

Chapter 4

Climate and Spatial Patterns of Wildfire in North America

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4.1 Introduction

Climate interacts with wildfire at a range of spatial and temporal scales. In this chapter I describe a conceptual model that describes how climate (a top-down control) interacts with processes of vegetation development and topography (bottom-up controls) to give rise to characteristic disturbance regimes and observed patterns of wildfire throughout North America. At the shortest timescales (synoptic to seasonal), climate influences fine fuel moisture, ignition frequency, and rates of wildfire spread. At intermediate timescales (annual to interannual), climate affects the relative abundance and continuity of fine fuels, as well as the abundance and moisture content of coarser fuels. At longer timescales (decadal to centennial) climate determines the assemblage of species that can survive at a particular location. Interactions between these species' characteristics and the influence of climatic processes on wildfire activity give rise to the characteristic disturbance regime and vegetation structure at a given location. Large-scale modes of climatic variability such as the El Niño – Southern Oscillation and the Pacific Decadal Oscillation affect patterns in wildfire by influencing the relative frequencies of shorter scale processes. Because the importance of these processes varies depending on topographic position and the ecology of the dominant vegetation the effects of these modes varies both between and within regions. Global climatic change is effectively a centennial to millennial scale process, and so its effects can be understood as resulting from interactions between the observed patterns of higher frequency processes, as well as processes of vegetation change whose temporal evolution exceeds the length of the observational record. Statistical models of future fire that are based on historical fire climate relations and regionally downscaled climate forecasts suggest that in most regions of North America wildfire will increase in frequency over the next several decades. Predictions beyond this interval are probably

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unreliable as vegetation structure and composition will be changing rapidly in response to changing climatic conditions and fire regimes.

Spatial variability in the structure and composition of vegetation occurs as a legacy of interacting processes that are biological, geological, geomorphological, climatic, and anthropogenic in origin. Of the processes that shape ecosystems, none is more dramatic or more important (at least in temperate regions) than fire. The behavior of individual fires is largely determined by the nature of the fuels, weather, and topography that characterize the site of ignition (Johnson 1992; Agee 1993). Of these factors, weather is the most variable over time (Bessie and Johnson 1995), and is the most poorly understood (e.g., Gedalof et al. 2005). Because the vast majority of area currently burned by wildfire is caused by relatively few fires that burn under extreme weather conditions (Strauss et al. 1989; Gedalof et al. 2005), it is important to understand the causes of variability in extreme fire weather.

The effects of fire weather on fire behavior do not appear to be consistent across space (Jones and Mann 2004). Rather, weather interacts with other factors to give rise to the specific fire regime of a given location. These factors can be generally characterized as being either top-down or bottom-up (Chaps. 1 and 3). Top-down controls include those that originate outside the ecosystem. Of these climate is the most important, although anthropogenic influences are locally important. Bottom-up controls include those that originate inside the ecosystem, such as topography and vegetation dynamics.

Assessments of the relative contributions of top-down and bottom-up controls on wildfire are complicated by many interacting factors, including:

- The climatic history of the earth has not been static at any scale of variability, and will continue to change over the coming decades to centuries (Karl 1985; Meehl et al. 2005)
- Intensive land use by people, including forestry and road building, and grazing by sheep and cattle changed forest conditions in many regions (Madany and West 1983; Belsky and Blumenthal 1997; Heyerdahl et al. 2001)
- Many landscapes may be a legacy of fire use by indigenous peoples, although the pre-settlement fire regime is not well known in most cases (Brown and Hebda 2002; Keeley 2002; Williams 2002; Gedalof et al. 2006)
- Records of fire history are generally short, often lack detailed location information, and are not easily reconciled (Westerling et al. 2003; Gedalof et al. 2005)
- The effects of fire suppression on area burned are uncertain, and controversial in many forest types (Keeley et al. 1999; Johnson et al. 2001; Ward et al. 2001; Bridge et al. 2005).

Despite these challenges, emerging data sets and analytical methods have allowed important insights into the processes that give rise to spatial patterning in severe wildfire years at regional scales (10^3 – 10^6 km²), and a coherent conceptual model is emerging. The purpose of this chapter is to summarize recent developments in

understanding the role of top-down controls, and in particular climate, on variability in area burned by wildfire. Specifically, I summarize the mechanisms by which top-down controls give rise to widespread severe wildfire years, describe several important patterns of climatic variability and assess their role in giving rise to regional patterns of wildfire, and discuss how vegetation cover and other bottom-up controls modulate the response of a given region to climatic variability to give rise to landscape-scale responses (10^0 – 10^4 km²) to these top-down controls. This context is used to understand how climatic change may affect fire frequency over the next several decades to centuries.

4.2 Mechanisms of Top-down Control

Variability in the Earth's climate system represents the most important source of variability in the fire regime of most regions (e.g. Stahle et al. 2000). Properties of the climate system that can affect wildfire include temperature, precipitation, wind speed, relative humidity, and lightning activity. These properties fluctuate in space and time across many orders of magnitude, ranging, for example, from a sunfleck that might dry a few square meters for a minute or two, to a mega-drought that might persist throughout a given region for decades or more (Schroeder 1969; Strauss et al. 1989; Johnson and Wowchuk 1993). The effects of these fluctuations are similarly variable, depending on their characteristic scale and properties of the ecosystem they are incident upon. In the following sections I summarize the main mechanisms by which climatic variability can affect fire, focusing on how the scale of the climatic process involved influences the impact on the fire regime.

4.2.1 Ignition Events

Lightning is the most important natural cause of wildfire ignitions throughout North America (Morris 1934; Rorig and Ferguson 1999; Malamud et al. 2005). Lightning is caused by convection within clouds acting to separate positive and negative charges (Uman 2001; Burrows et al. 2002; van Wagtenonk and Cayan 2008). The convection that gives rise to these charge differentials is most commonly associated with unstable air masses associated with differential surface heating, or diurnal variability in surface temperatures (Uman 2001). Most lightning associated with electrical storms is contained within the cloud (i.e., occurs as intracloud lightning), and is not associated with fire ignitions, but a small proportion of lightning occurs as cloud-to-ground lightning. Surprisingly, most cloud-to-ground lightning strikes are so brief that while they may cause considerable damage to trees they rarely generate sufficient heat to ignite fuels (Latham and Williams 2001). However, about

30% of these strikes are associated with sustained current flows that do generate sufficient heat to potentially ignite fires. Of particular importance are the approximately 10% of strikes that are positive in polarity, which much more commonly sustain the currents needed to ignite fires (Latham and Williams 2001). The causes of positive-polarity lightning are still unknown, but there are distinct regional patterns that may be an important cause of variability in fire frequency. For example, positive-polarity strikes occur most commonly in North America over northern Minnesota and adjacent parts of Ontario and Manitoba, Canada (Lyons et al. 1998). Podur et al. (2003) found evidence for regional increases in lightning ignitions in this region of Canada, which they attributed to localized dry weather and lightning storm occurrence, but increased frequency of positive-polarity lightning could also contribute to this region's anomalously frequent ignitions.

