DISTURBANCE CLIMATE EVENTS IN THE COLUMBIA RIVER BASIN

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INTRODUCTION

Climate patterns that disrupt ecosystem processes are common and often critical components in the natural cycle of events. Although the Columbia River Basin is not known for extreme climate events (like hurricanes, tornados, intense lightning, or severe drought) many climatic features that are ordinary to the Basin (like summer cold fronts fanning wildfires or rain-on-snow floods) cause significant disruption of ecosystem processes. This report summarizes several aspects of climate that affect disturbance patterns in the Columbia River Basin.

FRONTAL PASSAGE AND THE SPREAD OF WILDFIRE

During the 1988 fire season, it was observed that the frequency of dry cold fronts passing over the northern Rockies in August and early September caused ongoing wildfires to make major runs (National Weather Service 1988). This was the most severe fire season in the Basin since the turn of the century. Therefore, an understanding of the pattern and frequency of cold frontal passage during summer, which could contribute to severe fire spread, may help anticipate future extreme fire seasons.

Weather fronts often are associated with strong, gusty winds. This especially is true of cold fronts because cold air is heavier and denser than warm air. Approaching cold air displaces warm air by moving underneath, causing strong vertical lifting that promotes turbulence. The frontal boundary often marks a sharp contrast between air masses. Distinct pressure patterns between the two air masses cause a rapid change in air pressure as the front progresses. In addition, gradients in pressure associated with storms often are strongest near the cold frontal boundary that coincides with a surface trough of low pressure. The changes in pressure, increased gradients in pressure, and turbulence all combine to cause strong, gusty winds around cold fronts. Warm fronts are much less dramatic because approaching warm air, being more buoyant, can ride up over more stable cold air causing a gradually sloping frontal boundary. Winds may increase slightly and shift direction during the passage of a warm front, but changes often are unremarkable.

Identifying a front with available weather observations requires careful analysis of surface and upper-level patterns of pressure, temperature, and humidity. Unfortunately, there are no precise formulas because there are an infinite number of patterns in all atmospheric variables that define fronts. To simplify the problem, however, only significant changes in atmospheric pressure measured at the surface were considered. Because pressure changes are well correlated to strong winds, this simple analysis proved adequate for interpreting a simple pattern of cold frontal passage.

Surface pressure observations from the National Weather Service synoptic observation sites were analyzed to determine the number of times strong fronts passed the area. Pressure differences greater than 4 mb over the 12 hour observation period were considered significant enough to cause a period of strong, gusty winds. The value of 4 mb was somewhat arbitrarily chosen, based on forecaster experience. Only data from June 1 to August 31, the primary fire season, were analyzed.

Figure 1 shows the number of events occurring with pressure changes greater than 4 mb during a 12 hour period at Spokane, Washington. Highlighted on the plot are 3 years that were chosen as characteristic climates in the Columbia River Basin (1982 -- cool, wet; 1988 -- warm, dry; and 1989 -- typical). During typical years (like 1989), there are five to seven significant frontal events during the fire season. In cool, wet years (like 1982) only two to four cold fronts appear to pass over the area. This may be because a steady stream of Pacific storms inhibits continental warming. Because maritime and continental air masses are similar, fronts between the two are unremarkable. This is most common when seasonal upper-level flow patterns have a dominant zonal, or westerly component.

In warm, dry years (like the 1988 fire season) eight to 20 gusty wind events can occur. During these seasons, the continental air mass has plenty of opportunity to heat and dry out. The cool, moist air masses from Pacific storms, which progress eastward are dramatically different than the hot, dry continental air masses. As the air masses meet, associated fronts can be very strong. This effect is most significant when the seasonal upper-level flow pattern includes frequent meridional, or northerly flow, over the Basin.

To develop a planning tool, it would be appropriate to group months and years into low, medium, and high gust event categories. Monthly composite maps of upper-level flow patterns can then be created to help distinguish the categories. If possible, these maps may be correlated with long-range monthly forecasts that now are available from the National Weather Service, Fire

Weather Forecast Offices. Unfortunately, this task was beyond the scope of the current project.

BLOW-DOWN

Although the Pacific Northwest is devoid of hurricanes, strong winds that cause forest damage are common. For example, in the Columbia Basin around the outflow of the Columbia Gorge and Snake River plain, sustained wind speeds of 20 meters per second (about 45 mph) have 2 year return intervals and gusts with higher speeds are even more prevalent (Wantz and Sinclair 1981). In addition, many steep walled valleys and canyons can channel and accelerate winds to damaging speeds.

