examined the role of climate in altering fire patterns on the landscape.

The fire regimes of the SBW are directly linked to the particular forest species found in different elevation zones. In the lower elevation ponderosa pine-Douglas-fir zone, fires have burned relatively frequently at approximately 20 year intervals (e.g., Arno 1980; Arno & Petersen 1983). Historically, these fires were low to moderate severity surface fires, but infrequent stand-replacing fires possibly occurred under certain climatic and fuel conditions or over limited areas (Barrett 1988; Morgan et al. 1996). In contrast to the lower elevation forest types, subalpine forests of the higher elevations are characterized by infrequent stand-replacing fires (Arno 1980). Low to moderate severity fires have not been well documented in subalpine forests in this area but do occur (Habeck & Mutch 1973). Mean fire intervals of stand-replacing fires in subalpine forests of this region are typically thought to be between 50-300 years (Arno 1980).

Throughout the 20th century the suppression of fires has had an important impact on the structure and composition of forests within the SBW, particularly at lower elevations (Arno & Petersen 1983; Arno & Gruell 1983; Arno & Gruell 1986; Barrett 1988; Barrett & Arno 1991; Brown et al. 1994; Rollins et al. 2001). Wildfires were
consistently and effectively suppressed for most of the 20\textsuperscript{th} century in and around the SBW (Rollins et al. 2001). Since 1974, however, the U.S. Forest Service has allowed lightning ignited fires to burn within prescribed conditions in the SBW (Barrett & Arno 1991; Brown et al. 1994).

Rollins et al. (2001) suggest fire suppression in the wilderness has been very efficient during various portions of the 20\textsuperscript{th} century. Between 1880-1934, fire rotations (the time required to burn an area equal in size to the study area in question) were relatively short (110 years), encompassing some historically important fire years known as the Great Fires of 1910, 1919, 1934. These large fires were in part one reason for the dramatic increase in suppression efforts throughout the United States (Moore 1996; Pyne 2001). Fire rotations lengthened considerably during the modern suppression period between 1935-1974 to nearly 3,900 years (Rollins et al. 2001). The wildfire use era that began in 1974 resulted in a dramatic decrease in fire rotation to 218 years, though still not equal to the earliest time period. However, Rollins suggests that the changing rotations are variable among forest types with the most dramatic increases in rotation time occurring in western red cedar potential vegetation types and the subalpine zones experiencing the smallest increase. Nonetheless, Rollins et al. (2001) indicate fire rotations have increased in all areas from a low prior to effective fire suppression of 41-295 years to 1,957->396,00 years during the modern fire suppression era. Burning has increased since the prescribed fire era began, but fire rotations are still long relative to pre-modern suppression periods.
It is important to note, however, that fire rotation statistics can be misleading, particularly in the subalpine zone where fires are typically infrequent. Fires occurred sporadically, usually under drought conditions, and grew to large sizes. During a single year fires could burn over a large area while fire might not return again for more than a century. The limitation of fire rotation statistics is particularly apparent where the time period of interest is small relative to the fire return interval and study area size (Johnson et al. 2001). For example, Rollins et al. (2001) divide a 116 year record into three periods ranging from 22-54 years. It is not surprising that large areas burned during some of these periods but not others. Fire rotations in upper elevation forests might be more similar during the modern period if the fire year of 2000 (that burned ≈11,000 ha in the SBW) is included in determination of fire rotation. In other words, just the fact that rotations are longer in some areas does not necessarily suggest that human impacts are the main cause. In lower elevation forests where fire return intervals are generally shorter these statistics are probably more reliable.

**Climate of the Selway-Bitterroot Region**

Finklin (1983) provides a detailed summary of the climate of the SBW. He examined the historical record of climate from stations located within and around the SBW. Most of the station records from this area are short and discontinuous in time, beginning only in the late 1950s. For the most part weather records are only available for the summer months. This is most likely due to the interest in protecting the forested region from fires. Many of the records are from fire lookouts scattered among the many
peaks of the SBW. While short and limited by missing data, Finklin provides a remarkably thorough discussion of the climate characteristics of the region with particular emphasis on the steep gradients related to elevation and topography.

The climate of the SBW is characterized as transitional between the northern-Pacific Coastal climates usually associated with the Pacific Northwest and a continental climate more typically found in the interiors of continents (Finklin 1983; Barry & Chorley 1998; Ferguson 1999). The Pacific coastal influence is due to the movement of air originating in the north Pacific moving inland while a relatively cold, dry polar air mass moves south along the Rocky Mountains. Midlatitude cyclones forming along these air mass boundaries, typically lasting a week or less, can migrate through the Selway and bring significant amounts of precipitation, and dramatic changes in air temperature. In summer, precipitation may be minimal, but high wind speeds and dry lightning associated with these low-pressure centers are an important determinant in wildland fire initiation and growth.

The variable influences of both the Pacific climate and aspects of continentality are reflected in the annual climate regime of the Selway region. Annual precipitation varies from approximately 100 cm in the western portion of the SBW and declines to the south and east to about 62.5 cm, although the amounts vary considerably due to elevation and other topographic factors (Finklin 1983). Most of the region’s precipitation falls as snow between the months of November and May, with January typically being the wettest month. The highest elevations in the SBW can receive as much as 170 cm, mostly in the form of snow during winter and early spring (Finklin 1983).
peak of precipitation in May and June may be partially in response to the weakening Aleutian low that allows late forming Pacific Storms to move into the region rather than being deflected northward.

July and August are typically the driest months in the Selway region. High pressure in the subtropical eastern Pacific Ocean extending into the Pacific Northwest results in large scale ridging that steers midlatitude cyclonic storms north of the region. Some precipitation does fall during the summer, but it is usually due to convective storm development. These large convective complexes move under the influence of the prevailing westerly airflow and produce scattered, but locally intense precipitation over mountain regions. In addition, subtropical moisture related to the North American monsoon does occasionally reach as far north as the SBW and can result in widespread thunderstorms (Brunelle-Daines 2002). The fire season lasts from late June through early September with activity peaking between mid July and early September coincident with the development of convective storms with enhanced lightning activity and dry summer conditions (Habeck 1972; Brown et al. 1994).

Temperatures around 2,100 m are generally cool in the subalpine zone with average temperatures in July of about 12°C, and -8°C in January (Finklin 1983). Temperatures along the Lower Selway River at about 600 m averages about 21°C in July and -2°C in January. The highest temperature recorded within the Wilderness was 44°C (112°F) at Moose Creek Ranger Station in August 1961 (Finklin 1983, p. 32). Available climate data suggests much less spatial variability in temperature patterns than is exhibited by precipitation patterns (Figure 3.2, see Table 3.1 for climate stations).
Figure 3.2. Climographs for climate stations near the Selway-Bitterroot Wilderness Area. Average monthly temperature (°C) is shown by the line and bars represent total monthly precipitation (cm). Averages were calculated using the period of record for each station (see Table 3.1). Black dots show the location of each station. Note that precipitation data is missing for January-March for Sula, MT.
Table 3.1 Climate stations located near the Selway-Bitterroot Wilderness Area.

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Data Source</th>
<th>Lat./Lon.</th>
<th>Elev. (m)</th>
<th>Time Span</th>
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<td>Missoula, MT^</td>
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<td>46.92N 114.10W</td>
<td>967,973</td>
<td>1958-1998</td>
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<tr>
<td>Stevensville, MT</td>
<td>NCDC</td>
<td>46.52N 114.10W</td>
<td>1,029</td>
<td>1911-1998</td>
</tr>
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<td>Western Ag. Center, MT</td>
<td>NCDC</td>
<td>46.32N 114.12W</td>
<td>1,096</td>
<td>1965-1998</td>
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<td>HCN</td>
<td>46.23N 114.17W</td>
<td>1,076</td>
<td>1900-1998</td>
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<td>1,183</td>
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<td>Shoup, MT</td>
<td>NCDC</td>
<td>45.38N 114.28W</td>
<td>1,036</td>
<td>1966-1998</td>
</tr>
<tr>
<td>Dixie, ID</td>
<td>NCDC</td>
<td>45.55N 115.48W</td>
<td>1,713</td>
<td>1952-1998</td>
</tr>
<tr>
<td>Elk City, ID</td>
<td>NCDC</td>
<td>45.83N 115.45W</td>
<td>1,237</td>
<td>1951-1998</td>
</tr>
<tr>
<td>Fenn RS, ID</td>
<td>HCN</td>
<td>46.10N 115.53W</td>
<td>475</td>
<td>1908-1994</td>
</tr>
<tr>
<td>Pierce, ID</td>
<td>NCDC</td>
<td>46.50N 115.80W</td>
<td>939</td>
<td>1963-1998</td>
</tr>
<tr>
<td>Headquarters, ID</td>
<td>NCDC</td>
<td>46.63N 115.82W</td>
<td>965</td>
<td>1959-1998</td>
</tr>
<tr>
<td>Bungalow RS, ID</td>
<td>NCDC</td>
<td>46.63N 115.50W</td>
<td>698</td>
<td>1931-1964</td>
</tr>
<tr>
<td>Powell RS, ID</td>
<td>NCDC</td>
<td>46.52N 114.72W</td>
<td>1,076</td>
<td>1962-1998</td>
</tr>
</tbody>
</table>

1Station number corresponds to the numbers in Figure 2.5. 2The Missoula station was moved in 1966.

Most of the variability in temperature is due to the effects of elevation, but there is pronounced spatial variability in the patterns of precipitation, mostly owing to the rainshadow effect of the Bitterroot Range.

Spatial features of precipitation in the SBW

Precipitation amount and seasonality differs from the western side of the Bitterroot Mountains to the east. For the purposes of this description, precipitation
variability is described based on the climatological water year between the months of October-September, rather than based on the annual (January-December) calendar. The SBW straddles Idaho Climate Division 4 and Montana Climate Division 1 and reflects aspects of each. Divisional climate data clearly depict the strong rainshadow effect of the Bitterroot Mountains with precipitation greatly reduced in Montana Division 1 relative to Idaho Climate Division 4 (Figure 3.3). The effect of the Bitterroot Range on reducing precipitation from west to east is most marked during the winter months and the period of generally low rainfall is similar in both divisions (Figure 3.3). The increase in spring precipitation in Montana Division 1 is visually accentuated due to the relative dryness of the winter but the actual amounts were only slightly greater than those of Idaho Division 4 (Figure 3.3). Divisional climate data, while useful for summary purposes, is derived from many stations lying a long distance from the SBW itself. However, the reduction in winter rainfall is also clearly evident when considering only those stations closest to the SBW boundary (within ≈50 km of the wilderness boundary) and is again most evident during the cool season (Figure 3.2).

Several authors have suggested that the importance of summer precipitation in the annual water supply increases from west to east across the Rocky Mountains (Baker 1944; Whitlock & Bartlein 1993; Brunelle-Daines 2002). Considering the relative contribution of each month's total precipitation to the total precipitation accumulated during the water year suggests that the eastern portion of the SBW, represented by Montana Climate Division 1, receives a higher proportion of its annual water budget from late spring and early summer precipitation than areas to the west (Figure 3.4). In
Figure 3.3. Spatial and temporal patterns of precipitation variability from climate divisions bordering on the SBW. Precipitation data is presented in centimeters for the water year (October-September). Data are smoothed using a running average to highlight general trends.
Figure 3.4. Percentage of precipitation occurring in different months in climate divisions bordering the SBW. Percentages are based on the water year (October-September). Data are smoothed using a running average to highlight general trends.
both climate divisions, a notable dry period occurs in the months of July and September. However, this dry period is preceded by a peak in precipitation in Montana Division 1. While the relative amount of precipitation during late spring and early summer is similar between the two climate divisions, the relative contribution to the overall water supply is higher in Montana Division 1 mostly due to the lower water year total resulting from diminished precipitation during the winter months.

Examination of individual climate stations around the SBW also seem to suggest this is the case even though it occurs over a much shorter distance than the divisional data generally includes (Figure 3.5A). Further, May 1 snow water equivalent (SWE) also appears to diminish from west to east in the region (Figure 3.5B). The effect of this seasonality is not clear within the SBW itself however because few stations exist in the SBW making comparisons limited.

*Synoptic Climatological features of the SBW*

Winter storms are an important source of moisture in the SBW. Snowfall amounts can be heavy and its persistence into the dry summer months represents an important moisture source for many upper elevation tree species. These winter storms are primarily due to variations in the geographic position of the Polar Jet stream that, under normal conditions crosses the United States north of the Selway-Bitterroot region. In some respects, the Polar Jet Stream is "anchored" to the north of the Rocky Mountains and begins its shift southward in Alberta, CA. The position of the Pacific Jet is closely linked to the characteristics of the Aleutian low. When the Aleutian low is intense, counter-clockwise
Figure 3.5. Spatial pattern of precipitation and snowpack at climate stations near the SBW. Each circle represents the average water year precipitation or average May 1 snow water equivalent (SWE) with its size scaled relative to the other stations to indicate the overall amount received at each station. (A) Water year precipitation at different stations. The black pie piece depicts the percentage of the water year total that is received during the summer months. Numbers to the upper right of each circle correspond to the station numbers listed in Table 3.1. (B) Average May 1 SWE (cm) between 1961-1999 from NRCS snotel stations.
circulation deflects storm tracks northward and around the Selway region. Weakening of the Aleutian low permits storms developing in the Pacific Ocean to move onshore and inland to the region.

Ocean–atmosphere interactions occurring in the Pacific Ocean can produce important modifications to the Aleutian low with implications for winter climate in the Selway region. Teleconnections are defined as dynamic linkages between the ocean and atmosphere that occur over large areas and often involve alterations to large scale circulation patterns (Wallace & Gutzler 1981). Pacific Basin teleconnections have important effects on climate patterns in the Selway region primarily by altering the strength and position of the Aleutian low during the winter months. Although many of these interactions take place a great distance away from the Selway region, the dynamic ocean-atmosphere connections often result in cascading downstream effects.

Considerable attention has been focused on teleconnections occurring in the Pacific Basin with global and regional climate implications. El Niño-Southern Oscillation (ENSO), the Pacific/North American pattern (PNA), and the Pacific Decadal Oscillation (PDO) have all been implicated in climate variations in the Pacific Northwest, but their expression further inland on climate patterns are more difficult to discern.

Of the Pacific Basin teleconnections, ENSO has received the most attention. ENSO is well known for its dramatic, seasonal weather effects such as floods (Andrade, Jr. & Sellers 1988), forest fires (Swetnam & Betancourt 1990), and drought. Warm ENSO events result from the buildup of a pool of warm water along the South American coast and disruption of the normal patterns of the Hadley circulation (Barry & Chorley
Air pressure in the eastern Tropical Pacific decreases and the east-west pressure gradient diminishes leading to a weakening of the trade winds and a subsequent reduction in cold water upwelling along the coast of South America which further increases warm water accumulation (Barry & Chorley 1998). These warm events are termed El Niños, while the opposite events (cool water buildups along the South American coast) are considered La Niñas. El Niño events typically last around one year, with a frequency of occurrence around 3-10 years, but they have shown considerable variability through time (Michaelsen 1989; Cole & Cook 1998).

Though ENSO relationships with climate variability have been well established for some areas of North America, particularly the southwestern United States, the effects are not as well documented in the Pacific Northwest or in the vicinity of the SBW. Harrison and Larkin (1998) suggest that during average El Niño warm events the Pacific Northwest is generally warmer than much of the rest of the United States during fall. Dry conditions may occur as a result of El Niño events in some portions of the Pacific Northwest during fall and winter.

ENSO has been perhaps the most widely studied of the three patterns, but is directly related to the PNA pattern. The PNA pattern is essentially the dominant extratropical feature of ENSO and has important affects on circulation patterns over North America during the winter months (Leathers et al. 1991; Keables 1992; Leathers et al. 1992). The PNA is a circulation feature that is defined using an index that characterizes the degree to which air flow exhibits a zonal or meridional circulation pattern (Wallace & Gutzler 1981; Leathers et al. 1991). The PNA index has been defined
in various ways but typically includes pressure anomalies at several centers of action including points representing the center of the Aleutian low pressure, Alberta, Canada, and the southeastern United States (Leathers et al. 1991). When PNA is positive strong ridges will be present over Canada (meridional flow) due to a deep Aleutian low. Negative values represent zonal flow patterns over the northern United States.

The positive values of the PNA index that indicate a deepening Aleutian low pressure center and meridional flow deflect Pacific storms northward away from the Selway region. This effectively results in warmer temperatures and lower precipitation during the winter months in the SBW (Leathers et al. 1991). Cayan (1996) has identified similar relationships between winter circulation, snowpack accumulations, and streamflow. Cayan (1996) found strong relationships between snowpack and streamflow in Idaho resembling the PNA pattern, but found no strong relationships with temperature in the region. The high positive correlations he identified are almost directly centered on the Selway region during the winter months. Cayan (1996) refers to this pattern as the Idaho Pattern.

The Pacific Decadal Oscillation is the least well understood teleconnection pattern of the three discussed here, but has also been shown to have important winter climate effects in the Pacific Northwest (Mantua et al. 1997; Zhang et al. 1997; Bond & Harrison 2000; Mantua & Hare 2002). Much like ENSO, the Pacific Decadal Oscillation is related to the alternating movements of warm and cold ocean water but in the northern Pacific Ocean rather than in the tropics. However, Mantua and Hare (Mantua & Hare 2002) suggest there are distinguishing features of PDO events relative to ENSO events. First,
PDO events are more persistent than ENSO events (20-30 years versus 6-18 months). Second, PDO affects regional climate further to the north than does ENSO, with the strongest impacts on climate in the North Pacific region. The PDO does impact climate across much of the western United States, but its affects on regional climate are strongest further north relative to ENSO. However, PDO may have a strong moderating influence on ENSO events that are particularly evident with respect to precipitation delivery in the southwestern United States during El Niño years (Brown and Comrie, in review).

While the climatic effects of ENSO are strongest in the tropics, PDO has a greater effect on the North Pacific/North American sector (Mantua et al. 1997; Mantua & Hare 2002). Like the PNA pattern, PDO affects the intensity of the Aleutian low pressure center and alters the pattern and frequency of ridge and trough formation in the Pacific Ocean (Bond & Harrison 2000). When PDO is in its positive phase, the Aleutian low pressure is enhanced, acting as an important blocking mechanism for the movement of Pacific Storms into the interior of the North American continent, though cool season precipitation is generally higher along the Gulf Coast during these events (Mantua et al. 1997; Bond & Harrison 2000; Mantua & Hare 2002). Mantua et al. (1997) showed there were significant positive relationships between PDO and wintertime temperature in the Northern Rocky mountains but negative relationships with precipitation, similar to the signature expressed during the positive phase of PNA. Comparisons with the PDO index (November-March) developed by Mantua et al. (1997) and a regionally averaged temperature dataset (see appendix A for details) and total monthly precipitation from Montana Climate Division 1 support some of the general patterns observed with PDO in
the Pacific Northwest. November to March PDO is positively related to March-May temperatures ($r=0.40$, $p<0.0001$) and negatively correlated with December-February precipitation ($r=-0.41$, $p<0.0001$) from Montana Climate Division 1 over the period 1902-1998. In addition, November to March PDO between 1937-1998 was negatively correlated with April 1 snow pack at Savage Pass, near the interior of the SBW ($r=-0.68$, $p<0.0001$).

Although PDO has important relationships with climate, the primary importance is related to its persistence. Unlike ENSO, PDO is a long-lived feature, so conditions can persist at decadal time-scales and alter climate patterns for comparatively long periods. Mantua et al. (1997) indicated two important shifts in PDO polarity had occurred during the 20th century. PDO was generally in its positive (warm, enhanced Aleutian low) between about 1925-1947 and since about 1977. Negative anomalies occurred between 1947-1976. Recent investigations appear to indicate we have now entered a period where the PDO is negative again, since about the late 1990s (Chavez et al 2003). This suggests there may be more frequent winter storms and cooler temperatures around the Selway region.

The importance of these teleconnections to climate variability in the Selway region is not entirely understood. However, these features can alter weather patterns throughout a season resulting in conditions that deviate from that of the expected climate patterns. Shorter-term weather patterns will often be superimposed upon these large-scale atmospheric patterns, but overall they do not generally disrupt the prevailing conditions brought about due to ENSO, PNA, or PDO.
Dettinger et al. (1998) suggest most of the interannual variability in precipitation in the northern United States is more consistent with respect to local circulation patterns than large scale features of atmospheric circulation such as those described by Pacific Basin teleconnections. However, longer-term, decadal variability in precipitation appeared to reflect conditions further away. This suggests that PDO may be an important control on decadal-scale drought patterns in the Pacific Northwest depending on the manner in which local scale factors are superimposed from year-to-year. PDO seems to have a particular relevance to the Selway region because it appears to mirror many of the effects brought about by both ENSO and PNA (e.g., dry/wet winters associated with the strong/weak Aleutian low). The persistence of the PDO, at decadal scales, could diminish the short term importance of both ENSO and PNA patterns on climate variations in the Selway region. Further, the decadal variability expressed by PDO could disrupt many ecosystem processes and have important, long lasting effects. For example, winter snowpack in the Bitterroot Mountains is an important source of irrigation water in the Bitterroot Valley. Long-term disruptions to winter snow pack conditions could lead to economic problems in the region. In addition, snowpack is an important source of moisture for subalpine forest communities and persists into July in many high elevation areas of the Selway.

**Human Impacts on Fire Regimes**

The extent of human activity within the SBW prior to settlement by Euro-Americans in the Bitterroot region is not well documented. It has been previously noted
that members of the Nez Perce tribe passed through the SBW seasonally from the Idaho portion of the SBW enroute to bison hunting grounds located on the plains of Montana on a road still used for recreational traffic today (Leiberg 1899; Moore 1996; Koeppen 1998). Lewis and Clark traveled through the area (with much help from their Native American guides) during the winter and summer of 1805 and 1806 (Ambrose 1996). The SBW was increasingly exploited during the late 19th century for extractive purposes such as mining, trapping, and timber harvest (Moore 1996). These uses continued in the SBW until its designation as a wilderness area in 1964 putting an end to resource extraction. However, hunting, fishing, and trapping are still permitted. It also remains a popular destination for backpackers, rafters, and kayakers but most use is along the periphery of the wilderness boundary.

The degree to which Native Americans altered natural fire regimes remains the subject of considerable debate (Denevan 1992; Baker 2002; Vale 1998). For the most part native populations were somewhat concentrated and localized in particular geographic areas, and therefore native populations’ effects on fire regimes is also likely localized (Baisan & Swetnam 1997; Vale 1998; Kaib 1998; Allen 2002). Increased fire frequency due to Native American burning practices has been identified in particular areas and times when native Americans were present and burning for a specified purpose. However, at broader spatiotemporal scales the impact of Native Americans on fire regimes, and the extent to which the fundamental ecological character of the landscape was altered by this burning remains ambiguous (Baisan & Swetnam 1997; Vale 1998; Kaib 1998, Kaye and Swetnam 1999; Baker 2003).
Prior to Euro-American settlement, Native American populations are thought to have had an important influence on fire regimes in some parts of western North America (Pyne 1982; Denevan 1992; Kaib 1998). In some areas, and during some parts of the year, Native Americans set fires to facilitate hunting, manage croplands, and improve travel corridors (Lewis 1985). The ignition of fires by Native Americans in certain locales and at particular times may be responsible for higher fire frequencies in some areas as compared to others with similar physical characteristics (Pyne 1982; Kaib 1998).

In the vicinity of the Selway-Bitterroot Wilderness, four important groups of Native Americans were present: the Nez Perce, Flathead, Pend d’Oreilles, and Kootenai tribes. These groups did reportedly ignite fires, but the timing and frequency of these fires is somewhat obscure. Most fires were apparently ignited in the fall and spring when weather conditions were more conducive to non-lethal surface fires (Barrett 1980). Barrett (1980) reports higher fire frequencies in areas known to have been occupied by Native Americans. Additionally he reports that these fires occurred in the area of low elevation forests and at the grassland-forest ecotone. Barrett’s results are speculative, however, as microclimatic and topographic differences between paired sites may be sufficient to effect the small changes in fire frequency between sites he observed. Barrett also provides no statistical test off the patterns of fire in areas used by Native Americans to indicate the significance of these differences.

There is little evidence of fire use by Native Americans in upper elevation environments. At upper elevations in many mountain ranges Native American populations are thought to be transient (Barrett 1980; Vale 1998). Native Americans
regularly traveled across the SBW from present day Idaho enroute to hunting grounds in Montana (Leiberg 1899; Moore 1996), and Leiberg suggests the abundance of second growth forest surrounding these routes was evidence of Native burning. Overall, however, direct evidence of Native American burning is scarce. High elevation signal fires were sometimes ignited, but are considered to be relatively rare events (Barrett 1980). Lewis and Clark report the ignition of a fir tree by their native guide during their trek eastward across the Bitterroots (Ambrose 1996).

Mehringer (1977) speculates that increases in charcoal abundance over the last 2,000 years in a pollen core taken from within the SBW could be due to the burning practices of Native Americans. He discounts the effect of climate variability as a possible explanation for increased charcoal abundance because the time period was relatively mesic. However, it is possible that short-term (i.e., seasonal) drought could cause widespread fires, which could produce abundant charcoal, even though the overall period is relatively wet.

In some areas of the Rocky Mountain west, increases in fire frequency are thought to be associated with EuroAmerican settlement activities in the late 19th and early 20th centuries (Goldblum & Veblen 1992; Honaker 1995; Kipfmueller & Baker 2000). Activities such as mining and tie hacking (harvest of timber for railroad cross ties) were regularly carried out in many places. In the SBW, Leiberg (Leiberg 1899) identifies the occurrence of several large fires though to have been either intentionally ignited by miners to clear undergrowth or the result of untended campfires. However, Leiberg attributes all the fires in the SBW region to anthropogenic causes, discounting lightning
completely. Moore (1996) also discusses the intentional ignition of fires to clear underbrush by miners in the SBW during the early 1900's, but suggests it was not widespread.

Of the potential impacts of humans on fire regimes in the western United States, fire suppression likely has had the largest impact. Fire suppression effects take on three important forms: 1) active suppression of fires using modern fire fighting techniques (e.g., Heinselman & Wright 1973; Arno 1976; Arno 1980; Barrett 1988; Barrett et al. 1991; Swetnam 1993); 2) indirect suppression of fires due to the reduction of fuel through activities such as grazing or fragmentation of contiguous forest cover (Savage & Swetnam 1990; Swetnam 1993; Baker 1995); and 3) increased frequency of stand-replacing fires due to increases in fuel abundance as a result of reduced fire activity which would have otherwise removed fuels (Covington & Moore 1994; Fulé et al. 1997; Mast et al. 1999). Declines in fire frequency since ca. 1900 have been reported throughout the western United States in numerous fire history investigations, and in some cases a nearly complete absence of 20th century fire is evident. The reduction of fire since ca. 1900 is most pronounced in low elevation forests that have historically experienced frequent non-lethal surface fires (Covington & Moore 1994).

Some debate remains over the role fire suppression has played on the fire regime of subalpine forests, (e.g., Habeck & Mutch 1973; Romme & Despain 1989; Arno & Brown 1991; Johnson et al. 2001). The effects of fire suppression may not be as evident in only 100 years of suppression activity because these forests have historically experienced comparatively long fire return intervals (ca. 100 years). However,
investigations in the Pacific Northwest and some places in the Rocky Mountains have found evidence that fire suppression in the 20th century has impacted high elevation environments by lengthening fire and fire return intervals (Hemstrom & Franklin 1982; Wadleigh & Jenkins 1996; Kipfmuller & Baker 2000; Rollins et al. 2001). Fire suppression could result in the suppression of fires at lower elevation, that, left to burn may have spread to higher elevation forests. In addition, advancements in suppression technology such as the incorporation of aerial suppression techniques and smoke jumpers, have resulted in rapid response to reported fires even in more remote upper elevation forests. For example, small spot fires reported from a remote fire lookout can be rapidly suppressed, which may otherwise have led to large areas burned in subalpine habitats.

Summary

The SBW has played an important role in developing fire policy as well as increasing our understanding of fire in western forests. The unique geographic setting of the SBW provides a diverse environment conducive to fire. The variability in climate across the SBW as well as the dramatic elevational changes yields diverse forests with a broad spectrum of fire regimes. While human impacts on pre-20th century fire regimes were likely small when compared with the role of climate, lower elevations probably experienced burning by Native Americans. However, only limited and fragmentary evidence suggests anthropogenic burning took place in the past. Fire regimes have been heavily impacted by fire suppression during the 20th century. This is particularly evident
in lower elevation forests. However, the effects of fire suppression on upper elevation forests and fire regimes have likely not been impacted to any great degree given that fire intervals are long and fire suppression has only been effective for about the last 50 years.
REFERENCES


Leiberg, J.B. 1899. The Bitterroot Forest Reserve. USGS 20th Annual Report, Part V.


APPENDIX A

CLIMATE-TREE GROWTH RELATIONSHIPS IN SUBALPINE CONIFERS IN THE NORTHERN ROCKY MOUNTAINS, U.S.A.

Kurt F. Kipfmüller

This paper will be submitted to the journal *Arctic, Antarctic, and Alpine Research*

Abstract

The relationship between temperature, precipitation and radial tree growth were examined for whitebark pine (*Pinus albicaulis* Englem.) and subalpine larch (*Larix lyallii* Parl.) collected in the Selway-Bitterroot Wilderness Area, located on the border of Idaho and Montana. Correlation analysis was used to compare climate variables with radial growth for three time periods, the full instrumental record (1901-1998), the early period (1901-1949), and a late period (1950-1998). Results indicated that over the entire time period the strongest relationship between climate variables and growth occurred during the summer months, with strong positive associations with temperature and negative associations with precipitation. The association between radial growth and temperature was strongest during the early period of the record but weakened in both species during the late period. The weakening is more evident in whitebark pine than in subalpine larch. During the later period, spring temperature and precipitation appeared more important and leads to a weakened summer response. Decadal scale changes in temperature and
precipitation variability, particularly with respect to snowpack and the Pacific Decadal Oscillation may be the cause of the weakened growth climate relationship observed.

Introduction

Understanding climatic variability at long time scales is becoming increasingly important as predictions of rising global temperatures continue to be refined. Since the late 19th century, average global temperatures have increased between 0.2°-0.6°C (IPCC, 2001). Recent estimates suggest that global temperatures may rise 1.4°-5.8° by 2100, coincident with a projected doubling of CO₂ (IPCC, 2001). While variability is an inherent part of the climate system, the rise in global temperatures over the last 100 years has occurred at a faster rate than at any time in the last 1000 years (Mann et al. 1998). Increases in average global temperatures may have important implications for natural systems including increases in fire activity (Price and Rind 1994; Flannigan et al. 2001), loss of timberline habitat (Bartlein et al. 1997), and more frequent and severe outbreaks of insect pests (Logan et al. 1995).