The factors that cause convection, and consequently the frequency of lightning strikes, vary diurnally, by time of year, and between years. At large spatial scales, lightning occurs more frequently in continental than maritime regions, and more frequently at intermediate elevations than at higher elevations (which in turn occurs more frequently than at low elevations). Across North America, the greatest density of lightning strikes occurs in central Florida, and decreases toward the northwest. Relatively few lightning strikes occur west of the Western Cordillera in either the United States or Canada (Huffines and Orville 1999; Burrows et al. 2002). Superimposed on this large-scale pattern there are important regional differences in the frequency of lightning strikes. Topographic variability and land-water temperature differences influence patterns of atmospheric convection, resulting in sub-regional patterns of lightning variability. For example, in Colorado lightning strikes occur most frequently just east of the Continental Divide (Lopez and Holle 1986). Similarly, in Canada, a regional increase in lightning strikes is found in the foothills region, east of the Rocky Mountains (Burrows et al. 2002).

Surprisingly, however, most researchers have found a poor correspondence between the frequency of lightning strikes and the frequency of ignition events, suggesting that lightning is a necessary but not sufficient condition for wildfire to occur (Morris 1934; Nash and Johnson 1996; Rorig and Ferguson 1999; Latham and Williams 2001). The factors that cause spatial variability in lightning frequency include atmospheric humidity, topography, and surficial properties. These same factors are associated with changes in vegetation type that in turn influence the flammability and continuity of fuels. Consequently, patterns of lightning frequency alone are poor predictors of patterns in wildfire occurrence.

Part of this discrepancy is because for ignition to occur, lightning needs to strike a fuel bed that is sufficiently dry to maintain combustion, and sufficiently continuous for fire to spread. The percentage of successful ignitions per cloud-to-ground lightning strike, termed lightning ignition efficiency, typically ranges between 1 and 4% (Meisner 1993; Latham and Williams 2001). Because the processes that generate lightning require moisture, and are usually associated with precipitation, successful ignitions occur most efficiently under fairly specific conditions (Nash and Johnson 1996). In particular, ignitions occur when fuels are particularly dry

due to antecedent weather conditions (see below), and when lightning strikes are not accompanied by precipitation. This “dry” lightning occurs most frequently when the lower atmosphere is particularly unstable, resulting in intense convection and often dry conditions in the lower atmosphere that cause precipitation to evaporate before it reaches the ground (Rorig and Ferguson 1999). These same conditions are also associated with gusty winds that contribute to rapid fire spread. Although dry lightning is probably the most effective cause of ignitions, it is not the only type of lightning that ignites wildfires. Ignitions can also occur in cases where the fuel bed is exceptionally dry, and the precipitation associated with the thunder storm is not sufficient to inhibit burning, during small fast-moving storms that deliver little precipitation to any single location, or when the lightning strikes outside the main plume of the storm (Rorig and Ferguson 2002).

Ignition efficiency differs between various land cover types. Meisner (1993) examined lightning strike and ignition frequency in southern Idaho as functions of the dominant vegetation type. He found that ignition efficiencies ranged from 0.3% (for agricultural crops) to 10% (for logging slash). Mature forests ranged from about 2–4%. Latham and Williams (2001) reached similar conclusions for a more extensive region, and indeed found that some areas of exceptionally high strike density had actually experienced no fires over the duration of their analysis. In California, desert regions experience the most lightning per unit area, but ignitions are very rare due to the discontinuous nature of the fuel bed (van Wagtenonk and Cayan 2008). Krawchuk et al. (2006) found that conifer forests were more likely to burn than nearby deciduous forests. Ignition efficiency also differs between locations within the same basic vegetation type. For example, Díaz-Avalos et al. (2001) found that in the Blue Mountains in Oregon ignition efficiency was higher at lower elevations despite the lower frequency of lightning strikes, and peaked within the central portion of the range—although they were unable to explain the reason for this spatial pattern.

4.2.2 *Fire Spread*

It is generally recognized that the great majority of area burned by wildfire is caused by relatively few fires that occur under extreme weather conditions (Schroeder 1969; Strauss et al. 1989; Johnson and Wowchuk 1993). For example, one commonly repeated statistic suggests that 99% of the area burned is caused by 1% of the fires. Although the actual figure is probably closer to 90% (Strauss et al. 1989), the importance of relatively few fires causing the bulk of the variability in area burned remains the same. These fires are usually associated with high temperatures, exceptionally low relative humidity, and strong winds (Schroeder 1969; Flannigan and Harrington 1988; Crimmins 2006).

The relationship between fire spread and short-term variations in meteorological variables is reasonably well understood. Early work by Fons (1946) was built on, in particular, by Rothermel (1972, 1983) to develop empirical models of fire spread based

on fuel characteristics, slope, and wind speed. Several fire spread simulators based on these mathematical models are now used operationally and in the development of management plans (Finney 1998, 1999; Hargrove et al. 2000; Andrews 2007; Tymstra et al. 2007). These models explain how fire spreads across a given landscape in response to critical fire weather, but do not offer insights into how fire is synchronized across landscapes to give rise to characteristic years of exceptionally high or low fire activity. Schroeder (1969) undertook the first systematic effort to identify meso-scale patterns of atmospheric pressure associated with extreme fire hazard. The patterns most strongly associated with extreme fire hazard are characterized by anomalous high surface pressure. These systems, commonly called blocking ridges, divert moisture away from the region (Wiedenmann et al. 2002). Along their margins (or during their passage) strong pressure gradients contribute to strong winds that cause rapid spread. When blocking ridges are particularly intense the passage of cyclonic storms may produce strong wind and lightning, but little precipitation (Rorig and Ferguson 1999).

