The Relationship between the Polar Jet Stream and Extreme Wildfire Events in North America

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ABSTRACT: Northern Hemisphere midlatitude weather is strongly influenced by the polar jet stream (PJS), which dictates the position of storm tracks; this influence also extends to weather patterns conducive to the ignition and growth of large wildfires. We examined the role of the PJS on extreme wildfire events in North America (NA) between 40° and 70°N latitude, using fire spread events (FSEs) for 2002–16. Climatologies of the 300-hPa wind components and derived quantities show that the PJS weakens and moves northward in the boreal summer coincident with the fire season. We use spatiotemporal compositing of 300-hPa winds and 500-hPa geopotential height anomalies to show that FSEs are associated with an upper-level ridge and high centered over events, except eastern Canada where patterns are displaced westward. Ridge patterns persist longer for FSEs in western NA compared with eastern NA, as well as for May–August compared with April, September, and October. These tropospheric patterns also occur concomitantly with surface weather drivers of fire spread including positive daily mean temperature and vapor pressure deficit anomalies and negative precipitation anomalies. Distributions of maximum and minimum latitudinal jet stream peak, and ridge and trough positions, relative to FSEs, confirm that events occur predominantly southward of the jet stream core and near a ridge for low split flow configurations but not necessarily for high split flow configurations. These findings have wide-reaching implications for better understanding NA fire regimes and potential fire management strategies (e.g., resource prepositioning and tactical suppression) through improved forecasting of fire weather.

KEYWORDS: Atmospheric circulation; Synoptic climatology; Wildfires; Boreal meteorology

1. Introduction

Synoptic-scale upper-air atmospheric circulation patterns are a major driver of surface weather variability, including the occurrence of extreme weather events. In general, the environmental and societal impacts of extreme events are even greater when weather patterns persist for several days or even weeks. In particular, persistent tropospheric ridging has been found responsible for temperature extremes that have led to drought events (Teng and Branstator 2017) or heatwaves (Horton et al. 2016). Likewise, persistent upper-air trough patterns have led to major flooding events (Blackburn et al. 2008).

In the Northern Hemisphere midlatitudes, the polar front jet stream [or polar jet stream (PJS)] is a key feature of upper-level flow that determines the position and movement of weather systems. The PJS consists of fast-moving westerly flow with maximum speeds in the upper troposphere (between 100 and 400 hPa and between 30° and 60°N) with typical wind speeds greater than 30 m s^{−1}. It is itself driven by baroclinic eddies and the corresponding restoring force, assuming quasigeostrophic flow. Long wavelength meridional oscillations in the PJS are equivalent to *planetary-scale waves*, also known as

Rossby waves. This meridional flow transports energy (heat and moisture) between the tropics and Arctic or vice versa, leading to the existence of ridges or troughs (i.e., cyclogenesis) or blocking features that can dominate surface weather conditions and lead to surface extremes such as heat waves, droughts, or extreme fire weather. The dynamical mechanisms leading to these jet stream configurations and extreme events are still not well understood and are an active area of research. Some research suggests that drivers of jet stream variability may include climate oscillations (Hall et al. 2015, 2017), although Teng and Branstator (2017) found that internal dynamics of the midlatitude atmosphere played a larger role than ENSO variability for persistent ridging associated with drought events in California. Nakamura and Huang (2018) considered a mathematical analogy to traffic congestion to relate the onset of blocking with jet stream dynamics. Quinting and Vitart (2019) further explored the relationship of Rossby wave packets (RWPs) and blocking and found the decay of RWPs is associated with the onset of blocking in the European-Atlantic sector.

a. Synoptic weather and wildfires

At the landscape scale, wildland fire activity is determined by a combination of factors including ignitions, fuel availability, weather/climate, and anthropogenic factors. Notably, climate has been found to be the dominant control of wildfire activity for very large fires (Barbero et al. 2014; Viedma et al. 2015) as well as the main driver of area burned at large scales (i.e., continental and global) in landscapes that are not fuel-limited (Aldersley et al. 2011; Liu et al. 2013; Mansuy et al. 2019). In general, fire ignition and spread depend on high fuel aridity (i.e., above average surface temperatures and/or below average precipitation) (Littell et al. 2009), and in the case of

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lightning-caused fires, atmospheric conditions conducive to the development of convective systems responsible for lightning production (Rorig and Ferguson 1999).

For fire management purposes fire weather indices have been developed to describe surface or lower-atmospheric conditions conducive to fire occurrence and spread (Van Wagner 1987; Haines 1989). However, surface weather conditions are highly variable, making them difficult to forecast at the spatial resolutions and lead times required by fire management agencies. Conversely, synoptic-scale weather is more predictable than mesoscale weather in current numerical weather prediction models (Yano et al. 2018). In one study, numerical errors for 10-day synoptic-scale forecasts were found to be equivalent to those of a 1-day convective-scale forecast (Hohenegger and Schär 2007).