While most attention is focused on increasing trends in global or hemispheric temperatures, the warming also has important spatial variability. Some areas appear to have experienced more dramatic increases in temperature, such as high latitude and inland regions (IPCC, 2001). Further, past large scale climate anomalies such as the Medieval Warm Period and Little Ice Age, while initially thought to represent near global phenomena, have been found to have varying spatial signatures (Hughes and Diaz 1994; Jones and Briffa 2001). Determination of localized responses to climate variability is
crucial to developing a more complete picture of past climate variations and forcing mechanisms, and it gives the potential impacts of rising temperatures a spatial context.

The annual nature of tree-ring formation and the ability to statistically link growth with climate is an important tool in understanding climate variability at long time scales. Tree-rings are an important part of establishing a baseline by which recent climate variability and suspected increases can be gauged. Tree rings have been widely used to reconstruct a variety of climatological phenomena including precipitation patterns (Fritts et al. 1979; Grissino-Mayer 1995; Hughes and Graumlich 1996), annual and seasonal temperature variations (Graumlich and Brubaker 1986; Briffa et al. 1992, 1994; Luckman et al. 1997; Hughes et al. 1999; Salzer 2000), atmospheric circulation patterns (Villalba et al. 1997; Woodhouse 1997; Ni 2000), drought indices (Cook et al. 1996), and variations in streamflow (Meko and Graybill 1995).

In the northern Rocky Mountains, conifers growing near upper treeline have tremendous potential for developing long term, high resolution proxies of climate variability. Both whitebark pine (*Pinus albicaulis* Engelm.) and subalpine larch (*Larix lyallii* Parl.) are found in upper elevation environments of Idaho and Montana and have been used successfully elsewhere in developing reconstructions of past climate variability (Graumlich and Brubaker 1986; Luckman et al. 1997). These species are long lived, sometimes exceeding 700 years in age (Luckman et al. 1984; Perkins and Swetnam 1996). Slow rates of decomposition also enable the extension of the tree-ring record through the use of remnant materials (Tessier et al. 1997). Further, the life histories of these two species are different, with whitebark pine being evergreen and subalpine larch
deciduous. Used in concert, these species may complement one another due to differing physiological traits.

The extraction of paleoclimate signals from tree rings requires the development of regression equations or neural networks based upon the dominant mode of tree response to climate, or those factors that are most limiting to tree growth (Fritts 1976). This typically involves calibrating a regression equation using a subset of observed climate data and comparing the relationships with data withheld from the model to evaluate reconstruction quality. Recent studies have raised questions concerning the validity of some climate reconstructions recognizing that the statistical relationships in many elevational and latitudinal treeline chronologies has weakened or changed over the course of the 20th century (Briffa et al. 1998a; Briffa et al. 1998b; Vaganov et al. 1999; Biondi 2000; Lloyd and Fastie 2002). If the relationships between tree growth and climate are not temporally stable, then this presents new challenges for the development of reconstructions, and interpretations of past climate based on features of the present climate system.

The aim of this study was to describe the general patterns of whitebark pine and subalpine larch growth to climate and to determine the feasibility of climate reconstructions in Selway-Bitterroot region of the Northern Rocky Mountains. This research addressed three main questions. First, what were the general patterns of tree-growth response to climate in these treeline conifers? Using correlation and response function analysis, I analyzed the association between tree growth, and monthly temperature and precipitation using a twelve month window to identify the dominant
mode of response of whitebark pine and subalpine larch radial growth to climate variation. Daily temperature averaged over five day periods (pentads) were also examined to identify the fine scale climate-tree growth relationships during the growing season. Second, were tree growth-climate relationships stable over the 20th century? Using a split sample approach and moving correlations I assessed the strength of the tree growth-climate associations through time to identify possible reductions in sensitivity to climate, or switches in the dominant mode of climatic response. This procedure helped identify those patterns of tree growth-climate relationships that were temporally consistent and could be reconstructed using traditional dendroclimatological procedures. Finally, if reductions in sensitivity were evident, are there identifiable mechanisms that explain these changes over the 20th century?

Study Area

Tree-ring collection sites were located in upper treeline environments within and near the Selway-Bitterroot Wilderness Area (SBW, see Table 1 for a list of acronyms) located on the border of Idaho and Montana (Figure 1, Table 2). These sites had limited fuel continuity to support spreading fires and they were in remote locations with limited access, thus reducing anthropogenic activities and impacts. However, some evidence existed of past mortality in some whitebark pine stands due to mountain pine beetle epidemics, mostly occurring around the late 1920's (Evenden 1944; Ciesla and Furniss 1975; Kipfmueller et al. 2002). Whitebark pine killed by mountain pine beetle were present as dead standing snags with external J-shaped galleries typical of mountain
Table 1. List of frequently used acronyms and their descriptions.

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Definition</th>
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</thead>
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<tr>
<td>SBW</td>
<td>Selway-Bitterroot Wilderness Area.</td>
</tr>
<tr>
<td>SMMWB</td>
<td>Salmon Mountain whitebark pine chronology.</td>
</tr>
<tr>
<td>SMMLY</td>
<td>Salmon Mountain subalpine larch chronology.</td>
</tr>
<tr>
<td>BKWB</td>
<td>Baker Lake whitebark pine chronology.</td>
</tr>
<tr>
<td>BKRLY</td>
<td>Baker Lake subalpine larch chronology.</td>
</tr>
<tr>
<td>CTRWB</td>
<td>Carlton Ridge whitebark pine chronology.</td>
</tr>
<tr>
<td>CTRLY</td>
<td>Carlton Ridge subalpine larch chronology.</td>
</tr>
<tr>
<td>PIALPCS</td>
<td>Whitebark pine principal component growth chronology constructed using the standard whitebark pine chronologies.</td>
</tr>
<tr>
<td>LALYPCS</td>
<td>Subalpine larch principal component growth chronology constructed using the standard whitebark pine chronologies.</td>
</tr>
<tr>
<td>PIALPCR</td>
<td>Whitebark pine principal component growth chronology constructed using the residual (prewhitened) whitebark pine chronologies.</td>
</tr>
<tr>
<td>LALYPCR</td>
<td>Subalpine larch principal component growth chronology constructed using the residual (prewhitened) whitebark pine chronologies.</td>
</tr>
<tr>
<td>PIALPCLV</td>
<td>Whitebark pine principal components chronology constructed using the standard chronologies from living trees.</td>
</tr>
<tr>
<td>PSEP</td>
<td>September climate data prior to the year of tree ring formation.</td>
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<tr>
<td>POCT</td>
<td>October climate data prior to the year of tree ring formation.</td>
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<tr>
<td>PNOV</td>
<td>November climate data prior to the year of tree ring formation.</td>
</tr>
<tr>
<td>PDEC</td>
<td>December climate data prior to the year of tree ring formation.</td>
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<tr>
<td>JAN</td>
<td>January climate data for the current year of growth.</td>
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<tr>
<td>FEB</td>
<td>February climate data for the current year of growth.</td>
</tr>
<tr>
<td>MAR</td>
<td>March climate data for the current year of growth.</td>
</tr>
<tr>
<td>APR</td>
<td>April climate data for the current year of growth.</td>
</tr>
<tr>
<td>MAY</td>
<td>May climate data for the current year of growth.</td>
</tr>
<tr>
<td>JUN</td>
<td>June climate data for the current year of growth.</td>
</tr>
<tr>
<td>JUL</td>
<td>July climate data for the current year of growth.</td>
</tr>
<tr>
<td>AUG</td>
<td>August climate data for the current year of growth.</td>
</tr>
</tbody>
</table>
Figure 1. The Selway-Bitterroot study region. (A) The location of the Selway-Bitterroot Wilderness Area along the Idaho/Montana border. (B) Tree-ring sampling sites (gray diamonds) and climate stations used for the regional temperature series (numbers). Station numbers correspond to the stations listed in Table 3.
Table 2. Chronology statistics for tree-ring chronologies developed in the Selway-Bitterroot Wilderness Area.

<table>
<thead>
<tr>
<th>Location (lat./lon.)</th>
<th>Salmon Mountain</th>
<th>Baker Lake</th>
<th>Carlton Ridge</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>45.6° N, 114.8° W</td>
<td>45.9° N, 114.2° W</td>
<td>46.7° N, 114.2° W</td>
</tr>
<tr>
<td>Elevation (m)</td>
<td>2,725</td>
<td>2,670</td>
<td>2,650</td>
</tr>
<tr>
<td>Species'</td>
<td>PIAL</td>
<td>LALY</td>
<td>PIAL</td>
</tr>
<tr>
<td>No. Trees (Series)</td>
<td>31 (50)</td>
<td>30 (45)</td>
<td>50 (81)</td>
</tr>
<tr>
<td>M.S.L.²</td>
<td>267</td>
<td>298</td>
<td>494</td>
</tr>
<tr>
<td>I.S.C.³</td>
<td>0.562</td>
<td>0.680</td>
<td>0.518</td>
</tr>
<tr>
<td>M.S.⁴</td>
<td>0.206</td>
<td>0.317</td>
<td>0.216</td>
</tr>
<tr>
<td>1st PC⁵</td>
<td>42</td>
<td>47</td>
<td>25</td>
</tr>
<tr>
<td>1st A.C.F.⁶</td>
<td>0.78</td>
<td>0.63</td>
<td>0.75</td>
</tr>
</tbody>
</table>

¹PIAL = Whitebark Pine, LALY = Subalpine Larch; ²Mean segment length (years); ³Mean inter-series correlation; ⁴Mean Sensitivity (Fritts 1976); ⁵Percent variance explained by the first principal component of the tree-ring chronology. ⁶Mean first order autocorrelation in the tree-ring series before detrending.
pine beetles (Wood 1982). The impact of mountain pine beetles on whitebark pine ring-width signature was limited within these chronologies because trees were typically killed outright rather than experiencing a reduction in ring-width, as is common with other insect defoliators (e.g., Swetnam and Lynch 1993).

The climate of the SBW is transitional from a wetter Pacific Northwest type climate to a continental dry rainshadow climate (Finklin 1983; Ferguson 1999). Cold winters and mild summers typify the temperature regime of the region with winter temperatures averaging -2.7°C in winter and 17.4°C in summer recorded by nearby lower elevation stations. Precipitation ranged from 625 mm at lower elevations (500 m) near the southern end of the wilderness to about 1000 mm in the western portion and at higher elevations (Finklin 1983). Most of the precipitation falls as snow during the winter months. However, there is a gradient in the seasonality of precipitation with proportionally more precipitation falling in winter in the western portion of the study area (Finklin 1983; Ferguson 1999; Brunelle-Daines 2002).

Whitebark pine and subalpine larch coexist near timberline with subalpine fir (Abies lasiocarpa (Hook.) Nutt.) and Engelmann spruce (Picea engelmannii Parry). Subalpine larch are often found in nearly pure stands at elevations where other trees are relegated to a krummholz form (Arno and Habeck 1972; Richards 1984). The deciduous nature of subalpine larch enables their persistence at these extreme sites in the absence of other upright conifers where winter dessication and needle abrasion constitutes an important limiting factor to establishment and survival (Arno and Habeck 1972; Tranquillini 1979; Arno and Hammerly 1984; Richards and Bliss 1986).
Methods

Chronology development

Tree-ring series were collected from whitebark pine (PIAL) and subalpine larch (LALY) growing near alpine treeline using an increment borer and standard dendrochronological collection techniques (Fritts 1976). Whitebark pines exhibiting multi-stemmed growth form were omitted from climatological analysis. Two increment cores were extracted from each tree. All increment cores were crossdated to ensure the precise assignment of calendar dates. Crossdating accuracy was verified using program COFECHA. Ring-widths were measured to the nearest 0.01 mm on an incremental measuring bench. Increment cores with less than about 250 annual rings were omitted to preserve as much low frequency variability in the final tree-growth chronology as possible (Cook et al. 1995).

Dimensionless indices of tree-ring width were developed by detrending each series with a line fit to the measured values, then dividing the measurements by the values of the curve using ARSTAN (Cook and Holmes 1986). The detrended series were combined using a biweight robust estimate of the mean to minimize the effect of outlier values (Cook 1985). Measured ring widths were detrended using a 250 year cubic spline to remove trends related to small scale disturbances such as stem death or changes in bole geometry with increasing age.

A relatively stiff spline was selected for detrending individual tree-ring series rather than the more conservative negative exponential or linear detrending methods used in other dendroclimatological reconstructions. This approach was followed for two
reasons: First, sampled trees were usually extremely old and typically the innermost rings were not collected due to rotten centers. Because increment cores often did not reach the center of the trees, ring-widths did not often exhibit the negative exponential decline in ring widths commonly found with increasing tree age. Instead, plots of raw ring-width measurements indicated samples were typically already in the nearly horizontal “tail” of growth with little or no evidence of rapid juvenile growth near the inner rings of the samples.

Secondly, with respect to whitebark pine, it was likely that some of the sampled trees germinated from unused seed caches, even though they are single-stemmed today (Hutchins and Lanner 1982; Tombback et al. 1990). Clark’s nutcrackers (*Nucifraga columbiana*) are the primary dispersal agent of the large, heavy seeds of whitebark pines. These birds cache the seeds in groups of about 3–4 where uncollected seeds can germinate in a multi-stemmed growth form (Tomback et al. 1990). This morphological trait can have important influences on tree-growth patterns due to competitive interactions and mortality of the multiple stems. As individual stems die, the ring-widths of those remaining may record these deaths as a release from suppression, a disturbance unrelated to climate. Smoothing splines have been used successfully to detrend interior forest species that may also have experienced some form of disturbance (Cook and Peters 1981) and are therefore a good alternative to detrending whitebark pine that may have germinated from seed caches.
Historical climate records

Few long-term climate stations were present in the area and most were located at relatively low elevations and a considerable distance away from the tree-ring collection sites (>200 km). National Climate Data Center State Divisional Data is also likely biased in this respect and incorporates a majority of stations located well away from the study region. Further, complex topography in the area has important influences on climate at a finer scale than that represented by divisional data. Luckman (1997) suggests a useful approach for calibrating tree-growth to climate where limited station data exists by averaging the climate data from the longest and nearest station records available. Following this approach I selected the four climate stations nearest the tree-ring collection sites with the longest records and created a regional average monthly temperature record (Figure 1, Table 3). These stations are found primarily along the eastern front of the Bitterroot Mountains in the Bitterroot Valley (Figure 1) and are relatively close to the tree-ring collection sites (average distance between climate stations and tree-ring sites ≈68 km). They are, however, considerably lower in elevation than the tree-ring collection sites, but no high elevation stations exist in the region with sufficiently long records for the reconstruction of climate.

I used a simple averaging procedure to generate regional temperature series of mean minimum, average, and mean maximum monthly temperatures. While these stations were also generally located in low elevation valley bottoms, their geographic location nearer the tree-ring sites probably better reflects the general patterns of climate in the area better than the full Divisional data set that covers a much larger geographical
area. This subregional temperature time series is not statistically different from the Montana Division 1 temperature record (p>0.05, t-test), but localized differences in these records may be important to tree-growth response. Many of these same stations had incomplete precipitation records so Montana Climate Division 1 (Western Montana) total monthly precipitation was used in its place.

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Data Source</th>
<th>Lat./Lon.</th>
<th>Elev. (m)</th>
<th>Time Span</th>
</tr>
</thead>
<tbody>
<tr>
<td>Missoula WNW, MT</td>
<td>NCDC</td>
<td>46.88N 114.03W</td>
<td>967</td>
<td>1893-1965</td>
</tr>
<tr>
<td>Stevensville, MT</td>
<td>NCDC</td>
<td>46.52N 114.10W</td>
<td>1,029</td>
<td>1911-1998</td>
</tr>
<tr>
<td>Hamilton, MT</td>
<td>HCN</td>
<td>46.23N 114.17W</td>
<td>1,076</td>
<td>1900-1998</td>
</tr>
<tr>
<td>Darby, MT</td>
<td>NCDC</td>
<td>46.02N 114.18W</td>
<td>1,183</td>
<td>1948-1998</td>
</tr>
</tbody>
</table>

1Station number corresponds to the location numbers in Figure 1.

**Tree-growth response to climate**

To simplify the analysis and identify the general relationships between climate and the two tree species examined, the first principal component (PC) from the three standard whitebark pine growth chronologies, and the first PC of the three standard subalpine larch growth chronologies were extracted (hereafter referred to as PIALPCS and LALYPCS). The impetus for using Principal Components Analysis in this study was to extract the common variance and thus enhance the common climate signal from the two species. This procedure reduces any local effects within individual sites. It is important to note, however, that differences between sites could have important effects
on the climate-tree growth relationship but the aim here is to determine the main features of tree-growth response of these species. The first PCs from the standard chronologies explain 68% and 75% of the variance for whitebark pine and subalpine larch respectively. A similar principal components procedure was used on the residual (prewhitened to order 1) chronologies for comparison purposes. Residual chronologies of tree growth were used in addition to the standard chronologies to limit the effect of temporal autocorrelation in the growth chronologies. First order autocorrelation is greater in the whitebark pine chronologies than the subalpine larch chronologies (Table 4). While this autocorrelation may be an important feature of the growth of these conifers related to climate, it presents statistical challenges to identifying the months that are most important to growth processes for climate reconstruction purposes due to a reduction in the degrees of freedom available for most statistical tests. The first PC of the residual chronologies (PIALPCR and LALYPCR) explained 78% and 76% of the variance respectively.

<table>
<thead>
<tr>
<th>Time Period</th>
<th>Whitebark Pine (PIALPCS)</th>
<th>Subalpine Larch (LALYPCS)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1250-1998</td>
<td>0.323</td>
<td>0.288</td>
</tr>
<tr>
<td>1901-1998</td>
<td>0.498</td>
<td>0.220</td>
</tr>
<tr>
<td>1901-1949</td>
<td>0.519</td>
<td>0.299</td>
</tr>
<tr>
<td>1950-1998</td>
<td>0.387</td>
<td>0.130</td>
</tr>
</tbody>
</table>

Table 4. First order autocorrelation remaining in the principal component derived tree-ring chronologies constructed from standard growth indices. The subalpine larch chronology extends only through 1997.
Since the ultimate goal of the tree growth-climate modeling was to identify variables useful for climatic reconstruction using linear regression techniques, correlation analysis was used to identify tree growth-climate relationships. Pearson’s correlations were calculated between monthly climate data and the first PC of both the whitebark pine and larch standard and residual chronologies. The climate data included average, average minimum, and average maximum monthly temperature, and total monthly precipitation for twelve months. The climate window examined included September through December of the years prior to tree ring formation, and January through August of the growth year.

An important limitation of correlating tree growth with climate results from a high degree of collinearity in climate data. This can reduce the effectiveness of identifying meaningful relationships between tree-growth and climate. Response function analysis (Fritts 1976; Fritts and Xiangding 1986), a multiple regression technique that uses the principal components of the climate data to predict tree growth to overcome the issue of multicollinearity, was not used to examine tree-growth climate relationships. While this technique does reduce the problem of multicollinearity in the climate data, the inclusion of higher order PCs can make interpretation of the climate window difficult. In addition, statistical significance is often easier to achieve using response functions than using simple correlations, which are intuitive (Blasing et al. 1984). Initial exploratory analysis using response functions indicated that the PC regressions of the climate data usually explained little of the variance (typically less than 40%) even if 19 PC regressors were
used as predictors of tree growth. Moreover, the main patterns captured by correlations were nearly always the same as response functions.

Initially, the relationship between temperature and tree growth was examined at a fine temporal scale using daily temperature data from Hamilton, Montana for the months of April through September of the current growing season. The daily Hamilton, MT temperature data covers the period 1912-1998. The average temperature over five consecutive days (pentads, Vaganov et al. 1999) was compared with PIALPCS and LALYPCS tree growth chronologies to assist in identifying the onset of growth processes in these species. No reliable daily precipitation data is available for this region, so only temperature variables were examined in this manner.

Tree-growth-climate relationships were also examined for the entire time period, split time periods (analogous to traditional dendroclimatological investigations), and using moving correlations. For the split time period analysis, the time series of the climate data was split exactly in half, with the early period covering 1901-1949 and the later half 1950-1998. One subalpine larch chronology only extended to 1997, and therefore 1997 was used as the end date in the subalpine larch analyses.

Moving correlations were calculated using a centered 31-year window successively shifted forward one year. The centered 31-year window resulted in a symmetric window but unfortunately also results in a 15 year truncation of information from both ends of the time series of tree growth. Moving correlations were used as an aid in identifying the temporal patterns of tree growth-climate relationships over the entire 20th century.
Results

Tree-ring chronology characteristics

These well replicated tree-ring chronologies are the longest tree-ring chronologies collected in the Selway-Bitterroot Wilderness region. Previous tree-ring collections from Engelmann spruce located within 100 km of the study area extend only to 1758 (International Tree-Ring Data Bank, 2002). However, other whitebark pine chronologies reported by Perkins and Swetnam (1996) extend to about AD 726, but are about 250 km to the south.

The chronology statistics suggested there was good potential in these chronologies for producing long-term reconstructions of climate (Table 2). Although there was a strong common signal within the sites as illustrated by the inter-series correlations, the relationship between sites is weaker (Table 5). Inter-series correlations (within sites) ranged from 0.518-0.574 for whitebark pine and 0.653-0.696 for subalpine larch. However, the variance explained by first PC of the individual site chronologies was reasonably good, ranging from 25-50% (Table 2). Between site correlations of the standard chronologies ranged from 0.454 to 0.577 for whitebark pine and 0.618 to 0.626 for subalpine larch (Table 5). Between site correlations of the residual chronologies was somewhat higher ranging between 0.644-0.704 for whitebark pine and 0.634-0.655 (Table 5).

When the relationships between the two species are considered, correlations are reduced considerably, ranging from 0.247 to 0.459 for standard chronologies and 0.333-0.431 for the residual chronologies. The variance explained by the standard and residual
principal component chronologies is still relatively high, suggesting there was some intersite similarity within the species examined. Variance explained by the first PC was higher in the chronologies derived from the residual series than that of the standard PC chronologies.

Table 5. Pearson correlations among the site, species, and PC chronologies over the period 1250-1998. Correlations using standard chronologies shown in the lower left, those from residual chronologies shown in the upper right. All correlations statistically significant at $p<0.001$.

<table>
<thead>
<tr>
<th></th>
<th>SMMWB</th>
<th>BKRWB</th>
<th>CTRWB</th>
<th>PIALPC</th>
<th>SMMLY</th>
<th>BKRLY</th>
<th>CTRLY</th>
<th>LALLYPC</th>
</tr>
</thead>
<tbody>
<tr>
<td>SMMWB</td>
<td>--</td>
<td>0.644</td>
<td>0.668</td>
<td>0.870</td>
<td>0.387</td>
<td>0.375</td>
<td>0.352</td>
<td>0.425</td>
</tr>
<tr>
<td>BKRWB</td>
<td>0.454</td>
<td>--</td>
<td>0.704</td>
<td>0.886</td>
<td>0.342</td>
<td>0.399</td>
<td>0.333</td>
<td>0.410</td>
</tr>
<tr>
<td>CTRWB</td>
<td>0.536</td>
<td>0.577</td>
<td>--</td>
<td>0.896</td>
<td>0.419</td>
<td>0.431</td>
<td>0.429</td>
<td>0.488</td>
</tr>
<tr>
<td>PIALPCS</td>
<td>0.800</td>
<td>0.820</td>
<td>0.860</td>
<td>--</td>
<td>0.433</td>
<td>0.455</td>
<td>0.420</td>
<td>0.499</td>
</tr>
<tr>
<td>SMMLY</td>
<td>0.415</td>
<td>0.296</td>
<td>0.405</td>
<td>0.451</td>
<td>--</td>
<td>0.655</td>
<td>0.634</td>
<td>0.874</td>
</tr>
<tr>
<td>BKRLY</td>
<td>0.321</td>
<td>0.368</td>
<td>0.414</td>
<td>0.447</td>
<td>0.618</td>
<td>--</td>
<td>0.641</td>
<td>0.877</td>
</tr>
<tr>
<td>CTRLY</td>
<td>0.302</td>
<td>0.247</td>
<td>0.459</td>
<td>0.410</td>
<td>0.626</td>
<td>0.619</td>
<td>--</td>
<td>0.868</td>
</tr>
<tr>
<td>LALLYPCS</td>
<td>0.400</td>
<td>0.351</td>
<td>0.493</td>
<td>0.504</td>
<td>0.865</td>
<td>0.862</td>
<td>0.866</td>
<td>--</td>
</tr>
</tbody>
</table>

Whitebark pine response to climate

The results of the correlation analysis suggested the strongest relationships between the PIALPC chronologies and climate occurred primarily during the summer months (June-August, Figures 2 & 3). Daily temperature data indicated the strongest relationship between tree growth and temperature occurred during July and August (Figure 2). The relationship was positive for maximum July temperature indicating that warm daytime temperatures were conducive to whitebark pine growth. However, strong inverse correlations were evident between whitebark pine growth and minimum temperatures during the month of August and into September. The strong negative
Figure 2. Correlations between PIALPCS and the average five day temperatures (pentads) during the growing season. Climate data from Hamilton, MT (1912-1998). Dotted line indicates the approximate 95% confidence limit.
Figure 3. Correlation coefficients for standard and residual PIALPC chronologies. (A) Full time period (1901-1998). (B) Split time periods. Dark circles represent the early period (1901-1949), white circles the late period (1950-1998). Dotted line is the approximate 95% confidence limit.
relationship between growth and average minimum temperature yields correlations near zero for average temperature.

Over the full time period (1901-1998) July and August appeared to be the most consistent months with significant correlations (Figure 3A). In general, correlations between climate and PIALPCR were weaker, exhibiting fewer significant relationships (Figure 3A). The strongest individual correlation between the PIALPCS and climate over the twelve month period examined was with average maximum July temperature \((r=0.31, p<0.05)\). Average maximum July temperature also exhibited the highest correlation with PIALPCR, although somewhat lower than the PC developed using the standard chronology \((r=0.26, p<0.05)\). Summer average and average minimum temperature also had significant positive relationships with tree growth over the full time period, varying somewhat between PIALPCS and PIALPCR (Figure 3A). The only significant correlation between total monthly precipitation and the PIALPCS was negative and occurred during the month of July, although both July and August were significantly related in the PIALPCR chronology.

**Temporal relationships of whitebark pine-climate**

The climate-tree growth relationship had a weakened relationship with summer temperature from the early half of the climate record (1901-1949) to the later half (1950-1998, Figure 3B). Maximum July temperature remained the strongest correlation during the early period in both the standard and residual PC chronologies. July temperature was significant for all three temperature variables (maximum, average, and minimum) and
PIALPCS during the early period, but was not significant in the residual chronologies (Figure 3B).

While the summer temperature patterns remained similar during the late period, they were reduced and were no longer significant. Instead, there was a general trend toward strong inverse relationships between whitebark pine growth and spring temperatures (Figure 3B). Interestingly, spring temperature correlations switch sign from positive during the early period, to negative during the later period for average and minimum temperature respectively (Figure 3B). The strongest relationship during the late period occurred between maximum March temperature and PIALPCS (-0.38, p<0.05). Similar patterns of tree growth-climate response from the early to late period occurred when the PIALPCR was compared.

In general, early period correlations were more strongly related to summer temperatures while summer temperature, while still positive, appeared less limiting to growth during the later half of the record. Winter and spring conditions had a stronger relationship to growth than did summer conditions during the later period. July precipitation was significant and negatively correlated with growth for all time periods examined while May precipitation was inversely related only during the early period. This latter relationship was significant only with the PIALPCS chronology (r=-0.33, p<0.05).

The 31 year window of running correlations between monthly climate variables and PIALPCS generally supported the weakening of summer temperature after about 1940 (Figure 4). Spring temperatures become more strongly related to growth processes
Figure 4. Thirty-one year moving correlations between monthly climate variables and PIALPCS. Correlations < -0.30 or >0.30 are significant at p<0.10.
during the later half of the 20th century, although this relationship switched sign from positive to negative (Figure 4). A weakening of the summer precipitation relationship is also evident in the 31 year running correlations centered around 1950, coincident with the emergence of a positive relationship during late spring or early summer (Figure 4).

**Subalpine larch response to climate**

The results of the correlation analysis suggest the strongest relationship between growth of subalpine larch and climate also occurred primarily during the summer months, particularly during June and July (Figures 5 & 6). Daily temperature data indicated the strongest relationship between tree growth and temperature occurred somewhat earlier than in the whitebark pine, with a positive association with June temperatures being dominant (Figure 5). The strong inverse relationship in August identified in the whitebark pine growth relationships is not present in the subalpine larch chronology and likely reflects the physiological differences of the two species.

Over the full time period (1901-1998) June and July appear to be the most consistent months with significant correlations. The strongest individual month correlation between both the standard and residual PC chronologies and climate over the twelve months examined is with average June temperature ($r=0.45$, $p<0.05$, Figure 6A). Average maximum and average minimum temperature also have significant positive relationships with tree growth over the full time period for various summer months (Figure 6). Significant correlations between total monthly precipitation and the subalpine
Figure 5. Correlations between LALYPCS and the average five day temperatures (pentads) during the growing season. Climate data from Hamilton, MT (1912-1998). Dotted line indicates approximate 95% confidence limit.
Figure 6. Correlation coefficients for standard and residual LALYPC chronologies. (A) Full time period (1901-1998). (B) Split time periods. Dark circles represent the early period (1901-1949), white circles the late period (1950-1998). Dotted line is the approximate 95% confidence limit.
larch PC chronologies were negative and occurred during the months of prior October, January (not significant in the residual chronology), and July.

**Temporal relationships of subalpine larch growth-climate**

The temperature relationship between the LALYPC and summer climate appeared more temporally stable than did that of the whitebark pine, particularly for the month of June (Figure 6). The strongest correlation for the early period is minimum June temperature and the LALYPCS (0.50, p< 0.05), while maximum June temperature is the strongest for the late period when compared to LALYPCS (0.45, p< 0.05). Comparisons between climate and the residual chronologies suggested similar relationships. Spring temperature relationships also indicated an apparent switch from positive relationships in the early period to negative during the late period, similar (but weaker) to the switch observed in whitebark pine.