A second common set of patterns was associated with air masses that cross mountains (Schroeder 1969). Moisture is lost from these systems due to orographic precipitation along the windward slopes. As the (now) dry air descends the lee slopes it warms by compression, and relative humidity decreases further. Along the eastern slopes of the Rocky Mountains these winds are called Chinooks. Fire danger is greatest when Chinooks are associated with ridges west of the Rocky Mountains that enhance drying due to subsidence, and contribute to strong pressure gradients and the resulting surface winds. An analogous but more severe fire-weather pattern occurs when winds are easterly, i.e. from the continental interior to the coast. In these cases, the air mass is typically dry to start with, and is exceptionally dry when it reaches the coast. These winds are generally termed foehn winds, but often have local names such as Diablo, sundowner, or Santa Ana winds. In southern California and northern Baja California Santa Ana winds are associated with some of the most extreme wildfires (Keeley et al. 1999; Keeley and Fotheringham 2002) many of which spread into the urban wildland interface resulting in losses of structures and human lives (Keeley et al. 2004).

The synoptic circulation patterns that Schroeder (1969) identified have since been validated using more extensive data and objective analytical techniques (e.g., Skinner et al. 1999, 2002; Gedalof et al. 2005; Crimmins 2006), and confirmed in a large number of case studies (e.g. Countryman et al. 1969; Sando and Haines 1972; Finklin 1973; Street and Alexander 1980). Little work has been done, however, to explicitly link this variability in severe fire *weather* to large-scale ocean-atmosphere interactions. Such a linkage is implicit in analyses that identify climatic patterns at timescales longer than about 10 days (Flannigan and Harrington 1988; Johnson and Wowchuk 1993; Skinner et al. 1999, 2002; Gedalof et al. 2005; Trouet et al. 2006), but none of these analyses discriminated the factors that contribute to fire hazard (such as fuels production and fuels drying) or ignition efficiency from those that contribute to rapid spread. There is evidence that the frequency of extreme events differs depending on the state of large-scale modes of variability, suggesting that such an analysis might prove fruitful. For example, Thompson and Wallace (2001) found that strong winter winds in coastal Washington and Oregon

occur approximately three times more often during the positive phase of the Northern Hemisphere Annular Mode than during the negative phase. Similarly, blocking ridges in the Pacific Region occur more commonly during the cold (La Niña) phase of the El Niño Southern Oscillation than during the warm (El Niño) phase (Wiedenmann et al. 2002).

4.2.3 *Fuel Moisture*

Most land cover types are not so flammable that the above processes alone can explain the regional synchrony of severe fire years. Depending on the dominant vegetation structure, some period of antecedent drought is needed to dry the fuel bed so that fire will spread rapidly (Johnson and Wowchuk 1993; Bessie and Johnson 1995; Meyn et al. 2007; Littell et al. 2009). The relative importance of antecedent drought varies, depending on both mean regional climate and the structure and composition of the fuel bed. In particular, the relative abundance of fine vs. coarse fuels, and the continuity of the fuel bed, determine the importance of antecedent drought in preconditioning stands to burn (Schoennagel et al. 2004, 2005; Gedalof et al. 2005). Because the relative abundance and arrangement of fine fuels differs between land cover types, their drying rate and capacity to carry fire will also differ (Westerling et al. 2003; Gedalof et al. 2005).

Two basic factors regulate the moisture of fuels. First, plant functional type determines the phenology of vegetation (which determines whether foliage and shoots are metabolically active, dormant, or dead) as well as its structure and physiology (which determines its capacity to maintain high moisture levels in either plant tissue or in dead organic matter). Second, antecedent weather determines the moisture available to vegetation, as well as the rate of evaporative and transpirative losses. At their extremes, these processes support the ideas of ignition-limited ecosystems (those with abundant fuels, but that do not burn due to the infrequency of ignition events or the high moisture of fuels) vs. fuel limited ecosystems (those that experience frequent potential ignitions, but that often do not have sufficiently abundant or continuous fuels to allow fire to spread). Meyn et al. (2007) provide a useful conceptual framework for synthesizing these processes, and also identify a third type of ecosystem that does not fit neatly into this dichotomy. They characterize these ecosystems as “biomass poor, rarely dry,” and they are both fuels and ignition limited. Examples include subalpine forests, temperate savannas, many wetlands, and some types of chaparral.

High temperatures, low relative humidity, and strong winds in the days to months preceding a potential ignition dry living and dead fuels, and can cause vegetation to senesce. The relative importance of antecedent drying varies by ecosystem type, with some ecosystems requiring much longer periods of time to dry sufficiently to carry fire than others (Westerling et al. 2003). For example, Gedalof et al. (2005) analyzed the relative importance of drought in the months preceding extreme wildfire years in the Pacific Northwest, USA, and found that

coastal temperate rainforest experienced large area burned only during years of exceptional drought persisting throughout the winter and spring preceding the fire season. In contrast, dry forest types such as those found in eastern Washington and Oregon experienced extreme wildfire years even in the absence of persistent drought. These differences in sensitivity to antecedent climate can be explained in part by the relative abundances of fine fuels. Many wet forests are characterized by abundant standing and down woody debris that retains moisture effectively (Franklin et al. 1981). These large fuel classes require prolonged dry periods before they become flammable, and they also buffer surface vegetation against prolonged soil-moisture deficits by providing a reservoir of moisture. Closed-canopy forests further buffer surface fuels from drying by reducing insolation, temperature, and windspeeds at the surface, and by helping to maintain high relative humidity (Chen et al. 1999).

In ecosystems with higher relative abundances of fine fuels, such as grasslands, savannas, and chaparral, shorter periods of dry weather are sufficient to precondition ecosystems to burn (Westerling and Swetnam 2003; Gedalof et al. 2005). This difference can be explained in part by the faster drying rate of fine fuels, but is also enhanced by the tendency for these ecosystems to have lower canopy cover, and thus greater evaporation and transpiration from the surface. Many ecosystems dominated by fine fuels also have greater proportions of annual vs. perennial vegetation – meaning that there is more dead fuel at the surface (Knapp 1995). These fuels dry more readily than living vegetation, because they do not maintain internal moisture by using groundwater or resisting transpiration through adaptive measures.

Xeric ecosystems dominated by fine fuels respond to shorter-term variations in fire weather, and are generally more sensitive to the availability of a continuous fine fuel bed (see below), but they are also responsive to seasonal patterns of moisture availability. For example, in southwestern ponderosa pine forests regionally synchronous fire years are strongly associated with drier than average spring conditions (Swetnam and Betancourt 1990). In addition to synchronizing fire activity, large-scale drought may also increase the severity of fire in landscapes more commonly associated with low-severity fire – leading to a complex mosaic of snags, and patches of living trees with heterogeneous age structures (Agee 1998; Baker et al. 2007).