There has therefore been an impetus to relate wildfire activity and surface fire weather with upper air and synoptic-scale features (Flannigan and Wotton 2001), as evident by the large number of studies linking the two. Of particular note, studies have shown a robust correlation exists between area burned and positive 500-hPa geopotential height anomalies (Johnson and Wowchuk 1993; Nash and Johnson 1996; Gedalof et al. 2005; Macias Fauria and Johnson 2006; Hostetler et al. 2018; Skinner et al. 1999, 2002). In addition, a feature of midtropospheric atmospheric flow commonly associated with extreme fire weather is atmospheric blocking, as evident in persistence of 500-hPa geopotential height anomalies (Johnson and Wowchuk 1993; Skinner et al. 1999; Gedalof et al. 2005; Hoinka et al. 2009). Newark (1975) showed that a persistent longwave ridge in 500-hPa geopotential heights played a role in the 1974 Ontario fire season. Skinner et al. (2002) also showed that ridging in 500-hPa heights and the associated strong meridional flow were linked with large fires in Canada.

Several studies also found wildfire events were often associated with advection of warm dry air (Hoinka et al. 2009; Trigo et al. 2016) or convective atmospheric instability (García-Ortega et al. 2011). More recently, Lagerquist et al. (2017) used sea level pressure and 500-hPa geopotential heights to predict extreme fire weather events up to 8 days in advance. In contrast to the above studies that served to highlight common synoptic atmospheric patterns associated with wildfire activity, several studies also found important regional differences (Crimmins 2006; Kassomenos 2010; Wastl et al. 2013; Pollina et al. 2013; Papadopoulos et al. 2014; Labosier et al. 2015). For example, Abatzoglou et al. (2013) considered a synoptic-scale analysis of the Santa Ana winds, which are responsible for extreme fire weather conditions in Southern California. They found that the mean sea level pressure combined with lower tropospheric cold air advection was able to characterize these wind events. By examining sea level pressure and 925-hPa winds Ruffault et al. (2017) found most large fires occurred under the Atlantic Ridge and were associated with increased winds, with few events linked with blocking episodes. Nauslar et al. (2019) examined the effect of displacements of the subtropical ridge and the onset of the North American monsoon on the fire season in the southwest United States.

b. The jet stream and wildfires

While the majority of research investigating the connection between synoptic weather patterns and wildfires has

focused on lower to midtropospheric flow, there is also interest in the influence of upper troposphere atmospheric flow. In particular, a few papers have directly considered the role of jet streams and planetary wave dynamics in wildfire activity. Early work by Schroeder et al. (1964) found synoptic weather patterns associated with extreme fire weather conditions in the contiguous United States, which included long wave patterns in the meridional flow of the jet stream as well as blocking patterns. Schaefer (1957) also considered the role of the jet stream, atmospheric instability, and surface winds on wildfire occurrence, citing several case studies in the United States. Recently, Reeder et al. (2015) found that wildfires in southeastern Australia were associated with development of an anticyclonic anomaly in the potential vorticity (at 350 K) followed by the passage of a cold front, as attributed to Rossby wave breaking (RWB). Hayasaka et al. (2016) also related Rossby wave breaking and increased meridional flow of the PJS to active fire periods in Alaska. Two recent studies used paleoclimate proxies to reconstruct jet stream dynamics and to show how interannual poleward shifts in storm track positions are related to annual area burned (Dannenberg and Wise 2017; Wahl et al. 2019). More recently, Petoukhov et al. (2018) linked the 2016 Horse River fire (in Fort McMurray, Alberta) with a persistent high-amplitude planetary wave oscillation of wavenumber 4.

Recent interest in the connection between planetary wave dynamics and extreme weather events has been further motivated by several developments. One major avenue of investigation has been the connection of amplified circumglobal planetary wave patterns with extreme events (Blackburn et al. 2008; Screen and Simmonds 2014). In particular, quasi-resonant amplification (QRA) has been proposed as a mechanism for persistent planetary wave patterns that lead to extreme surface weather events (Petoukhov et al. 2013). Here quasi-stationary planetary waves become trapped in a mid-latitude waveguide, which leads to strong amplification. QRA events with wavenumbers between 4 and 8 have been linked to extreme weather in several studies (Coumou et al. 2014; Petoukhov et al. 2016; Stadtherr et al. 2016; Kornhuber et al. 2017; Petoukhov et al. 2018; Kornhuber et al. 2019, 2020). Screen and Simmonds (2013) caution that more work is needed to test the QRA hypothesis. Apart from these studies of global-scale planetary wave dynamics, there are several papers that advocate for a regional analysis of planetary waves (Röthlisberger et al. 2016; Fragkoulidis et al. 2018; Ghinassi et al. 2018). For example, Fragkoulidis et al. (2018) suggest that Northern Hemisphere temperature extremes are more strongly correlated with the local amplitude of Rossby wave packet envelopes than with global wavenumber amplitudes. In general, a greater understanding of the link between planetary wave dynamics and atmospheric blocking is needed although some progress has been made (Madonna et al. 2017). In addition, Röthlisberger et al. (2019) identified recurrent Rossby wave packets (RRWPs) as an alternative mechanism to atmospheric blocking for persistent cold and hot spells.