The 31 year window of running correlations does not show the reduction in the relationship between growth and climate through time that was evident in the whitebark pine (Figure 7). Summer temperature-growth relationships appeared to be generally stable throughout the entire time period. The relationship between summer maximum and minimum temperatures appears to weaken slightly around the late 1940s and early 1950s, coincident with the emergence of significant winter/spring precipitation variables (Figure 7).
Figure 7. Thirty-one year moving correlations between monthly climate variables and LALYPCS. Correlations < -0.30 or >0.30 are significant at p<0.10.
Discussion

Chronology characteristics

Crossdating in these species was generally good with considerable coherence in the common signal over large areas as evidenced by the variance explained by the PC chronologies. The whitebark pine and subalpine larch chronologies presented here had similar dendrochronological characteristics to chronologies constructed using these species elsewhere. Perkins and Swetnam (1996) produced whitebark pine chronologies at sites approximately 250 km to the south of the SBW. They report inter-series correlations ranging from 0.56-0.63, slightly higher than those reported here, but with mean sensitivities between 0.12-0.17, lower than in these chronologies. Colenutt and Luckman (1991) sampled subalpine larch sites approximately 600 km to the north. Inter-series correlations in their chronologies are around 0.80 with mean sensitivity between about 0.3 and 0.4, both slightly higher than in these subalpine larch chronologies. This discrepancy may be due to differences in climate, with the Perkins and Swetnam sites being much drier and the Colenutt and Luckman sites somewhat wetter than the SBW.

Considering the close proximity of the sampled sites in the SBW, correlations between sites was lower than expected, but nonetheless significant (p<0.001). While the crossdating of the individual tree-ring series mostly relied on the same narrow marker years from site-to-site, there were apparent differences in the ring-width chronologies. Correlations would be expected to be highest between the two closest sites (SMM and BKR), approximately 50 km apart (Figure 1). However, correlations between these two sites were actually lowest for both the whitebark pine and subalpine larch standard
chronologies (Table 3). This discrepancy may be due to the fact that the Baker Lake site is located to the east of the Bitterroot Range and likely experiences a strong rainshadow effect while both the Salmon Mountain and Carlton Ridge sites lie to the west of the crest. This might impart a stronger moisture signature in the Baker Lake chronology leading to differences between these sites. However, during years in which a narrow ring was formed (consistent from site-to-site for the most part), temperatures may be cool throughout the region and the Baker Lake site may be limited more by temperature.

Tree-growth response to climate

Upper elevation trees are usually considered to be most sensitive to summer temperatures (e.g., Fritts et al. 1965; LaMarche and Stockton 1974; Fritts 1976; Tranquillini 1979; Graumlich and Brubaker 1986; Kienast et al. 1987), as is the case in these chronologies. Low temperatures limit respiration and photosynthesis as well as other physiological processes. The onset and duration of the growing season, and timing of snowmelt, is also controlled by low temperatures (Tranquillini 1979). Growing season length is more limited by temperatures at the beginning of the growing season than by conditions near the end (Tranquillini 1979). The growing season at upper treeline in the SBW region is very short, with only about 90-110 frost-free days in stands containing whitebark pine and subalpine larch (Arno and Habeck 1972; Arno and Hoff 1990).

Given that the growing season is of limited duration, it is not surprising that temperatures during the summer season are the primary limiting factor of growth for
these subalpine species. Correlations clearly supported this relationship over the period 1901-1998. In the SBW region both whitebark pine and subalpine larch growth benefited from warm and dry conditions during the summer months, while cool and moist conditions were detrimental.

The strong inverse relationship between growth and July precipitation in both whitebark pine and subalpine larch was more difficult to explain. The physiological mechanism for this inverse relationship is unclear. Soil moisture availability is generally good in treeline environments due to the abundance of snow that can persist late into the summer months (Tranquillini 1979; Weaver 1990). Precipitation during summer in this area was usually low (about 12 cm) accounting for about 20% of the annual total. However, the summertime total can be misleading because it is driven primarily by June precipitation, accounting for an average of more than 50% of the three month summer (June, July, and August) precipitation total between 1901-1998. It seems unlikely that the small amount of precipitation falling in July and August would have much effect on tree-growth given that the relationship is inverse. The most plausible explanation for the inverse relationship observed with July precipitation is the reduction of incoming solar radiation due to cloud cover and associated cool temperatures (Fritts 1976; Tranquillini 1979; Rolland 1996). July temperature and July precipitation have a strong inverse correlation with one another ($r = -0.575$, $p < 0.001$), suggesting that it is indeed cooler when precipitation is higher even at the coarse scale of a month.

The prior October and July precipitation relationships found in the LALYPC chronologies are also more likely related to reduced temperatures due to cloud cover than
a hydrologic limitation. The January relationship may be related to the accumulation of snowpack that could carryover to the growing season and induce water stress. If snowpack persists late into the growing season soil water will remain frozen and unavailable to trees when temperatures are warm enough for respiration to take place resulting in a net loss of photosynthates (LaMarche and Stockton 1974; Graumlich and Brubaker 1986; Vaganov et al. 1999).

While strong similarities exist between the subalpine larch chronologies and climate presented here and those found elsewhere, there were important differences with respect to the relationships between whitebark pine and climate. The whitebark pine chronologies presented by Perkins and Swetnam (1996) were more strongly limited by winter-spring precipitation than summer temperature. In their sites July temperature has an inverse relationship to growth, the opposite effect of that observed in the SBW. This is not surprising given that their sites were considerably more arid, whereas the SBW region has climate conditions strongly influenced by maritime Pacific airmasses (Mitchell 1976; Finklin 1983; Ferguson 1999). In the SBW sites water stress is probably infrequent during most years due to snowfall at these upper elevation sites, particularly late in the spring. Snowpack was often greater than 1 m in depth in late June and early July during data collection in these areas

Luckman et al. (1997) have also described positive associations between tree growth and summer temperatures in southern Alberta and British Columbia, Canada in upper elevation tree species including whitebark pine. The relationships identified here between subalpine larch and growth are similar to findings from other areas with strong
summer relationships, particularly in June, being the norm (Graumlich and Brubaker 1986; Peterson and Peterson 1994; Colenutt and Luckman 1995).

**Potential causes of shifting climate-growth responses**

The strong positive relationship between growing season temperature and tree growth over the entire 20\textsuperscript{th} century broke down considerably when shorter periods were examined. This was particularly evident in whitebark pine. Whitebark pine growth response appears to be more limited by spring conditions during later portions of the 20\textsuperscript{th} century (Figures 3 and 4).

Recent observations of the reduction or alteration in climate-tree growth relationships in the 20\textsuperscript{th} century have been identified by many authors (e.g., Briffa et al. 1998a; Briffa et al. 1998b; Vaganov et al. 1999; Biondi 2000; e.g., Lloyd and Fastie 2002). A number of potential mechanisms have been explored to explain this including increased drought stress due to higher temperatures or exceedence of some upper threshold where warm temperatures become detrimental to growth (Lloyd and Fastie 2002), changes in growing season length or moisture availability due to the timing of snow melt (Vaganov et al. 1999; Biondi 2000), or anthropogenic changes to atmospheric chemistry (Briffa et al. 1998a; Briffa et al. 1998b).

Lloyd and Fastie (2002) and Briffa et al. (1998b) have noted that tree growth in treeline environments (alpine and latitudinal) has not kept pace with temperature increases during the late 20\textsuperscript{th} century, leading to marked reductions in sensitivity to summer temperature. Lloyd and Fastie (2002) found growth generally increased in tree-
ring chronologies in Alaska during the early part of the 20th century, then declined after about 1950. Summer temperatures in their study region increased throughout the 20th century. Lloyd and Fastie (2002) suggest that summer temperatures have exceeded a threshold over the most recent 50 years and observed increases in temperature were no longer beneficial to tree growth, but instead have likely led to increased rates of water loss and reduced growth. McKenzie et al. (2001) found significant increases in growth of high elevation conifers in maritime environments of the western United States between 1850-1980, but do not interpret a specific cause due to the limited number of high elevation chronologies used in their analysis.

Growth rates of whitebark pine sampled in the SBW region also show a significant decline since 1950 (Figure 8, linear correlation with time = -0.56, p< 0.001), but no apparent trend in subalpine larch. The mechanisms responsible for the decline in whitebark pine growth in the SBW appear to be different than those reported by Lloyd and Fastie (2002). In the SBW region, there is no evidence that summer temperatures have been increasing. Individual climate stations in the region do not show significant trends and other investigators have found summer temperatures have either declined somewhat (Ferguson 1999) or have experienced decadal scale fluctuations (Diaz and Bradley 1997). Further, July precipitation at 8/12 stations with sufficient precipitation records contained a significant positive trend since about 1950 (p< 0.05). This suggests that increased water stress is probably not an important factor contributing to the decline in whitebark pine growth. However, this does not rule out the possibility that summer cooling, even if slight, has played some role in reducing growth rates. The difference in
Figure 8. Principal component chronologies 1850-1998. PCS chronologies constructed from standard tree-ring chronologies, PCR chronologies constructed from residual tree-ring chronologies. Heavy black line is a 10 year cubic smoothing spline fit to the data.
average summer temperatures between the early part of the record and the later half is small, with an average difference of only 1.5° C between the two periods. If cooling temperatures were indeed responsible for the decline in tree growth it would be expected that subalpine larch would show a similar reduction in growth, but there is no significant trend in subalpine larch growth since 1950 ($p > 0.05$).

A number of authors have pointed out the complex response of tree growth to climate variability through the interactions of various factors (Fritts 1976; Graumlich and Brubaker 1986; Graumlich 1993; Vaganov et al. 1999). A complex pattern of tree growth-climate response may also be the cause of reduced sensitivity in the chronologies presented here. Vaganov et al. (1999) observed a shift in the seasonality of tree growth near latitudinal treeline in Russia. They observed a reduction in early summer temperature and a trend toward later snowmelt in their study regions. They conclude that the shift in seasonality observed in that region was due to a shortening of the growing season due to increasing winter precipitation and delayed snowmelt. They suggest that later snowmelt coupled with decreasing early summer temperatures leads to delayed cambial activity and a shorter effective window for growth processes.

Changes in the growing season may at least partially explain the changes in climate-growth response of these conifers. Warm spring temperatures coupled with increasing spring precipitation that arrives as snow may result in a water stress effect. Short periods of warm weather when a tree is effectively dormant, coupled with frozen soil water, can lead to a net loss of photosynthates from respiration and transpiration when photosynthesis is not occurring (Fritts 1976; Graumlich and Brubaker 1986).
Examination of the individual climate stations within about 125 km of the SBW indicated that temperatures have risen in the spring (March-May). Individual stations with more than 30 years of data between 1958-1998, the period when most climate stations were in operation, indicated average spring temperatures increased at 10/11 stations \( p < 0.05 \). This warming appears most prevalent in minimum spring temperatures (10/11 stations) than changes in maximum spring temperature (6/11 stations), leading to a smaller diurnal temperature range. Short periods of warm weather may allow trees to respire without the benefit of the replacement of reserves because photosynthesis processes are suppressed (Fritts 1976). There is only limited evidence of increasing spring time precipitation with only April indicating a significant increase between 1901-1998 \( r = 0.22, p < 0.05 \). However, much of this precipitation may be arriving as snow at the treeline sites.

A potential climatic mechanism that could be partially responsible for a shift or weakening of tree-growth response is the Pacific decadal oscillation (PDO, Mantua et al. 1997). The PDO is like the better known El Niño-Southern Oscillation (ENSO) in that it is manifested by changes in Pacific Ocean sea surface temperature (SST), however, it is much more persistent than ENSO and involves the north Pacific Ocean (north of 20° latitude) rather than the more tropical regions. The warm (positive) phase of the PDO is characterized by a warm pool of water near the Gulf of Alaska with relatively cooler waters in the central and western north Pacific (Mantua et al. 1997; Zhang et al. 1997). During the PDO cool (or negative) phase, SSTs in the north Pacific are reversed. Mantua et al. (1997) suggests that when the PDO is in its positive (negative) phase the Pacific Northwest is subject to relatively warm (cool) winter temperatures and reduced
(enhanced) winter precipitation. The PDO index was generally positive between 1925-1947, negative from 1948-1976, and positive again through the middle 1990s (Mantua et al. 1997).

Comparisons with the PDO index (November-March) developed by Mantua et al. (1997) and the climatic data used here (average temperature and total precipitation) support this relationship. November to March PDO is positively related to March-May temperatures ($r=0.40$, $p<0.0001$) and negatively correlated with December-February precipitation ($r=-0.41$, $p<0.0001$) over the period 1902-1998. However, the correlation between PDO and the PC chronologies of tree growth is weaker (Table 6). The relationship between PDO and tree growth suggests that when PDO is in its positive phase, tree growth is slightly enhanced. It may be that the phase of the PDO index modulates to some degree the tree-growth response to climate. With respect to both whitebark pine and subalpine larch (Figures 4 & 7) it appears that during the PDO phase switches, the response of growth to both temperature and precipitation are different.

<table>
<thead>
<tr>
<th>PC Chronology</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>PIALPCS</td>
<td>0.09&lt;sup&gt;ns&lt;/sup&gt;</td>
</tr>
<tr>
<td>PIALPCR</td>
<td>0.24</td>
</tr>
<tr>
<td>LALYPCS</td>
<td>0.30</td>
</tr>
<tr>
<td>LALYPCR</td>
<td>0.28</td>
</tr>
</tbody>
</table>
During the early period when PDO is generally positive, relationships between summer temperatures appear stronger (Table 7). However, after the late 1940s when PDO is largely in its negative phase there is a reduction in sensitivity to summer temperatures. With respect to larch, spring (March-May) temperatures appear to become more important as does precipitation. It is important to note however, that temperatures have apparently been increasing during the spring in this region (see above) since about 1958. This contradicts the relationship between temperature and PDO to some degree because PDO is negative for much of the same time period which should result in lower spring temperatures. It may be that the increased snow pack associated with the negative phase of the PDO has an overriding influence on tree growth patterns than spring time temperatures. The relationships between PDO and tree-growth require a more detailed examination to better understand the physiological impacts on these species.

<table>
<thead>
<tr>
<th>PDO Phase (Warm):</th>
<th>Years</th>
<th>Climate Relationship</th>
<th>PIAL Growth Response</th>
<th>LALY Growth Response</th>
</tr>
</thead>
<tbody>
<tr>
<td>Positive</td>
<td>1925-1946</td>
<td>Warmer springs, Reduced DJF precipitation</td>
<td>Stronger response to summer temperature</td>
<td>Stronger response to summer temperature</td>
</tr>
<tr>
<td></td>
<td>1977-mid 1990s</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>1998(?)-?</td>
<td>Cooler springs, Increased DJF precipitation, Increased snowpack</td>
<td>Weakened response overall to climate</td>
<td>Weakened response overall to climate</td>
</tr>
</tbody>
</table>

Table 7. Summary of PDO-climate relationships and the general response of tree growth during different phases.
Examination of the correlation between the standard PC chronologies and average
monthly temperature and total precipitation during different phases of the PDO suggest
climate responses could moderated by the phase of the PDO (Figure 9). The relationship
between whitebark pine growth and temperature is greatly reduced when the PDO is in its
cool phase, and the relationships switches sign from positive to negative for the month of
August. The reduction in sensitivity from the warm phase of the PDO to the cool phase
is not as pronounced in subalpine larch (Figure 9). While PDO is not strongly related to
tree growth, it appears that it may alter the climate system thereby mediating the strength
of the tree-growth response. PDO-climate interactions may indeed be responsible for
reductions or shifts in sensitivity to climate but further analysis is needed to fully
understand the potential impacts.

The observed differences between the reductions in sensitivity between subalpine
larch and whitebark pine was more difficult to explain. Reduced sensitivity to summer
temperatures in whitebark pine were more pronounced than in subalpine larch (Figures 3
and 6). Only limited phenology data exists for either species, and this is mostly restricted
to winter relationships in subalpine larch or related *Larix* species (Richards 1984;
note that subalpine larch typically experiences bud burst in early May and completes the
growth of its needles by early June while snowpack is still in place. Other evergreen
associates apparently do not produce new needles until after the snow pack has melted
(Arno and Habeck 1972). In this respect, the subalpine larch may experience a longer
effective growing season than whitebark pine and therefore the relationship between
Figure 9. Climate-tree growth relationships during different phases of the Pacific Decadal Oscillation.
summer temperature remains relatively stable, though weakened. The shorter effective growing season due to snow depth, coupled with warm spring temperatures may have an overriding importance for whitebark pine as indicated by the more pronounced reduction in sensitivity. Because needles are present throughout the year on whitebark pine, it may experience a more severe reduction in photosynthates when soil water is frozen but temperatures are warm enough for respiration to take place for short periods of time.

Another plausible explanation for the more dramatic reduction in sensitivity observed in whitebark pine could be related to the mortality whitebark pine has experienced from mountain pine beetle epidemics and white pine blister rust (Cronartium ribicola). Mountain pine beetle and white pine blister rust have had a substantial effect on whitebark pine survival and have presented important concerns over the long term existence of whitebark pine at upper elevations (Keane and Arno 1993). While the chronologies were adjusted to remove potential impacts of mountain pine beetles by omitting rings near their end, it is possible that the trees killed by mountain pine beetles or blister rust were more sensitive to their surrounding environment. Environmental stress is widely regarded as an important factor that predisposes conifers to mortality by colonizing bark beetles (Craighead 1925; Thomson and Shrimpton 1984; Safranyik 1989). Extreme stress can result in reduced resin flow, limiting the host’s capacity to resist attack. Trees that are growing vigorously, while representing a higher quality food source, are apparently able to repel most attacks so long as beetle populations are low and tree-growing conditions are favorable.
It is possible that the whitebark pines killed by mountain pine beetles or blister rust during the 1920s-1930s were more sensitive to their surrounding environment and experienced a higher degree of stress during this outbreak due to moisture deficits during the outbreak period. Examination of the mean sensitivity (a proxy for climate sensitivity) of whitebark pines thought to have been killed by mountain pine beetle during the 1920s-1930s outbreak (based on field observations) and those that survived revealed that mean sensitivity is significantly lower in the surviving trees (t-test, \( p < 0.05 \)).

If the reduced sensitivity to climate evident in the whitebark pine is related to differential mortality due to mountain pine beetles killing climatically sensitive trees it would follow that a chronology constructed using only living trees would have a different tree growth response to climate during the early period than would the full chronology. Surviving whitebark pines would theoretically be less susceptible to mortality and less sensitive to climate. To test this I constructed a chronology using only living whitebark pine following the same procedures used to construct the PIALPC. The chronologies were constructed using between 13 trees (23 series) and 18 trees (28 series). Inter-series correlations for the living tree chronologies were somewhat lower than that of the full chronology (0.518-0.532). However, the climatic response of the whitebark pine chronology with only living trees was nearly identical to the full chronology for the early period (results not shown).
Local influences on growth response

It is important to note that while the species scale-response to climate identified by extracting the common variance using PCA identifies summer temperature as being one of the strongest variables related to growth processes, site specific relationships may be somewhat different. For the purposes of this paper PCA was used to uncover the general patterns of subalpine tree growth-climate relationships at the expense of presenting the individual response of each site. There do, however, appear to be important differences in the response of individual sites to climate interactions.

Simplified response surfaces (Graumlich 1991; Graumlich and Brubaker 1986) generated by using a simple running average to extrapolate between the observed standard growth indices, average summer temperature (June-August), and total spring precipitation (March-May) clearly illustrate this point. Over the 1901-1998 period the general response to climate in these subalpine conifers was with summer temperature with growth generally better when temperatures were warm (Figure 10). However, Baker Lake whitebark pine had a response to climate with a structure different from that of the other treeline chronologies (Figure 10). For the most part, the whitebark pine and subalpine larch chronologies, including the PC chronologies experienced increased growth with warmer summer temperatures (Figure 10). There is a slight decrease in growth of whitebark pine as spring precipitation increased even when temperatures are warm, supporting the drought induced effect of potentially longer residence time of snowpack. The Baker Lake whitebark pine chronology, however, appeared to reach a threshold at which temperatures above about 18°C resulted in reduced growth,
particularly when spring precipitation was also relatively high (Figure 10). Moreover, growth also apparently decreases when spring precipitation reaches about 14 cm.

The use of a reduced set of tree-growth indices, while a useful method for understanding the general patterns of growth-climate relationships, overlooks important fine scale patterns that may help explain various aspects of climate variability. The Baker Lake site may moderate growth-climate relationships due to its unusual substrate characteristics. While soil development is minimal at all sites, trees growing at the Baker Lake site were often found growing on stabilized talus slopes where water storage capacity is likely very minimal. During dry years above normal temperatures may no longer be beneficial to growth and water availability is instead more limiting. Site selection has long been a crucial aspect of dendroclimatic studies (Fritts 1976). Exploitation of similar species growing at different locations may provide more robust reconstructions, so long as the site effects are recognized.

**Implications of reduced sensitivity**

The development of tree growth-climate models for reconstructing climate has been an important part of understanding high-frequency climate variability at long time scales. Moreover, the low frequency features of tree growth are now being exploited to enhance the understanding of climate variations occurring over longer periods of time (Esper et al. 2002).

The reduction of sensitivity to climate, or switches from one limiting growth factor during one time period to a different limiting factor during another presents
Figure 10. Simplified response surfaces illustrating the interactive effects of summer temperature and spring precipitation on growth processes. The standard tree-ring chronology was used to generate the response surfaces. The surfaces were smoothed using a simple running average to extrapolate between observed values.
challenges to developing robust estimates of past climate variability using tree-rings. This is particularly the case when climate data is split in half, calibrated and verified using these partial periods, then reconstructed using a model developed from the full climate dataset. If a switch or reduction is evident in the data then the validity and strength of these reconstructions could be seriously reduced. For example, in the case of whitebark pine, the mechanism leading to reduced ring width during the early part of the century appears to be summer temperature. During the late period however summer temperature may still play a role, but spring temperature conditions might modify these relationships to some extent.

High quality reconstructions using whitebark pine may be difficult to produce from the chronologies presented here. It is possible that aggregating months together into a season that encompasses the growth window could result in a climate-growth relationship that is stable with time. An additional possibility would be selecting a variable for reconstruction that represents an integrated measure of temperature and moisture such as the Palmer Drought Severity Index.

With an ever increasing number of researchers identifying reductions in tree growth-climate responses, particularly in high latitude or high elevation temperature sensitive trees, experimental approaches to better understand the physiological relationships between tree growth and climate are imperative. This would require the development and testing of specific hypothesis rather than post hoc explanations of observed relationships. In addition, examinations of species phenology is imperative to link variations in climate with tree growth. Moreover, identifying the specific climatic
cause of reduced sensitivity (if it is indeed related to climate at all) requires climate data that is representative of the tree-ring sites, which is often difficult to if not impossible to acquire. Climate modeling with the aim of extrapolating climate data over a gridded area to account for elevation and topographic differences (e.g., Daly et al. 1994) is a logical first step but this process requires further work to improve estimates of climate parameters and account for local variations that are likely important for tree-growth climate relationships.

**Conclusions**

Whitebark pine and subalpine larch show good potential for the development of high resolution, millennial length climate reconstructions based on their strong response to summer temperature. There are strong relationships between the sites in both whitebark pine and subalpine larch, although relationships between species are somewhat weaker. July temperatures appear to be most important to whitebark pine growth while June temperatures appear most strongly related to growth of subalpine larch. A strong July precipitation relationship is also evident but is most likely the result of reduced effective solar heating due to cloud cover. The strong relationships between July temperature and precipitation with tree growth leads to the possibility of reconstructing an index of drought such as PDSI or some other proxy of moisture availability.

Split sample, running correlation, and multiple regression analyses suggest some of the relationships observed over the full time period are not temporally stable. Reduced sensitivity appears to be most problematic in the PIALPC chronology. Further research
is necessary to determine the true nature of reduced sensitivity to climate in these chronologies, but several potential mechanisms have been identified. The most likely cause of reduced sensitivity seems to be interactions due to increases in spring temperature and precipitation which may have led to alterations to the growing season window. Further, changes in climate sensitivity may be mediated by the PDO.

While there are important reductions in sensitivity over the modern period, valid reconstructions can still be developed using a split sample approach using a careful selection of variables and screening of tree-ring series to remove any possible non-climatic effects. Seasonalized or integrated temperature-moisture variables may prove to have stronger, temporally stable relationships than single-month variables. The inclusion of lagged predictors, particularly in the case of whitebark pine, may provide improved estimates of past climate variability in this region.

Acknowledgements

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Literature Cited


APPENDIX B

A 748-YEAR RECONSTRUCTION OF SUMMER TEMPERATURE IN THE
NORTHERN ROCKY MOUNTAINS, USA

This paper will be submitted to the journal *Holocene*

Abstract

Average summer (June-August) temperatures were reconstructed in the vicinity of the Selway-Bitterroot Wilderness Area, on the border of Idaho and Montana. Ring-width series from whitebark pine (Pinus albicaulis Englem.) and subalpine larch (Larix lyallii Parl.) collected at three high elevation sites, comprising six-tree growth chronologies, were used. A multiple regression procedure was used to screen predictor variables for the development of a transfer function to reconstruct temperature. The resulting regression equation explained approximately 36% of adjusted variance in the instrumental temperature record. Product mean and sign tests used to verify fidelity of the reconstruction were significant ($p<0.05$). Reduction of error (RE) and coefficient of efficiency (CE) values were both positive. Notable cool periods occurred during the mid to late 1300s, the early 1700s, and the mid 1900s. Warmer periods were present in the mid 1600s, the early 1300s, and early 1400s. The early 1700s cool period was more pronounced and persistent than any other period of the reconstruction. The reconstruction is similar in many respects to others from the general area but also
contains some important differences. Spectral analysis indicated significant ($p<0.05$) peaks in variance around 2, 15, and 90 years.

**Introduction**

Recent observations by the Intergovernmental Panel on Climate Change (IPCC) indicate that 20\textsuperscript{th} century temperatures increased more rapidly and have remained above the 1961-1990 average for longer than at any other time in the last millennium (IPCC 2001). Further, the IPCC assessment suggests extreme precipitation events have occurred with greater frequency than in the past. That 20\textsuperscript{th} century climate is different now than in the last 1000 years is the subject of little debate (but see Esper et al. 2002), however, the timing and magnitude of these changes is typified by important spatial variability. While the general characteristics of temperature increases have a near global scale pattern, there are important differences in the magnitude of these changes at finer spatial scales (IPCC 2001). In general high latitudes have experienced more dramatic warming than lower latitudes.

Assessing the range of variation in climate is an important part of understanding the scale and magnitude of 20\textsuperscript{th} century climate shifts and the behavior of the climate system in the future. Tree rings are a unique natural archive that can aid in assessing climate variation at long time scales. The annual nature of tree growth coupled with precise dating control makes them valuable recorders of climate variability at interannual to centennial time scales (Douglass 1920; Fritts 1976).
Whitebark pine (*Pinus albicaulis* Engelm.) and subalpine larch (*Larix lyallii* Parl.) are two subalpine conifers found in the northern Rocky Mountains that seem to have, at first sight, tremendous potential for dendroclimatic reconstruction. These treeline species attain great age (Arno & Habeck 1972; Luckman et al. 1984; Colenutt & Luckman 1991; Colenutt & Luckman 1995; Perkins & Swetnam 1996), and slow decomposition rates permit the inclusion of remnant materials that can dramatically lengthen tree-growth chronologies. The oldest living whitebark pine sampled in this study (and perhaps the oldest yet collected) spans 1,278 years with an inner date of AD 721. The oldest living subalpine larch sampled in this study has an inner ring date of AD 987 and spans 1,011 years, also among the oldest known trees of this species.

While these conifers have tremendous potential for developing millennial length chronologies that may be suitable for climate reconstruction, they have been little used. Several studies have investigated the relationship between whitebark pine (Perkins & Swetnam 1996; Biondi et al. 1999), subalpine larch (Peterson & Peterson 1994; Colenutt & Luckman 1995) and climate variability, but few have produced reconstructions using transfer functions developed from this relationship. 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 whitebark pine (Luckman et al. 1997; Biondi et al. 1999). Positive relationships between summer temperatures and tree growth have also been noted in subalpine larch (Colenutt & Luckman 1995). In addition, Graumlich and Brubaker (1986) have produced reconstructions of average annual temperature using various
subalpine species including subalpine larch collected in the Cascades of northern Washington.

Here I present a ≈750-year reconstruction of average summer (June-August) temperature near the Selway-Bitterroot Wilderness Area (SBW) near the border of Idaho and Montana, USA (Figure 1). Chronologies of tree growth constructed from upper treeline whitebark pine and subalpine larch were used to develop a regression equation to backcast climate to AD 1250. This reconstruction was compared with other previously published reconstructions of growing season temperature to assess the coherence of summer temperature variability in this region and at a broader spatial scale. These chronologies were also examined for evidence of warm and cool periods, such as the Medieval Warm Period (≈9th-14th centuries) and Little Ice Age (≈15th-19th centuries). While at one time considered to be nearly global phenomena, these periods exhibit substantial spatial variability (Hughes & Diaz 1994; Bradley & Jones 1995; Briffa 2000). Potential forcing mechanisms of summer temperature were also examined based on the spectral characteristics of the reconstruction.

Study Area

Tree-ring collection sites were located in upper treeline environments within and near the Selway-Bitterroot Wilderness Area (Figure 1, Table 1). These sites had limited
Figure 1. The location of the Selway-Bitterroot Wilderness Area and the three tree-ring collection sites (gray diamonds). CTR=Carlton Ridge, SMM=Salmon Mountain, BKR=Baker Lake. Black dots show the location of the climate stations used to develop a regional temperature record. Numbers near climate stations refer to station numbers listed in Table 2.
Table 1. Chronology statistics for tree-ring chronologies developed in the Selway-Bitterroot Wilderness Area.