Significantly strong winds, and storm winds that occur other than prevailing directions, are the dominant climate components to blow-down potential (Boose and others, in review). Trees most susceptible to blow-down usually are exposed to recently created up-wind fetches like the edge of clear-cuts made while harvesting timber or for agriculture and land development. Tree health and soil stability also can affect blow-down potential (Lohmander and Helles 1987; Miller and others 1987; Schaetzl and others 1988). Recent work in the Columbia River gorge also suggest that snow and ice accumulation on trees aid in blow-down potential.[1]

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[1] Personal communication. 1994, 1995, and 1996. Diana Sinton, Ph.D.

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To evaluate blow-down potential in the Columbia River Basin a brief survey of blow-down events was accumulated from forests around the region. The survey requested information on date of occurrence, average surface wind speed, wind gust speeds, species, stand height and density, if altered root environment or stand isolation (by fire scar, harvest, road, or water) existed, topographic characteristics, and percent of rot, bug kill, and dead. Unfortunately, such detailed records are not usually maintained. Scattered data on 15 events, however, were acquired. These events occurred on 13 different days; 2 days in winter, 3 in spring, 6 in summer, and 2 in autumn (Table 1). Because this was not a thorough survey, few inferences about the distribution of blow-down events can be made. It is surprising to note, however, that the greatest number of events appear to occur in the summer.[2] This is somewhat contrary to the Columbia gorge study where most blow-down events occur during winter with strong east winds.[1] This may be because gorge winds are channeled in only two directions: strong west winds during most of the growing season, and periodically strong east winds ahead of storms in winter.

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[2] Although care was taken to request the date of actual blow-down, instead of the date of observation, the number of events on June 15th (1987, 1989, and 1993) are suspicious.

To identify places where strong winds are common and then determine the direction and speed at which trees become "wind hardened," mean surface winds for each month were modeled.[3] The model output shows growing season winds (Appendix Q-7) are predominantly controlled by diurnal temperature variations that cause weak upslope winds during the day and relatively strong downslope winds during the night. In addition, onshore flows (marine pushes) are common during the summer as the land heats causing westerly winds to prevail through the Cascade passes. Trees at all elevations in the Cascade passes can become wind-hardened to the prevailing westerlies. Elsewhere around the Basin, trees at high elevation are wind-hardened to prevailing southwesterly upper-level flow patterns. Trees at lower elevation, which are exposed to weaker downslope flows, usually are less wind-hardened.

During winter (Appendix Q-5), winds are predominantly controlled by passage of storms from the Pacific. Prevailing winds are southwesterly at the higher elevations, which are exposed to the upper-level jet stream. At low elevations, however, east to southeasterly winds prevail where prefrontal surface pressure gradients are strongest. The offshore flow is channeled through the Cascade passes and strong easterly winds result. Because the east winds are stronger and completely opposite to the wind-hardening westerly direction, blow-down is common in the passes during winter. Winds in autumn

[3] Ferguson, S.A.; Peterson, M.R.; Hayes, P.S.; Akram, T. [In preparation]. Surface wind patterns in the Pacific Northwest.

(Appendix Q-8) and spring (Appendix Q-6) represent a shifting pattern between winter and summer.

A gap flow similar to that in the Cascade passes may play an important role for blow-down potential in northeastern Washington and northern Idaho. Here, deep river valleys funnel and help accelerate pressure gradient winds through the forests. The direction of these gap winds depend on whether storms pass over north or south of the area. Because trees are wind-hardened with dominating downslope flow in summer, the strength and direction of gap winds in the Colville, Okanogan, and Panhandle forests are significant enough to cause blow-down.

Strong winds in the upper Snake River plain result from converging air as the valley turns and narrows. The combination of high mountains and steep valleys contribute to strong wind patterns in far western Montana.

A simple model was developed to help determine the location of potential blow-down events by comparing storm winds with prevailing winds.[3] The model assumes that blow-down is possible if the storm wind is greater than 5 m / s and at least 45 degrees away from prevailing directions. During storm days in which blow-down was observed (Table 1), daily surface winds were compared against mean monthly winds in each 5 minute latitude / longitude grid. A few of the blow-down sites were correctly identified with the model.[3] Therefore, some promise in this type of regional modeling may exist for showing spatial and temporal patterns of blow-down potential, especially if combined with other environmental information like soil condition, and stand structure and health.