Understanding the variability and dynamics of the polar jet stream is crucial to prediction of weather extremes in the midlatitudes, including the occurrence of extreme fire weather

and wildfires. In this paper, we investigate the relationship between the PJS and large wildfire events in North America, where we use the concept of *fire spread events* as an objective measure of extreme wildfire events. To quantify patterns in the jet stream we consider Northern Hemisphere 300-hPa wind speed components and 500-hPa geopotential height anomalies. We use a composite analysis to examine the average patterns of these fields relative to fire spread events and further stratify the results by month of year and regions to gain insight into the variability of the PJS and wildfire activity. Using time-lagged composites we are able to investigate dynamical aspects of the composite patterns such as buildup, persistence, and decay. To link the upper air patterns with surface fire weather, the composite analysis is also repeated for daily surface weather anomalies including temperature, vapor pressure deficit, and precipitation. We further make use of several metrics including the meridional circulation index (MCI), split flow index (SFI), maximum and minimum peak latitudes of the jet stream core, and longitudinal trough or ridge positions, relative to fire spread events.

2. Data and methods

a. Fire spread events

In this study we use the concept of *fire spread events* as an objective measure of moderate to extreme wildfire activity. Fire spread events are defined by days where the fire front advances 2 m min^{-1} for at least 4 h (i.e., at least 480 m in total) on a single day for a given fire (Parisien et al. 2013; Wang et al. 2014). Because, in general, daily fire perimeters are not available, fire spread events must be estimated from satellite active fire detections (or *hotspots*), and here we describe the algorithm. For each fire in our study we subset all MODIS hotspots for the corresponding year of the fire within a 1-km buffer of the final fire perimeter. We then divide the entire fire area into multiple polygons using Voronoi tessellation, where each polygon contains a single hotspot (Aurenhammer 1991). Assigning the time of burn of each Voronoi polygon to the corresponding time of detection for the corresponding hotspot then gives a time series of cumulative area burned. We then fit a cubic spline to the cumulative area burned using the `splinefun` function in the R stats package (using the “`hyman`” option to fit a monotonic curve) (R Core Team 2020). Evaluating the spline function for consecutive days gives the daily area burned, which can be used to determine fire spread events, with the additional assumption of circular fire growth (see Fig. S1 in the online supplemental material). The threshold of 480 m day^{-1} corresponds to linear spread and is thus less sensitive to daily area burned, particularly for fires that have already grown large. Although this method is necessarily approximate, we are only interested in relating large-scale atmospheric patterns with multiple fire spread events, and not the growth of individual fires.

We used hotspot data from the NASA MODIS C6 data product for 2002–16 (Giglio et al. 2016); although MODIS provides hotspot data from 2001 onward, we omitted the first year as one of the MODIS satellites (*Aqua*) was not yet online and this affected coverage. Fire perimeters were downloaded from the Canadian National Fire Database (CNFDB; <https://cwfis.cfs.nrcan.gc.ca/ha/nfdb>) for Canada; from the Monitoring

Trends in Burn Severity (MTBS) database (Picotte et al. 2020) for the contiguous United States (CONUS); and from the Alaskan Interagency Coordination Center (AICC; <https://fire.ak.blm.gov/predsvcs/maps.php>) for Alaska. We used active fire detections from the Moderate Resolution Imaging Spectroradiometer sensors (MODIS) downloaded from the NASA Fire Information for Resource Management System (FIRMS) (MCD14DL collection; <https://earthdata.nasa.gov/firms>).

We processed all large fires in North America between 40° and 70°N latitude and from 2002 to 2016 and for the months April–October, which included almost all fire spread events for the study region. Large fires were defined by a size threshold of 5000 ha, and these fires accounted for 85% of the total area burned for our study area. The choice of size threshold was further determined by two requirements. First, fires of at least this size correspond to fires that have evaded initial suppression attempts (i.e., escaped fires) and are generally not fuel-limited, or fires for which no suppression was attempted. This makes it likely that the main control of fire growth is weather. Second, the daily growth of large fires is better represented by MODIS hotspots since the relative number of false positives (false alarms) and false negatives (missed events) are low. Even so, with this threshold it was necessary to discard approximately 2% of fires because there were 5 or fewer hotspots detected over the entire duration of the fire, which was insufficient to determine daily fire growth. For the purposes of our analysis, the location of each fire spread event is defined by the geographic centroid of the corresponding fire perimeter. Figure 1 shows the spatial and monthly distributions of fire spread events for our study region and period. The majority of fire spread events occur in the months of June, July, and August.