4.2.4 Fuels Production

At seasonal and longer timescales, climatic variability can affect the wildfire regime by modifying the abundance and continuity of fuels, and the relative abundance of fine vs. coarse fuels. As with fuel moisture, the relative importance of antecedent climate in fuels production varies depending on the dominant vegetation present, and the climate of the region (Westerling et al. 2003; Gedalof et al. 2005; Littell et al. 2009). Because the rate of fire spread and intensity at the flaming front is determined primarily by fine fuels, it is largely this fuel component that limits the ignition and spread of fire in most ecosystems (Rothermel 1972; Bessie

and Johnson 1995). Most closed-canopy forests have abundant fine fuels, and do not increase in fire hazard with increased production beyond the point at which closed-canopy conditions are achieved, which typically occurs in the first two or three decades of development (Bessie and Johnson 1995; Schimmel and Granström 1997; Keeley et al. 1999; Johnson et al. 2001; Schoennagel et al. 2004).

In ecosystems that are characteristically dry enough that fuels are patchy, climatic conditions conducive to the growth of vegetation may increase the abundance and continuity of fine fuels, increasing the potential for fire during subsequent seasons (Swetnam and Betancourt 1998; Westerling et al. 2003; Collins et al. 2006). Open ponderosa pine forests that are characterized by short fire return intervals often show positive correlations to precipitation in the year(s) preceding regionally synchronous fire years (e.g. Swetnam and Betancourt 1998; Brown and Shepperd 2001; Kitzberger et al. 2007). This relationship is not constant throughout the species' range, however. For example, Sherriff and Veblen (2008) found that antecedent moisture increased fire occurrence in ponderosa pine forests in northern Colorado only at low elevations; at higher elevations this relationship was unimportant. Brown and Shepperd (2001) found that it occurred only in the southernmost portion of their study region in Colorado and Wyoming. In the northern portions of their study region they found that fire was associated only with drought during the year of fire. They also found that stand-replacing fires occurred frequently throughout the study region, even in the pre-suppression era. In the U.S. Southwest, relationships to antecedent moisture are more common due to the generally shorter fire-return interval and warmer mean climatic conditions, which limit fuel accumulation and production respectively (Swetnam and Betancourt 1998; Stephens and Collins 2004), but even there it is generally restricted to ponderosa pine forests and the relationship is not found for mixed conifer forests in the same region (Swetnam and Baisan 1996).

Antecedent moisture plays a particularly strong role in producing fuels and synchronizing fire in ecosystems dominated by annual grasses and herbs (Cable 1975; Knapp 1995; Brooks and Matchett 2006). Indeed, this relationship is sufficiently important in grass dominated ecosystems that in the Great Plains region at long time periods (decades to centuries), fires are more commonly associated with prolonged wet periods than dry ones (Brown et al. 2005) – although conditions are likely dry while fires actually burn. The importance of grasses in producing a continuous fine fuel bed that will carry fire has changed the fire regime of many arid and semi-arid ecosystems where exotic grasses have invaded. For example, in the Intermountain West the introduction of annual grasses, especially cheatgrass (*Bromus tectorum*) has increased the size and frequency of wildfires (Knapp 1995). This change to the fire regime has altered vegetation dynamics, as the affected communities have not evolved with frequent fire (Knapp 1998). The result has been reduced biodiversity, and economic losses associated with lost pasture and suppression efforts.

In North American deserts introduced grasses have caused fires to occur in regions that would historically have experienced little or no fire at all (Brooks and Pyke 2001; Brooks and Matchett 2006). These fires are disrupting regeneration of

desert vegetation that is not adapted to fire, and in some regions have converted desert scrub to grassland. In these regions the relationship between fire and climate has changed: whereas historically fire would have occurred very rarely, and only following multiyear to decadal pluvials, it is now occurring following short wet periods lasting perhaps a single season (Brooks and Pyke 2001).

4.3 Patterns of Top-down Control

A distinctive feature of the Earth's climate system is that it varies over time and space in characteristic patterns or "modes" of variability. Some of these modes, such as diurnal or seasonal temperature fluctuations, are readily observable and can be easily explained as a result of the Earth's rotation and revolution around the sun. Other modes are less readily observable, affect different regions of the Earth uniquely, and overlap in time and space (Namias and Cayan 1981). These modes of variability influence patterns of atmospheric pressure, temperature, and precipitation over spatial scales that exceed 10^6 km², and over timescales of months to decades or longer (Wallace 2000). They influence fire frequency mainly through their influence on rates of fuel production and drying, but may also influence the frequency of ignitions and the statistics of extreme winds. Their spatial imprint on the wildfire record is a result of interactions between the spatial expression of the mode of climatic variability and the response of individual ecosystems to variability in climate. There are a dizzying array of these modes documented in the literature, although many of them are probably related to each other (Dommenget and Latif 2002), or statistical artifacts rather than separate physical processes (Enfield 1989; McPhaden et al. 2006). Nevertheless, a few of these modes are emerging as fundamental. In this section I review four of the more important modes that impact North America, and summarize their influence on wildfire. I use the term *teleconnection* to explain how climatic variability in one region affects the climate of more distant locations (Wallace and Gutzler 1981).

4.3.1 *The El Niño Southern Oscillation*

The El Niño Southern Oscillation (ENSO) is the most important source of global climatic variability at interannual timescales. ENSO events result from feedback between the tropical oceans and atmosphere (Wyrski 1975). During non-ENSO years, the trade winds blow from east to west, and surface waters are pushed away from South America towards Indonesia. These waters warm as they are heated by the sun, and the height of the sea surface increases as water accumulates along the western margin of the Pacific Ocean. On average, the surface height is about 0.5 m higher along the Indonesian coast than along the South American coast (Enfield 1989). This warm water pool heats the air above it, causing it to rise—helping to

maintain the east-to-west flow of the trade winds, and bringing the high rainfall typical of Indonesia. Episodically, the trade winds weaken, and the pool of warm water “sloshes” eastwards towards South America. This process further weakens the trade winds, accelerating the eastward movement of the warm water. This system is coupled, in that either the ocean or the atmosphere can initiate the event, and the feedback between them will cause it to strengthen. These events are known as “warm ENSO events” due to the anomalous heating of the Pacific Ocean east of the International Date Line. They are also often simply called El Niño events, although strictly this term applies to only the oceanic component of the system. ENSO events typically initiate in September, are most strongly expressed from December to April, and then decay from September through to March of the following year (Namias 1976; Ropelewski and Halpert 1986; Yarnal and Diaz 1986; Hamilton 1988; Kiladis and Diaz 1989; Sardeshmukh 1990; Diaz and Kiladis 1992).