DROUGHT

Drought in the Columbia River Basin is common but tempered by an abundant source of moisture that spills over the coast ranges and through mountain gaps from the Pacific Ocean. Because most rainfall accumulates during winter and spring in the Basin, and because much of the summer water resources rely on winter snowpack, the most severe droughts occur with abnormally low winter and spring rainfalls (Namias 1983). For example, the 1977 and 1988 winter droughts were associated with an El Niño event that helped to strengthen a low center near the Aleutian islands. The Aleutian Low, coupled with a deep trough over the eastern United States, combined to strengthen and hold stationary a ridge of high pressure over the western United States (Namais 1978; Weber 1990). El Niño patterns occur every 1 to 7 years (Rasmusson and Wallace 1983, Philander 1983). Therefore, it may be reasonable to expect that drought in the Basin would have a similar return interval.

To help determine the spatial pattern of drought events, Palmer Drought Severity Index (PDSI) data (Palmer, 1965; Heddinghause and Sobol, 1991) were obtained from the National Climate Data Center (NCDC). The index generally ranges from -6 to +6, with negative values denoting dry spells and positive values indicating wet spells. There are a few values in the magnitude of +7 or -7. PDSI values 0 to -.5 are normal; -0.5 to -1.0 = incipient drought; -1.0 to -2.0 = mild drought; -2.0 to -3.0 = moderate drought; -3.0 to -4.0 = severe drought; and greater than -4.0 = extreme drought. Similar adjectives are attached to positive values for wet spells.

It should be understood that there is a 6 to 12 month response period inherent in the PDSI (Alley, 1984; McKee et al., 1993; McKee et al., 1995). In other words, it may require up to 6 to 12 months for periods of low rainfall to become evident in the PDSI. Many plants have a faster response time to water stress than 6 to 12 months. Therefore, PDSI is a very coarse approximation to drought. Data required to calculate other drought indices (for example, Keetch and Byram 1968; McKee and others 1993 and 1995), which are capable of higher resolution approximations, were not available in time for this study. Drought frequency patterns for each season were created by counting the number of times PDSI was less than -3 in each NWS climate division boundary for every season from 1895 to 1994. During winter (Appendix Q-9), drought is most common in places that typically have heavy winter precipitation, which is dependent on positioning of the jet stream and related storm tracks. During some years the winter storm track remains west to northwesterly and storms penetrating inland are too weak to maintain strength after crossing the coastal mountain barrier. This causes winter drought to be most frequent in areas that do not have enough topography to enhance lifting or convergence with weakened westerly winds: for example, on the east side of the Columbia Gorge (the coastal barrier's most significant gap), the Blue Mountains (which rely on abundant moisture from the southwest), and some of the lower elevation inland areas in Idaho, and western Montana and Wyoming.

In spring (Appendix Q-10), convective precipitation becomes important. Because convection is highly variable year-to-year, those places that depend on convective precipitation have more frequent periods of drought. This includes most of the Basin, except for some of the mountain areas near the Cascades, which benefit from spill-over Pacific moisture, and places in northern Washington and Idaho, which collect moisture from a few storms that pass over Canada during summer. Winds at the head of the Snake River commonly converge and this helps to enhance lifting and precipitation there.

During summer (Appendix Q-11), the reliance on convective precipitation becomes even more apparent with a spatial pattern of drought similar to spring but with somewhat higher frequencies. In Autumn (Appendix Q-12), the winter storm tracks begin to establish. This helps to relieve much of the Northwestern and mountainous parts of the Basin. The south and southeast portions, however, retain some reliance on the highly variable convective precipitation. Therefore, these regions of the Basin experience drought whenever autumn convection is minimal.

To determine the spatial nature of drought during characteristic climate years, total months of drought were tallied for the years 1982, 1988, and 1989. In 1982, a wet and cool year, drought was virtually nonexistent in the Basin (Appendix Q-13). In a warm, dry year (like 1988) drought can pervade the Basin except in areas that benefit from intrusions of marine moisture, such as near the outflow of the Columbia Gorge and where spill over occurs in the eastern Cascades (Appendix Q-14). During typical years (like 1989), drought can occur everywhere at times but typically is not severe except in the southeast, where water resources are dependent upon highly variable spring convection (Appendix Q-15).

By accumulating the number of months with significant drought for each climate division over time, periods of Basin-wide drought are apparent. Figure 2 shows the Basin has experienced frequent drought during the last 100 years. Most often, however, drought is limited to a small areas of the Basin or short periods. Severe, multiple year, Basin-wide droughts occurred in the 1930s and late 1980s.

LIGHTNING

Cloud-to-ground lightning strikes are the most common cause of wildfires in the western United States. Although the number of lightning strikes in the Pacific Northwest is low compared to elsewhere in the U.S. (Orville, 1994), their fire ignition potential is no less significant.