<table>
<thead>
<tr>
<th>Location (lat./lon.)</th>
<th>Salmon Mountain</th>
<th>Baker Lake</th>
<th>Carlton Ridge</th>
</tr>
</thead>
<tbody>
<tr>
<td>Elevation (m)</td>
<td>2,725</td>
<td>2,670</td>
<td>2,650</td>
</tr>
<tr>
<td>Species'</td>
<td>PIAL</td>
<td>LALY</td>
<td>PIAL</td>
</tr>
<tr>
<td>No. Trees (Series)</td>
<td>31 (50)</td>
<td>50 (81)</td>
<td>62 (117)</td>
</tr>
<tr>
<td>SSS (0.85)(^1)</td>
<td>1366</td>
<td>1223</td>
<td>1145</td>
</tr>
<tr>
<td>M.S.L.(^2)</td>
<td>0.562</td>
<td>0.518</td>
<td>0.653</td>
</tr>
<tr>
<td>M.S.(^3)</td>
<td>0.206</td>
<td>0.216</td>
<td>0.294</td>
</tr>
<tr>
<td>1(^{st}) A.C.F.(^6)</td>
<td>0.78</td>
<td>0.75</td>
<td>0.59</td>
</tr>
</tbody>
</table>

\(^1\)PIAL = Whitebark Pine, LALY = Subalpine Larch; \(^2\)SSS = Year in which subsample signal strength (Wigley et al. 1984) is attained at 0.85; \(^3\)Mean segment length (years); \(^4\)Mean inter-series correlation; \(^5\)Mean Sensitivity (Fritts 1976); \(^6\)Mean first order autocorrelation in the tree-ring series before detrending.
fuel continuity to support spreading fires and they were in remote locations with limited access, thus reducing anthropogenic activities and impacts. However, some evidence existed of past mortality in some whitebark pine stands due to mountain pine beetle (*Dendroctonus ponderosae*) epidemics or white pine blister rust (*Cronartium ribicola*), mostly occurring around the late 1920's (Evenden 1944; Ciesla & Furniss 1975; Kipfmueller et al. 2002). Whitebark pine killed by these agents were present as dead standing snags, some with external J-shaped galleries typical of mountain pine beetles (Wood 1982). The impact of mountain pine beetles on whitebark pine ring-width patterns was limited in decades and centuries prior to death because trees are typically killed outright rather than experiencing a reduction in ring-width, as is common with insect defoliators (e.g., Swetnam & Lynch 1993).

The climate of the SBW is transitional from a wetter Pacific Northwest type climate to a continental dry rainshadow climate (Finklin 1983; Ferguson 1999). Cold winters and mild summers typify the temperature regime of the region with winter temperatures averaging -2.7°C in winter and 17.4°C in summer recorded by nearby lower elevation stations (Table 2). Precipitation ranges from 625 mm at lower elevations (500 m) near the southern end of the wilderness to about 1000 mm in the western portion and at higher elevations (Finklin 1983). Most of the precipitation falls as snow during the winter months. There is a gradient in the seasonality of precipitation with proportionally more precipitation falling in winter and spring in the western portion of the study area (Finklin 1983; Ferguson 1999; Brunelle-Daines & Whitlock 2001; Brunelle-Daines 2002).
Table 2. Climate stations used for developing a regionally representative temperature dataset.

<table>
<thead>
<tr>
<th>Station Name</th>
<th>Data Source</th>
<th>Lat./Lon.</th>
<th>Elev. (m)</th>
<th>Time Span</th>
</tr>
</thead>
<tbody>
<tr>
<td>Missoula WNW, MT</td>
<td>NCDC</td>
<td>46.88N 114.03W</td>
<td>967</td>
<td>1893-1965</td>
</tr>
<tr>
<td>Stevensville, MT</td>
<td>NCDC</td>
<td>46.52N 114.10W</td>
<td>1,029</td>
<td>1911-1998</td>
</tr>
<tr>
<td>Hamilton, MT</td>
<td>HCN</td>
<td>46.23N 114.17W</td>
<td>1,076</td>
<td>1900-1998</td>
</tr>
<tr>
<td>Darby, MT</td>
<td>NCDC</td>
<td>46.02N 114.18W</td>
<td>1,183</td>
<td>1948-1998</td>
</tr>
</tbody>
</table>

1Station number corresponds to the location numbers in Figure 1.

Methods

Chronology Development

Tree-ring series were collected from whitebark pine (PIAL) and subalpine larch (LALY) growing near alpine treeline using an increment borer and standard dendrochronological techniques (Fritts 1976). All increment cores were crossdated to ensure the precise assignment of calendar dates and dating was checked using program COFECHA (Holmes 1983). Ring widths were measured to the nearest 0.01 mm on an incremental measuring bench. Increment cores with less than about 250 annual rings were omitted to preserve as much low frequency variability in the final tree-growth chronology as possible (Cook et al. 1995).

Dimensionless indices of tree-ring width were generated by detrending each series with a line fit to the measured values, then dividing the measurements by the values of the fitted curve using ARSTAN (Cook and Holmes 1986). The detrended series were combined using a biweight robust estimate of the mean to minimize the effect of outlier
values (Cook 1985). Measured ring widths were detrended using a relatively stiff 250 year cubic spline to remove trends unrelated to climate.

A stiff spline was used for detrending individual tree-ring series rather than negative exponential or linear detrending methods for two reasons: First, sampled trees were often extremely old and typically the innermost rings were not collected due to rotten centers. The negative exponential decline in ring widths commonly found with increasing tree age was often not present because increment cores often did not reach the center of the trees. Instead, plots of raw ring-width measurements suggested samples were typically already in the nearly horizontal “tail” of growth with little or no evidence of rapid juvenile growth near the inner rings of the samples.

Second, with respect to whitebark pine, it is possible that some of the sampled trees germinated from unused Clark’s nutcracker (*Nucifaga columbiana*) seed caches (the primary agent of seed dispersal for whitebark pine), even though they are single-stemmed today (Hutchins & Lanner 1982; Tombback et al. 1990). As individual stems die, the ring widths of those remaining may record these deaths as a release from suppression, a disturbance unrelated to climate. Detrending was necessary to remove these growth surges in individual trees. Smoothing splines have been used successfully to detrend interior forest species that may also have experienced some form of disturbance (Cook & Peters 1981) and are therefore a good alternative to detrending whitebark pine that may have germinated from seed caches.
Temperature data

Climate data in the region is sparse and National Climatic Data Center (NCDC) divisional climate data is biased by stations located great distances (>200 km) from the sample sites. As an alternative to NCDC divisional data I selected the four climate stations nearest the tree-ring collection sites with the longest records and created a regional averaged monthly temperature record (Figure 1, Table 2). These stations were located along the eastern front of the Bitterroot Mountains in the Bitterroot Valley (Figure 1) and are relatively close to the tree-ring collection sites (average distance between climate stations and tree-ring sites ≈ 68 km). They are, however, lower in elevation than the tree-ring collection sites, but no high elevation stations existed in the region with sufficiently long records for the reconstruction of climate.

Average summer temperature (June-August) was selected as the variable for reconstruction. Examination of the response to climate variability using correlation and response function analyses indicated these conifers were generally most limited by temperature during the growing season, particularly between June-August. Warmer temperatures during the summer months generally promoted positive growth in these species. However, the whitebark pine chronology for Baker Lake had a weakly negative response to summer temperatures during the growth year and a significant negative response to summer temperatures from the previous year (r = -0.29, p < 0.01).
Reconstruction methods

Stepwise linear regression was used to screen potential candidate predictors prior to their inclusion in the final model. The pool of potential predictors included 18 variables: the six standard tree-ring chronologies lagged 0, -1, and +1. Potential predictor variables were allowed to enter the regression model according to their linear relationship with the dependent variable until a maximum $R^2$ and a minimum root-mean-square-error was attained.

The procedures used for calibration and verification of the regression equation used to reconstruct climate followed two approaches. First, cross-validation of the reconstruction model was completed using the predicted residual sum of squares method (PRESS) that employed a leave-one-out cross-calibration procedure (Weisberg 1985; Michaelsen 1987). This method successively fits a regression model to the data with one observation removed. The resulting regression model was then used to predict the withheld value. This procedure is repeated for each year of data. Verification statistics were then generated from the residuals produced using this procedure. The PRESS method was used to identify problems such as overfitting of the regression model. The second method followed the more commonly used approach in dendroclimatology (e.g, Fritts 1976). The observational climate data were split into two equal halves, calibrating on one half and verifying the regression quality on the second half. This process is then reversed by switching the calibration and verification periods. The split sample approach was used as an additional test of the strength and validity of the climate reconstruction to ensure the climate-tree growth relationships were temporally stable. Past climate
variability was then estimated by applying the resulting regression equation from the full time period (1900-1997) to the tree-ring chronologies that entered the stepwise model.

Spectral analysis was applied to the resulting reconstruction of summer temperature to identify peaks in the variance structure possibly related to climate forcing at different timescales. The characteristics of the reconstructed summer temperature in the frequency domain were examined using a Fourier transform of the autocovariance of reconstructed temperature using the Blackman-Tukey lag window method. A 100 year lag window was selected to balance the resolvability of nearby peaks and minimize the variance of the estimated spectrum as much as possible (Chatfield 1996).

Results and Discussion

Reconstruction characteristics

Only three predictors entered the regression equation using the stepwise procedure (Table 3). Two subalpine larch chronologies and one whitebark pine chronology entered the equation, although the Baker Lake whitebark pine chronology included was lagged forward by one year (Table 3). The remaining chronologies added no additional prediction skill to the reconstruction so were not used in the reconstruction. Approximately 36% of the variance (adjusted for loss of degrees of freedom) of average June-August temperature was explained by the regression model. Residual plots (not shown) indicated no apparent violations of regression assumptions and the Durbin-Watson D = 1.64, indicating acceptable levels of first order autocorrelation in the residuals (first order autocorrelation of the residuals = 0.171).
Table 3. Regression model characteristics based on full time period calibration (1900-1997).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Estimate</th>
<th>Std Error</th>
<th>$R^2$</th>
<th>$R_{adj}^2$</th>
<th>F Ratio</th>
<th>Sig. F</th>
</tr>
</thead>
<tbody>
<tr>
<td>Intercept</td>
<td>16.53</td>
<td>0.59</td>
<td>0.38</td>
<td>0.36</td>
<td>18.79</td>
<td>0.000</td>
</tr>
<tr>
<td>CTRLY</td>
<td>1.40</td>
<td>0.36</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BKRWB_{t+1}</td>
<td>-1.18</td>
<td>0.47</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>BKRLY</td>
<td>0.68</td>
<td>0.51</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The model calibration/verification using the PRESS method resulted in a reduction of error (RE) similar to the calibration $R^2$ and root-mean-square error that was small relative to the variance of the observed temperature data (Table 4). Sign and product means tests were significant ($p<0.01$) and indicated the reconstruction was of good quality. Verification of the reconstruction using the split sample approach supported the PRESS verification results (Table 5). The coefficient of efficiency (CE), a more stringent test statistic than the RE, was positive for both verification time periods using the split sample approach indicating there is useful climate information in the reconstruction (Cook et al. 1994).

There was marked weakening of the climate signal between the early period and late period as evidenced by lower verification statistics using the split sample approach. The RE and CE statistics were only slightly positive when verifying the early period model on the second half of the data (Table 5). Fritts (1976) and Cook et al. (Cook et al. 1994) suggest only a few poor estimates can result in poor RE and CE values. That appears
Table 4. Verification statistics using the PRESS calibration procedure.

<table>
<thead>
<tr>
<th>Model</th>
<th>RE</th>
<th>RMS</th>
<th>Bias</th>
<th>APE</th>
<th>1st Diff. Sign Test (agree/disagree)</th>
<th>PMT</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.31</td>
<td>0.69</td>
<td>0.002</td>
<td>0.56</td>
<td>66/31</td>
<td>3.32</td>
</tr>
</tbody>
</table>

RE = Reduction of error; RMS = Root-mean square error; APE = Average prediction error (Weisberg 1985). PMT = Product Means Test (Fritts 1976, Cook et al. 1994). Sign test and PMT significant at p < 0.01.

Table 5. Summary of split sample calibration and verification of the regression model used to reconstruct summer temperature.

<table>
<thead>
<tr>
<th>Time Period</th>
<th>R²</th>
<th>r²</th>
<th>RE</th>
<th>CE</th>
<th>1st Diff. Sign Test (agree/disagree)</th>
<th>PMT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Calibration</td>
<td>1900-1948</td>
<td>0.51</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Verification</td>
<td>1949-1997</td>
<td>0.47</td>
<td>0.11</td>
<td>0.12</td>
<td>33/15</td>
<td>2.42</td>
</tr>
<tr>
<td>Calibration</td>
<td>1949-1997</td>
<td>0.29</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Verification</td>
<td>1900-1948</td>
<td>0.37</td>
<td>0.42</td>
<td>0.42</td>
<td>36/12</td>
<td>2.63</td>
</tr>
<tr>
<td>Full Model</td>
<td>1900-1997</td>
<td>0.37</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

RE = Reduction of error, CE = Coefficient of efficiency, PMT = Product Means Test (Fritts 1976, Cook et al. 1994). Sign and PM test significant at p < 0.01.

to have been the case here as the early period model predicts 1956 poorly. The summer climate conditions during 1956 were near normal but winter snowpack was unusually high during the winter (z-value = 2.0). April 1st snow water equivalent measured at Savage Pass (near the center of the wilderness) was the second highest value recorded between 1937-1998 at 103 cm. Increased snowpack may have kept soil frozen later in the growing season limiting water uptake and inducing a drought effect and narrow ring formation even when summer conditions were near normal later. Further, both the
Figure 2. Observed average summer (June-August) temperature for the regionally averaged climate stations (thin line) compared to estimated values (thick line).
highest and lowest values of summer temperature are found in the later half of the climate data (1961 and 1993 respectively, Figure 2).

June-August average temperature was reconstructed back to 1250 as sample depths before this time diminish considerably (Figure 3). Subsample signal strength (Wigley et al. 1984) indicates sample depth is large enough to reliably reconstruct temperature to 1250 for both chronologies collected in Baker Lake but is substantially weaker in the Carlton Ridge subalpine larch chronology (Table 1). This limits the reconstruction's reliability in earlier periods. The number of radii at 1300 is lowest for the Carlton Ridge subalpine larch chronology with only three series sampled from two trees. Five series from four sampled trees are present by 1350 at the Carlton Ridge site. The subsample signal strength for Carlton Ridge when four trees are present decreases to 0.75 and to 0.70 where only three trees are present. The decision to include data from before the 0.85 subsample signal strength cutoff was made to maximize the information available in the other chronologies, but periods before about 1350 should be interpreted with caution.

Comparison of observed versus reconstructed July-August temperature suggested the reconstruction did a better job at capturing low frequency variations in temperature than those of higher frequency (Figure 2). The reconstruction captured the increase in summer temperature during the 1930s-1940s as well as the cool period during the late 1940s and early 1950s. The reconstruction did a poor job capturing the most extreme warm or cool summers such as 1961 and 1993 (Figure 2).
Figure 3. Sample depth of tree-ring chronologies developed in the Selway-Bitterroot Wilderness. Black lines indicate the number of whitebark pine series present for a given year and dark gray represents subalpine larch. Note that the y-axis differs for Baker Lake. The black and gray vertical lines are at the 0.85 SSS cut-off for whitebark pine and subalpine larch respectively.
The characteristics of the regression model and calibration/verification tests were similar to other growing season temperature reconstructions produced elsewhere in the western United States. Briffa et al. (1992b) reconstructed climate using a gridded network of climate stations and maximum latewood density chronologies in the western United States. Variance explained in their reconstruction of temperature at grid point 13, the point nearest this study, was similar (early period verification = 0.39, late period verification = 0.22, Table 4 in Briffa et al., 1992b). Their reconstruction of summer temperature used a network of 23 chronologies that extended to 1600. However, variance explained in their reconstructions using a larger (albeit shorter temporally) set of maximum density chronologies was considerably higher at around 60%. Luckman et al. (1997) reconstructed summer climate using a combination of maximum density and ring-width chronologies for the Columbia Icefield in southwestern Alberta with similar calibration/verification results to those reported here. Calibration/verification statistics and model $R^2$ were somewhat better in this reconstruction than a reconstruction of July temperature from central Idaho that used chronologies from upper treeline whitebark pine and lower treeline Douglas-fir (Biondi et al. 1999). Biondi et al. (1999) report early period calibration $R^2 = 0.18$ and verification $R^2 = 0.35$. Variance explained by their calibration of the late period model was 43% with verification $R^2 = 0.13$.

Tree-growth chronologies from only two of three potential sampling sites were suitable for the reconstruction of summer temperature. Both the Baker Lake and Carlton Ridge sites were represented in the reconstruction model but no chronologies from Salmon Mountain. The instrumental climate data used to calibrate the reconstruction
model might not have been a good representative of climate at Salmon Mountain as it was for the other sites. The stations used here to generate the regional temperature were closer to the sites that entered the model using the stepwise screening procedure (Figure 1). The average distance between the climate stations and Salmon Mountain was around 105 km while the average distance between the other sites and the meteorological stations was only about 50 km. Salmon Mountain is also higher in elevation and further south and west within the region and might experience a different climatic regime given the strong precipitation gradients in the area (Finklin 1983).

The leading (BKLWBt+i) variable is somewhat unusual given the general positive relationship between summer temperatures and whitebark pine growth in the region. The most common physiological explanation for this relationship is the carryover effect of food storage from year-to-year (Fritts 1976). Leading relationships between temperature and growth have been noted by other authors, but the relationship observed is often direct (Salzer 2000). The correlation between BKRWBt+i and June-August temperature observed here, however, was inverse. This negative relationship may have been due in part to a difference in the site characteristics between Baker Lake and the other tree-ring sites. Whitebark pine sampled in Baker Lake were growing in areas where soil development is minimal and in many cases trees were growing directly on stabilized talus fields. Moreover, Baker Lake was located within the rainshadow of the Bitterroot Mountains while other sites are located either along the crest of the mountains or further west. In this respect water availability might be less in the Baker Lake site than that of other sites. Warm summer temperatures coincident with reduced moisture availability
could induce a drought effect and reduced growth with detrimental effects on growth processes spanning more than one year.

It was not surprising that the contribution of subalpine larch chronologies was greater than that of whitebark pine in the development of this reconstruction. Although the whitebark pine chronologies all possessed good crossdating characteristics and had in common most of the same marker years with the subalpine larch, their response to climate was weaker and less stable temporally (Appendix A). Further, the relatively high first order autocorrelation present in whitebark pine was not as prevalent in the larch chronologies. This high first order autocorrelation makes comparisons with annual climate data problematic by introducing lagging relationships between growth and climate variables. It is possible that an improved understanding of the phenology of whitebark pine could yield high quality reconstructions but further research into the physiology of this species is necessary.

Climate variability in the SBW, 1300-present

A fifty year cubic smoothing spline was fit to reconstructed summer temperature anomalies to highlight frequency characteristics at multidecadal time scales. Anomalies were based on the 1900-1997 long-term average summer temperature (17.4° C). The fifty-year smoothing spline preserves 50% of the variance in the reconstruction at 50 year periods, but almost none of the variance at the centennial scale (only about 5% at 104 years).
Figure 4. Reconstructed average summer temperature near the Selway-Bitterroot Wilderness Area, 1250-1997. Data are shown as anomalies from the 1900-1997 mean. The smooth black line is the reconstruction smoothed using a 50 year cubic smoothing spline. The solid vertical line indicates the SSS>0.85 cut-off of the most limiting chronology (CTRLY).
The nearly 750-year long reconstruction of summer temperature included several sustained cool periods as well as evidence of multidecadal warm events but was best characterized by its alternating pattern of relatively warm and cool conditions (Figure 4). In general, somewhat longer periods of warmth or near normal conditions were interrupted by relatively rapid shifts to shorter term cool conditions through most of the first \( \approx 500 \) years of the reconstruction. Since about 1800, however, both warm and cool periods appeared to persist for approximately equal lengths of time (Figure 4). For the most part, the magnitude of summertime warmth was similar throughout the \( \approx 750 \) year chronology but the magnitude of cooling appeared to vary.

The beginning of the 14th century was generally warmer than the calibration mean but experienced marked cooling during its later half. Several years during the 14th century were among the warmest and coolest years of the reconstruction including 1330, among the five coolest years, and both 1311 and 1314, both among the five warmest (Table 6). In addition, the 1310s and 1360s were among the warmest and coolest decades of the reconstruction respectively (Table 6). To some degree these large deviations could be due to limited sample depth during the early part of the time period, but for the most part sample depth during this period was good (Figure 3). The BKRLY chronology has 8 trees present at 1300 while BKRWB has 15. Sample depth is weakest in CTRLY with only two trees present at 1300. Conditions returned to near the long-term average from about the end of the 14th century through the middle 15th century followed by cooling in the mid to late 1400s. This cooling, which began around 1458 and persisted until about
1492, included the year 1458, the second coolest year in the reconstruction and 1451-1460, among the coolest ten year periods (Table 6).

<table>
<thead>
<tr>
<th>Year</th>
<th>Reconstructed JJA T</th>
<th>SDU</th>
<th>Year</th>
<th>Reconstructed JJA T</th>
<th>SDU</th>
</tr>
</thead>
<tbody>
<tr>
<td>1417</td>
<td>15.7</td>
<td>-2.13</td>
<td>1314</td>
<td>19.2</td>
<td>1.84</td>
</tr>
<tr>
<td>1458</td>
<td>15.7</td>
<td>-2.10</td>
<td>1790</td>
<td>18.9</td>
<td>1.53</td>
</tr>
<tr>
<td>1723</td>
<td>15.8</td>
<td>-1.98</td>
<td>1662</td>
<td>18.7</td>
<td>1.27</td>
</tr>
<tr>
<td>1330</td>
<td>15.9</td>
<td>-1.86</td>
<td>1831</td>
<td>18.6</td>
<td>1.23</td>
</tr>
<tr>
<td>1832</td>
<td>16.0</td>
<td>-1.74</td>
<td>1311</td>
<td>18.6</td>
<td>1.22</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Decade</th>
<th>Avg. JJA T</th>
<th>Average Anom. (°C)</th>
<th>Decade</th>
<th>Avg. JJA T</th>
<th>Average Anom. (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1367-1376</td>
<td>16.7</td>
<td>-0.66</td>
<td>1657-1666</td>
<td>18.2</td>
<td>0.77</td>
</tr>
<tr>
<td>1712-1721</td>
<td>16.8</td>
<td>-0.65</td>
<td>1311-1320</td>
<td>18.0</td>
<td>0.61</td>
</tr>
<tr>
<td>1947-1956</td>
<td>16.8</td>
<td>-0.55</td>
<td>1428-1437</td>
<td>18.0</td>
<td>0.59</td>
</tr>
<tr>
<td>1451-1460</td>
<td>16.8</td>
<td>-0.55</td>
<td>1931-1940</td>
<td>17.9</td>
<td>0.52</td>
</tr>
<tr>
<td>1702-1711</td>
<td>17.0</td>
<td>-0.46</td>
<td>1865-1874</td>
<td>17.9</td>
<td>0.50</td>
</tr>
</tbody>
</table>

Summer temperatures near the long-term average persisted throughout most of the 16th century until around 1581 when a switch to cool conditions occurred (Figure 4). This cool period lasted until about 1626. It is also interesting to note that sample depths for whitebark pine at both Salmon Mountain and Carlton Ridge decreased somewhat around this cool period (Figure 3). While many trees spanned this cool period completely, a large number at both sites had either their beginning or ending date within
or near this cool period. Warmer than normal conditions persisted for most of the remainder of the 1600s. The warmest decade identified in the reconstruction was 1657-1666, which included 1662, one of the five warmest years (Table 6).

One of the coolest and most protracted cool departures occurred near the beginning of the 18th century. Summer temperatures averaged 16.9°C between 1702-1738 with only a few years reaching levels that were near or above the long-term mean (Figure 4). Two of the coolest decades occurred during this extended cool period (Table 6). This cool period appeared to both begin and end rapidly and is followed by generally sustained warmth into the early 19th century.

The early 19th century experienced a rapid shift from relatively warm summer conditions to a short period of cool temperatures (Figure 4). This shift was similar in structure to that which occurred around 1700, but was shorter in duration, lasting only from about 1809-1826, and not as cold. Average summer temperatures during this cold snap were about 17.1°C at elevations around 1000 m. Shortly after this cool period a dramatic interannual switch between warm and cool conditions occurred. The reconstructed summer temperature of 1831 was among the five warmest in the reconstruction but 1832 was among the five coolest. Reconstructed summer temperature in 1831 was estimated at 18.6°C and 1832 at 15.9°C, a reduction of about 2.7°C. This represents the largest interannual shift in temperature since 1300 in the reconstruction. Volcanic eruptions have been implicated as important agents that lead to large shifts from warm to cool conditions over short periods of time (e.g., Lough & Fritts 1987). Volcanic activity did not at present appear to have been the cause of this shift as no volcanic
eruptions of sufficient size were evident in the eruption record. Mount St. Helens, located about 700 km to the west of the study site, was active in 1831 but this eruption was small and apparently had only localized effects (Crandell et al. 1975; Foxworthy and Hill 1982). Temperature shifts of such as this, though rare, were not unprecedented in the reconstruction as shifts of nearly the same magnitude occurred between 1416 and 1417 (-2.5°C) and 1329-1330 (-2.4°C). High quality (i.e., well dated) records of climatically effective volcanic eruptions before 1500, however, were limited so it is possible these earlier shifts could be related to past, undocumented volcanic activity.

Summer conditions were near the long term average through most of the remaining 19th century with the exception of a cool period near the 19th-20th century boundary. This departure was relatively short, lasting only from about 1890-1903.

Warm temperatures associated with the drought of the 1930s that affected a broad area of the western United States (e.g., Woodhouse and Overpeck 1998) was also well represented in this reconstruction. The ten years between 1931 and 1940 were among the warmest ten year periods in the reconstruction (Table 6). Cooler summers during the middle 1900s followed by increased temperatures were also evident, consistent with past reconstructions in the region (e.g., Biondi et al. 1999). This latter increase toward the end of the 20th century suggests summer temperatures have been increasing since the most recent cool period in the 1950s and 1960s. However, the end of the 20th century did not appear warmer than earlier portions of the reconstruction (Figure 4).
Comparisons with other reconstructions

Comparisons between this smoothed reconstruction and others at varying spatial scales smoothed using a 20-year cubic spline illustrated many common features but also important differences (Figure 5). These reconstructions of temperature anomalies (based on the calibration period means of the individual reconstructions) represent various portions of the growing season. Mostly these examine a longer seasonal temperature window such as the maximum latewood density reconstructions of Briffa et al. (1992b) that reconstruct April-September temperatures. Biondi et al. (1999) reconstructed average July temperature using tree-ring series collected from upper treeline whitebark pine and lower treeline Douglas fir which have opposite responses to summer temperature (displayed as standard deviation units rather than temperature anomalies). In addition, the Briffa et al. (1992b) western United States and British Columbia/Pacific Northwest datasets were pieced together from two separate reconstructions based on their network. The early portion of these reconstructions (1600-1750) includes fewer chronologies and had lower explained variance but longer temporal depth. The 1750-1982 period used a larger number of chronologies as predictors resulting in better variance explained but was shorter. The regional chronologies displayed here were pieced together to maximize temporal length using the weaker reconstruction to represent 1600-1750 with the stronger reconstruction used for the period 1750-1982.
Figure 5. Low-frequency patterns of reconstructed growing season temperature anomalies from other reconstructions. Values are smoothed using a 20-year cubic smoothing spline fit to anomalies based on each reconstructions calibration period of the calibration periods (see text for details). Dashed lines near zero are used to highlight deviations from the calibration period mean for each chronology. Note that the central Idaho reconstruction is displayed as standard deviation units rather than anomalies.
Reconstructed summer temperature in the SBW contained several prominent cool periods common to other reconstructions (Figure 5). Generally cool conditions in the mid 1400s, around 1600, 1700, and near the beginning and end of the 1800s were present in other reconstructions at a variety of spatial scales. Warm or near long term average conditions common to the reconstructions are somewhat more difficult to identify perhaps due to the different calibration means used in developing chronologies as well as the different impacts of 20th century warming between sites. In general, warm or near normal conditions common between sites occurred in the early 1400s, middle 1600s, late 1700s and early 1900s also appeared to be widespread events. The timing of these warm and cool conditions, however, varied somewhat among the reconstructions. Moreover, some prominent cool periods in other reconstructions were not present in the SBW reconstruction.

There is a striking difference around 1300 between the Central Idaho temperature reconstruction and the SBW reconstruction developed here. The tree-ring sites used in the central Idaho reconstruction are only about 350 km apart. The Biondi et al. (1999) reconstruction suggested prolonged cooling around this period while temperatures in the SBW appeared to be somewhat warmer (Figure 5). Examination of the ring-widths of the individual chronologies sampled in the SBW are somewhat ambiguous around this period (Figure 6). Tree-growth chronologies at Salmon Mountain from both whitebark pine and subalpine larch record lower growth for a prolonged period around 1300, consistent with the Biondi et al. (1999) reconstruction (Figure 6). However, whitebark pine and subalpine larch at Carlton Ridge suggest warmer conditions because growth appeared to
Figure 6. Six tree-growth indices from the sites sampled in the SBW smoothed with a 50 year cubic spline. Note the y-axis for Baker Lake whitebark pine has a different scale than the other chronologies.
be above average. Whitebark pine sampled at Baker Lake had a reduction in growth that suggested warmer conditions given its inverse relationship with summer temperatures (Figure 6). The Baker Lake subalpine larch chronology had only a slight reduction in growth around 1300 but a more pronounced reduction later in the century.

The disparity between the Central Idaho reconstruction and the SBW may be the result of several factors. First, sample depth was poor during this period in the Carlton Ridge subalpine larch chronology used in the reconstruction. This chronology had high leverage in the regression equation that likely biased the reconstruction in the early periods due to poor sample depth. The sample depth of the central Idaho reconstruction was not reported so it is difficult to ascertain the overall effects of sample depth reduction in that reconstruction.

Second, the observed reduction in temperature around 1300 in central Idaho might be an effect of reduced precipitation rather than a reduction in summer temperature. Biondi et al. (1999) suggested that Douglas-fir growth is enhanced by cool summer temperatures because growth was more strongly related to moisture stress (Biondi et al. 1999, p. 1446). Further, the contribution of Douglas-fir to the reconstruction was reported to be about 60% greater than that of whitebark pine in the central Idaho temperature reconstruction (Biondi et al. 1999, p. 1446). Hughes and Graumlich (1996) reported a drought centered about 1299 occurred in the White Mountains of California extending into the Great Basin. It is possible this drought impacted tree-growth of lower elevation Douglas-fir in central Idaho but had limited impacts in the SBW because snowpack at upper elevations generally provides a reliable water source through most of
the summer months at the sites I examined. Snowpack was usually still present into early July at these sites. It seems plausible that the observed reduction in temperatures around 1300 in central Idaho is due to reduced precipitation rather than temperature limitations. However, an equally plausible explanation is that it was indeed cooler, but this cooling did not extend to the SBW. This might also explain why whitebark pine growth at Salmon Mountain, the most southerly site sampled for this study, also showed reduced ring-widths around the 1300s.