b. Synoptic weather fields

The polar jet stream maximum can be located in the upper part of the troposphere with elevation variations between the pressure levels of 400 and 100 hPa. Because of this spatial variation, the jet stream core position can in principle be defined using a mass weighted mean wind speed (Archer and Caldeira 2008) although it is more straightforward to use a single level (Rikus 2018). Here we consider jet stream configurations and metrics derived from the 300-hPa wind components, consistent with much of the literature examining the midlatitude jet stream (Teng et al. 2013; Coumou et al. 2017; Belmecheri et al. 2017).

The 300-hPa wind components and 500-hPa geopotential heights were obtained from the NCEP–DOE II Reanalysis (Kanamitsu et al. 2002), defined on a $2.5^\circ \times 2.5^\circ$ grid. The resolution of the reanalysis was sufficient to represent synoptic-scale patterns in the mid- to upper-troposphere—the full field consists of 144 points in the longitudinal direction, which is equivalent to applying a low-pass band filter with maximum sampling wavenumber of 72. The 6-hourly fields were further aggregated to daily fields where each day was offset to start at 1200 UTC to ensure a daily period, which covered the active burning window across all times in North America. To further eliminate transient signals, we sometimes applied a moving average to the fields, and this is indicated in the text where applicable. Climatologies were calculated using daily data

FIG. 1. Map and histogram (inset) of fire spread events for 2002–16, colored by month of occurrence. Black borders indicate the modified terrestrial ecozones from Skinner et al. (2002).

from 1979 to 2018 where in each leap year the daily data corresponding to February 29 were first removed.

c. Surface weather fields

Surface weather fields were also obtained from the NCEP–DOE II Reanalysis on a spectral Gaussian grid and were re-sampled to the same $2.5^\circ \times 2.5^\circ$ grid of the synoptic fields using bilinear interpolation. The surface weather variables of interest for fire spread were the 2-m temperature, 2-m vapor pressure deficit (VPD), precipitation, and 10-m wind speed. VPD was not directly available from the reanalysis so was calculated using temperature, specific humidity, and air pressure (Stull 2018). The 6-hourly fields were aggregated to mean daily values for temperature and VPD, and to daily sum for precipitation. Anomalies were calculated by subtracting the climatological mean of the corresponding field computed at each grid cell and day of year for the period 1979–2018. For wind speed, the maximum daily value was used instead of the daily mean since large wind speeds can drive rapid fire growth.

d. Jet stream metrics

To characterize different jet stream configurations and properties so that we can relate them to fire spread events, we considered various metrics that were calculated from the daily 300-hPa fields. Here we describe the different metrics used in this study:

1) MERIDIONAL CIRCULATION INDEX

The meridional circulation index (MCI) is one of several circulation indices that quantifies the north–south (meridional) meandering of the jet stream. Others include the sinuosity

index (Cattiaux et al. 2016) and the meandering index (Di Capua and Coumou 2016). The MCI is defined as $MCI = V|V|/(U^2 + V^2)$, where U and V are the zonal and meridional 300-hPa wind components, respectively (Francis and Vavrus 2015). $MCI = 0$ represents purely zonal flow whereas $MCI = 1$ (–1) represents purely northward (southward) flow. The MCI can be averaged over any region of interest. Here we consider the meridional mean over the latitudinal band 35° – 75° N, which encompasses the calculated fire spread events.

2) JET STREAM PEAK POSITIONS

Determination of the jet stream core position relative to Earth's surface (i.e., longitude and latitude) is challenging because of wind speed variability and the fact that there may be two distinct branches of the jet stream corresponding to the PJS and the subtropical jet. These two branches tend to be separated in latitudinal position during winter but can become closer and can even merge during the summer months (Molnos et al. 2017) (also see climatology results in the online supplemental material). Some authors use the zonally averaged zonal wind to determine the latitudinal position of the jet core (Archer and Caldeira 2008; Rikus 2018) or only identify the strongest band (Belmecheri et al. 2017). Determination of the position of distinct jet stream branches requires more sophisticated methods (Molnos et al. 2017). Here, we apply a simple peak finding algorithm at each longitude that can determine the presence of multiple peaks with a wind speed greater than a given threshold. Our method uses the “find_peaks” function of the scipy Python library (Virtanen et al. 2020) and further requires that each peak correspond to westerly flow and be separated by at least 15° in latitude. Unless otherwise specified the peak positions were restricted to

FIG. 2. Synoptic weather patterns during an active fire spread day with (a),(b) high split flow (24 Jun 2015) and (c),(d) low split flow (22 Jul 2007). Purple dots correspond to locations of fire spread events on given day; small dots correspond to location of jet stream peaks (for wind speeds $> 20 \text{ m s}^{-1}$).

the latitudinal band 30° – 80°N . Typically a velocity threshold of 30 m s^{-1} is chosen to detect the Northern Hemisphere polar jet stream maximum (Koch et al. 2006; Pena-Ortiz et al. 2013). In other studies (Melamed-Turkish et al. 2018) a seasonal adjustment is made to the wind speed threshold. Here we choose a lower threshold of 20 m s^{-1} to compensate for the weaker zonal component during summer but also because we apply a moving window mean (which can reduce the peak jet stream velocities).