Until 1983, the best lightning occurrence data were available from fire lookouts. These data remain in a hardcopy format and are difficult to obtain. The data from 1925 to 1931, however, were summarized in a classic study by Morris (1934). Morris analyzed data from more that 2600 storms that were observed from 404 fire lookouts in Washington and Oregon. He found that most lightning storms occurred on the south side of the Blue Mountains in eastern Oregon and in the Colville district of northeast Washington. Smaller areas of high frequency lightning occurred on the east slopes of the Cascade mountains in the Okanogan and Deschutes forest districts.

With the advent of an automated lightning detection system (Latham 1983; Rasch and Mathewson, 1984) more detailed and ongoing analysis of lightning strike frequency patterns became possible. Hill and others (1987) analyzed lightning frequency pattern in Idaho with automated lightning data from 1985 and 1986. The greatest number of strikes were found to occur in Idaho's southeast corner.

Lightning data for the Columbia River Basin were obtained from the Bureau of Land Management (BLM) lightning detection network (Krider and others 1980) and include the ground location (latitude and longitude), polarity, signal strength, number of flashes per stroke, and date and time of occurrence. It has been estimated that the efficiency of the lightning detectors is roughly 70 percent (Orville 1994). It is assumed the detectors randomly Amiss@ lightning strikes and there are no systematic errors in the data associated with detection efficiency. Therefore, the relative frequency of strikes between different areas should be consistent even though exact numbers are probably less than actually occurred. The years 1986 to 1990 were available for this study.

A summary of lightning frequency during the period 1986 to 1990 was generated for each 10 km grid in the Basin (Appendix Q-16). It can be seen that the highest lightning frequencies occur near the perimeter and outside the boundary of the Basin, in the higher elevations of Nevada, Utah, Wyoming, and Montana. Within the Basin, highest frequencies are observed in western Montana, the Idaho panhandle, and eastern Oregon. Very few lightning strikes were recorded throughout the period of record in most of Washington and the western half of Oregon, except a relatively high number of strikes occurred in the Oregon Cascades in 1989 and 1990. It is important to note that higher frequencies observed in regions where lightning is typically rare are likely the result of one or two individual storms, rather than a higher occurrence of lightning throughout the course of the year.

A degree of inter-annual variability is evident when comparing all years (Table 2). For example, some years have more strikes than others. Also, some variation in seasonality is seen from year to year. Although July typically has the most number of strikes, August showed higher amounts in 1986 and June had the most in 1988. Usually less than one percent of the annual total occurs in the winter.

The number of lightning strikes in a given area does not directly correspond to the number of wildfires in the Columbia River Basin. Most lightning caused fires in the Basin occur in the Blue Mountains of northeast Oregon, and Sawtooth-Bitterroot ranges of central Idaho and far-western Montana (Appendix Q-16a) whereas most lightning strikes occur in southern Idaho and the Montana Rockies. This is somewhat contrary to the 1934 work of Morris, who found that lightning-caused fire locations agreed relatively well with lightning storm locations reported for the same period in Washington and Oregon. Morris noted, however, that some storms ignited many more fires than others without further explanation.

The date that fires are reported can be several days different from the actual fire ignition because long-period smoldering or inaccessibility can prevent immediate observation. Therefore, direct correlation between lightning and fire ignition could not be made with the available data. To investigate a better correlation between summarized information on lightning strike frequency and fire occurrence, some effort to distinguish those types of lightning discharges that ignite fires from those that do not was made.

It has been suggested that positive strikes (those that release a positive charge to ground) ignite forest fires (Fuquay 1980 and 1982). There could be several reasons for this. For example, positive flashes may have longer continuing current and higher peak currents than negative flashes. Also, positive flashes are more likely to strike the ground away from precipitation associated with the parent cloud, and are associated with greater flash densities (Fuquay 1980 and 1982; Beasley 1985; Brook and others 1989; MacGorman and Burgess 1994; Stolzenburg 1994). Maps generated with only positive strike locations, however, showed similar spatial patterns to total strike locations in the Basin and did not correlate with fire locations.[4]

Precipitation can put fires out and wet fuel can prevent ignition. Therefore, lightning that occurs with little or no precipitation is more likely to cause wildfire than lightning that occurs with precipitation. Work to distinguish between "wet" and "dry" lightning began many years ago (for example, Gisborne 1931). A simple correlation between lightning strikes and days with precipitation was attempted. Unfortunately, the spatial and temporal resolution of available precipitation data (Chapter 17A this volume) was too

[4] Peterson, M.R.; Ferguson, S.A.; Latham D.J. [In Preparation]. Lightning and fire ignition in the Pacific northwest.

coarse for reasonable correlation, even when nearest neighbor methods were used.[4]

Another way to distinguish wet lightning from dry lightning is to categorize it by air mass characteristics. For example, wet lightning is more common in conditionally unstable air where the entire air mass is relatively moist. Conditional instability often requires an external trigger (like lifting over topography) for the onset of convection and resulting lightning. Therefore, wet lightning patterns may be more likely to follow topographic patterns. Dry lightning is more common in unstable air masses that are relatively dry in the lower layers. This type of lightning often is associated with cold fronts (Finklin, 1981) and requires little or no topographic influence to trigger convection. Therefore, dry lightning patterns may align along frontal bands and trough axes with little or no regard to underlying topography.