Several authors have documented widespread cooling around the 1600s possibly related to the eruption of Huaynaputina, Peru in 1600. Cooling during this period was well represented by the SBW reconstruction but appeared to occur somewhat earlier than in central Idaho (Figure 5). The greatest reduction in temperature in other reconstructions of growing season temperature often occurred in 1601 (Biondi et al. 1999; D’Arrigo & Jacoby 1999). The year 1601 was the coolest year of the 1585-1626 event but is not among the coolest five years, or even the coolest ten when the entire ≈750 year reconstruction was considered (Table 6). This cool event in the SBW apparently began even before a general period of high volcanic activity that began with the eruption of Ruiz, Colombia in 1595 and continued with the eruption of at least four additional volcanoes through about 1606, including the 1601 eruption of Huaynaputina, Peru (Simpkin & Siebert 1994). Volcanic eruptions around 1600 may have simply prolonged an already established cooling trend in the SBW.

The early 1700s in the SBW reconstruction was one of the most pronounced cool periods of the ≈750-year record. This departure in summer temperature was also evident
in the regional reconstructions of Briffa et al. (1992b) in both the western United States as a whole as well as in the British Columbia/Pacific Northwest reconstruction. A slight reduction in temperatures around 1700 was also present in Luckman et al.'s (1997) Columbia Icefield reconstruction. Reduced temperatures in the SBW chronology, however, appeared to begin a bit later than in other regions, and persist longer (Figure 5). Further, this departure was absent in central Idaho although there is a slight downward trend (though still warmer than average) in temperature around the same time. Generally warm conditions appeared to occur over a broad region near the end of the 1700s (Figure 5).

Two prominent cool periods in the 1800s, one near the beginning and one close to the turn of the century were evident in other reconstructions as well as in the SBW. Luckman et al. (1997) identified a nearly century long cool period between about 1780-1900, coincident with the most recent maximal extent of the Athabasca Glacier in 1840. Four of the coolest decades in their reconstruction occurred during this time period. Cool conditions were also found in the SBW chronology in the early and late 19th century, but these cool periods were interrupted by relatively warm conditions (Figure 5). Cool conditions around 1900 in the SBW appeared to be somewhat shorter than in other regions lasting only between about 1890-1903. Similar relationships of cool periods interrupted by relative warmth were present in the summer reconstruction in central Idaho (Biondi et al. 1999), and in the Briffa et al. (1992b) regional reconstructions.
The Medieval Warm Period, Little Ice Age, and 20th century warming

There is some debate regarding the magnitude of large scale warm and cool events over the last millennium (Hughes & Diaz 1994; Mann et al. 1998; Mann et al. 1999; Esper et al. 2002). Recently published reconstructions of Northern Hemisphere temperatures suggest that conditions between the 9th and 14th centuries, also known as the Medieval Warm Period (MWP) were possibly as warm or warmer than the 20th century (Esper et al. 2002). Further, these more recent reconstructions of Northern Hemisphere annual (January-December) temperature suggest prolonged cooling following the MWP persisted between about and 1500-1850 (Esper et al. 2002). While large scale warm and cool events reconstructed from networks of proxy data are extremely valuable for illuminating past climate variability, the expression of warm or cool events can be moderated at local scales.

In addition, the most widely cited tree-ring evidence of pronounced warmth during the MWP and cold during the LIA is from Esper et al. (2002). This reconstruction uses a standardization technique designed to preserve multicentennial variability in tree growth that was impossible to use in this reconstruction. The technique used by Esper et al. (2002) and also used in previous reconstructions (Briffa et al. 1992a; Briffa et al. 1998) aligns tree-ring measurement series to a common tree age and detrends the series using a single average growth function. This technique does preserve more of the low frequency variability in tree-ring measurements but is constrained by the identification of the exact cambial age of an individual sample. Because most trees sampled here lacked the innermost rings, and in some cases were likely centuries away from the trees true age,
these samples could not be aligned using this standardization technique. This makes direct comparisons with the Esper et al. (2002) reconstruction difficult. However, the chronologies used in the SBW reconstruction were detrended using a stiff spline of 250 years and were composed of long series (Table 1). For example 37 of the 81 series used in the development of the tree-ring chronology exceed 500 years and 21 exceed 600 years in length. In the Carlton Ridge subalpine larch chronology, with the shortest average segment length (Table 1), 75% of the series used are greater than 300 years in length, with 8 series greater than 400 years. Thus, the reconstruction presented here likely preserves most of the centennial scale variability.

The early part of this reconstruction overlapped slightly with the end of the Medieval Warm Period (MWP) and presented an opportunity for comparison. Several authors have suggested warming between the 9th and 14th centuries nearly matched or exceeded warming observed in the modern period (Esper et al. 2002; Graumlich 1993; Briffa et al. 1992a). Evidence of the MWP is varied, present in some places but absent in others (Hughes & Diaz 1994). Graumlich (1993) found evidence of sustained warmth between about 1100-1375 in reconstructed summer temperature in the Sierra Nevada. Relative warmth between 1073-1120 in the Canadian Cordillera was reported by Luckman et al. (1997), but sample depth during this period of their reconstruction was low (<5 samples), limiting its reliability in this time period.

The reconstruction presented here lies essentially between these areas and it would be reasonable to expect evidence of warming during similar periods. It does appear as if summer temperatures were warm during the early part of the reconstruction
that overlapped the end of the MWP for at least the late 1200s and early 1300s (Figure 4). The subalpine larch chronology from Baker Lake that extended through all of the MWP also seemed to suggest conditions were favorable or near normal for tree growth shortly after AD 1000 through about AD 1300, but important departures from this pattern are also evident (Figure 7). Further, two of the largest peaks in tree-growth emphasized by the 50 year spline occurred roughly during the MWP at about 1130-1200 and 1240-1280 (Figure 7). However, there were also marked reductions in growth during the MWP and much of the time period appeared unremarkable in terms of positive tree-growth. Upper elevations in the SBW may have experienced some warming during the MWP but the evidence presented here was limited to one tree-growth chronology with adequate sample depth during that time period. Even so, warmth near the end of the MWP in the reconstruction of temperature did not appear substantially greater in magnitude or more persistent than other warm periods in the reconstruction.

Several reconstructions of temperature have identified pronounced cooling between about the 15th and 19th centuries following the MWP. Evidence of widespread cooling during this period, commonly referred to as the Little Ice Age, is most often found in extratropical temperature reconstructions (e.g., Esper et al. 2002; Briffa et al. 2001) or inferred from glacial maxima (Grove 2001). There was no evidence of persistent cooling in the SBW between ≈15th and 19th centuries. While cool periods occurred throughout this time, none persisted for an extended period relative to the calibration mean. Longer periods of relatively cool growing seasons were present in the other chronologies presented for comparison, and were quite noticeable in Northern
Figure 7. Baker Lake subalpine larch tree-growth chronology AD 1000-1997. Gray line are annual tree-ring index values. Smooth black line is a 50-year cubic smoothing spline to emphasize at low frequency variability.
Hemisphere temperature reconstructions (Figure 5). Even in the chronologies representing a smaller region, the Western United States and British Columbia/Pacific Northwest, this cooling trend was more pronounced and appeared to be of longer duration. However, these reconstructions also showed a slight increase in growing season temperatures around the middle 1800s that was more pronounced in the SBW reconstruction and occurred somewhat later.

High interannual variations in temperatures have been suggested as an important feature of the LIA (Briffa et al. 1990). Examination of the first differences of the temperature reconstruction, a form of high pass filtering, further dispels the notion of a Little Ice Age signal in this reconstruction. Both the early and later portions of the reconstruction appeared to have higher interannual variability than the portion generally considered to fall within the time frame of the LIA (Figure 8). Periods of low interannual variability such as the long period between about 1620-1710 occurred during what is generally considered the Little Ice Age. Interannual variability did increase near the end of the so called Little Ice Age but overall the reconstruction included many shifts between high and low variance rather than any persistent departures (Figure 8).

Anomalous warming during the 20th century likely due to anthropogenic activities has been well established with sound scientific evidence (Mann et al. 1998; Mann et al. 1999; Jones et al. 1999; Huang et al. 2000; Esper et al. 2002). However, the expression of the global warming signature is both spatially and seasonally variable (IPCC 2001).

Anomalous summer warming during the 20th century was not as evident in this reconstruction as has been indicated elsewhere. Summer temperatures over the
Figure 8. First differences of June-August temperature reconstruction in the SBW region. First differences calculated as $t - t_1$, where $t$ is the reconstructed temperature at time $t$. 
calibration period (1900-1997) and the entire reconstruction both averaged about 17.4°C. There was also no apparent trend in the regionally averaged summer temperature record used for calibrating the reconstruction over the 20th century. The slope of regression lines for both the instrumental temperature data and reconstructed values was near 0 (0.001 and 0.004 respectively) with small explained variance. This suggested that overall, summer temperatures during the 20th century have not warmed to the level reported for other areas. This result was consistent with the findings of Biondi et al. (1999) in central Idaho as well as for the larger interior Columbia River Basin (Ferguson 1999). Temperatures did appear to be increasing near the end of the reconstruction and appeared similar to earlier reconstructed warm periods such as the middle 1600s and late 1700s. However, this apparent summer warming has not yet reached the magnitude of warm events such as the early 1400s. It remains to be seen whether this warming near the end of the reconstruction simply represents a short term oscillation from cool to warm conditions or if it is the beginning of a persistent trend of above normal summer temperatures.

The lack of a strong warming trend near the end of the reconstruction does not, however, imply that the 20th century climate of the SBW region has been devoid of anomalous warming in some form. The restricted seasonal reconstruction of summer temperature more likely did not capture slowly rising temperatures that have occurred in other seasons. For example, spring temperatures in this area have warmed over the 20th century, particularly in June (Kipfmueller unpublished data). The diurnal temperature range has decreased as well primarily as a result of warmer minimum temperatures.
While *summer* temperatures in this region may have been little affected by anthropogenic climate forcing, other seasons have likely experienced important impacts.

*Spectral characteristics and the role of forcing mechanisms*

The most prominent peak in variance in the estimated spectrum occurred at around 87 years (Figure 9). Upper elevation tree-ring chronologies such as those used to derive the reconstruction often contain most of their variance at relatively low frequencies (LaMarche 1974). Spectral peaks $\approx$90 years have also been identified in temperature reconstructions in the San Francisco Peaks area of northern Arizona (Salzer 2000) and Fennoscandia (Briffa et al. 1990). Smaller peaks in variance occurred around 15 years and 2 years.

This reconstruction preserved more high frequency variability than other chronologies developed using upper treeline conifers. This was most likely related to the biological relationships of subalpine larch, the primary constituent of the reconstruction. While most subalpine conifers have considerable food storage capabilities from year-to-year, subalpine larch is influenced less by autocorrelation (Table 1). Because subalpine larch loses its needles at the end of each growing season (with the exception of juvenile trees, Amo and Habeck 1972) it is more reliant on current climate conditions for the production of new needles for photosynthesis. Other subalpine conifers such as spruce and whitebark pine, two of the most commonly used species for dendroclimatology at upper treeline, have a full complement of leaves already in place at the beginning of the
Figure 9. Estimated spectrum from reconstructed June-August average temperature (solid line) using a truncation point of 100 years. Dashed line denotes the 95% confidence limit and the dotted line constitutes a "white noise" process.
growing season and thus do not need to expend as much energy in production of new growth before photosynthesis can occur.

The variance structure of reconstructed temperature did not suggest much periodicity in the time series and there is limited support for any well documented forcing mechanisms that correspond to observed peaks. Volcanic forcing has been suggested as a potential mechanism for temperature at both annual and multidecadal timescales (Scuderi 1990; Graumlich 1993; Briffa et al. 1998; D’Arrigo & Jacoby 1999; Salzer 2000). At the timescale of one to several years, increased stratospheric sulfur produced by erupting volcanoes can reflect incoming solar radiation resulting in a reduction of surface temperature. Scuderi (1990) and D’Arrigo and Jacoby (1999) suggest that some large individual volcanic eruptions can lead to short term cooling from one to several years following the eruption event. Scuderi (1990) and Briffa et al. (1998) further suggest that the eruption of a series of volcanoes can lead to cooling over longer periods and glacier fluctuations in the Sierra Nevada have been linked to these occurrences. However, there was little correspondence between individual cool summers reconstructed in the SBW and known volcanic activity at either interannual or longer timescales with the notable exception of the cool periods around 1600 and 1450. Volcanic eruptions, while not imparting significant reductions in temperature reconstructed from the width of annual rings in the SBW do provide other evidence of past eruptions in the form of frost or light rings. Frost or light rings were often present in 1453, 1601, 1666/67, 1784, 1801 and 1884, coincident with known (or suspected) volcanic eruptions (Briffa et al. 1998).
The peak in density at ≈15 years may be related to the Pacific Decadal Oscillation. Biondi et al. (2001) and Gedalof and Smith (2001) identify peaks in spectral density around 20 years in their reconstructions of the PDO. Biondi et al. (2001) identified a peak in the power spectrum between 17 and 28 years with a maximum amplitude at about 23 years using tree-ring series collected in southern California and Mexico. Using tree-ring chronologies collected around the Gulf of Alaska, Gedalof and Smith (2001), found peaks in the power spectrum between 10-20 years.

The Pacific Decadal Oscillation does have strong teleconnections to climate conditions in the Pacific Northwest (Mantua et al. 1997) and might at least partially explain the peak in relative density around 15 years evident in this reconstruction. The most important PDO-climate interactions in the SBW region occur during the winter and spring months but tree-ring chronologies used in this reconstruction have a limited response to conditions during that part of the year. It may be that winter climate conditions influenced by the PDO have some carryover effect from the winter/spring season to the summer such as a more persistent snowpack that alters growth climate relationships thereby moderating tree growth-climate relationships. More detailed analysis and testing of specific hypotheses regarding these relationships is necessary to fully understand the factors responsible for low frequency growth relationships in this area.
Conclusions

Calibration and verification statistics for this 748 year reconstruction of summer temperature in the SBW demonstrated its potential reliability as a proxy for summer temperature variability over long periods. The regression model developed from three upper treeline chronologies had qualities that were similar to reconstructions of temperature produced in other regions. Reconstructed summer temperature appeared to oscillate between relatively long periods of warm or near normal temperatures and shorter term cool events. There was weak evidence of warming near the end of the Medieval Warm Period but little evidence of widespread cooling during the Little Ice Age. Evidence of anomalous summer warmth during the 20th century was also not apparent in this reconstruction as others. Spectral analysis revealed peaks in the power spectrum at ±90, ±15, and 2 years.

Acknowledgements

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depth in its earliest periods. Franco Biondi and Brian Luckman provided access to their datasets for comparison. Valuable comments on earlier drafts of this manuscript were provided by Tom Swetnam and Malcolm Hughes.

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APPENDIX C

FIRE-CLIMATE INTERACTIONS IN SUBAPLINE FORESTS OF THE
SELWAY-BITTERROOT WILDERNESS AREA, IDAHO-MONTANA, USA

Kurt F. Kipfinueller & Thomas W. Swetnam

This paper will be submitted to the Canadian Journal of Forest Research

Abstract

The interactions between fire and interannual climate variability in subalpine forests are poorly understood. Fire history was reconstructed in four subalpine watersheds in the Selway-Bitterroot Wilderness Area (SBW), located on the border of Idaho and Montana. Crossdating techniques were used to assign exact calendar dates to fire events enabling comparisons with interannual climate variability. Fire scar and age structure information from nearly 2,000 trees were used to differentiate between widespread fires and those that were smaller. Fire dates were compared to tree-ring reconstructed summer temperature, Palmer Drought Severity Index (PDSI), and Southern Oscillation Index (SOI). Instrumental climate data were compared to the ten largest fire years since 1895. Mean fire intervals ranged from about 19-173 years in the four watersheds for any fires occurring anywhere within a watershed, and between about 40-350 years for widespread fires within a watershed. The mean interval between fires occurring within a small stand ranged from about 140-350 years. Only PDSI was related to tree-ring reconstructed fire years. Reconstructed fire years were significantly dry ($p<0.01$) as was the summer prior to fire occurrence ($p<0.05$). Both summer temperature
and summer PDSI were significantly associated with modern fire events (p<0.05).

Examination of the spatial patterns of during reconstructed fire years suggests multiple synoptic patterns are related to widespread fire occurrence in the SBW. One spatial pattern suggests El Niño conditions could be related to widespread fires in the region.

**Introduction**

The recent fire years in 2000 and 2002 in which 3.4 million hectares and 2.7 million ha burned respectively across the United States (National Interagency Fire Center 2002) have focused national attention on fire management activities in United States, particularly the west. Suppression costs for the 2000 fire season topped $1.3 billion (USD) with the costs of the 2002 season expected to exceed this amount (National Interagency Fire Center 2002). Further, efforts geared toward rehabilitating lands affected by large fires will likely add significantly to these financial costs.

While the factors leading to large areas burned are complex, accumulation of fuels due to successful fire suppression over the last century, and episodic drought are the most commonly cited agents by scientists, politicians and the news media. Effective fire suppression, either directly or via land use activities such as grazing have led to increases in fuel loadings in many forests that historically experienced infrequent fires burning at low severity (Covington & Moore 1994; Hann et al. 1997; Fulé et al. 2000). The development of drought conditions can then lead to large fires that burn with high severity.
While fire suppression surely played an important role leading to reductions in fire frequency in the 19th and 20th century and increases in severity in the late 20th century, these effects are likely more pronounced in lower elevation forests that had historically burned with relatively high frequency. In the past, these fires were mostly low severity surface fires that reduced fuel loads maintaining a relatively open forest with low fuel accumulations. However, upper elevation subalpine forests generally do not support surface fires but rather high intensity crown fires at intervals ranging from 100-300 years (Arno 1980; Romme 1982; Agee 1993). In these forests, fuel accumulation as a result of only 100 years of fire suppression may not be extraordinary, and climate and weather rather than fuel accumulations may play a dominant role in driving large fire events (Bessie & Johnson 1995).

A broad field of research has developed investigating the weather and climatic elements resulting in large fires in upper elevation forest communities and their related boreal counterparts (Larsen & Delavan 1922; Flannigan & Harrington 1988; Johnson et al. 1990; Larsen & MacDonald 1995; Potter 1996; Nash & Johnson 1996; Rorig & Ferguson 1999; Hely et al. 2001; Wierzchowski et al. 2002). The majority of this research has been confined to the relationships between area burned and short-term (i.e., intraseasonal) weather events. Severe fires in the subalpine and boreal zones have been related to persistent blocking high pressure systems that result in extended periods of high temperature and low relative humidity (Flannigan & Harrington 1988; Fryer & Johnson 1988; Johnson 1992; Johnson & Wowchuck 1993; Bessie & Johnson 1995; Nash & Johnson 1996). In addition, the breakdown of the high pressure system due to the
passage of cold fronts can result in periods of increasing wind coupled with storms that often have little precipitation but abundant lightning due the instability of the atmosphere (Brotak & Reifsnyder 1977; Finklin 1981; Finklin 1983; Rorig & Ferguson 1999).

The interactions between fire and interannual to decadal climate patterns have been well studied in many parts of the western United States including the sequoia groves of the Sierra Nevada (Swetnam 1993), ponderosa pine forests of the Southwest (Baisan & Swetnam 1990; Covington & Moore 1994; Grissino-Mayer 1995; Swetnam & Baisan 1996; Fulé et al. 1997), Oregon (Heyerdahl et al. 2002), and the Front Range of Colorado (Brown et al. 1999; Veblen et al. 2000; Donnegan et al. 2001). Comparisons between fire and climate have uncovered more complex interannual and longer time-scale relationships between fire and climate than the simplistic assumption that fires occurred during dry years. In sequoia groves, for example, Swetnam (1993) found high frequency relationships with precipitation and fire occurrence while low frequency trends in fire occurrence were related mostly to long-term temperature change. In contrast, Heyerdahl (Heyerdahl et al. 2002) found that both interannual and decadal precipitation were associated with fire occurrence in Oregon. In southwest and Colorado Front Range ponderosa pine forests fires typically occur during dry years preceded by one to several years of wet conditions promoting fuel build up (Swetnam & Baisan 1996; Brown et al. 1999; Veblen et al. 2000; Donnegan et al. 2001).

In addition, important relationships between fire occurrence and the El Nino-Southern Oscillation have been noted in Southwestern forests (Swetnam & Betancourt 1990), the Southeastern United States (Simard et al. 1985; Harrison & Meindl 2001),
interior Alaska (Hess et al. 2001), the Pacific Northwest (Heyerdahl et al. 2002) as well as forests in South America (Veblen et al. 1999; Kitzberger et al. 2001; Kitzberger 2002) and the tropics (Phulpin et al. 2002). La Niña (cool events) often result in dry conditions in the Southwest and have been linked to large area burned in that region (Swetnam & Betancourt 1990) and in Colorado (Donnegan et al. 2001). In contrast, El Niño (warm events) can produce enhanced spring time precipitation in the southwest (Andrade & Sellers 1988; Sheppard et al. 2002) resulting in increased production of fine fuels that burn readily during subsequent dry years (Swetnam & Betancourt 1998; Veblen et al. 2000). Teleconnections are weaker in other regions such as the northern Rocky Mountains and Pacific Northwest, but warm El Niño events have been linked to reduced spring precipitation (Harrison & Larkin 1998).

While many fire histories have been completed in the Northern Rocky Mountains, few have explicitly examined the interactions with interannual (or longer time-scale) patterns of climate (but see Barrett et al. 1997). For the most part, investigations in the northern Rockies have focused on changes in fire intervals resulting from human activities, particularly with respect to fire suppression. Crossdating, a technique that ensures the precise assignment of a calendar date to a fire scar, has been used in few of these studies limiting comparisons between climate and fire events which require annual precision not attained using ring counts.

In this study we examine the relationship between fire occurrence and interannual climate characteristics in subalpine forests of the Selway-Bitterroot Wilderness Area. Crossdated fire histories developed from a combination of fire scar and age structure
analysis were used to identify fire events in four subalpine watersheds. We used age structure information to differentiate between fires that burned relatively larger areas versus fires that were isolated. Fire-climate interactions were examined by comparing fire dates to various tree-ring reconstructed climatic time series, including summer PDSI, summer temperature, and an index of the Southern Oscillation. Interannual fire-climate relationships were assessed by comparing fire dates and climate time series using superposed epoch analysis (SEA). Finally, we examined relationship between climate and large fire occurrence between 1900-1997 using instrumental climate records and fire atlas data compiled by Rollins (2000).

Study Area

The ≈542,000 ha Selway-Bitterroot Wilderness Area (SBW) is located on the border of Montana and Idaho (Figure 1). Elevation ranges from 463 m along the Selway River in Idaho to 3,048 m on Trapper Peak in the southeast portion of the wilderness. Subalpine forests are comprised of varying compositions of lodgepole pine (Pinus contorta), whitebark pine (Pinus albiculis) Engelmann spruce (Picea engelmannii), and subalpine fir (Abies lasiocarpa). More extreme sites near upper treeline sometimes contain subalpine larch (Larix lyallii).

Finklin (1983) provides a detailed summary of the climate of the SBW. Temperatures around 2,100 m are generally cool in the subalpine zone with average temperatures in July of about 12°C and -8°C in January (Finklin 1983). Around 170 cm of precipitation is possible at the highest elevations in the SBW, with about 70% of this
Figure 1. Location of four sampled watersheds within the SBW. The approximate location of PDSI grid point 27 is shown by the black circle in Montana on the state map.
accounted for by snowfall (Finklin 1983). Summertime precipitation is from convective storms where precipitation from a single storm can be locally heavy. Summer drought is common throughout the SBW during July and August and is often associated with the occurrence of dry lightning storms. The fire season lasts from late June through early September with activity peaking between mid July and early September coincident with the development of convective storms with enhanced lightning activity (Habeck 1972; Brown et al. 1994).

Four subalpine watersheds were selected for fire history data collection (Figure 1, Table 1). Watersheds were composed primarily of lodgepole pine, and subalpine fir with smaller amounts of Engelmann spruce and whitebark pine (Table 2). Douglas-fir and subalpine larch were rare or absent in most watersheds (Table 2).

Watersheds were selected to include forests that had not been completely burned during the 20th century, and had substantial areas of old-growth forest to facilitate the reconstruction of a multi-century fire history. An additional site selection criteria included the existence of a lake suitable for palynological sampling as part of a companion study carried out by colleagues at the University of Oregon (Brunelle-Daines 2002). Finally, accessibility of each watershed was an important consideration to reduce costs associated with horse and helicopter transportation of equipment and samples. The selected watersheds for fire history examination are therefore located near the periphery of the wilderness boundary within one mile of foot trails.
Table 1. Locations and sampling information for watersheds used for fire history reconstruction.

<table>
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<tr>
<th>Watershed Name</th>
<th>Abbreviation</th>
<th>Lat./Lon.</th>
<th>Elevation range (m)</th>
<th>Approximate area (ha)</th>
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<td>1753-2240</td>
<td>415</td>
</tr>
<tr>
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<td>HOO</td>
<td>46.33°/114.53°</td>
<td>1688-2140</td>
<td>2005</td>
</tr>
<tr>
<td>Baker Lake</td>
<td>BKR</td>
<td>45.88°/114.25°</td>
<td>2050-2789</td>
<td>375</td>
</tr>
<tr>
<td>Burnt Knob Lake</td>
<td>BKL</td>
<td>45.72°/115.00°</td>
<td>1718-2490</td>
<td>438</td>
</tr>
</tbody>
</table>

Table 2. General forest compositional characteristics of the four sampled watersheds. Basal area (BA) is reported as m²/ha and stem density (Dens.) is the number of stems/ha. Data are from intensive age structure transects and grid points (N) averaged together for each watershed.

<table>
<thead>
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Methods

Field Sampling

Transects were established to collect age and forest structure data within each of the four watersheds. Transects for collecting age and forest composition data were located in forest stands identified from aerial photographs and high view points. Forest stands were delimited visually based on structural and compositional differences that were probably related to disturbance history. Nested belt transects were established to characterize the age, species composition, and structural characteristics of stands within each watershed. Transects were established within each stand more than 200 m away from the perimeter to minimize edge effects.
The size of each transect depended upon the structure of the stand. Transects 100 m in length were established in old-growth stands characterized by a high degree of complexity such as multiple canopy levels, variable tree size, and high species diversity. Transects 50 m in length were used in high density, even-aged stands. The width of each transect varied in a manner that ensured collection of trees from all species and size classes. Because large-sized trees occur at lower densities than those of smaller size classes, we varied transect width to include more large sized trees that would otherwise be missed by a relatively narrow transect. In these nested belts only trees larger than the predominant size class (estimated in the field) were sampled. The nested belt transect technique also ensured an adequate number of trees was collected to accurately document the age structure of a fire initiated patch (Abolt 1997).

In general, transect size was established to collect age data from approximately 50-100 trees from each stand. Along each transect we recorded the X-coordinate, species, diameter at breast height (dbh), and canopy position of all live trees >1m in height. Increment cores were collected from each living tree and the height of the sample on the tree bole was recorded. Trees were cored up to four times until either a sample near the pith was obtained or rot prevented the collection of the innermost rings. In some stands we collected additional samples from outside the transect to further augment age data. These samples were mostly from standing snags of whitebark pine and were sampled to estimate their date of mortality related to mountain pine beetle (*Dendroctonus ponderosae* Hopk.) outbreaks. In other cases trees of approximately the same size or larger than trees in a transect were sampled to identify possible survivors of the most
recent fire, or to provide additional age structure information to account for rotten centers in large transect trees. In all cases additional sampling was conducted within ≈25 m of the transect. These samples were excluded from stem density and basal area calculations and were within the identified fire initiated stand. Increment cores were labeled and stored in paper drinking straws for transport back to the laboratory for crossdating.

In addition to the intensive age-structure transects, we established a gridded network of points to collect additional age and forest structure data. Points were established approximately 500 m apart along a north-south and east-west grid centered around each lake. Data was collected at 15-31 grid points depending on the watershed size. Stem density and basal area for each tree species present was determined using a variable radius plot and angle gauge. Increment cores were collected from 5-10 trees at each point to estimate the time since last disturbance. In addition, snags were sometimes sampled to identify recruitment and mortality dates of trees apparently killed by the most recent fire or some other disturbance such as insect infestation. Tree-age data collected from grid points was used to supplement the intensive age structure information and was used to aid in interpreting the spatial scale of past fire events.

In fire regimes characterized by infrequent stand replacing fires, such as the subalpine forests of the northern Rocky Mountains, fire scars are somewhat rare (Arno et al. 1993). Nonetheless, fire scars were readily found along drainages, near lakes, ridgetops, or on the edges of fuel breaks. In addition, pockets of surviving, fire-scarred trees were sometimes found within a sampled stand. Such “fire-scar refugia” were probably places where trees survived stand-replacing fires and supported some limited
surface fire (Turner & Romme 1994). Intensive searches for fire scars were conducted in and around each transect. Additional systematic searches around probable fire-scar refugia such as ridgetops were also conducted. In addition, fire scars were collected opportunistically when encountered. For each sampled tree we recorded species, DBH, degree of preservation and location using a GPS. Using a chainsaw, or crosscut saw, a partial wedge was extracted from living trees, snags, and logs containing visible fire scars (Arno & Sneck 1977). The sampling of snags and logs enabled us to extend the fire history of the area further back in time than would otherwise be possible using only living trees. Due to its status as a wilderness area, we were limited by US Forest Service managers to collecting no more than 40 fire-scarred samples from each watershed.

Laboratory Sample Processing and Analyses

In the laboratory, all increment cores and cross-sections were surfaced using successively finer grades of sandpaper following standard techniques (Stokes & Smiley 1968). The process of preparing high quality surfaces is crucial to resolve annual ring structures in samples where there are often growth suppressions following fire scars. Samples were then crossdated to ensure the assignment of precise calendar dates to annual growth rings using skeleton plotting techniques and characteristic marker years (Stokes & Smiley 1968). Age structure, grid point, and fire scar samples were crossdated using master chronologies developed from climatically sensitive trees growing near or within each watershed.
Establishment dates for trees sampled from age-structure transects and grid points were estimated using a pith estimator when the center of the tree was missed by the increment borer (Applequist 1958). Given the difficulty of accurately assigning germination dates to trees cored above the root-shoot interface, and with missed pith, estimated germination dates were grouped into ten-year age classes by species to identify the age structure of each sampled patch (detailed age structure presented elsewhere). Core samples were omitted where the number of rings to a tree’s center could not be estimated within 20 years.