3) SPLIT FLOW INDEX

Split flow or double jet configurations may have implications for persistent midlatitude patterns. Rex (1950) identified a specific midtropospheric split flow configuration where a high is situated to the north of a low, which is known as a Rex block. More recently, double jet configurations have been given as a requirement for quasi-resonant amplification of planetary waves (Petoukhov et al. 2013; Coumou et al. 2014). In general, meteorologists also refer to split flow blocks where a high is situated between two westerly jet branches separated in latitude (Colucci et al. 1981). However, there are relatively few attempts to quantify split flow jet configurations with two studies that define split flow indices for the Southern Hemisphere (Bals-Elsholz et al. 2001; Babian et al. 2018). Here we define a split flow index (SFI) using the peak detection calculation described above. If two or more peaks are identified at longitude ϕ then $\text{SFI}(\phi) = 1$; otherwise $\text{SFI}(\phi) = 0$. We apply the calculation to only consider peaks between 30° and 80°N unless otherwise stated. Figure 2 demonstrates two active fire periods corresponding to either high or low split flow configurations.

4) RIDGES AND TROUGHS OF MERIDIONAL FLOW

Because in general the jet stream may have multiple branches or may fragment, it is not straightforward to determine the

latitudinal positions of ridges and troughs. Here we consider a definition that is straightforward to calculate. First, the wind speed component at 300 hPa (V_{300}) is averaged between the latitudinal band 35° and 75°N and the first derivative with respect to longitude is calculated using simple finite differencing. The minima and maxima of the resulting field (denoted V'_{300}) correspond to ridges and troughs respectively (see Fig. 3 for an example). Note that this definition applies to average meridional flow and therefore only corresponds with the usual meteorological definition of ridges and troughs when there is a single continuous jet stream branch.

5) SPATIOTEMPORAL COMPOSITES

Composite analysis is a widely used tool in environmental science to help understand relationships between events and covariates of interest. Compositing can be performed using time series data (i.e., superposed epoch analysis; Chree 1913), or, as is the case in climate science, using spatial fields [for an early application see Sanders and Gyakum (1980)]. Composite analysis has also been used in the context of the jet stream and planetary waves. For example, Woollings et al. (2010) found three preferred positions for the PJS; Hall et al. (2017) examined jet stream variability with respect to a number of potential predictors; Wolf et al. (2018) examined 300-hPa wind patterns associated with European temperature extremes.

Here we consider spatiotemporal composites where time-varying climate fields $F(x, t)$ are averaged relative to the location x and time t of N events $E(x_i, t_i)$, $i \leq N$. For each event, the relevant climate field is subset to a box $100^{\circ} \times 60^{\circ}$ (longitude $\times$ latitude) centered at the event location and the N such instances are averaged to give the final composite field. There are advantages to compositing over other correlational analyses of atmospheric variables and events (such as weather

FIG. 3. Example showing calculation of jet stream parameters for a single date. (top) 300-hPa wind speed. Small dots indicate peak detections (for wind speeds $\geq 20 \text{ m s}^{-1}$). The red circle represents a single fire spread event location. The yellow circles represent the maximum and minimum jet stream peaks at the longitude corresponding to the fire spread event. (bottom) The black curve shows meridional mean of V_{300} over $35^\circ\text{--}75^\circ$. Ridge/trough positions (red/blue vertical lines) are given by the minima/maxima of the first derivative (green dashed curve). The ridge and trough closest to the fire spread event are shown in the upper panel by the red and blue lines, respectively.

typing or regression analyses). First, compositing is a non-parametric method that only depends on observed patterns. Second, using lead-lag composites one can track the formation and decay of patterns associated with an event, which provides valuable dynamical information. Here we construct lead-lag composites by compositing fields from 20 days before to 20 days after each event (i.e., 41 daily composites in total). Last, one can stratify events by categorical variables and therefore examine the patterns associated with different categories of an event of interest. In this paper, fire spread events are further stratified by seasonality (month of year) or location (zones) and composites are formed for each stratification (see Table 1 and Fig. 7). We can further investigate the temporal dynamics of spatiotemporal composites by using a similarity metric. The structural similarity index measure (SSIM) is an objective method for comparing two digital images using their covariance properties (Wang et al. 2004). Here SSIM values are calculated between each composite map for $t \neq 0$ and the map at $t = 0$ to track the development and decay of the composite

TABLE 1. Number of fire spread events used for each composite by month of year for results shown in Fig. 6.