To help categorize days based on atmospheric stability, a variety of stability indices were investigated. The Haines Index (HI) appeared most useful for this study (Haines, 1988). It is an index of atmospheric stability and low-level moisture based on upper-level temperatures, dewpoint temperatures, and heights of constant pressure surfaces. High instability, which often results in strong, gusty surface winds, and a dry lower atmosphere combine to produce a high potential for fires to spread and become large (Werth and Ochoa, 1990). Similar conditions should allow fires to ignite more readily. An unstable air mass is required for convection and associated lightning. Dry lower layers are needed to maintain dry fuels or evaporate any precipitation formed at the upper levels.

The HI was computed for Spokane and Boise, the only two locations within the Basin for which upper-air data are available. An example of a high HI day (13 August 1988) is shown in Appendix Q-17. While lightning is seen in the mountainous areas of western Montana and western Wyoming, a long band of strikes, suggestive of frontal forcing, is seen stretching across south-central Idaho, with no real connection between strike locations and terrain elevation. Light amounts of precipitation were reported on this day (less than 1.5 cm) but it may have been associated with a system that brushed the area within the previous 24 hours (National Weather Service 1988).

An example of a low HI day (August 1, 1989) is shown in Appendix Q-18. Most of the lightning strikes occurred at higher elevations in eastern Oregon, eastern Idaho, and western Montana. Few lightning strikes were seen in the Snake River Valley of Idaho or in eastern Washington. Rainfall on this day occurred over much the same area, with daily amounts near 4 cm.

A definite pattern emerges on the low-HI days, with lightning corresponding to the terrain features, while no such pattern is evident on the high-HI days. Current work is underway to develop map classification schemes that identify dry lightning patterns, which appear to have definite synoptic signatures. Similar efforts have appeared promising in Florida (Reap 1994) and Idaho (Hill and others 1987) but failed to be conclusive because wet lightning, which may have no preferred synoptic pattern, was included in their analyses.

COLD DAMAGE POTENTIAL

Very cold temperatures can damage vegetation and cause surface and ground water to freeze. Short-term freeze primarily affects new growth. Long-term freeze can affect lake and stream systems and cause sustained damage to mature vegetation.

Previously stressed trees are more likely to sustain damage from temperatures that normally cause no damage (Levitt, 1980). Micro-scale topographic effects also are factors in cold damage. Trees on north and west facing slopes receive more damage during winter storms than south and east facing slopes (Porter, 1959). On cold, clear nights saplings in hollows and valleys are much more prone to frost damage due to pooling of frigid air (Blennow, 1992).

There are few cases of documented cold damage in the Columbia River Basin. Therefore, in order to gain insight into the potential for cold damage, a model was designed to include many of the known causes for cold damage to trees.[5] In winter, cold damage primarily occurs during a cold snap that immediately follows an unseasonably warm period. Spring damage occurs to trees when freeze follows bud-break. During autumn, damage occurs if freeze precedes cold hardiness. By applying this model with distributed temperature data that was generated by Thornton and Running (1996) the spatial extent of cold damage potential during three characteristic years becomes apparent.

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[5] Westrick, K.J.; Ferguson, S.A.; Peterson, M.R.; Jewett D.S. [In Preparation]. Cold damage potential in northwest forests.

During a characteristic cool, wet year (1982), low to moderate cold damage is possible for many areas around the Basin, mainly above 2500 meters (Appendix Q-19). Some damage to peach trees in east-central Oregon was reported during this time, but most other areas showing potential damage were above tree-line. A moist, zonal flow pattern commonly is associated with this type of climate in the Basin, causing moderately varying temperatures with few extremes. There was one spring blizzard and two moderately cold weather systems in autumn during this time that may have caused some cold damage.

A warm, dry year (1988), spawned a few more cold damage events (Appendix Q-20). Weather patterns in this type of climate commonly are associated with a split upper-level flow that prevents frequent storms from crossing the Basin. Warmer overall temperatures allow a greater percentage of trees to come out of dormancy earlier in the spring season and go into dormancy later in autumn. This makes them more susceptible to cold storms that can break through the split flow and pass over the area.