Several lines of evidence were used to classify isolated, or local fires from those in the watershed that were widespread. Widespread fires were defined as those fires recorded by a fire scar with 1) an associated post-fire recruitment pulse identified in transects and grid point data, or 2) a fire year recorded by more than one fire-scar sample separated by > 500 m. Additional evidence of widespread fires was provided by surviving trees or dead snags within a fire-initiated patch of more recent origin (i.e., representing survivors) and with inner ring dates closely matching a known fire event recorded elsewhere in the watershed. In these instances only whitebark pine, lodgepole pine, and subalpine larch germination dates were considered. The logical basis of this inference are the facts that Clark’s nutcracker’s (Nucifraga columbiana) preferentially cache whitebark pine seeds in areas that have been recently burned (Tomback 1982; Tomback et al. 1993) and lodgepole pine in the region often have serotinous cones providing a readily available seed source following fire events. Moreover, subalpine larch is among the most shade intolerant conifers in the northern Rockies and requires
open conditions such as those created by stand-replacing fires or avalanches for successful germination (Arno & Habeck 1972). In contrast, Engelmann spruce and subalpine fir are shade tolerant and often do not germinate immediately after a fire event because they are readily out competed by fast growing lodgepole pine or other shade intolerant conifers (Smith & Fischer 1997).

Isolated fires were defined as fire events recorded as a single fire-scarred tree or group of fire scars in a small area (<500m apart). In addition, a lack of tree regeneration associated with these localized fire scars suggested these fires burned at a lower intensity. However, it is important to note that some of these fires could have been widespread but evidence of their former extent has been erased from the landscape by subsequent fires or decomposition processes.

**Fire-Climate Relationships**

The dates of fire events, as determined by fire-scar analysis, were entered into FHX2 fire history software (Grissino-Mayer 2001) to quantify the descriptive characteristics of the fire regimes of the sampled watersheds including mean fire interval (MFI) and fire frequency. Estimated stand establishment dates that lacked an associated fire scar were not included in the analyses due to the imprecise nature of assigning exact fire dates. Fire events were analyzed for all fire years and for widespread fire events separately. Mean fire interval is a commonly used statistic which describes the average length of time between successive fires within an area of interest (Agee 1993). Note that mean fire interval should not be confused with “fire rotation” or “fire cycle”. The latter
terms refer to typical intervals required for entire areas to burn over, or average fire interval at any given location. Mean fire intervals, in contrast, refer to the occurrences of particular fires anywhere within a specific study area and time period (Romme 1980). An estimate of the fire interval at any given point on the landscape was calculated by examining the age structure and fire evidence for each individual age structure transect and grid point and considering the existence of survivors and snags within a site.

Superposed epoch analysis (SEA) was used to examine the interannual association of fire events with various climate variables. SEA is a commonly used statistical technique in fire-climate analyses that computes the mean value of the climate parameter using each observed fire year and the years prior to fire occurrence (Baisan & Swetnam 1990; Swetnam & Betancourt 1992). This assists in the assessment of antecedent climate conditions that may be related to fire occurrence as well as determination of the conditions during the fire year. Confidence intervals were estimated using 1000 randomly generated Monte Carlo simulations. We note that results tend to be more erratic in cases where sample size is small (<10 events). There were generally <10 widespread fire events at the watershed scale, so the four watersheds were aggregated for a single analysis rather than examining each watershed individually. SEA was performed using all fire years, local fire years, and widespread fire years.

Tree-ring reconstructed climate variables that were compared with the fire events using SEA included summer Palmer Drought Severity Index (PDSI) (Cook et al. 1999), the Southern Oscillation Index (SOI), an index of El Niño-La Niña conditions (Stahle et al. 1998a), and tree-ring reconstructed average summer (June-August) temperatures
(Appendix B). The SOI was the reconstructed December-January sea level pressure difference between Tahiti and Darwin, Australia from 1706-1977 (Stahle et al. 1998b). Negative values of the SOI indicate warm events (El Niño conditions) and positive values signify cool events (La Niña conditions).

Comparisons between fire events and PDSI were conducted using Cook et al. (Cook et al. 1999) grid point #27, located near the southeastern end of the SBW (Figure 1). This grid point likely includes the climate stations and tree-ring sites closest to the SBW and is not weighted by the more northerly climate stations associated with the wetter Glacier National Park area, but it may incorporate stations located on the plains. Calibration of the tree-ring data with instrumental PDSI was good with an $R^2=0.64$. Verification statistics were weaker, however, with reduction of error (RE) equal to about 0.49 and coefficient of efficiency (CE) slightly negative (-0.20) (see Cook et al. 1994 for description of verification statistics).

The relationship between modern fires and climate was also examined using SEA. The ten largest fire years in the SBW based on area burned between 1895-2000 from digital fire atlas data (Rollins 2000; Rollins et al. 2001) were compared to instrumental climate data to identify interannual relationships. Montana Climate Division 1 average temperature and average PDSI for the months of June-August obtained from the National Climatic Data Center were compared to the 10 modern fire dates. An instrumental index of SOI was obtained from the National Oceanic and Atmospheric Administration’s Climate Prediction Center to examine 20th century SOI-fire interactions. The same 6 year
window and Monte Carlo simulation procedure was employed to determine the climatic relationships with modern fire events.

The spatial patterns of PDSI over the coterminous United States in relation to widespread fire years in the SBW were evaluated using a principal components analysis using the years of widespread fires since 1700 and the gridded PDSI dataset. The shorter temporal scale was necessary because some grid point reconstructions do not begin until 1700. The entire 154 point PDSI reconstruction of Cook et al. (Cook et al. 1999) was reduced to its most important principal components (eigenvalue >1) using the Kaiser-Guttman criterion (Legendre & Legendre 1998). The principal components were rotated using varimax rotation and the rotated scores were then plotted onto the grid of reconstructed PDSI to identify drought patterns related to widespread fires in the SBW. Rotation of the axes results in individual variables (fire years) loading most strongly on a single component, and facilitates interpretation of the relationships between drought patterns and widespread fire years (Reyment & Jöreskog 1993). Rotated factor loadings are the correlation coefficients between the factor and the original variables, in this case the fire year. Contour maps of the principal components were constructed by using an inverse distance weighting method and a 250 km search area for interpolation using the factor scores of the individual grid points.
Results

Sample Size

A total of 96 fire-scar samples were dated in the four watersheds yielding 45 fire years (Table 3). Widespread fires within watersheds were identified in seventeen years, fourteen of which occurred since 1700. Estimated germination dates were determined for 1,443 trees collected from 23 intensive age structure transects in the four watersheds. An additional 461 estimated germination dates were obtained from 78 grid points in the four watersheds. Some samples, both fire-scar and age-structure, were not used because complacent ring width patterns prevented crossdating. Additional age structure samples were omitted due to the presence of rot near the center of a tree making estimations to the center impossible. Despite the omission of these samples, this data set remained large by most fire history standards, particularly since sampling density in these relatively small watersheds is high.

Table 3. Fire history statistics for four sampled watersheds in the SBW, full time period. The mean fire interval (MFI) is the average number of years between fires occurring anywhere within a watershed. MFIws is the mean fire interval between widespread fire events in years. MFIsite represents the average fire interval at a given point within the watershed determined from age structure transects and grid points and includes the interval from the most recent fire to the present.

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<th>No. Fires</th>
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<th>Fire frequency</th>
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³Excludes the 534 year fire interval between 1204-1738.
Fire history characteristics of individual watersheds

The spatial and temporal pattern of fire occurrence are described in this section, beginning with the most northerly site (Figure 1).

**Beaver Lake (BVR):** Twenty-four fire-scarred samples were successfully dated from samples collected in the BVR watershed (Table 3). These samples recorded 12 fire years with the earliest occurring in 1639 and the most recent in 1919 (Figure 2A, Table 3). In addition, a large fire in 1919 was identified from historical fire atlas and dendrochronological data. Mean fire interval (MFI) of a fire in any part of the watershed was approximately 26 years while the mean interval between widespread fires was approximately 43 years (Table 3). For small stands, MFI is approximately 212 years.

Of the twelve fires identified, five had evidence suggesting they were widespread (Figure 3A). The 1748 fire was recorded by only a single fire scar sample but is also recorded as an old stand of mostly lodgepole pine with inner-ring dates near 1750 in one transect within the watershed. The fire in 1801 was also restricted largely to areas near the bottom of the cirque basin (mostly around the lake) though subsequent fires may have erased evidence of its full extent. The 1901 fire burned areas that appeared to have been previously burned in an 1864 fire, particularly in the southern part of the watershed along the ridge line (Figures 1 & 3A). Evidence along the ridge suggested this fire burned as a low intensity surface fire, scarring only a few widely scattered whitebark pine, but there were also stands of mixed subalpine fir and lodgepole pine that appeared to have germinated in response to the fire. The 1919 fire burned in the lower (northern) part of the basin and was documented in the fire atlas (Rollins 2000).
Figure 2. Master fire chronologies for each watershed. Each horizontal line represents one sampled tree, and inverted triangles represent fire events. Filled triangles depict stand-replacing fire events. Composites of all fire years in each watershed are shown along the bottom. Watersheds are arranged from north (top) to south. (A) Beaver Lake, (B) Hoodoo Lake, (C) Baker Lake, (D) Burnt Knob Lake.
(C) Baker Lake

- Local fire event
- Widespread fire event
- Non-recording years
- Recorder years
- Inner or outer ring date
- Pith or bark date

(D) Burnt Knob Lake
Figure 3. Locations of evidence recording widespread and local fires in the four sampled watersheds. (A) Beaver Lake, (B) Hoodoo Lake, (C) Baker Lake, and (D) Burnt Knob Lake. The two types of evidence used for inferring fire size are shown—fire scars (filled circles) and age data (open triangles).
(B) Hoodoo Lake

1794

1851

1889

1934

Local Fires

● = Fire Scar Date
△ = Age Data
--- = Streams
--- = Sampling Area

1 km
Local Fires

\[ \text{Fire Scar Date} \]
\[ \text{Age Data} \]
\[ \text{Streams} \]
\[ \text{Sampling Area} \]

1 km

N
Burnt Knob Lake

1527
1580
1709
1719
1729
1883

Local Fires

- = Fire Scar Date
\(\Delta\) = Age Data
--- = Streams
---- = Sampling Area

1 km

\(\uparrow\) N
Hoodoo Lake (HOO): Thirty-seven fire scar samples were successfully crossdated yielding fifteen fire dates (Figure 2B, Table 3). The earliest dated fire in the watershed occurred in 1660 and the most recent in 1934. The mean fire interval in the Hoodoo Lake watershed was about 20 years with MFI of widespread fires near 50 years, and MFI at a small site 139 years (Table 3).

Widespread fires within the watershed occurred in 1794, 1851, 1889, and 1934 within the watershed. The 1851 fire was recorded as a fire scar on only one sample but age structure evidence suggested this was a widespread fire (Figures 2B & 3B). There was considerable spatial overlap between the 1851 fire and 1889 fire in the watershed, and many of the fire-scarred trees recording the 1889 fire appeared to have germinated shortly after the 1851 fire (Figure 2B). The 1934 fire is recorded by four trees far from one another. There is little age structure evidence related to the fire year suggesting that it may have been many small spot fires or burned at low intensity through recently burned areas.

Baker Lake (BKR): There was little evidence of fire in the Baker Lake watershed. Only nine of thirteen fire-scar samples that were collected were able to be reliably crossdated (Figure 2C). Many samples had too few rings to establish a convincing crossdating sequence and others were too complacent. While the sample size was very low for this watershed it reflects a lack of available sample trees more than a reduced effort (see discussion below).

Only five fire years were identified in the fire-scar record and three of these fires were apparently widespread (Figures 2C & 3C). The earliest fire was recorded in 1204,
and this was the earliest fire recorded in any of the watersheds (Figure 3C & Figure 4). The 1204 fire in BKR was recorded on only two fire scars that were near one another. However, near-pith dates of eight Douglas-fir trees scattered throughout the watershed provided additional evidence that this fire was widespread (Figure 3C). It is interesting to note that these Douglas-firs were growing within stands dominated today by subalpine larch and whitebark pine, species that were normally found at much higher elevations than Douglas-fir. Moreover, several subalpine larch and whitebark pine sampled in intensive age structure transects and at grid points had pith dates following 1204 very closely (within 10-20 years).

Figure 4. Selway-Bitterroot four site composite fire chronology. Solid vertical lines at the bottom represent widespread fires and dashed vertical lines depict local fires.
The 1748 fire was recorded by two trees growing in close proximity, however age structure evidence suggested this fire occurred over a larger portion of the watershed (Figure 3C). The 1896 fire burned mostly in the eastern end of the watershed and is present today as a relatively contiguous lodgepole pine forest.

Mean fire interval in BKR was long at 173 years for all fire events and 346 years for widespread fires (Table 3). The much longer fire return interval is mostly a product of the long gap between fires in 1204 and 1738. Excluding the 1204 gap results in a mean fire return interval of \(~53\) years. The MFI\textsubscript{ws} was also very long at 346 years, also driven primarily by the long gap in the fire-scar record. Omitting this long gap leads to an MFI\textsubscript{ws} of 148 years. Mean fire intervals for a given stand in the watershed was \(\approx340\) years including the long interval from the 1204 fire, or 233 when it was omitted.

*Burnt Knob Lake (BKL):* Thirty-four dated fire-scar samples recorded sixteen fire events in the BKL watershed. The earliest recorded fire occurred in 1527 with the most recent fire in 1910. Of these sixteen fire events, age structure and fire scar evidence suggested six were widespread. MFI in the watershed was about 26 years with a MFI of widespread fires near 71 years (Table 3). The mean interval between fires in a stand within the watershed was 235 years.

The 1527 and 1580 fires in the BKL watershed were identified by several widely dispersed fire scars and also scattered whitebark pine individuals (usually snags), with inner dates closely following the fire dates (Figure 3D). The fires in 1709, 1719, and 1729 fires were represented by fire scars without an associated age class. The areas
burned by these fires appear to have also burned during the 1883 fire that likely
eliminated any regeneration initiated by the earlier fires (Figure 3D). Even if
regeneration pulses did exist for these fires it would be difficult to determine which of the
three fires they belonged to given the coarse temporal resolution of the age structure data.

The 1883 fire was extensive, burning over nearly the entire watershed and leaving
only a few, isolated, old-growth stands. A large patch of primarily even-aged lodgepole
pine that germinated shortly after this fire was evident throughout much of the watershed.

Temporal patterns of fires

For the most part, fire occurrence appears random with respect to the temporal
pattern of occurrence (Figure 4). There are no obvious clusters of fire years during any
period in the fire history, with the possible exception of the late 19th or early 20th century
(i.e., 1883, 1901, 1896, 1901). Three fires also occurred in Burnt Knob Lake at 10 year
intervals between 1709-1729. A widespread fire near the end of the 19th or early 20th
century is present in each of the four watersheds (Figures 3 & 4). These most recent fires
have possibly eliminated some evidence of past fires.

Only four years were coincident in more than one watershed, 1715 (BVR and
BKL), 1748 (BVR and BKR), 1842 (HOO and BKR) and 1812 (BVR and HOO, Figure
3). Of these fire years 1748 was the only year that was also identified as widespread
within a watershed (BVR & BKR, Figures 3A & 3C).

The MFI for all watersheds combined was approximately 20 years, but there was
considerable variation between watersheds. Widespread fires within individual
watersheds occurred at intervals ranging from 43-346 years. Consistent periods of no detected fires occurred between 1749-1793 and 1802-1850 (Figures 4 & 5). Taken together, a nearly 100 year gap in widespread fire occurrence in the four watersheds occurred, interrupted only by widespread fires in 1794 and 1801 in Hoodoo Lake and Beaver Lake respectively. PDSI during this time period appeared to have been less variable and drought may have been infrequent for much of this time period (Figure 5). Small fires however, as indicated by single fire-scarred trees, did occur throughout this period (Figures 4 & 5).

The largest (1910) and third largest (1934) fire years between 1885-1996 in the SBW fire atlas, occurred in Burnt Knob Lake and Hoodoo Lake. A 1910 fire was restricted to a small area around Burnt Knob Lake and does not appear to have been widespread. Neither of these two fires were mapped in the fire atlas in these watersheds (Rollins 2000). Further, the 1889 fire identified in Hoodoo Lake was assigned the year 1895 in the SBW fire atlas and the 1883 fire in Burnt Knob Lake was reported to be 1900. In Beaver Lake the fire atlas perimeters a 1919 fire correspond well to the tree-ring record, but a 1901 was not cataloged in the atlas. No fires were documented in the atlas in Baker Lake, but the 1896 occurrence was clearly recorded in the tree-ring evidence.
Figure 5. Reconstructed PDSI in relation to fire events in the SBW. Heavy dark line is a 20 year spline used to emphasize lower frequency variability. White triangles represent local fires, gray circles widespread fires, and white squares modern fires.
Fire-climate associations

No statistical association was detected between local fires and the climate variables that were tested (Figure 6). However, fires in all watersheds combined and widespread fires occurred during years that were significantly dry during both the fire year and the year prior to fire occurrence (Figure 6). Large fires during the modern period were related to warm-dry summers but lagging relationships were unimportant (Figure 6). There appeared to be no relationship between SOI and fire events in the reconstructed fire record or to modern fires when compared using SEA.

The spatial pattern of drought related to widespread fires varied. Dry conditions in the Pacific Northwest appeared to consistently coincide with widespread fires in the SBW. Prior year droughts in the region were sometimes present and sometimes not (Figure 7). With few exceptions, reconstructed widespread fires and the ten largest fires during the modern period occurred during years when regional conditions were drier than average (Figure 7). Local fires were almost evenly divided between wet (15) and dry (13) years (Figures 5 & 7). In addition, 9 tree-ring reconstructed fire years occurred during a dry summer that was also preceded by a dry summer (Table 4). In contrast, only two widespread fires occurred during a wet summer and only one widespread fire occurred during a wet summer that was also preceded by wet summer conditions (Table 4).

It is interesting to note that reconstructed summer PDSI during 1910, one of the largest fire years in the western United States (and the largest in the SBW fire atlas, Table 4), was wetter than average. Instrumental summer PDSI during 1910 also indicated
Figure 6. SEA results between climate variables and different fire event types. Gray bar is the departure of the actual measured variable minus the simulated value estimated from 1000 Monte Carlo simulations. Solid horizontal lines are the 95% confidence limits and dashed lines are the 99% confidence limits. Note that sample size is smaller for SOI comparisons because the reconstruction does not begin until 1706.
Figure 7. Tree-ring reconstructed PDSI since 1700 during widespread fire years and the year prior. Maps are presented as couplets with PDSI during the fire year shown on the bottom and the PDSI from the year prior (lag 1) shown on top. (A) Reconstructed fires. (B) Modern fires (note 2000 is absent because it is not included in the gridded PDSI data set). Black dot shows the approximate location of the study area. Data from Cook et al. 1999.
Table 4. Summer Palmer Drought Severity Index during the fire year and one year prior to widespread fire occurrence in the SBW. Reconstructed PDSI for grid point 27 (Cook et al. 1999) is shown for dendrochronologically dated fires and instrumental PDSI is presented for modern fires. Patterns are summarized by indicating the combination with “W” representing wet conditions and “D” dry conditions. Asterisks indicate the significance level of individual years: * p<0.05, **p<0.01, ***p<0.001.

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</tr>
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<td>-0.68*</td>
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<td>W/D</td>
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<tr>
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<td>-2.54***</td>
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<td>-0.72*</td>
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Dendrochronologically Dated Fire Events

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<th>Fire Year (rank)</th>
<th>Prior Year PDSI</th>
<th>Fire Year PDSI</th>
<th>Area Burned (ha)b</th>
<th>Combination (t1/t2)</th>
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<td>-2.92***</td>
<td>11,172c</td>
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Modern Fire Events

Note: aYears also in tree-ring record of fires. bArea burned represents the area within the Selway-Bitterroot Wilderness boundary, but most fire years were larger due to areas burned outside the boundary. (Data from Rollins 2000). cData from Missoula Fire Lab.
conditions were relatively wet with a PDSI value of 1.40. The calculation of PDSI includes persistence from prior months that may have resulted in an overestimation of the index for 1910 (Palmer 1965; Heddinghaus & Sabol 1990).

Larger-scale patterns of continental drought during widespread fire years also had substantial spatial variability (Figure 8). The PCA analysis suggested four spatial modes of drought related to large fire occurrence in the SBW. Four PCs were retained based on the Kaiser-Guttman criterion of considering only principal components with eigenvalues >1 (Legendre & Legendre 1998). The four rotated PCs explained ~69% of the variance in PDSI during widespread fire years (Table 5). Factor 1 explained 15.4% of the variance in PDSI during widespread fire years and suggests weak drought conditions from the central US stretching eastward, as indicated by the negative loadings. Positive loadings in the region of the Pacific Northwest indicate generally wet conditions (Figure 8). The largest negative loadings were around the southwestern United States. It is interesting to note that none of the reconstructed fire years were strongly related to this component, with the possible exception of 1709, but that several years (1794, 1889, 1919) had strong inverse relationships suggesting that a reversal of this pattern (dry northwest, wet east and southwest) was important (Table 5). Dry conditions in the northwest in the widespread fire years of 1889 and 1919 were well documented in relation to fire activity during these years (Larsen & Delavan 1922; Larsen 1925).

Approximately 22% of the variance in PDSI during widespread fire years was explained by PC 2 and was characterized by negative loadings and dryness centered on the southwest and California, and the eastern seaboard. Relatively wet conditions were
Figure 8. Rotated factor loadings of widespread fire years in the SBW. Variance explained is shown in parentheses. Heavy dark lines differentiate positive loadings (wet conditions) from negative loadings (dry conditions).
Table 5. Results of PCA analysis of widespread fire years in the SBW. Note that only fire years since 1700 are considered because some grid points do not extend beyond that year.

<table>
<thead>
<tr>
<th></th>
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<th>Component 3</th>
<th>Component 4</th>
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<td>1719</td>
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<td>1934</td>
<td>-0.351</td>
<td>-0.068</td>
<td>0.796</td>
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<sup>a</sup>Variance explained by the varimax rotated PCs.
present in the central US indicated by positive loadings (Figure 8). Only a weak drought is suggested in the SBW region. Widespread fires in 1729, 1748, and 1883 appeared to match this pattern (compare Figures 7 & 8), and 1851 had a slight relationship.

PC 3 explained only about 13% of the variance and indicated drought conditions mostly restricted to the southwest and northern Great Plains (Figure 8). Wet conditions were present in the Pacific Northwest and eastern states. Widespread fires in 1864 and 1934 were most strongly related to spatial patterns captured by this component but reconstructed PDSI seemed to suggest drought was much more extensive during these years than is evidenced by the PC (Figure 7, Table 5).

Approximately 18% of the variance in PDSI during widespread fire years was explained by PC 4. PC 4 suggests drought conditions were centered around the four corners region of the southwest (Figure 8). Widespread fires in 1801 and 1896 seemed to be most strongly related to this pattern (Figure 7, Table 5).

**Discussion**

*Subalpine fire regime*

The use of multiple lines of evidence-including fire-scar and age-structure information-highlighted the heterogeneity of the fire history in upper elevation forests of the SBW. While the individual watersheds were relatively small compared to most studies that examine subalpine (or boreal) fire history, the amount of detailed information collected in the watersheds was large. About 2,000 crossdated samples were used to reconstruct the fire history of these four watersheds.
The prevailing fire regime paradigm in subalpine forests is that of infrequent stand-replacing fires occurring every 100-300 years (Arno 1980; Romme 1982; Barrett et al. 1991; Barrett & Arno 1991; Barrett 1994). Smaller, non-lethal fires, though often considered less important to the subalpine fire regime (Romme 1982), occurred more frequently, at intervals of about 10-40 years in other study areas of the Rocky Mountains (Barrett et al. 1991; Brown et al. 1994; Wadleigh & Jenkins 1996; Kipfmueller & Baker 2000). Although lightning ignitions in these forests are generally common, moist fuel conditions prevent fires from reaching large sizes during most years (Renkin & Despain 1992).

The fire regime of these watersheds was clearly mixed, with many local fires that appeared to have been limited in both size and severity, punctuated by relatively infrequent widespread fire events with a more extensive stand-replacing component. This is in general agreement with the notion of few large fires, but many small fires, suggested by most examinations of subalpine fire regimes.

Some widespread fires had evidence of mixed severity within the watersheds. For example, the 1889 fire in Hoodoo Lake had a stand-replacing component in some parts of the watershed, while age structure information from other areas indicated the fire burned as a low intensity surface fire. In contrast, there were few indications that the 1883 fire in Burnt Knob Lake was anything but stand-replacing in most of the watershed examined. Small pockets of trees survived this fire but they were mostly restricted to rocky areas around the lake, near stream courses, or along dry ridge tops, where limited fuel
continuity probably decreased the fire severity. With the exception of two 1883 fire-scar samples located near the lake most scars were near the probable boundary of fire.

In this study, MFI for fires occurring anywhere within a watershed ranged from about 20-170 years. These watersheds were relatively small by most standards but this result was comparable to MFI reported elsewhere in much larger subalpine study areas (Tande 1980; Kipfmueller & Baker 2000; Barrett 2000). Fire return intervals for smaller areas, typically summarized from data collected over larger regions, has been reported to be between about 90-300 years in subalpine forests of the Rocky Mountains (Arno 1980; Hawkes 1980; Romme 1982; Barrett et al. 1991; Barrett 1994). MFI determined based on the age structure and fire scar evidence from individual transects and grid points ranged between 139-341 years and is similar to other study areas (Arno 1980; Hawkes 1980; Romme 1982; Kipfmueller & Baker 2000).

The interval between widespread fires in these watersheds was between about 40-70 years in three watersheds, but was much longer in the Baker Lake watershed. The long interval between fires in Baker Lake was most likely due to the unique physical setting of Baker Lake rather than the result of missed fires. Baker Lake lies in a steeply dissected glacial trough. Forests in this watershed are discontinuous and separated by large rock outcrops and talus slopes that act as barriers to fire spread. More contiguous forests are found on ridge tops but fires igniting here would have difficulty spreading down slope into forests within the watershed. Beaver Lake and Burnt Knob Lake are also located in glaciated valleys with prominent headwalls but forests in these watersheds are more contiguous with forests adjoining on slopes and ridges than those in Baker Lake.
Most striking in the dendrochronological record is the high frequency of local fires even in these relatively small watersheds (Figures 2 & 4). With the exception of Baker Lake, the interval between all fires was between about 20-25 years, consistent with other examinations of fire history in subalpine forests in the region (Barrett 2000).

The high density of lightning in these forests can result in many fire ignitions that never reach large size because climate conditions are not conducive to spreading fire. Lightning is common in upper elevation forests of the SBW. Between 1989-1999 nearly 70,000 lightning strikes were recorded within the wilderness boundary (Rollins, unpublished data). Thus, there exists an abundant ignition source and fires can frequently start but may not spread to appreciable size unless climate conditions result in low fuel moisture. Renkin and Despain (1992) suggest fires can occur in subalpine forest habitats dominated by lodgepole pine at a wide range of fuel moisture conditions. However, fuel moisture conditions were found to limit the spread of fires when thousand-hour time lag fuel moisture was above about 13% (Renkin & Despain 1992).

Local fires were evident throughout the record in most of the watersheds but the dramatic reduction in fire occurrence after the early 1900s was obvious. The nearly complete cessation of fire since the late 1800s and early 1900s is evident in most western US fire histories (Hemstrom & Franklin 1982; Barrett 1988; Baisan & Swetnam 1990; Swetnam 1993; Kipfmuller & Baker 2000). The most recent fire recorded in any watershed examined in this study was 1934.

The study design may at least partly explain the decrease in fire occurrence in the 20th century. We selected watersheds for this study to maximize the temporal length of
the fire history record to improve our understanding of fire-climate relationships. We avoided watersheds that had been burned over by fires occurring in the 20th century, particularly 1910 and 1934 that would have likely eliminated much of the evidence of past fires. There may also be physical differences in these watersheds that are limiting to ignition by lightning for some unknown reason. However, these explanations seem unlikely since fires are common throughout most of the earlier portion of the record.

A more plausible explanation for reduced fire evidence since ca. 1934 is that fire suppression efforts have limited fire activity in these watersheds. Three of four study areas examined (Beaver Lake, Hoodoo Lake, and Burnt Knob Lake) have fire lookouts overlooking the watershed. These fire lookouts were all constructed and staffed since at least 1922 (National Historic Lookout Registry, http://www.firetower.org), with the wilderness guard station at Hoodoo Lake in operation by 1912 (Moore 1996). The Baker Lake study site is only about 5 km from a highway. Given the policy of rapidly suppressing fires in the SBW through most of the 20th century it seems likely that the lack of 20th century evidence of fires is due to suppression and not the result of diminished fire starts. Fires may have started during the 20th century but suppression kept them small.

Because of the complex, mixed severity of fires in this landscape, stand age/landscape scale approaches to fire history (Heinselman 1973; Van Wagner 1978; Johnson et al. 1990; Reed et al. 1997) is probably impractical. Further, this approach limits comparison with climate variability because it is often difficult to assign post-fire regeneration to a single year. Coupling age structure and fire scar information for the
reconstruction of fire history is an important approach for making comparisons of the relative importance of various factors controlling fire occurrence.

*Fire-climate interactions*

The long perspective provided by precisely dated tree-ring fire histories expands the temporal scale of analyses and permits comparisons of climate at interannual timescales. Moreover, the watersheds examined here likely have fire regimes that were unaffected by human impacts such as logging and fire suppression, at least in the early portion of the record. Comparisons of fire-climate interactions using data from the modern period may be biased by extremely large events because most smaller fires were likely suppressed even if climate conditions were generally conducive to fire spread.

*Drought*—A previous investigation of drought-fire linkages in the Interior Columbia River Basin of the Pacific Northwest found only limited relationships between drought and fire, mostly restricted to fire activity during 1860s and 1920s (Barrett et al. 1997). A major limitation of this comparison, however, is that fire events were examined based on five-year periods rather than using precisely dated fire years as was done here. Drought periods were identified by Barrett et al. (1997) using tree-ring reconstructed precipitation, but the details of this reconstruction were not provided. Further, the comparison used fire occurrence data from a variety of forest types that may obscure fire-climate interactions important to different forest cover types. For example, wet years several years prior to fire occurrence are well documented in dry ponderosa pine forests of the Southwest (Baisan & Swetnam 1990; Swetnam & Baisan 1996).
In the SBW, drought conditions appear to be quite important to the development of widespread fires. While this is logical and intuitive, the lagging relationship identified here was not present in the modern fire record, and this pattern has not been previously recognized in the northern Rocky Mountains. Fifteen of seventeen widespread fires had dry summers during the fire year and ten of seventeen widespread fire years had dry summers one year preceding fire occurrence (Table 4). While PDSI varied widely during these fire years the pattern was remarkably consistent. Further, reconstructed PDSI in widespread fire years was significantly different from reconstructed PDSI during local fire years (two tailed t-test, $t = 3.34, p < 0.01$), a finding also evident in the SEA results (Figure 6).