Month No.	Month	No. of events
4	April	35
5	May	464
6	June	2166
7	July	2460
8	August	1307
9	September	180
10	October	19

patterns relative to the time of fire spread events. To calculate the SSIM we used the scikit-image Python library (van der Walt et al. 2014).

3. Results

a. Climatologies

We characterize the climatology of the PJS using the 300-hPa wind components and derived quantities as described in the methods (section 2). We compare these metrics to the climatology of fire spread events found by averaging the number of fire spread events for each grid cell of the NCEP reanalysis for 2002–16. Figure 4 shows Hovmöller diagrams for the meridional and zonal 300-hPa wind speeds, the meridional circulation index, the minimum and maximum jet stream peak positions, and the split flow index. The PJS moves north in the boreal summer and weakens with respect to both zonal and meridional wind components (see also Fig. S2). Interestingly, the seasonal variation in the 300-hPa meridional wind speed spectrum shows that the dominant wavenumber changes from $k = 3$ during the boreal winter to $k = 5\text{--}7$ in the boreal summer (see Fig. S3).

An increase in the climatological density of fire spread events coincides with minima in both the zonal and meridional wind components. The density of fire spread events is also associated with positive (i.e., northward) displacements in the latitudinal position of both the minimum and maximum peak positions. The climatology of the absolute value of meridional circulation index ($|MCI|$) also shows that in summer there is a preferred position for a synoptic-scale ridge approximately

FIG. 4. Hovmöller diagrams showing climatology (1979–2018) of various jet stream related quantities: (a) zonally averaged zonal 300-hPa wind, (b) meridionally averaged meridional 300-hPa wind (35° – 75°), (c) meridional mean of MCI (35° – 75°), (d) minimum meridional peak position (0° – 90°), (e) maximum meridional peak position (0° – 90°), and (f) meridional mean of SFI (30° – 80°). The superimposed heat map shows the density of fire spread events for the period 2002–16. See Fig. S2 for circumpolar full-year climatologies of the same quantities.

situated between -120° and -90° longitude and approximately between days 150 and 250, corresponding to a region of positive meridional flow to the west of a region of negative meridional flow. Split flow is also smaller on average for this region and time period as shown by the split flow index. This region accounts for the majority of fire spread events in the

study area, with the exception of the events that occur to the west in Alaska.

b. Composite analysis

The climatological analysis only relates mean quantities for the jet stream and fire spread events. To further investigate the

FIG. 5. (left) Composite of 300-hPa meridional wind speed (V_{300}) for all fire spread events where V_{300} is composited relative to longitude and latitude for each fire spread event, and with individual composites ranging from 20 days before each fire spread event (lag = -20) to 20 days after each fire spread event (lag = +20). (right) As in the left panels, but for 500-hPa geopotential height anomalies (ΔZ_{500}).

relationship between the jet stream and fire spread events we also consider intra-annual variations in these quantities to better determine their relationship with respect to synoptic patterns that manifest on the scale of days or even weeks. Figure 5 shows composite analysis of the mean 300-hPa meridional wind speed (V_{300}) and 500-hPa geopotential height anomalies (ΔZ_{500}) corresponding to fire spread events, relative to the location of each event and composited from 20 days before to 20 days after each event. The composites show the emergence of a clear pattern on the day of the fire spread event, corresponding to positive meridional (northward) flow to the west of the event and negative meridional (southward) flow to the east of the event. Such meridional flow is equivalent to the formation of an upper-tropospheric ridge. The pattern is also concomitant with positive values of ΔZ_{500} centered at the event location. Two asymmetries are apparent, one spatial and one

temporal. The spatial asymmetry is manifest as the meridional flow is tilted at an angle from the meridian (i.e., relative longitude = 0°). A temporal asymmetry is evident in the time series plots of the structural similarity at different day lags relative to the day of each fire spread event (Fig. 8); the timing of the mean pattern is such that the onset of the ridge occurs at least 20 days before $t = 0$ but decays faster after $t = 0$.

We also stratified the composites by month of year and spatial zone as shown in Figs. 6 and 7 respectively, for $t = 0$ days relative to each event. The monthly composites show stronger patterns in the “shoulder” fire season (i.e., April, May, September, and October) compared with the main fire season (June, July, and August). The composites by zone also show clear regional differences. Specifically, the ridge pattern is centered around event for the central region (zone 2) but is displaced to the east for the western zone (zone 1) and to the west for the eastern region (zone

FIG. 6. Composites of synoptic and surface weather patterns by month for day of fire spread event, composited relative to longitude and latitude of each event. From left to right: composites for 300-hPa meridional wind speeds (V_{300}), 500-hPa geopotential height anomalies (ΔZ_{500}), 2-m temperature anomalies (ΔT), 2-m vapor pressure deficit anomalies (ΔVPD), and daily precipitation anomalies (ΔP).