A more typical year is represented by temperatures in 1989 (Appendix Q-21). Typical northwest weather occurs when the upper-level flow pattern fluctuates between moist zonal flow and cold meridional flow. Although seasonal conditions allow normal timing of dormancy and bud-break, extremes in temperature also are possible. During this year, significant cold damage was reported in western Montana (Klein, 1990).

RAIN-ON-SNOW FLOODS

When heavy rain falls on and penetrates an existing snow cover, intense and damaging floods are possible. Harr (1981) indicated that 85 percent of landslides in western Oregon are associated with snow-melt during rainfall. Similar studies have been conducted in British Columbia (Beaudry and Golding, 1985), California (Hall and Hannaford, 1983), and western Washington (Harr and Cundy, 1990), but little is known about the severity and frequency of rain-on-snow (ROS) floods east of the coastal mountains.

There have been few, if any, studies of ROS floods east of coastal mountains because it is commonly believed that fall and winter rain in continental areas is rare. The Columbia River Basin, however, has a unique topographic configuration that allows frequent incursion of warm, moist air from the Pacific Ocean. There is anecdotal evidence of significant ROS floods in the Blue Mountains of eastern Washington and Oregon during a November 1990 event that also affected many west-side rivers. Also, the 1995 to 1996 season caused significant rain-on-snow flooding in northern Idaho. Paleo-climate evidence suggests that ROS events caused significant flooding in the Columbia River Basin during several periods between 18000 and 9000 yr B.P. (Chatters and Hoover, 1992). Since 120 A.D., major floods (similar to the 1948 event) along the Columbia River system appear to occur every 140 years, except for a period between 1020 and 1390 A.D. when floods occurred at rates 3 to 4 times more frequently (Chatters and Hoover 1986).

To help understand the frequency and severity of ROS floods in the Basin, this study began by reviewing unregulated stream gauges. All of the stations in the Wallis-Lettenmaier-Wood data set (1991) in the Columbia Basin above The

Dalles were selected. Maximum daily flow for each station for the 40 year period (1948 to 1987) was identified. Next, all of the floods that occurred during the months of October to February were classified as ROS events and those that occurred in other months as pure snow-melt (that is, dominantly radiation melt) events.

Appendix Q-22 shows the percentage of floods that were classified as pure snow-melt at each stream gage site. Twenty-three of the stations had no floods that were classified as rain-on-snow. These stations generally either represent streams that head in the Rockies or represent relatively high elevation catchments. Those stations in the interior Basin that were classified with a high percentage of ROS flood represent relatively low elevation catchments. This corroborates evidence that snow accumulations at low elevations can significantly enhance ROS floods (Harr and Cundy, 1992). Even with a low percentage of ROS floods, it is interesting to note that several stations experience their heaviest floods during the infrequent ROS events.[6]

There are a number of snow melt equations (for example, U.S. Army Corps of Engineers 1956; Brun and others 1989) that are used to estimate snow melt and ROS events. Accurate calculations of these equations require a number of weather parameters, which include; air temperature, air pressure, precipitation, incoming radiation, relative humidity, and wind speed. Unfortunately, only a few of these measurements are available historically.

[6] Colburn, K.J., S.A. Ferguson, D.P. Lettenmaier, M.R. Peterson, and D.S. Jewett. [In Preparation]. Rain on Snow Flood Potential.

In addition, geographically distributed models of snow melt usually describe mean or average conditions and omit extreme events like those caused by rain on snow.

Without precise measurement data or extreme statistics, ROS flood potential can be estimated using threshold criteria (e.g., Brunengo, 1990). To develop a criterion approach that would be applicable to all regions of the Columbia River Basin, an empirical model using available temperature and precipitation was designed.[6] The model was developed by analyzing weather observations during incidents where river flow increased over 75 percent within a period of 2 days. It was found that if a shallow snow cover exists, then on any given day when air temperatures rise above 0 °C with accompanying precipitation, rain-on-snow flooding is possible.

The rain-on-snow model was applied over the Basin with distributed temperature and precipitation data (Thornton and Running, 1996) for characteristic climate years of 1982, 1988, and 1989 (Appendix Q-23, 24, and 25, respectively). A comparison of model output with stream gage data shows good agreement. In all years, the model indicates rain-on-snow events are possible in many areas of the Basin. On closer inspection, areas most susceptible to ROS events are those where topography allows incursion of relatively warm, moist marine air that flows into the Columbia plateau and up the Snake River valley from the Pacific Ocean. These areas include the Cascade mountains; northern Idaho, northeastern Washington, and Northwestern Montana with valleys that open into the Columbia plateau; the Blue Mountains of northeastern Oregon; and western Wyoming and central and Idaho adjacent to the Snake River.