The two year lag pattern associated with widespread fires suggested that prolonged drying of fuels could be necessary to develop fuel conditions in upper elevation forests suitable for fire spread. The suggestion by Renkin and Despain (Renkin & Despain 1992) that 1000-hour time lag fuel moisture seems to limit fire spread in lodgepole pine stands may support this to some degree. However, moisture recovery during the winter months would likely offset any previous year drying of fuels, resetting the system in some regard. Anderson (1990) examined moisture recovery in fine fuels (e.g., conifer needles) and found the response time required for lodgepole pine needles to return to high humidity (90% RH) to be about 37 hours, much longer than the National Fire Danger Rating System (NFDRS) used in its models. McCammon (1976) found that by the time of emergence from snowpack, 3.8-7.6 cm dia. fuels had only 32% moisture content. It may be that drier than average summers preceding a dry fire year lead to
limited moisture recovery in forest fuels, but limited research is available that examines this effect.

The two year drought relationship may also simply reflect the occurrence of fires during drought periods extending for more than one year simply by chance. In other words, droughts may extend for multiple years and the probability that a fire occurs within the period is high. The examination of a larger spatial scale might better indicate the patterns of drought-fire relationships in these forests by providing a larger set of events to explore.

The pattern of two successive dry summers was remarkably consistent with respect to widespread fire years, but there were important deviations from this pattern. For example, widespread fire years in 1204, 1719, and 1910 all lacked drought relationships during the summer months (Figures 5 & 7, Table 4). Wet conditions were present in 1204 and in the preceding year (Table 4). A fire burning in the watershed may have grown to large proportions with the passage of a cold front that brought high winds even where conditions were not anomalously dry, possibly due to the passage of a cold front (Brotak & Reifsnyder 1977; Barrett et al. 1997). Drought conditions were also not present in 1719, but both 1717 and 1718 (PDSI= -3.60 and -3.15 respectively) were extremely dry.

The 1910 fire year, one of the most notable fire events in the western United States, represents an important caveat to longer-term climatic causes of large fires since PDSI suggested wet conditions. Winter conditions during the 1910 were near or above average in the region. Beals (1910) indicated snowfall between January and March was
above average and resulted in a number of large avalanches in the northern Rocky Mountains later in the winter, possibly coincident with rapidly warming temperatures. The beginning of March marked the onset of a seasonal drought, characterized by early snow melt, with the drought persisting through September (Donnel 1910). Fires burned throughout the dry summer of 1910, but most of the area burned did so over only about four days between August 19-21 (Beals 1914; Koch 1942). The unusually large areas burned have been attributed to high wind events that may have been associated with the passage of a cold front (Pyne 2001), as well as topographic effects that concentrated and increased these winds to drive fires to large size (Beals 1914). This highlights the importance of multiple (and sometimes interactive) factors that can lead to large fires at shorter time scales, those generally considered to be weather rather than climate.

The lack of a significant dry antecedent year in comparison of the modern fire record to instrumental PDSI is intriguing. One possible explanation for this is that area burned estimates from the modern period include all forest cover types. While lagging relationship between fire and drought conditions maybe especially important in subalpine forests, it may be less important in other cover types, such as lower elevation ponderosa pine or cedar-hemlock (Thuja plicata-Tsuga heterophylla) forests. The largest areas burned in the SBW during the 1910 fire year occurred in cedar- hemlock forests (Rollins 2000). Fire-climate interactions vary by forest cover type and this could obscure relationships to some extent.

In addition, the effects of fire suppression were likely appreciable in the modern record, particularly between about ≈1945-1970 (Rollins et al. 2001). Fires may have
started under ideal burning conditions but were quickly suppressed before spreading. However, due to the remoteness of the area, under the most extreme conditions suppression was mostly ineffective and many fires did burn larger areas. Although the SEA results showed drought conditions were significant ($p<0.05$) during the fire year examination of instrumental PDSI for the preceding years indicates drier conditions were present at least one year in advance in 8/10 of the largest modern fire years (Table 4).

*ENSO variability*—The absence of any link identified by SEA with El Niño did not come as a complete surprise for a number of reasons. First, teleconnections between climate patterns and ENSO were not as strong in this region as in others (Barry et al. 1981, Harrison & Larkin 1998). Harrison and Larkin suggest El Niño conditions can lead to dry conditions in the spring stretching from the Pacific Northwest eastward across Montana. However, these results were based on only ten El Niño events since 1946. Dry spring conditions could result in an early onset of drought conditions that may not be alleviated by usually sparse and local summer rains. Larsen and Delevan (1922) observed very dry spring conditions prior to the 1919 fire season; conditions that were exacerbated by an equally dry summer. El Niño conditions were in place during the winter of 1919 (Quinn & Neal 1992) that may have led to the dry spring conditions. The strength of the 1919 El Niño is surpassed by only six events since 1876 (Allan et al. 1996). It is plausible that El Niño conditions could lead to large fires if the following summer also remains dry. Moreover, the pattern of drought in 1919 seems to follow the classic pattern of a wet southwestern U.S. and dry northwest often associated with El Niño (Figure 7).
An additional problem investigating the relationship between ENSO events is that the SEA examines averaged conditions based on all the fire years. It is unlikely that all widespread fires would be related to ENSO conditions, so there is likely some mixing of climate-fire signals using this approach. This is particularly important since only small areas were inventoried for fire history. It might be expected that an ENSO teleconnection would synchronize fire events over a broad region. A much larger region of study with more long reconstructions is needed to reliably link ENSO relationships to fire occurrence in these subalpine systems.

*Spatial patterns*—The continental-scale drought patterns associated with widespread fire in the SBW also display a degree of spatial variability. Four PCs were identified with differing patterns of drought across the United States. For the most part these suggest dry conditions over broad regions prevailed during widespread fire years in the SBW. In three of the four PCs (PC 2-PC 4), drought conditions were present near the SBW, but usually the SBW was on the drought margins (Figure 8).

In contrast to PCs 2-4 that indicated drought conditions near the SBW, PC1 suggested conditions were wet. The spatial pattern illustrated by PC 1 is roughly similar to moisture patterns during La Niña years and it was interesting to note that several fire years (1794, 1889, and 1919) all have strong negative loadings on PC 1, but that only 1709 had a relatively strong positive loading (Figure 8, Table 5). The opposite pattern from PC 1 would result dry conditions in the Pacific Northwest while much of the remainder of the continental US would be relatively wet during those years that were inversely related to PC1 (Figure 8).
It seems reasonable that this pattern could be related to El Niño events and reconstructed SOI (Stahle et al. 1998) and the El Niño record of Quinn et al. (1992) suggests this could be the case for 1889 and 1919. Reconstructed values of SOI in 1889 and 1919 ranked as the 17th and 11th smallest values respectively. The instrumental SOI record of Allan et al. (1996) also supports an El Niño linkage. Instrumental SOI in 1889 (SOI = -11.72) and in 1919 (-13.42) were both strongly negative. During the period 1876-1996 only six years had SOI values less than that observed in 1919 while only nine years were less than observed SOI in 1889 (Allan et al. 1996).

A compositing of instrumental PDSI at the 154 grid points for the ten years with the smallest SOI index suggests that during El Niño years conditions were dry in the Pacific Northwest (Figure 9). The composite PDSI was nearly the exact opposite of PC 1, except for some minor differences around the Great Lakes region.

The identification of a potential El Niño connection to fire events in subalpine forests was hampered by a limited number of major fire events in the modern and paleo periods and a lack of data that differentiates fire size or the occurrence of fire events over a broader region. However, the possibility is intriguing and suggests a potential area for future research. In particular, a broader network of fire history reconstructions with annual resolution would be useful to identify regional fire patterns in the northern Rockies.

The spatial patterns of drought suggest there are multiple synoptic regimes that might lead to widespread fire years in the Pacific Northwest. While instructive, these patterns need to be evaluated in light of the particular climatologies associated with these
Figure 9. Composite average of summer PDSI at 154 grid points for ten El Niño years with the smallest SOI Index (PDSI data from Cook et al. 1999). Dashed contours signify dry conditions and solid contours wet conditions. The heavy black line separates regions of relative wetness from drier regions.
patterns using additional climate data beyond the scope of this manuscript. Analysis of climatic conditions could highlight important synoptic conditions leading to widespread fires in subalpine forests of the SBW region and may aid in fire management planning. However, the association of regional fire years with short-term phenomena (e.g., weather events, wind, multiple ignitions) such as 1910 highlight the need to consider short-term events as well.

**Conclusions**

Multiple lines of evidence and large data sets used in developing the fire histories of four subalpine watersheds indicated fires were relatively common. Most fires remained small, but widespread fires anywhere within the relatively small watersheds were more frequent than have been previously reported. The dramatic reduction in fires after about 1934 evident in all four watersheds was likely the result of effective fire suppression. While fires may have started during this period, the close proximity to fire lookouts and guard stations likely resulted in their effective suppression before reaching appreciable size.

Only PDSI had consistent relationships with widespread fire years in subalpine forests of the SBW. The year of fire occurrence was conspicuously dry but the preceding year was also dry in most cases. Spatial patterns of drought show that multiple climatic regimes can be associated with widespread fire years. While the SEA analyses did not identify a consistent relationship between ENSO and widespread fires in these study areas, El Niño conditions could be related to broader regional events.
Further research using precisely dated fire histories are needed across a larger area and in diverse forest types of the northern Rocky Mountains to better understand fire-climate interactions in this region.

Acknowledgements

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References


Koch, E. History of the 1910 forest fires in Idaho and western Montana. 1942. Unpublished manuscript on file, USDA For. Serv. Region 1, Missoula, MT.


APPENDIX D

STRUCTURAL AND COMPOSITIONAL CHARACTERISTICS IN RELATION TO TIME SINCE FIRE IN SUBALPINE FORESTS OF THE SELWAY-BITTERROOT WILDERNESS AREA, USA

Kurt F. Kipfmüller and John A. Kupfer

This paper will be submitted to *The Professional Geographer*

Abstract

The patterns of forest change following stand replacing fires are explored in four subalpine watersheds in the Selway-Bitterroot Wilderness Area, Idaho and Montana. Dendrochronological techniques were used to identify the timing of establishment of four major forest species at sites that had experienced different intervals since the last fire and forest structural and compositional traits. Lodgepole pine (*Pinus contorta* var. *latifolia* Dougl.) dominated most overstories for the first 100-200 years. As forests increased in age Engelmann spruce (*Picea engelmannii* Parry) and subalpine fir (*Abies lasiocarpa* (Hook.) Nutt.) became more prominent. The initial composition is dependent upon the time between fire events. Short fire intervals may limit the development of lodgepole pines capable of producing serotinous cones leading to young forests dominated by spruce or fir. However, intervals longer than the lifespan of lodgepole pine (as long as 350 years) may lead to early dominance by spruce or fir following fire. A conceptual model is presented highlighting the multiple successional pathways.
Introduction

The structure and composition of forest stands are the product of complex and interactive processes occurring at various scales of space and time and involving species adaptations, environmental conditions and disturbance history (Halpern 1988; Aplet et al. 1988; Huff 1995). Following a disturbance, life history traits and physiological adaptations such as seed dispersal, shade tolerance and germination requirements are important determinants of a forest’s initial composition (Noble and Slatyer 1980). As a stand ages, changing microenvironmental conditions influence the growth and establishment patterns of colonizing species, selecting for the establishment of some species over others (Donnegan and Rebertus 1999). Properties of the disturbance regime such as size and intensity exert important constraints on patterns of establishment within forested environments. Disturbance size and intensity can affect seed availability through the creation of large patches that exceed the dispersal distances of the propagules of some species (Turner et al. 1997), or the reduction of the seedbank due to high disturbance intensity (McCune and Allen 1985). Other factors such as the physical characteristics of available germination sites (aspect, slope, and elevation), climate variability (Villalba and Veblen 1997), and legacy effects (McCune and Allen 1985) can also have important consequences on species establishment and successional patterns.

In subalpine forests of the northern Rocky Mountains, fire is the most dominant disturbance influencing the structure and composition of forests (Peet 1988). The fire regime is typified by large (sometimes exceeding 10,000 ha) stand-replacing fires with intervals often greater than 100 years (Arno 1980, Kipfmueller and Swetnam,
unpublished manuscript). Large fire events are punctuated by smaller, relatively frequent fires, usually less than 2 ha in size, that may burn at low intensity creating smaller openings within the larger forest matrix (Renkin and Despain 1992). Both the large and small fire events have an important impact on forest patterns across the landscape (Farris et al., in review).

The spruce-fir (*Picea engelmannii* Parry-*Abies lasiocarpa* Hook. Nutt.) forest type is considered the end point of successional processes in subalpine forests of the northern Rocky Mountains in the absence of intervening disturbances (Larsen 1929; Pfister et al. 1977; Pfister and Arno 1980; Antos and Habeck 1981; Kessell and Fischer 1981; Arno et al. 1985; Donnegan and Rebertus 1999; Keane 2001). While a number of conceptual frameworks have been developed for successional processes in subalpine forests (Davis et al. 1980; Kessell and Fischer 1981; Arno et al. 1985) they are often based on assumptions concerning adjacent, lower elevation habitats or a heuristic understanding of successional dynamics in these forests. Keane (2001) presents a useful successional model for upper elevation habitats based on a synthesis of the limited available literature. This model, however, is grounded on research that often only indirectly focuses on successional dynamics but instead centers on fire regime characteristics. Capturing the variations in successional processes is difficult because old stands are rare on the landscape due to return intervals for disturbances that are often shorter than the time required for succession to proceed from beginning to end (Connell and Slatyer 1977).
The effects of nearly 50 years of fire suppression on subalpine habitats remain ambiguous. It is clear that fire suppression activities have reduced area burned and fire frequency in many subalpine habitats (Veblen et al. 1994; Barrett 1994; Kipfmueller and Baker 2000; Rollins et al. 2001). But because succession in these areas occurs over centuries, changes to overstory structure and composition of these forests may not be evident over such a short period (Habeck and Mutch 1973; Johnson et al. 2001). An understanding of the temporal development of subalpine communities is needed to refine our understanding of fire effects, and to define and implement long range management goals.

The main goal of this research was to identify demographic patterns, structure, and composition in subalpine forests in the northern Rocky Mountains. Fire histories and associated forest age structures were determined for forests experiencing different intervals since the last stand-replacing fire event. Species composition, age structure, and fire histories were developed in 23 sampled stands in four subalpine watersheds near the Selway-Bitterroot Wilderness Area. Nonmetric multidimensional scaling (Legendre and Legendre 1998) was used to ordinate sampled forest stands based on structural and compositional traits. Ordination results were compared with environmental variables including time since fire, topographic relative moisture index, slope, aspect, and elevation to assess the relative contributions of each to the character of subalpine forests in the area.
Study Area

The Selway-Bitterroot Wilderness Area (SBW) is located on the border of Montana and Idaho and encompasses approximately 542,000 ha (Figure 1). Elevation ranges from 463 m along the Selway River in Idaho to 3,048 m on Trapper Peak in the southeast portion of the wilderness. Subalpine forests contain lodgepole pine (Pinus contorta var. latifolia Dougl.), whitebark pine (Pinus albicaulis Engelm.) Engelmann spruce, and subalpine fir (Abies lasiocarpa (Hook) Nutt.). More extreme sites near upper treeline sometimes contain subalpine larch (Larix lyallii Parl.). Average temperatures around 2,100 m are generally cool (ca. 12°C in July and -8°C in January) (Finklin 1983). Approximately 170 cm of precipitation falls at upper elevations in the SBW, with about 70% of this as snow (Finklin 1983). Summertime precipitation is from convective storms, and precipitation from a single storm can be locally heavy. Summer drought is common throughout the SBW during July and August and is often associated with the occurrence of lightning storms with limited precipitation that ignites fires.

Four subalpine watersheds were selected for fire history data collection (Figure 1, Table 1). Watersheds were selected to include forests that were not completely burned during the 20th century, and had reasonable amounts of old-growth forest to facilitate the reconstruction of a multi-century fire history. An additional site selection criteria included the existence of a lake suitable for palynological sampling as part of a
Figure 1. The Selway-Bitterroot Wilderness Area on the border of Idaho and Montana. White circles are locations of the watersheds sampled for fire history and successional patterns.
Table 1. Locations and approximate sampling area of the four watersheds used for the fire history reconstruction and determination of successional patterns.

<table>
<thead>
<tr>
<th>Watershed Name</th>
<th>Abbreviation</th>
<th>Lat./Lon.</th>
<th>Elevation range (m)</th>
<th>Approximate area (ha)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beaver Lake</td>
<td>BVR</td>
<td>46.55°/114.28°</td>
<td>1753-2240</td>
<td>415</td>
</tr>
<tr>
<td>Hoodoo Lake</td>
<td>HOO</td>
<td>46.33°/114.53°</td>
<td>1688-2140</td>
<td>2005</td>
</tr>
<tr>
<td>Baker Lake</td>
<td>BKR</td>
<td>45.88°/114.25°</td>
<td>2050-2789</td>
<td>375</td>
</tr>
<tr>
<td>Burnt Knob Lake</td>
<td>BKL</td>
<td>45.72°/115.00°</td>
<td>1718-2490</td>
<td>438</td>
</tr>
</tbody>
</table>

companion study (reported in Brunelle-Daines 2002). Lake level elevations of the watersheds range from ≈1770 m at Hoodoo Lake to ≈2400 m at Baker Lake placing these habitats primarily in the lower to upper subalpine “potential vegetation habitat types” (PVT) as defined by Pfister et al. (1977). Subalpine fir is considered to be the climax vegetation in these areas with lodgepole pine serving as an important seral species.

**Methods**

*Data Collection and Processing*—Within each of the four watersheds, we identified stands with relatively homogenous structural and compositional characteristics. We then established nested belt transects that were at least 200 m away from the estimated edge of the stand. We collected information to characterize the age, species composition, and structure of the stand. The size of each transect varied depending on the structure of the forest stand. Transects of 50 m in length were used in stands that appeared to be even-aged and were composed of predominantly a single early colonizing species (e.g., young lodgepole pine stands). In forest stands exhibiting high complexity (i.e., a multi-storied canopy, and high species diversity), we used 100 m transects. The
width of each transect varied in a manner that ensured collection of trees from all species and size classes present within the stand. In general, transect size was set to collect age data from at least 50 trees from each stand. Because large-sized trees occur at lower densities than those of smaller size classes, we varied transect width to include more large sized trees that would otherwise be missed by a relatively narrow transect. In these “nested” belts only trees larger than the predominant size class (estimated in the field) were sampled.

Along each transect we recorded the X-coordinate along the transect, species, diameter at breast height (DBH), and canopy position (emergent, canopy, subcanopy, or understory) of all live trees >1.5 m in height. Increment cores were collected from each living tree, and the height of the sample on the tree bole was recorded. Trees were cored up to four times until either a sample near the pith was obtained or rot prevented the collection of the innermost rings.

Fire scars were collected within each watershed to develop an annually resolved fire history to aid in the determination of time since fire. Intensive searches for fire scars were carried out in and around each transect. In general, all fire-scarred trees from which suitable samples could be collected were sampled. However, US Forest Service concerns restricted sampling to areas that would not be detrimental to wilderness aesthetics (i.e., away from lakes and trails). Additional systematic searches around probable fire-scar refugia such as ridgetops and drainage bottoms were also conducted. Fire scars were also collected opportunistically when encountered. For each fire-scarred sampled tree we recorded species, DBH, and location within the watershed using a GPS. Using a
chainsaw (if outside the wilderness boundary) or crosscut saw, a partial wedge was extracted from living trees, snags, and logs containing visible fire scars (Arno and Sneck 1977). The sampling of snags and logs enabled us to extend the fire history of the area further back in time than would otherwise be possible using only living trees.

**Analysis**--The present ("static", c.f. Johnson et al. 1994) age structure of each stand was determined from living trees sampled within the transects. Increment cores from each tree were crossdated against master chronologies developed from climatically sensitive trees using standard dendrochronological procedures (Stokes and Smiley 1968). Age data was summarized by the estimated decade of establishment. In cases where a tree’s center was missed by the increment borer we used pith estimators composed of concentric circles representing different growth rates (Applequist 1958). Missed tree-centers (pith) and associated errors with estimates to a tree’s center and tree age to coring height limit the resolution of age structure data to the relatively coarse decadal time scale. Trees that could not be reliably estimated to center (e.g., more than 20 rings estimated to center), and trees that lacked curvature of the innermost rings (negating the estimation procedure, mostly as a result of rot near a tree’s center) were not included in the age structure analysis. In some cases understory subalpine fir or Engelmann spruce could not be reliably crossdated due to suppressed growth, limiting the identification of wide and narrow ring-width patterns. If annual rings were readily visible then simple ring counts were used to estimate their age.
Stem density and basal area was determined for each sampled stand in the following categories: 1) the forest stand as a whole; 2) individual tree species; and 3) overstory and understory trees on a per ha basis. Measures for the overstory class were determined using trees with canopy positions determined to be emergent or dominant while the understory class included subcanopy or understory trees. Species importance values (Bray and Curtis 1957) were determined for each sampled plot by calculating the relative basal area and relative stem density of each species: 1) in the stand as a whole, and 2) for the understory and overstory separately.

The time since the last stand replacing fire (the number of years that have passed since the last stand-replacing fire or a fire that resulted in the oldest regeneration cohort; TSF) was estimated based on the age structures collected in each stand along with fire scars located near each sample plot. Based on fire-scar and age structure evidence, each stand was assigned to one of four TSF classes: 1) 75-125 years; 2) 126-200 years; 3) 201-300 years; and 4) greater than 300 years (Tables 2 & 3). These four classes were used to generate descriptive statistics for comparison of forest composition and structural characteristics and are analogous to a chronosequence of stand development following a stand initiating fire.

In some cases, a regeneration cohort associated with a known fire event was evident but a stand also had a small number of trees that apparently survived the fire event. In these cases a qualitative assessment of the age structure patterns was made based on the number of surviving trees relative to the regeneration cohort. If no direct evidence of TSF was available for a stand (e.g., no fire scar date associated with a
Table 2. Time-since-fire categories (TSF) and their representative sample stands.

<table>
<thead>
<tr>
<th>TSF Category</th>
<th>Time-Since-Fire (years)</th>
<th>Representative Stands</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>75-125</td>
<td>BVR3, BVR4, BVR6, HOO1, HOO4, BKR1, BKL2, BKL4</td>
</tr>
<tr>
<td>2</td>
<td>126-200</td>
<td>HOO2, HOO3, HOO5, HOO7, HOO8</td>
</tr>
<tr>
<td>3</td>
<td>201-300</td>
<td>BVR1, BVR5</td>
</tr>
<tr>
<td>4</td>
<td>&gt;300</td>
<td>BVR2, HOO6, HOO9, BKR2, BKR3, BKL1, BKL3, BKL5</td>
</tr>
</tbody>
</table>

regeneration cohort) we assigned the TSF of the stand to the age of the oldest seral tree species (either lodgepole pine or whitebark pine). In three stands early colonizers were not present, and age structure evidence was indeterminate as to which fire year was most likely related to regeneration; TSF was then assigned based on the oldest trees in the stand.

Nonmetric multidimensional scaling (NMDS) was used to group stands based on their degree of compositional similarity. NMDS is an indirect gradient analysis technique that defines coordinates for observations in such a way that the actual distances between observations is preserved as much as possible (Faith et al. 1987; Legendre and Legendre 1998). In this case, sites with similar species compositions would lie close to one another in ordination space. NMDS assumes monotonicity rather than assuming any particular form of relationship with respect to ecological distance. This results in a comparatively robust ordination technique with respect to other methods (such as principal components) that assume linearity (Fasham 1977; Faith et al. 1987).
Table 3. Site characteristics of each sampled stand. Individual sampled stands are numbered below each listed watershed.

<table>
<thead>
<tr>
<th>Beaver Lake</th>
<th>Fire Yeara</th>
<th>TSF Class</th>
<th>Principal Tree Speciesb</th>
<th>Overstory</th>
<th>Understory</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>≈1760</td>
<td>3</td>
<td>Engelmann Spruce</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>≈1500</td>
<td>4</td>
<td>Engelmann Spruce</td>
<td>Subalpine Fir</td>
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</tr>
<tr>
<td>3</td>
<td>1919</td>
<td>1</td>
<td>Engelmann Spruce</td>
<td>Engelmann Spruce</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>1919</td>
<td>1</td>
<td>Lodgepole Pine</td>
<td>Lodgepole Pine</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>1748</td>
<td>3</td>
<td>Lodgepole Pine</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>1901</td>
<td>1</td>
<td>Subalpine Fir</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Hoodoo Lake</th>
<th>Fire Year</th>
<th>TSF Class</th>
<th>Principal Tree Speciesb</th>
<th>Overstory</th>
<th>Understory</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1889</td>
<td>1</td>
<td>Lodgepole Pine</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>1851</td>
<td>2</td>
<td>Lodgepole Pine</td>
<td>Lodgepole Pine</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>1851</td>
<td>2</td>
<td>Lodgepole Pine</td>
<td>Lodgepole Pine</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>1889</td>
<td>1</td>
<td>Lodgepole Pine</td>
<td>Lodgepole Pine</td>
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</tr>
<tr>
<td>5</td>
<td>1851</td>
<td>2</td>
<td>Lodgepole Pine</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>≈1650</td>
<td>4</td>
<td>Lodgepole Pine</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>1851</td>
<td>2</td>
<td>Subalpine Fir</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>1851</td>
<td>2</td>
<td>Lodgepole Pine</td>
<td>Lodgepole Pine</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>≈1680</td>
<td>4</td>
<td>Subalpine Fir</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Baker Lake</th>
<th>Fire Year</th>
<th>TSF Class</th>
<th>Principal Tree Speciesb</th>
<th>Overstory</th>
<th>Understory</th>
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</thead>
<tbody>
<tr>
<td>1</td>
<td>1896</td>
<td>1</td>
<td>Lodgepole Pine</td>
<td>Lodgepole Pine</td>
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<tr>
<td>2</td>
<td>1204</td>
<td>4</td>
<td>Engelmann Spruce</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>1204</td>
<td>4</td>
<td>Whitebark Pine</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Burnt Knob Lake</th>
<th>Fire Year</th>
<th>TSF Class</th>
<th>Principal Tree Speciesb</th>
<th>Overstory</th>
<th>Understory</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1580</td>
<td>4</td>
<td>Whitebark Pine</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>1883</td>
<td>1</td>
<td>Lodgepole Pine</td>
<td>Whitebark Pine</td>
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</tr>
<tr>
<td>3</td>
<td>1529</td>
<td>4</td>
<td>Subalpine Fir</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>1883</td>
<td>1</td>
<td>Lodgepole Pine</td>
<td>Whitebark Pine</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>≈1670</td>
<td>4</td>
<td>Subalpine Fir</td>
<td>Subalpine Fir</td>
<td></td>
</tr>
</tbody>
</table>

*aApproximate fire years are based on the age of the oldest trees in the sampled stand with the exception of probable survivors. bPrincipal species is based on the overstory and understory species with the highest importance value (see text for details).
NMDS was first applied to a matrix of distances that used the importance values of the four most common species for both the overstory and understory (Approach 1). Subsequent ordinations were then performed for just overstory and understory observations (Approach 2). Similarity matrices were calculated using the Bray-Curtis coefficient (Bray and Curtis 1957) using Systat 8.0 (SPSS 1998). In experimental studies the Bray-Curtis coefficient has been found to be relatively robust with respect to ecological distance (Faith et al. 1987). A two dimensional monotonic NMDS ordination was applied using Systat 8.0 and stress was measured using the Kruskal stress function for each method. The starting configurations of the NMDS were derived from an initial run using 50 iterations in two dimensions.

NMDS axes were interpreted with respect to species importance values. Pearson correlations between the axes from Approach 1 and environmental variables were calculated to identify relationships between forest characteristics and environmental variation. Environmental variables included elevation, slope, aspect, time since fire category, and a topographic moisture index (Parker 1982). Low values of TRMI indicate relatively dry conditions whereas higher values indicate relative wetness based on elevation, slope, aspect, topographic position, and slope curvature (Parker 1982). Aspect was transformed using a cosine transformation that scaled the values between 0 and 2 (Beers et al. 1966). Spearman’s rank correlation coefficient was calculated for the TSF category comparison.
Results

Summary of fire history results--A detailed description of the fire history of the four watersheds is presented in Kipfmueller and Swetnam (Appendix C). In summary, fires within the watersheds are relatively common but most remain small (Figure 2). Mean fire interval (MFI) was around 20 years for all but the Baker Lake watershed, where MFI was about 50 years. The MFI between stand-replacing fires within a watershed was shortest at Beaver and Hoodoo Lakes at around 45 years. The MFI between stand-replacing fires was ≈70 years at Burnt Knob Lake and ≈150 years at Baker Lake. These MFIs refer to the interval between fire events occurring anywhere within an individual watershed but do not reflect the interval at any given point within the watershed. MFI based on small stands within the watersheds varied from ≈140 years at Hoodoo Lake to ≈340 years at Baker Lake. MFI for small stands was 212 years at Beaver Lake and 235 years at Burnt Knob Lake.

Stand age and composition--Age structures for twenty-three sampled stands were generated from more than 1100 estimated germination dates (Figure 3). The age structures of younger forests (TSF 1) are mostly dominated by lodgepole pine that germinated shortly after the fire event (Figure 3). However, species colonization patterns are highly variable with some early colonization by all of the four principal species present in some stands. In many cases species colonize a site in a nearly synchronous manner. For example, BKL2 (Figure 3D) has high proportions of whitebark pine that appeared to establish rapidly following lodgepole pine and establishment persisted over
Figure 2. Composite fire chronology for four subalpine watersheds in the Selway-Bitterroot Wilderness Area. Each composite represents the fire history of a watershed developed from 13-37 fire-scarred trees. Solid vertical lines at the bottom represent a composite of all the detected widespread fires and dashed vertical lines depict smaller fires.
several decades. Lodgepole pine establishment in the same stand seemed to decline somewhat. In BKRI (Figure 3C), however, the only tree-sized species present was lodgepole pine even after almost 100 years of succession. In Beaver Lake, subalpine fir or Engelmann spruce established either without lodgepole pine completely (BVR3, Figure 3A) or it was present to only a minor extent (BVR6, Figure 3A). Other relatively young stands were intermediate between these patterns.