At the peak of warm events, the weakening of the tropical Pacific trade winds and the redistribution of heat along the equator disrupts the global climate system, redistributing energy and moisture (Trenberth et al. 1998). Although there is considerable variability in the effects of warm ENSO events, the average response during the boreal winter (December to February) includes anomalous dry conditions in the western Pacific, including Indonesia, southeast Asia, and Australia, and warm wet conditions in the central and eastern tropical Pacific (Ropelewski and Halpert 1986; Trenberth et al. 1998). In North America, winter conditions are typically warmer throughout southern Canada and the northern United States. Precipitation effects are more variable, but the Pacific Northwest, USA, is typically drier than normal, while Alaska and the southwestern United States are wetter than normal (Trenberth et al. 1998).

Cool ENSO (or La Niña) events are largely opposite to warm events. They are associated with enhanced trade winds, a larger temperature and height gradient between the western and eastern Pacific Ocean, and approximately the opposite teleconnections. For example, regions of the Earth that are droughty during warm events often are exceptionally wet during cool events (McCabe and Dettinger 2002).

The effects of ENSO events on wildfire vary regionally and sub-regionally, depending on the sign and magnitude of the individual event’s effect on climate (Fig. 4.1), and properties of the local vegetation. As the most important mode of global climatic variability, its effect on wildfire spans the globe, with significant effects on every continent except Antarctica (Nkemdirim and Budikova 1996; Kitzberger et al. 2001). Because the ENSO teleconnection to North America is strongest during the boreal winter, in many regions the strongest climatic impact is on total winter snow accumulation (Cayan 1996; Moore 1996). Winter snow accumulation can affect wildfire behavior through two contrasting mechanisms. First, in regions where snow persists into the summer, such as at high elevation, higher than normal snow accumulation will shorten the length of the fire season and help to maintain high moisture levels. These processes collectively decrease the likelihood of fires occurring (Westerling et al. 2006; Heyerdahl et al. 2008). Second, in

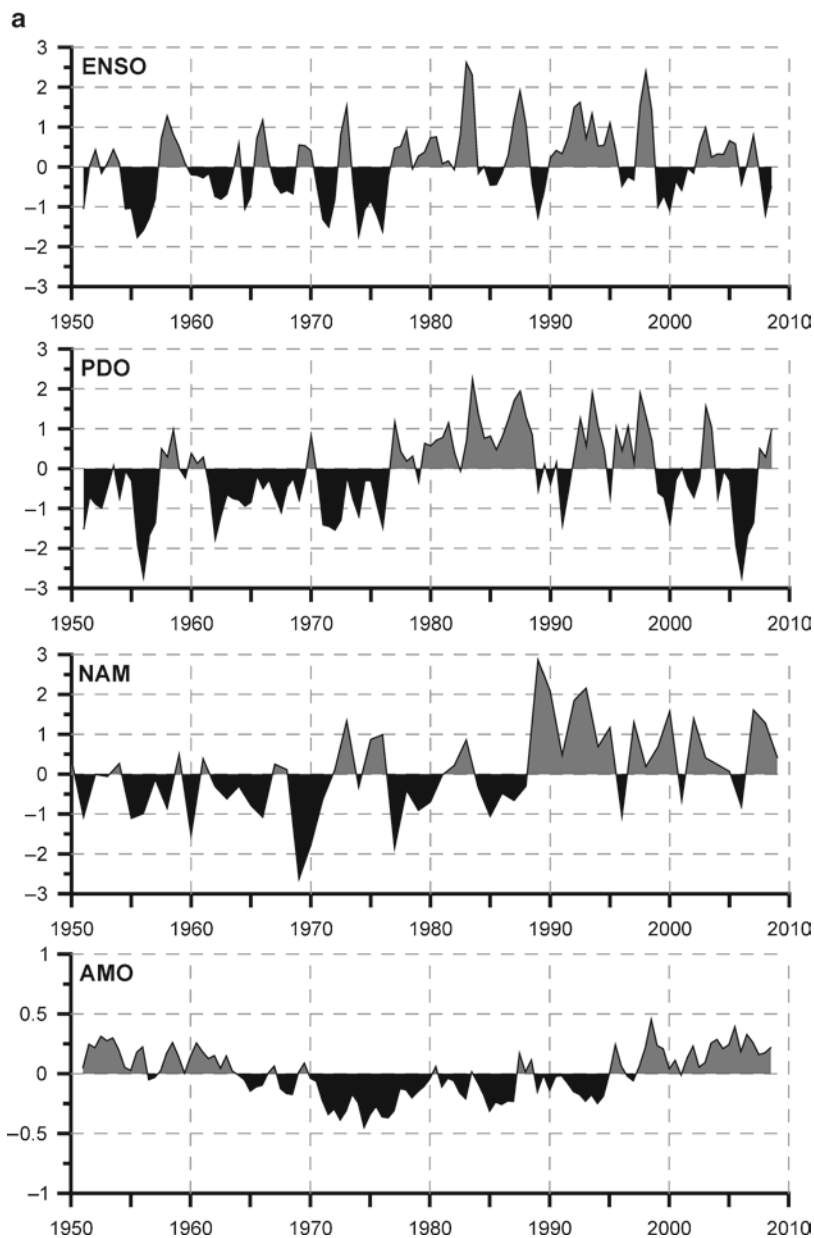
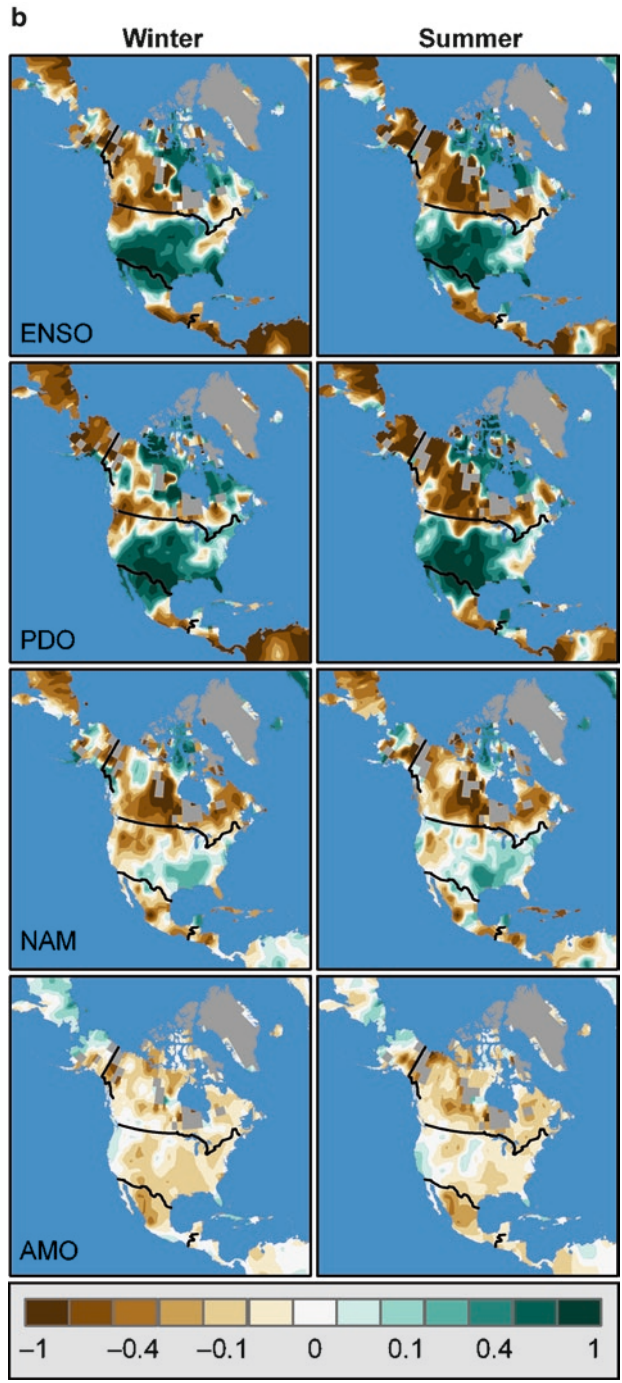