3). This is consistent with the preferred position of the synoptic-scale ridge as shown in the climatologies (Fig. 4c).

The time series of the structural similarity index measure (SSIM) for both V_{300} and ΔZ_{500} , shown in Fig. 8, are also stratified by month of year and zone. Two interesting observations are apparent: first, for the monthly composites, patterns persist the longest for months May–August whereas patterns are less persistent (i.e., more transient) for April, September, and October; second, for the zonal composites,

the patterns persist longer for zones 1 (west) and 2 (central) relative to zone 3 (eastern), which exhibits a more transient pattern.

To link the upper-level synoptic patterns with surface weather conditions for fire spread events, we also repeated the composite analysis for surface weather anomalies of 2-m temperature (daily mean), VPD (daily mean), and precipitation (daily sum). Figure 6, which shows the composites stratified by month, demonstrates that fire spread events in all months are

FIG. 7. (top) Three spatial zones used in this study, shown with North American ecozones. (bottom) Composites of synoptic and surface weather patterns by spatial zone for day of fire spread event, composited relative to longitude and latitude of each event. From left to right: composites for 300-hPa meridional wind speeds (V_{300}), 500-hPa geopotential height anomalies (ΔZ_{500}), 2-m temperature anomalies (ΔT), 2-m vapor pressure deficit anomalies (ΔVPD), and daily precipitation anomalies (ΔP). Zone 1 (west) is composited from 1909 fire spread events, zone 2 (central) from 3988 events, and zone 3 (east) from 383 events.

associated with positive temperature and VPD anomalies and negative precipitation anomalies. Similarly to the synoptic field composites, these patterns vary by month with the strongest anomalies occurring in the shoulder season months. Composites stratified by zone (Fig. 7) also show patterns in temperature and VPD anomalies consistent with ΔZ_{500} composites. Interestingly, positive VPD anomalies are smallest for the zone 2 (central) and also are weakest for

early season fire events, increasing as the fire season progresses. Composites of these variables for all fire spread events are shown in Figs. S4 and S5. The time series of the surface variables for zero relative longitude and latitude are shown in Fig. 9. Here SSIM time series were not used because in general these surface weather variables have much greater variability than synoptic fields and one is interested in the surface weather conditions at the specific location of each fire event. We also

FIG. 8. Time series of structural similarity index (SSIM) measuring similarity of composite at lag $t = 0$ (days) as a function of lag t (days) relative to date of fire spread event for 300-hPa meridional wind speed (V_{300}) composites stratified by (a) month and (b) zone, and 500-hPa geopotential height anomaly (ΔZ_{500}) composites stratified by (c) month and (d) zone.

examined monthly and zonal composites for maximum daily 10-m wind speed (not shown), but these did not exhibit a consistent pattern, although the composite for all fire spread events (Fig. S5) did indicate a modest reduction in wind speed during each fire spread event. Due to the difficulty of resolving wind speeds at the scale necessary, the influence of surface winds on fire growth is outside the scope of the results presented here.

c. Meridional and split flow metrics

To further characterize the relationship between the positions of jet stream peaks and fire spread events we show in Fig. 10 the distributions of relative latitudinal positions of the maximum and minimum peak latitudes and the relative longitudinal positions of the closest ridge and trough. The distributions are shown for both low and high split flow configurations, corresponding to $SFI \leq 0.625$ and $SFI > 0.625$, respectively. In the low split flow case, the relative latitude of the jet stream peaks generally appears northward of fire spread event. In contrast, for the high split flow case the minimum latitude peak generally appears southward of the maximum peak latitude and northward of the minimum peak

latitude. The relative longitudinal position of the closest trough has a bimodal distribution for low split flow, with the trough located either to the west or east of the fire spread event. However, in the case of high split flow the longitudinal trough position can also occur near the fire spread event. That is, the extreme fire weather associated with split flow configurations is not necessarily due to a ridge like pattern centered on the active fire.

We also examine the regional means of the MCI and SFI for North America. Figures 11a and 11b also respectively show the $|MCI|$ and SFI anomalies averaged between -170° and -50° longitude, approximately representing the extent of the study area. Both metrics show that for extreme fire days—defined by days with more than 5 fire spread events—that both $|MCI|$ and SFI are statistically different from the corresponding values for nonextreme fire days (5 or fewer events). It should be noted that although the differences are statistically significant for July and August, the effect is small, particularly for $|MCI|$. Finally Fig. 11c shows anomalies of the mean latitudinal position of the zonal 300-hPa wind $[\overline{L}(U_{300})]$ averaged between -170° and -50° longitude; the differences in this quantity between extreme and nonextreme fire days are not significant from June to August.