Lodgepole pine is the most prominent overstory species in the younger TSF classes (Figures 4 & 5). Stem density, basal area, and importance values are clearly dominated by lodgepole pine in the overstories of TSF 1 and 2 (Figures 4 & 5). Understory composition was variable in the stands that burned more recently (Figures 4 & 5). Understories with a strong whitebark pine component were present only at Burnt Knob Lake. One potentially complicating factor is that all of the TSF2 stands were found in the Hoodoo Lake watershed. These stands likely originated following a fire in 1851 but subsequently experienced a surface fire in 1889 (Figure 3B). In these stands there are two prominent peaks in lodgepole pine regeneration. The earliest peak likely is the result of stand initiation following the 1851 fire, while the second peak around 1890 is probably the result of the 1889 fire freeing up available resources by reducing understory density. In addition it is likely some serotinous cones were opened by the 1889 fire providing a ready source of lodgepole pine seeds to develop a second regeneration peak.
Figure 3. Age structures of 23 sampled forest stands in 10-year establishment classes. Age structures and diameter distributions are presented as relative frequencies. The time since fire category is indicated for each stand. The number of trees successfully dated for age structure is presented for each stand. The percentage indicates the percentage of the total samples obtained that were successfully dated. (A) Beaver Lake, (B) Hoodoo Lake, (C) Baker Lake, and (D) Burnt Knob Lake.
Figure 4. Stem density and basal area relationships with time-since-fire. The number of stands in each TSF category is shown in parentheses. Error bars are 1 standard deviation from the mean.
Figure 5. Stem density, basal area, and importance values for 23 sampled stands. Data is presented for the four principal species and overstory and understory trees. Species codes are as follows: PICO=lodgepole pine, PIAL=whitebark pine, ABLA=subalpine fir, and PIEN=Engelmann spruce. (A) Beaver Lake, (B) Hoodoo Lake, (C) Baker Lake, and (D) Burnt Knob Lake. The overstory is represented by black bars and gray bars depict the understory. TSF category is indicated for each stand.
As stand age progressed beyond about 200 years (TSF 3 & 4) whitebark pine, subalpine fir, and Engelmann spruce became more prevalent (Figure 3). The density of each of these species was variable depending on the individual site and is likely related to a variety of environmental factors as well as important disturbances such as mountain pine beetle and blister rust that affect whitebark pine. Stands in TSF 3 & 4 have overstories composed mostly of either subalpine fir, Engelmann spruce, or whitebark pine (Table 3, Figure 5). Lodgepole pine became essentially absent from both the overstory and understory in most stands beyond about 200 years of age. Two notable exceptions to the increasing dominance of spruce, fir, and whitebark pine at the expense of lodgepole pine are BVR5 (a ≈250 year old stand) and H006 (a ≈350 year old H006 stand). Both of these stands contained large numbers of old lodgepole pines in the overstory. Eight lodgepole pines in H006 had estimated establishment dates between 1647 and 1652 and are probably near the maximum age limit of the species.

As forests increase in age, total stem density initially appears to decline slightly before increasing in the oldest stands, though this pattern is not significant (Figure 6). A significant decrease in overstory density occurred between TSF 1 and TSF 4 stands (two-tailed t-test, t=3.314, p<0.01, 14 d.f.). Basal area remained similar for the different TSF classes (Figure 6). The significant decline in overstory density coupled with the general stability of overstory basal area reflects increases in the stem diameter of trees in the older TSF classes.

Understory density and basal area is largely dominated by subalpine fir for all TSF classes except TSF 2 (Figure 4). However, examination of individual stands
Figure 6. Mean values for structural variables for the four principal tree species related to time since fire. Error bars are 1 standard deviation from the mean.
indicated that the understories of most young stands were dominated by either lodgepole or whitebark pines but that high densities of subalpine fir in a few stands associated with overstories of spruce or fir density subalpine fir obscured this pattern (Figure 5).

*Nonmetric dimensional scaling*—The two NMDS approaches used here yielded similar results. Approach 1 (combined overstory and understory importance values) explained approximately 99% of the variance in two dimensions in the similarity data and Approach 2 (separate overstory and understory importance values) had an explained variance around 97% in two dimensions. Overall stress of the configurations was around 0.06 for Approach 1 and 0.09 using Approach 2.

In both NMDS approaches Axis 1 differentiated stands based on the importance values of lodgepole pine and subalpine fir, although Axis 1 using Approach 2 that considered the overstory and understory separately also captures the importance of Engelmann spruce to a limited extent (Table 4). Axis 2 primarily separated stands based on the importance values of Engelmann spruce and whitebark pine but Approach 2 also captured overstory lodgepole pine and understory subalpine fir (Table 4).

In general, the arrangement of the stands in species space reflected changes related to time since fire with stands composed of lodgepole pine in TSF 1 or 2 clustered close together along Axis 1 (Figure 7). However, BKL2, which contained high amounts of whitebark pine, and the two older lodgepole pine stands were somewhat distant from the main lodgepole pine grouping. Stands in TSF 3 or 4 generally were aligned along Axis 2 with only minor differences in Axis 1.
Table 4. Pearson correlations with NMDS axes and overstory and understory importance values. Correlations in italics are significant at $p<0.05$, those in bold are significant at $p<0.01$ (two-tailed test).

<table>
<thead>
<tr>
<th>Species Importance Values</th>
<th>Overstory/Understory Combined (Approach 1)</th>
<th>Overstory/Understory Separate (Approach 2)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Axis 1</td>
<td>Axis 2</td>
</tr>
<tr>
<td>Overstory</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lodgepole Pine</td>
<td>-0.962</td>
<td>-0.013</td>
</tr>
<tr>
<td>Whitebark Pine</td>
<td>0.325</td>
<td>-0.615</td>
</tr>
<tr>
<td>Subalpine Fir</td>
<td>0.814</td>
<td>-0.147</td>
</tr>
<tr>
<td>Engelmann Spruce</td>
<td>0.599</td>
<td>0.642</td>
</tr>
<tr>
<td>Understory</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lodgepole Pine</td>
<td>-0.903</td>
<td>0.187</td>
</tr>
<tr>
<td>Whitebark Pine</td>
<td>-0.314</td>
<td>-0.509</td>
</tr>
<tr>
<td>Subalpine Fir</td>
<td>0.863</td>
<td>-0.190</td>
</tr>
<tr>
<td>Engelmann Spruce</td>
<td>0.314</td>
<td>0.695</td>
</tr>
</tbody>
</table>

Groupings of overstories with similar species were generally more compact using Approach 2 than when considering overstory and understory importance values (Figure 8). The ordination of understories also resulted in compact groupings for stands with understories composed primarily of lodgepole pine and subalpine fir. In addition, with the exception of most relatively young stands (TSF 1 & 2), understories were very similar and were mostly composed of subalpine fir. Most young stands of lodgepole pine had overstories and understories that were dominated by lodgepole pine, but the two oldest lodgepole pine stands (H006 and BVR5) had understories of subalpine fir (Figure 8).

Changes in both direction and magnitude between the understory and overstory suggested that broad differences in TSF resulted in similar directions of change (Figure 9). For example, understory to overstory shifts in TSF 1 and 2 were in roughly the same
Figure 7. Two-dimensional ordination of overstory and understory importance values using nonmetric dimensional scaling (see text for details). Circles indicate axis scores for individual stands. Shading and shapes denote the TSF category for each stand. Stands are grouped based on the dominant overstory species based on importance values.
Figure 8. Two dimensional NMDS ordination of overstory and understory importance values (Approach 2 see text for details). Circles indicate the axis scores for the overstory and squares the understory. Shading indicates TSF categories for individual stands. The line connects the overstory axes scores for a stand to its understory scores. Each stand is labeled at its overstory symbol. Solid lines enclose overstories dominated by the individual species based on importance values and dashed lines delimit understories (indicated by four letter codes, see Figure 5 caption for explanation).
Figure 9. Angle (degrees) and distance (ordination space) of difference from the understory axes scores to the overstory scores for each sampled stand. The time since fire category is indicated for each stand. Angles and distances were calculated from the results of NMDS Method 2 (see text for details).
direction. Stands in the older age-classes behave similarly to one another but are offset almost 90° from the ≈100-year old stands (Figure 9), primarily due to convergence of the understories that are mostly dominated by subalpine fir (Figure 8). The actual magnitude of change between TSF categories is significant only between TSF 1-3, and 2-3 (two-tailed t-test, \( p<0.05 \)).

Comparison of the axis scores from NMDS Approach 1 to environmental variables showed that TSF category was the only variable with a significant correlation (Table 5). Multiple regression equations between axis scores and the environmental variables also did not reveal any significant relationships (regression coefficients \( p>0.05 \)).

<table>
<thead>
<tr>
<th>Variable</th>
<th>Axis 1</th>
<th>Axis 2</th>
</tr>
</thead>
<tbody>
<tr>
<td>TRMI</td>
<td>0.335</td>
<td>0.150</td>
</tr>
<tr>
<td>Elevation</td>
<td>0.151</td>
<td>-0.377</td>
</tr>
<tr>
<td>Slope</td>
<td>0.058</td>
<td>-0.009</td>
</tr>
<tr>
<td>Aspect (transformed)</td>
<td>0.126</td>
<td>0.039</td>
</tr>
<tr>
<td>TSF Category (^a)</td>
<td>0.616</td>
<td>-0.113</td>
</tr>
</tbody>
</table>

\(^a\)The association between TSF and the axes was calculated using Spearman’s rank correlation coefficient.
Discussion

**Successional sequence**—In the upper elevation forests of the northern Rocky Mountains successional change proceeded slowly. The general patterns observed in these forests suggested that lodgepole pine dominated most early stands but that other species could colonize a site in a nearly synchronous manner (Figure 3). Stands of ≈100-150 years old likely originated following stand-replacing fires in lodgepole pine stands that contained appreciable amounts of serotinous cones. The serotinous characteristic of some lodgepole pines enable seed to be stored in closed cones until fires open the cone scales to release the seed (Lotan and Critchfield 1990). This provides abundant seed for early establishment of lodgepole pine and a competitive advantage over species that need to disperse from outside the burned areas to a site (Lotan et al. 1985; Lotan and Critchfield 1990; Turner et al. 1997). The lags observed in establishment by spruce and fir that largely rely on wind dispersal may be at least partially related to distance to an available seed source (Alexander et al. 1984; Alexander and Shearer 1984). However, site conditions in early post-fire environments may be too harsh for successful establishment (Noble and Alexander 1977; Cui and Smith 1991).

The role of whitebark pine as an early colonizer in the relatively young forests remains somewhat uncertain. In some environments whitebark pine is known to be an early seral species that colonizes a site soon after a fire event following fire (Keane et al. 1990; Tombback et al. 1990; Tombback et al. 1993; Keane 2001). Given that whitebark pine possess a bird dispersed seed that preferentially caches seeds in burned areas, it seems likely that whitebark pine could be among the first colonizers of burned sites.
Tomback et al. (2001) suggest successful germination of whitebark pine seedlings may be delayed until environmental conditions permit their emergence, particularly with respect to soil moisture. In post-fire stands conditions can be harsh due to the higher incidence of solar radiation and limited plant cover (Larsen 1925; Stahelin 1943; Noble and Alexander 1977; Cui and Smith 1991). The lag between whitebark pine establishment and fire events that was evident in several sampled stands (Figure 3) is more likely related to delayed germination of the seeds rather than delayed arrival at a site. If germination is delayed, other species that arrive earlier may be able to establish and out compete whitebark pine seedlings for resources.

Whitebark pine, though present in only a few young stands, often occurred at relatively high densities, but established slightly later than the lodgepole pine with which it coexisted (Figures 3 & 4). However, the presence of whitebark pine in young stands was most notable in the Burnt Knob Lake watershed where living whitebark pines were abundant on ridges surrounding the site and provided an abundant seed source for Clark's nutcrackers.

The ubiquitous presence of subalpine fir through most of the early (and later) stages of succession results from its germination and light requirements (Alexander et al. 1984). Lodgepole pine is intolerant of shading whereas subalpine fir and Engelmann spruce tolerate shaded conditions. This allows both spruce and fir to persist in the understory of relatively young stands with closed canopies as is typical of the ≈100-year old lodgepole pine forests sampled here.
Stem densities are high in both the overstory and understory in forests around 100 years of age and are often dominated by either lodgepole or whitebark pine (Figures 4 & 6). As stands increased in age, Engelmann spruce and subalpine fir gradually reached an overstory position, with subalpine fir dominating most mid- to late-successional stands (Figures 4 & 5). The shift from lodgepole pine dominance of the canopy to that of spruce-fir or whitebark pine appeared to generally take more than 200 years in these sites. Understories of older stands were largely composed subalpine fir with varying amounts of other species, most notably spruce (Figures 4 & 5). The ordination results suggested the final sequence of succession can result in canopy dominance by either Engelmann spruce, subalpine fir, or whitebark pine (Figures 7 & 8). No environmental variables were significantly related to axis 2, which generally segregates the older forests. However, examination of the TRMI values of the few old forests suggests wetter sites are dominated by spruce (avg. TRMI=38, n=3) and drier sites by whitebark pine (avg. 30, n=2) with subalpine fir being somewhat intermediate (avg. TRMI=32, n=5). Verification of this relationship would require many additional sites and these results should be interpreted with caution.

While generalizations can be made regarding successional patterns in these forests there were important exceptions that suggested multiple pathways of succession were important. Lodgepole pine was clearly an early colonizer indicated by its strong presence after about 100 years. The serotinous nature of lodgepole pine, though a trait that is variable across space (Lotan and Critchfield 1990; Tinker et al. 1994), can ensure early colonization of recently burned areas (Turner et al. 1997; Turner et al. 1999). In order for
lodgepole pine to attain dominance early in the successional sequence several conditions must be met. First, lodgepole pine with serotinous cones must have been present within a site prior to fire occurrence to provide a reliable seed source for post-fire regeneration. This suggests that fires in the oldest forests, where mature lodgepole pine was often absent from a stand would likely result in early post-fire recovery that is dominated by those species arriving at a site first and can establish a competitive position with respect to resource use.

A second limitation on lodgepole pine dominance is that the fire return interval can not be shorter than the time required for lodgepole pines to produce seed. Under ideal conditions lodgepole pines have been observed to produce seed at a very young age, between 4-8 years (Lotan and Critchfield 1990). However, most of the areas sampled are near the upper limits of lodgepole pine so the age-to-seed production could be considerably longer. The recurrence of fire in a stand with young lodgepole pine that have not yet begun producing seed may result in forest recovery due to seed dispersal rather than the direct seed rain from the burned canopy associated with serotinous lodgepole pines.

Where these two conditions are not met, regeneration following fire may skip the lodgepole pine stage and instead stand initiation could ensue with dominance from other species (Cattelino et al. 1979; Noble and Slatyer 1980). This suggests that there are at least two successional pathways in these subalpine forests, one that begins with stands dominated by subalpine fir or Engelmann spruce and a second that begins with lodgepole
pine. A third pathway that originates with whitebark pine is also possible but was not observed.

Two ≈100 year old stands, BVR3 and BVR6, possibly resulted due to successive fires occurring at intervals shorter than the time required for lodgepole pines to produce seed. Stand initiation at BVR3 followed a fire in 1919 that is well documented by age structure and fire history data and also mapped in the area’s fire atlas (Rollins et al. 2001). However, BVR3 is composed largely of young Engelmann spruce and subalpine fir rather than lodgepole pine (Figures 3 & 5). A dead, fire-scarred whitebark pine within 500 m of the stand yielded an additional fire date of 1901. While the exact composition of stand conditions prior to the fire event is impossible to determine, the lack of snags and logs within the stand suggested that most of the trees were relatively young and possibly consumed (or since decomposed) when the 1919 fire occurred. It seems more likely that two successive fires, the first in 1901 followed by a second in 1919, burned the area at an interval shorter than the time required for lodgepole pines to produce viable. This would lead to the regeneration patterns observed in BVR3 resulting from dispersal that favors subalpine fir or Engelmann spruce (Table 6).

Similarly, BVR6 regenerated following the fire 1901 but dead fire-scarred whitebark pine within the stand (killed by mountain pine beetle ≈1930) recorded fires in both 1864 and 1901. While this regenerating stand does have some lodgepole pine, subalpine fir and Engelmann spruce dominate the stand’s overstory and understory. Lotan et al. (1985) suggests lodgepole pine is capable of producing seed at a very young age but that seed production peaks at 40 years. The low amounts of lodgepole pine
present in the 1901 regeneration cohort suggested seed production may have been limited, allowing subalpine fir and Engelmann spruce to establish early in the successional sequence.

Stand-replacing fires in forests lacking mature lodgepole pine would also lead to early regeneration favoring spruce and fir. In H007 for example, there was no evidence of a previous lodgepole pine stand, yet the age structure clearly suggests this stand regenerated ≈150 years ago and likely resulted from the 1851 fire that initiated extensive lodgepole pine forests throughout the watershed. Moreover, there was no evidence in this stand that fires burned at short intervals and no fire scars were found within 300 m of the stand boundary. There were a few lodgepole pine in the stand and most were younger than the spruce and fire that occupied the canopy (Figure 3). This stand is located near a riparian area and the generally wetter conditions as well as a large meadow nearby may have provided some protection from the subsequent 1889 fire that that is adjacent to the stand.

The principal effect of early colonization by spruce and fir appeared to be a more rapid progression to a forest with “old-growth” characteristics than those that followed the lodgepole pine successional pathway (McCune and Allen 1985). When lodgepole pine is absent from a stand the normally suppressed subalpine fir and Engelmann spruce can grow more quickly. In H007 for example, ≈150 year old spruce and fir were commonly greater than 50 cm in diameter, and sometimes as large as 70 cm. In contrast subalpine fir that were growing with lodgepole pine in H006, a ≈350 year old lodgepole pine stand experiencing canopy senescence, rarely exceeded 20 cm even though some
were more than 200 years old. In addition, the ordination results suggested the structure and composition of H007 were more closely associated with forests around 500 years of age (Figures 7 & 8).

Periodic fires maintain lodgepole pine forests, likely at a landscape scale, so long as the return interval is neither too short or too long. If fire returns to a stand before lodgepole pines are able to reach sexual maturity, regeneration will largely be from dispersed seed. This may favor the establishment of spruce and fir since their seeds are dispersed over a greater distance than lodgepole pine, though dispersal information for these species is scarce (Table 6).

Fire return intervals that exceed the life span of lodgepole pine would similarly result in forests that experience early colonization by spruce, fir, or whitebark pine but where regeneration of lodgepole pine is essentially absent due to a scarcity of seed (McCune and Allen 1985). The fact that lodgepole pine is present (indeed dominant) at the early stages of most stands of 100-200 years old suggests that the forests giving rise to these also contained considerable amounts of mature lodgepole pine. This suggests that fire-free intervals long enough to permit forests to succeed to old growth spruce-fir are rare and may be restricted to particular site environments such as sheltered drainages or particular slope aspects. Consideration should be given to landscapes where lodgepole pine is maintained as a “fire climax” rather than considering the vegetation that could potentially grow at these sites since it appears that at least some of these areas may never reach the end state of succession before a fire returns.
Table 6. Summary of life history attributes for four principal species based on a model of succession (Noble and Slatyer 1980). Relevant citations for each life history trait are referenced by number and are listed at the bottom of the table in the table. Age to seed production represents the age at which seed production is reported to be reliable. Longevity is based on the oldest individual of a given species identified in this study.

<table>
<thead>
<tr>
<th>Dispersal Agent</th>
<th>Lodgepole Pine (PICO)</th>
<th>Whitebark Pine (PIAL)</th>
<th>Subalpine Fir (ABLA)</th>
<th>Engelmann Spruce (PIEN)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Canopy storage/Wind</td>
<td>Bird</td>
<td>Wind</td>
<td>Wind</td>
</tr>
<tr>
<td>Dispersal Distance</td>
<td>60 m&lt;sup&gt;(1)&lt;/sup&gt;</td>
<td>8-12 km&lt;sup&gt;(2)&lt;/sup&gt;</td>
<td>15 m&lt;sup&gt;(3)&lt;/sup&gt;</td>
<td>100-200 m&lt;sup&gt;(4)&lt;/sup&gt;</td>
</tr>
<tr>
<td>Shade Tolerance</td>
<td>Intolerant</td>
<td>Moderately Inolerant</td>
<td>Tolerant</td>
<td>Tolerant</td>
</tr>
<tr>
<td>Noble and Slatyer (1980) Species Code</td>
<td>CI/DI</td>
<td>DI</td>
<td>DT</td>
<td>DT</td>
</tr>
<tr>
<td>Age to Maximum Seed Production (years)</td>
<td>20-30(CI)&lt;sup&gt;(1)&lt;/sup&gt;</td>
<td>60-100&lt;sup&gt;(6)&lt;/sup&gt;</td>
<td>100-150&lt;sup&gt;(7)&lt;/sup&gt;</td>
<td>150-200&lt;sup&gt;(4)&lt;/sup&gt;</td>
</tr>
<tr>
<td>Maxim. Longevity (years)</td>
<td>352</td>
<td>544</td>
<td>468</td>
<td>655</td>
</tr>
</tbody>
</table>


1 Lodgepole pine is depicted as both storing seed in the canopy due to cone serotiny (CI) and as intolerant dispersed species (DI) because non-serotinous cone bearing trees were also frequently observed.

The existence of these two different starting points for post-fire forest recovery and the subsequent changes in compositional characteristics through time is highlighted graphically in Figure 10. This conceptual model builds on earlier work by Noble and Slatyer (1980) that outlined a framework for describing succession in plant communities subject to disturbance. Their model is based on the delineation of only a few "vital" attributes for the pool of species available to colonize an area. In their scheme they define three key life history attributes that can be used to explain species change in a
given environment: 1) method of arrival and persistence at a site (dispersal characteristics); 2) establishment characteristics (tolerant versus intolerant), and 3) the timing of critical life stages (maturity, longevity, and propagule extinction). The application of these few vital traits have become an important part of developing successional models in many landscapes including those of the northern Rocky Mountains (Cattelino et al. 1979; Davis et al. 1980; Kessell and Fischer 1981; Crane and Fischer 1986; Fischer and Bradley 1987; Smith and Fischer 1997; Keane 2001).

Many of the key differences in the life history traits of the four most common subalpine species have already been mentioned, such as the serotiny of lodgepole pine, dispersal of whitebark pine seeds by the Clark’s nutcracker, and the relative shade tolerance of different species (Table 6). In addition, longevity, or in this case the maximum age of a species observed in this study, also represents and important aspect of life history especially with respect to lodgepole pine because its shade intolerant seeds are poorly dispersed and generally will not germinate well beneath the canopy of older forests (Lotan and Critchfield 1990). In this case, longevity is crucial because lodgepole pine seeds will only be readily available within a stand if living, mature lodgepole pine are present. The oldest lodgepole pine sampled in this study was ≈350 years old and suggests that without fire lodgepole pine will be functionally extinct from a stand beyond this point (Figure 10).

The upper portion of the figure represents the critical life history stages in terms of the time required to reach maturation and seed production. Species denoted by “DT” rely on dispersal to a site and are tolerant of shaded conditions whereas “DI” lodgepole
Figure 10. Conceptual model of stand development based on the age structures and disturbance history of sampled stands and the life history characteristics presented by Noble and Slatyer (1980) and Cattelino et al. (1979). The upper figure (A) shows the relevant life history traits for each of the four common species along a timescale. A lower case “p” denotes the presence of propagules at a site, “m” time to maturity to seed producing age, “l” the upper age limit of a species, and “e” functional extinction from a site due to a loss of trees capable of producing seeds. The lower figure (B) is a flow chart showing stand development with increasing time since fire. Black lines in the flow chart indicate progression with time while gray lines denote resetting a stand due to fire occurrence. Examples of each stage are indicated by the stand numbers. In parentheses next to each stand are listed the dominant overstory and understory species in each stand.
Lodgepole Pine (PICO) CI:
Lodgepole Pine (PICO) DI:
Whitebark Pine (PIAL) DT:
Subalpine Fir (ABLA) DT:
Engelmann Spruce (PEN) DT:

Stand Initiation
Post-fire colonization dependent on seed dispersal from PICO, PIAL, ABLA, & PEN

Serotinous lodgepole present to colonize post-fire stand plus dispersal of other species Lodgepole Pine Pathway

TFS Categories 1 & 2

Stand-replacing fire

TSFS Categories 3 & 4

Stand-replacing fire

Stand-replacing fire

Stand-replacing fire

Stand-replacing fire
pine denotes the possibility of dispersed lodgepole pine propagules but also indicates that it is intolerant of shade. Lodgepole pine denoted by “CI” represents serotinous lodgepole pines. Seeds are present at the earliest stage of succession for all four species due to dispersal from outside the disturbed area. The age to maturation, defined as the age at which viable seeds are produced is depicted by an “m” with a “p” indicating propagules present during the sequence. The age that propagule production or migration ceases is denoted by an “e” and represents extinction from a site. This likely only occurs with respect to lodgepole pine with potential reproduction from serotinous cones (CI) since it is limited by its longevity, but species that are dispersed to a site are never extinct from the seed pool (Cattelino et al. 1979). The longevity of a species is denoted by “l”. The solid horizontal line for each species represents the presence of propagules throughout the sequence.

The bottom graph serves as a flow chart through the temporal sequence of recovery from fire. The two potential pathways observed in the field are shown as parallel horizontal lines. Stands that conform to different periods within the sequence are listed along with their dominant overstory and understory species in parentheses after their identification number. Black arrows trace the sequence of stand regeneration and gray arrows depict the hypothetical return to either of the two initial conditions following a stand-replacing fire, depending on forest characteristics at the time of the fire.

The figure depicts the two pathways as merging together at the later stages of succession in the absence of fire (Figure 10). As lodgepole pine senesces and opens gaps in the overstory, understory species will replace them in the canopy. The timing of
canopy break up in lodgepole pine forests appears to be somewhat long. Two stands, BVR5 (≈250 years) and H006 (≈350 years), contain old lodgepole pine in the canopy with appreciable amounts of subalpine fir in the understory (Figure 5A and B). Most other stands of lodgepole pine (≈100-150) continue to have understories with abundant lodgepole pine or whitebark pine, suggesting canopy closure occurred relatively recently because these species have not been shaded out. Comparison of the understory importance values with overstory values indicated that the largest changes occurred primarily in these oldest lodgepole pine stands (Figure 9). This suggests that the transition from lodgepole pine to subalpine fir probably occurred after stands are at least 200 years old. Since understories of these mid to late sequence stand are dominated by subalpine fir (Figure 4) it seems reasonable that these pathways will converge from a composition stand point with stands following the spruce-fir pathway. Beyond this point most stands were relatively similar to one another suggesting little further change with time. In addition, the point at which overstory lodgepole pine likely begins disappearing from the canopy is shown as a gray box to represent the transition period.

The role of whitebark pine in subalpine forests—Whitebark pine has declined dramatically across half of its range due to an introduced pathogen, changes in fire regimes, and advancing succession (Keane and Arno 1993). Whitebark pine populations have declined by as much as 42% in the past 20 years in parts of the northern Rocky Mountains (Keane and Arno 1993). The area dominated by whitebark pine has declined by more than 45% in the US portion of the Interior Columbia Basin (Keane 2001).
Whitebark pine has declined more than 98% on sites where it is seral to other conifers (Keane et al. 1996). Twentieth-century die-offs of whitebark pine are evident today as “ghost forests” of standing dead trees in broad swaths across the highest elevations in Idaho and Montana. Many trees were likely killed by white pine blister rust (*Cronartium ribicola*) (Lachmund 1926; Hoff and Hagle 1990) and mountain pine beetle (Bartos and Gibson 1990). The blister rust is a Eurasian fungus that was introduced to the Pacific Northwest in 1910 (Hoff and Hagle 1990), whereas the mountain pine beetle is a native forest insect.

Living whitebark pine is rare in most stands although a few scattered individuals are present and in some old forests it serves as the dominant species in the overstory. However, many whitebark pine snags were present in old forest stands. Mountain pine beetle outbreaks have been documented in the 1930s in Baker, Beaver, and Burnt Knob Lakes and also in the 1960s in Beaver Lake and 1980s in Hoodoo and Burnt Knob Lakes. Mortality could have also been exacerbated by blister rust. In any case, it seems likely that the overstory composition of older forests of the region have changed as a result of the reduction in whitebark pine. It is also possible that the high density of small-sized subalpine fir in these stands could be related to overstory mortality improving establishment of understory trees. This appears possible in BVR2, H009, BKR3, and BKL1 & 5 all of which contained dead overstory whitebark pine killed during the 1930s outbreak (Kipfmueller et al. 2002).

Periodic fires are thought to help maintain whitebark pine habitats by reducing its fire-sensitive competitors, principally subalpine fir. In the whitebark pine stands we
examined we found no evidence of repeated surface fires that would have maintained whitebark pine dominance. The reduction in whitebark pine dominance observed here was probably most likely the result of blister rust and mountain pine beetles that killed overstory trees. It seems possible that competition from associated conifers could reduce the resistance of whitebark pine to these pathogens. Further, regeneration of young whitebark pines in these stands would also be affected by blister rust because their low stature allows them to capture spores to a higher degree than adult trees (Tomback et al. 1995). However, replacement of whitebark pine by subalpine fir does not appear to be related to reduced fire activity in these stands since fire incidence appeared to be low in these areas already.

Conclusions

Subalpine forests in the SBW appear to progress through early colonization and dominance by lodgepole pine toward forests composed primarily of subalpine fir. Two successional pathways were identified that have important ramifications for the rate of successional processes. These pathways are largely determined by characteristics of the fire regime related to the length of fire-free periods and the ability of lodgepole pine to reach seed bearing age. If fires recur before lodgepole pine reaches seed bearing age, early succession will be dominated by those species easily dispersed to a site. Similarly, if lodgepole pine is absent from an old stand, after perhaps ≈300 years, recovery will follow the spruce-fir pathway.
Comparison of the similarity of the overstory and understory of sampled stands revealed that both the youngest and oldest forests had characteristics overstories with species similar to those in the understory. The large differences in overstory/understory similarity appeared most pronounced in lodgepole pine stands >200 years in age and probably represents the transition and canopy break up stage related to lodgepole pine longevity.

Whitebark pine was a commonly occurring early colonizer in many ≈100 year old stands but was nearly absent in middle to late successional stands. The effects of mountain pine beetle and white pine blister rust may result in a shift toward stands more frequently dominated by subalpine fir. Whitebark pine was dominant in a few of the oldest sampled stands but does not appear to have been maintained by periodic surface fires that would have reduced competition by spruce or fir.

Acknowledgements

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References