Fig. 4.1 (a) Time evolution of the four dominant modes of variability affecting the climate of North America: the El Niño Southern Oscillation (*ENSO*), the Pacific Decadal Oscillation (*PDO*), the Northern Hemisphere Annular Mode (*NAM*), and the Atlantic Multidecadal Oscillation. (b) Spatial regressions of each index onto the winter and summer Palmer Drought Severity Index (*PDSI*) records for North America. The maps show the typical effect of a one-standard deviation increase in the associated index on regional *PDSI*



particularly dry regions higher winter precipitation can promote the growth of grasses and shrubs, which increases the abundance and continuity of fine fuels (Westerling et al. 2003; Brown et al. 2005; Gedalof et al. 2005).

The best documented ENSO effects on wildfire occur in the U.S. Southwest, where warm (El Niño) events are associated with increased precipitation and reductions in area burned; cool (La Niña) events exhibit the opposite relationship (Swetnam and Betancourt 1990). Often the most widespread fire years occur when cool (dry) events follow warm (wet) events, and fuel production associated with the warm event is dried and burns during the subsequent cool event (Swetnam and Betancourt 1998). Similar effects occur in the U.S. Southeast, where cool events were found to be associated with decreased rainfall, increased lightning strikes, and consequently more and larger fires (Beckage et al. 2003; Dixon et al. 2008). Because the fire season in the Southeast occurs during winter months ENSO effects are particularly strong, explaining up to 50% of the variability in area burned (Brenner 1991). ENSO teleconnections to the Pacific Northwest are approximately opposite to those in the U.S. Southwest and Southeast, and warm events are associated with drier than normal conditions (Kiladis and Diaz 1989). Because the strongest impact is on winter conditions, though, the effect on fire frequency at most forest types in the Pacific Northwest is small. For example, Norman and Taylor (2003), Gedalof et al. (2005), and Heyerdahl et al. (2008) all found no significant associations between wildfire occurrence and ENSO (but see Hessl et al. 2004), although they did find that it interacted with other processes (see below) significantly. Surprisingly few analyses have been undertaken on the effect of ENSO variability on wildfire activity in the boreal forest. Macias Fauria and Johnson (2006, 2008) found an association between ENSO-like conditions and fire weather for the boreal forest of North America. They did not explicitly separate ENSO conditions from Pacific Decadal Oscillation conditions (see below), however, and they focused on fire weather rather than regionally synchronous fire events. They found that warm ENSO-like conditions were associated with *reduced* wildfire hazard west of the Canadian Rocky Mountains, and *increased* wildfire hazard over the western prairies.

4.3.2 *The Pacific Decadal Oscillation*

The Pacific Decadal Oscillation (PDO) is an ENSO-like mode of variability that is most strongly expressed in the North Pacific Ocean (Mantua et al. 1997). The precise mechanisms that force it are still unclear, but it does appear to be distinct from ENSO (Zhang et al. 1997; Barlow et al. 2001; Gedalof et al. 2002). The temporal variability in the PDO is characterized by intervals of anomalously warm or cool water in the central North Pacific Ocean that persist for 20–30 years, punctuated by abrupt shifts between phases (Mantua et al. 1997; Minobe 1999; Gedalof and Smith 2001). Superimposed on this low-frequency pattern there is considerable year-to-year variability. The effect on the climate of North America is strongest over the

coastal Pacific Northwest, and during winter (Fig. 4.1b). The warm phase of the PDO is associated with slightly elevated winter temperatures, and reduced precipitation, especially west of the Cascade Range. The PDO also affects other regions of North America, most notably the Canadian prairies, where the warm phase is associated with increased drought (Shabbar and Skinner 2004), and the U.S. Southwest, where it is associated with increased precipitation.

The effects of the PDO on wildfire activity are generally weak but significant throughout the Pacific Northwest. Gedalof et al. (2005) found no significant difference in the area burned by wildfire in the Pacific Northwest between the warm and cool phases of the PDO, although they did find a significant correlation between the PDO index and area burned. They also found that seven of the ten largest-fire years followed winters when the PDO index was positive, whereas eight of the ten smallest-fire years occurred following winters when the PDO index was negative. These results together suggest that the PDO exerts important controls on wildfire at annual to interannual scales, if not at interdecadal scales. Similarly, Hessl et al. (2004), examining fire activity in central and eastern Washington, found that the six most regionally coherent fires since 1650 all occurred when the PDO index was positive.

Few analyses of PDO-fire interactions for regions outside the Pacific Northwest have been undertaken to date, although the PDO has been examined for its modulating effect on ENSO. Brown (2006) found an inverse correspondence between wildfire activity in ponderosa pine forests in the Black Hills, South Dakota, and a tree-ring reconstruction of the PDO (cf. Biondi et al. 2001). Similarly, in Mississippi the cool phase of the PDO is associated with increased wildfire activity (Dixon et al. 2008). However there are two good reasons to believe that it may be an important determinant of fire activity in at least the arid Southwest and possibly other grass-dominated ecosystems. First, the influence of the PDO is primarily on winter precipitation, which has more of an effect on fuels production than on fuel moisture during the fire season. Second, several analyses of grass-fire interactions have noted a correspondence between wildfire and several wet years followed by one dry one. The persistent nature of the PDO should affect the frequency of these types of events.