Cool, wet years (1982) are likely to have more rain-on-snow events. The cool temperatures allow low elevation snow to accumulate; whereas the frequent precipitation brings the possibility of mid-winter rain. Warm, dry years (1988) are less likely to experience rain-on-snow flood events. There is little low elevation snow at these times and only occasional precipitation.

SUMMARY

Although the Columbia River Basin is not noted for extreme climate, there are many climatic events that can disrupt ecological processes. These include: summer cold fronts, which fan and enlarge wildfire; strong winds that blow down or damage trees; periodic drought, especially during winter and spring, which deplete the snowpack; lightning that ignites wildfire; extreme or untimely cold temperatures, which damage vegetation and cause freezing water problems; and rain falling on snow, which can significantly enhance flooding. Although all disturbance climate events in the Columbia River Basin can be relatively common, little work has been done to synthesize or model the event characteristics in ways that can help land managers better plan for their destructive influence.

A few simple methods were developed to help summarize even structures for this report, but much work remains. Our understanding of climatic trends and our ability to forecast long-term weather patterns are improving rapidly. These skills should be included in seasonal and multiple-year management plans.

Increasing dialogue between climatologists and those who manage and study land resources, however, is needed to better define climate-based management tools.

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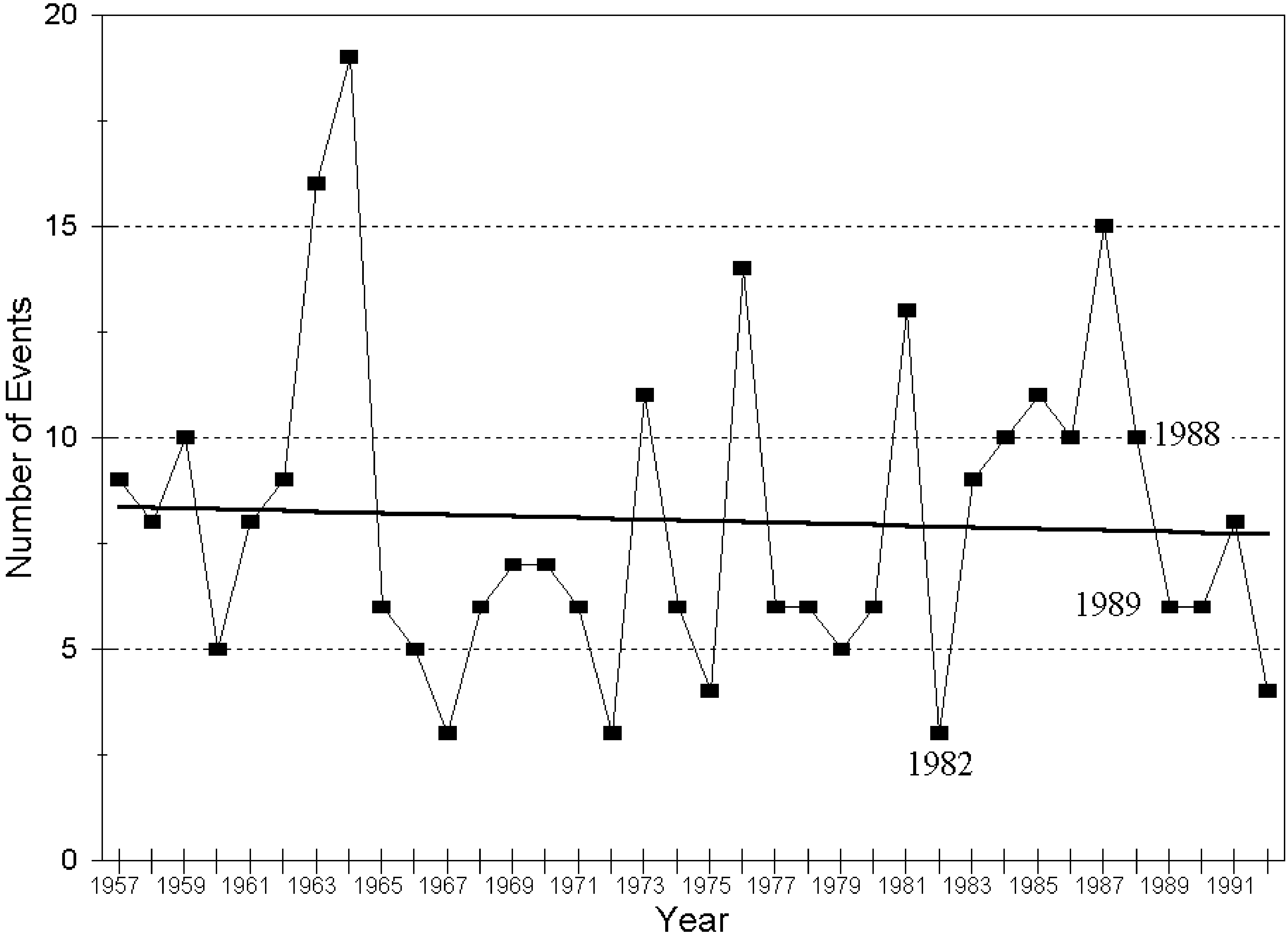
LIST OF FIGURES