The PDO has been found to interact with other modes of climatic variability to influence the fire regime. Gershunov and Barnett (1998) examined climatic conditions for various combinations of ENSO and PDO phases, and found that the two modes enhance each other when they are in the same phase (e.g., an El Niño event during the warm phase of the PDO), and they offset each other when they are in opposite phases (e.g., a La Niña event during the warm phase of the PDO). A number of fire-climate studies have found these interactions to be an important source of variability in regionally synchronous fire years. Corresponding warm phases are associated with increased wildfire activity in pine forests in northeastern California (Norman and Taylor 2003), in subalpine forests in Yellowstone and Jasper National Parks, in the central United States and southern Canadian Rocky Mountains, respectively (Schoennagel et al. 2005), and in Douglas-fir and ponderosa pine forests in the Pacific Northwest (Heyerdahl et al. 2008). Corresponding cool phases

are associated with increased wildfire activity in subalpine forests in Rocky Mountain National Park, in the southern U.S. Rocky Mountains (Schoennagel et al. 2005), in ponderosa pine forests in northern Colorado (Sherriff and Veblen 2008) and South Dakota (Brown 2006; Kitzberger et al. 2007). An analysis of fire weather in the Canadian boreal forest suggests that west of the Rocky Mountains, increased fire hazard is associated with the cool phases of ENSO and the PDO; the opposite relationship was found for regions east of the Rocky Mountains (Macias Fauria and Johnson 2006).

4.3.3 The Northern Hemisphere Annual Mode

The Atlantic Ocean exerts a smaller influence on the climate of North America than the Pacific Ocean does, due to the prevailing westerly circulation in mid latitudes. Nevertheless, several related modes of climatic variability may exert important controls that are relevant to patterns of wildfire. The Northern Hemisphere Annular Mode (NAM) is a pattern of variability in atmospheric pressure that is characterized by out-of-phase differences between the polar and subpolar sectors (Thompson and Wallace 1998, 2000). Although some disagreement exists as to which is the fundamental process (Ambaum et al. 2001), the NAM is very closely related to the North Atlantic Oscillation (NAO) and the Arctic Oscillation (AO), and they are assumed to be the same process for the discussion here.

In North America the positive phase of the NAM is associated with anomalous warm temperatures in the eastern half of the continent, south of the Great Lakes (Thompson and Wallace 2001; Visbeck et al. 2001) (Fig. 4.1b). Precipitation relationships are weak, but slightly drier than normal conditions occur through much of central North America. A wide range of winter climatic extremes have been found to be associated with the NAM, including short-term events (rather than mean conditions), and events in western North America (Thompson and Wallace 2001). The summer climate of North America has not been analyzed in this fashion, but it seems likely that similar relationships would exist for variables that could affect fire such as temperature, lightning, strong winds, and precipitation.

The role of the NAM in forcing wildfire has not been widely studied. However Macias Fauria and Johnson (2006; see also Le Goff et al. 2007) found that the positive phase of the NAM was associated with almost 70% of large fires in eastern Canada; the negative phase was associated with increases in fire in Alaska and the Northwest Territories. Dixon et al. (2008) found complex relationships between variability in the NAM and area burned in Mississippi: Significant negative correlations were found between total area burned in March and April (the height of the fire season) and the February state of the NAM. Curiously, area burned in October, November, and December was positively correlated to the NAM during September. They attribute these opposing relationships to differences in the seasonal expression of the NAM. In the late winter, the negative phase of the NAM is associated with

drier than normal conditions, which would reduce fuel moisture. In the summer, the positive phase is associated with increased convection, dry lightning, and strong winds, which would increase the frequency of ignitions and the rate of spread. Given that these processes have distinct effects that are modulated by vegetation structure and composition it is possible that they have distinct spatial expressions within this region.

4.3.4 The Atlantic Multidecadal Oscillation

The Atlantic Multidecadal Oscillation (AMO) is a slowly changing pattern of variability in North Atlantic surface temperatures. It was first identified by Schlesinger and Ramankutty (1994), and later named by Kerr (2000). It is characterized by alternate warming and cooling in the Atlantic Ocean north of the equator, with a period of approximately 65–80 years (Delworth and Mann 2000; Enfield et al. 2001). The warm phase of the AMO is associated with decreased rainfall throughout most of central North America, but increased rainfall in Florida and in some regions of the Pacific Northwest (Enfield et al. 2001).

Because of its slowly changing nature the AMO does not exert a strong control on year-to-year variability in wildfire (Fig. 4.1b). However it may influence patterns of vegetation structure and composition, fuel production and accumulation, and the frequency of ignitions, which in turn feed back to modulate the fire regime at longer timescales (Sibold and Veblen 2006; Schoennagel et al. 2007). The AMO has also been found to interact with other modes of variability to influence regional synchrony of wildfires. Throughout most of the U.S. Western Interior, years when the positive phase of the AMO corresponds to negative phases of ENSO and the PDO are associated with the most regionally synchronous wildfire (Kitzberger et al. 2007). This same pattern has been found in subalpine forests (Schoennagel et al. 2007) and ponderosa pine forests (Sherriff and Veblen 2008) in Colorado. South of central Colorado this association shifts, and wildfires are most commonly associated with combined negative phases of the AMO, ENSO, and the PDO (Kitzberger et al. 2007). In the Pacific Northwest fires are associated with the negative phase of the AMO combined with the positive phases of ENSO and the PDO (Kitzberger et al. 2007). The AMO has not been found to influence fire in the boreal forest (Macias Fauria and Johnson 2006; Le Goff et al. 2007). Relationships in the U.S. Southeast are inconsistent, but weakly negative (Guyette et al. 2006; Dixon et al. 2008).

4.4 Fire in the Future

The Earth's climate is changing in response to the actions of people (Solomon et al. 2007), and fire regimes will change in response to changing climates. The effects of climate change will differ regionally due to variability in the magnitude and

seasonality of climatic changes, as well as differences in how vegetation and fire respond to climate. One approach to predicting how fire will change in response to climate change is to model the current relationship between fire and climate, and then use future climate projections to assess change. Nearly all of these efforts predict a substantial increase in wildfire activity over the next century, including analyses of the Canadian boreal forest (Stocks et al. 1998; Gillett et al. 2004) and the western United States (McKenzie et al. 2004). Exceptions to this general pattern are rare, but include regions where fine-fuel abundance and continuity are more important than flammability, such as in the deserts of eastern California (Westerling and Bryant 2008).

A second approach to forecasting fire activity in the future involves using down-scaled climate projections to assess fire hazard based on operational guidelines or process based models (e.g. Torn and Fried 1992). These models have the advantage of reducing the complexities introduced by land use change, fire management practices, and vegetation change and focusing simply on the climate-associated fire hazard compared to today (Brown et al. 2004). They also have the advantage of forecasting fire hazard indexes in use by managers, providing a recognized “currency” for planning purposes. Flannigan and Van Wagner (1991) examined seasonal fire severity for the Canadian boreal forest under a range of climate projections and determined that annual area burned would increase by 46% with a doubling in carbon dioxide (CO₂) (see also Flannigan et al. 2000, 2001). Brown et al. (2004) evaluated fire hazard in the western United States, and determined that the number of days of severe fire weather will increase throughout their study region by up to 2 weeks per year by 2089. These effects are strongest in the northern Rocky Mountains, the Great Basin, and the Southwest. Although all of these models predict an overall increase in wildfire there are considerable regional differences in the magnitude and even the sign of the change, depending on the projection used (Flannigan et al. 2001).

Most analyses of the effects of climatic change on wildfire have focused on temperature and precipitation as the driving variables, but other approaches are possible. For example Miller and Schlegel (2006) modeled the occurrence of Santa Ana winds under a range of future climate scenarios, and concluded that Santa Ana occurrence “may significantly increase the extent of California coastal areas burned by wildfires, loss of life, and property.” Price and Rind (1994) modeled thunderstorm activity and concluded that lightning activity in the United States will increase by 26% and annual area burned would increase by 78%.

These efforts may provide insights into fire activity for the next several decades, but they assume that vegetation structure and composition are static. However, as vegetation responds to more frequent fire and changing climate, there may be rapid changes to ecosystem structure and composition (Bachelet et al. 2001b). Consequently these forecasts are probably unreliable beyond a few decades. There have been a number of efforts to model the interaction between climatic change, wildfire, and vegetation (e.g. Neilson and Drapek 1998; Bachelet et al. 2001a, 2003; Thonicke et al. 2001). To date, these efforts have produced variable results, depending on assumptions about the role of atmospheric carbon, nitrogen limitations, and

disturbance (Running 2008). Neilson and Drapek (1998) provide one perspective on why this exercise is both critically important and particularly challenging. In a comparison of short-term CO₂ effects on vegetation distribution vs. long-term combined temperature and CO₂ effects, they found that in many locations the trajectory of ecosystems reversed direction over time. For example, near-term expansion of grasslands into arid lands and westward expansion of eastern temperate mixed forest reverse in the long-term, and substantial area is ultimately lost. Fire and other disturbances are the most likely mechanism of vegetation dieback.

These predictions seem dire, but there is a growing body of evidence that fire is already increasing in severity or frequency. For example, Westerling et al. (2006) found that annual area burned by wildfire in the western United States has increased by a factor of more than 6.5 relative to 1970. They attributed this change to earlier snowmelt and longer fire seasons. Similarly, Kurz and Apps (1999; see also Kurz et al. 2008) concluded that fire and insect disturbance have caused the Canadian boreal forest to become a source of carbon since 1979, in contrast to the preceding 60 years. There is also a growing body of evidence to suggest that climatically induced forest dieback is underway at many locations in western North America: Breshears et al. (2005) found evidence for drought-induced dieback of two-needle pinyon pine (*Pinus edulis*) in the southwestern United States. van Mantgem et al. (2009) documented increased mortality among a wide range of species and age classes throughout the western United States and Canada that they attributed to climate-induced water deficits. Logan and Powell (2001) found that recent warming has allowed the mountain pine beetle (*Dendroctonus ponderosae*) to expand its range to higher elevations and thereby attack whitebark pine (*Pinus albicaulis*), leaving behind “ghost forests.” These changes alter forest habitat quality, but also the abundance and moisture of fuels, and the likely development of forest ecosystems over the coming decades.

4.5 Summary and Conclusions

Climatic processes act as top-down controls on regional patterns of fire ignition, rate of spread, fuel moisture, and fuel abundance and continuity. Lightning is the most important natural cause of fire ignition. Lightning frequency varies at continental, regional, and local scales, with areas of convergence and convection experiencing the highest frequency of lightning strikes. In mountainous regions the greatest frequency of lightning strikes often occurs at intermediate elevations. Lightning frequency alone is a poor predictor of the number of ignitions or total area burned, partly because lightning is often accompanied by precipitation, but also because not all vegetation types are equally flammable and ignition efficiency varies between land cover types. Following ignition, rates of spread are most rapid when strong winds, low humidity and high temperatures coincide. Regionally synchronous conditions conducive to rapid rates of spread are associated with several specific synoptic circulation types. In particular, persistent blocking ridges often

contribute to the development of strong pressure gradients and intense cyclonic activity that contributes to rapid fire spread. Another important set of circulation patterns is associated with air masses that cross mountains. These masses lose moisture as they are pushed up the windward side of mountains, and warm and dry rapidly as they descend the leeward side. The most severe of such patterns is probably the Santa Ana winds, which are associated with extreme wildfire hazard in southern California.

Patterns of ignition and spread depend on slower varying patterns in fuel moisture and fuel abundance. Fuel moisture is a function of climate over the days to years preceding ignition. In regions such as the coastal temperate rainforest where there are abundant coarse fuels, a seasonally wet climate, high soil water-storage capacity, and dense canopy cover, extended periods of antecedent drying are a necessary precondition to wildfires. At the other extreme, arid and semiarid ecosystems frequently have conditions conducive to ignition and spread, but require anomalously wet conditions over the preceding seasons in order to produce the continuous fine fuels required for fire spread. In between these extremes these processes interact, depending on such factors as long-term changes in mean climate, which determine dominant vegetation types; slope, aspect, and soil properties, which influence soil moisture and microclimate, contributing to variability in vegetation structure and composition; and recent disturbance history, which affects the abundance of fuel and the developmental stage of vegetation.

These processes operate at different scales, and interact to give rise to regionally synchronous wildfire years that differ depending on properties of the affected ecosystems. The oft-cited dichotomy of “fuel vs. climate” fails to incorporate the full range of possible relationships between top-down and bottom-up processes in regulating the fire regime. For example, ponderosa pine forests in the U.S. Southwest are both fuel- and ignition-limited, and respond to interannual variability in fuels production but to sub-seasonal variability in drought. Several important patterns of climatic variability influence the processes that control wildfire across a range of temporal and spatial scales. Globally, ENSO is the most important such pattern of variability, but over North America it interacts with the PDO, the NAM, and the AMO to produce regionally synchronous variability in patterns of wildfire. In the coming decades, fire is likely to be an important agent of ecosystem change, as climatic change and exotic species increase the frequency and magnitude of wildfire nearly everywhere in North America.

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