LIST OF TABLES



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1991
Year
1988
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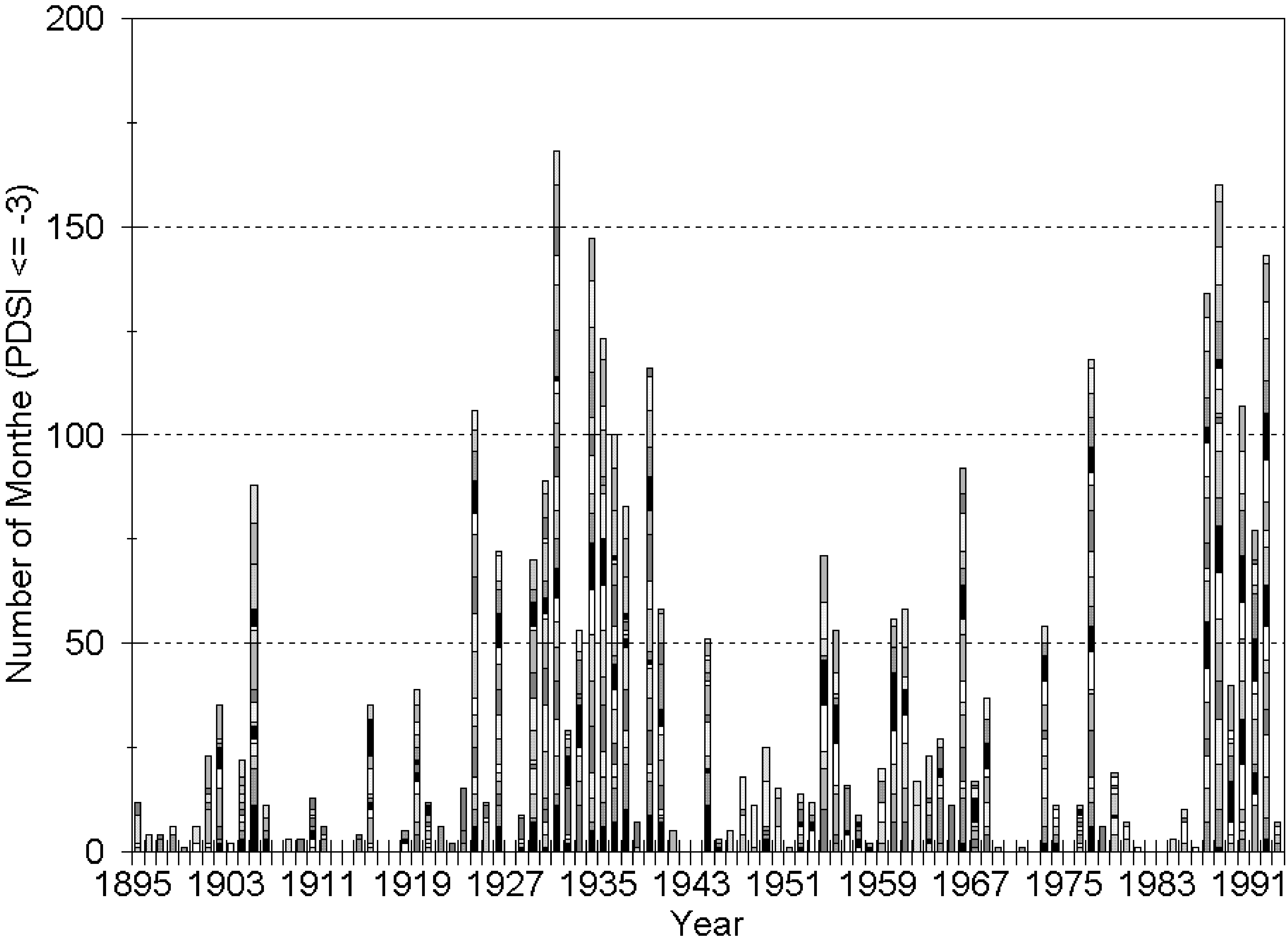

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0
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1911
1919
1927
1935
1943
1951
1959
1967
1975
1983
1991
Year