FIG. 9. Time series of surface weather composites at lag $t = 0$ (days) as function of lag t (days) relative to date of and location of fire spread event for mean daily 2-m temperature anomaly (ΔT) composites stratified by (a) month and (b) zone; mean daily 2-m vapor pressure deficit anomaly (ΔVPD) composites stratified by (c) month and (d) zone; and daily precipitation anomaly (ΔP) composites stratified by (e) month and (f) zone.

4. Discussion and conclusions

In this study, we have investigated the relationship of the polar jet stream with fire spread events—an objective measure of extreme wildfire activity. Fire spread events for North America (2002–16, between 40° and 70° latitude) occurred

predominantly in June and July for the Alaskan and Canadian boreal forest and between June and September for the western United States (Fig. 4). This seasonality is consistent with observed wildfires in Canada (Coogan et al. 2020), Alaska (Abatzoglou and Kolden 2011), and the northwestern forested mountains and North American deserts of the western United

FIG. 10. Histograms of jet stream metrics relative to position of fire spread events. Fire spread events (FSEs) are further partitioned by split flow index (SFI), where the threshold of 0.625 has been chosen to give approximately an equal number of events occurring above and below the threshold. Here the SFI is defined as the mean split flow within 40° latitude centered on each event.

States (Balch et al. 2017). In the North American boreal forest, lightning is a major driver of fires in June and July (Veraverbeke et al. 2017), whereas in the northwestern United States the lightning-caused fire season occurs between May and September (Balch et al. 2017).

The climatological position of the PJS has a strong seasonal variation, moving farther north and weakening in the summer (Hall et al. 2015; Rikus 2018). The subtropical jet (STJ), which can form south of 30° latitude, preferentially forms a distinct branch from the PJS during winter, although the two branches frequently merge during the summer, especially in the western hemisphere (Hall et al. 2015; Molnos et al. 2017). The winter pattern is also associated with a spiral-like structure that becomes a weak quasi-annular structure in the summer (Koch et al. 2006). These seasonal variations in the PJS were consistent with the climatologies shown here (see Fig. 4 and Fig. S2). We further found that the climatology of fire spread events over North America for 2002–16 was associated with minima in both the climatological zonal and meridional wind speeds, as well as the northward displacement of the jet stream core peaks, which occur during the boreal summer. However, while the PJS seasonal variability can therefore be associated with

seasonal variation in wildfire activity, covariability of the PJS and fire spread events also occurs on the time scale of synoptic-scale pattern development (and decay), which is typically from days to weeks.

To probe the subseasonal relationship further we considered spatiotemporal composite analyses. Compositing using all fire spread events (Fig. 5) showed on average that events are associated with formation of an upper-level atmospheric ridge and corresponding positive geopotential height anomalies centered on each event. This finding is consistent with the synoptic-scale patterns that are well known to lead to above average surface temperatures (Horton et al. 2016; Teng and Branstator 2017) and that are suitable for the drying of fuels and large wildfire events (Johnson and Wowchuk 1993; Skinner et al. 1999, 2002; Brewer et al. 2013; Hostetler et al. 2018; Wahl et al. 2019). We verified that these patterns occurred concomitantly with anomalous surface temperatures and other surface weather variables with composites that showed positive anomalies in mean temperature and VPD and negative anomalies in daily precipitation (Figs. S4 and S5). These surface weather conditions are well-known drivers of fire weather conducive to fire occurrence and spread (see, e.g.,

FIG. 11. (a) Meridional circulation index anomalies. (b) Split flow index anomalies [averaged between 30° and 80° latitude and between -170° and -50° longitude (North America) and over 5 days preceding and including each fire spread event (FSE)]. (c) Mean latitude of zonal 300-hPa wind speed averaged between -170° and -50° longitude. For all cases, a two-sample Kolmogorov–Smirnov test indicates where the two distributions [for $n(FSE) \leq 5$ and $n(FSE) > 5$] can be considered different with statistical significance of $\alpha = 0.1$ (*) or $\alpha = 0.05$ (**). The number of days per month with $n(FSE) > 5$ for 2002–16 is shown in Fig. S4.

Flannigan and Wotton 2001). The time series derived from the structural similarity index shows that the development and decay of the upper-level pattern is asymmetric in time with a longer mean development time compared with the mean decay time. The same asymmetry is also evident in the time series composites of temperature, VPD, and precipitation anomalies composited relative to the longitude and latitude of each fire event (Fig. 9). This is likely due to the drying (i.e., priming) of vegetative fuels in advance of a fire spread event. The notion of time lags in fuel moisture are encapsulated by fuel moisture codes or indices with equilibrium drying times determined by fuel layer depth (deeper layers or heavier fuels have longer drying times); see for example the Canadian Fire Weather

Index System (Van Wagner 1987) or 1000-h time lag fuel moisture (Fosberg et al. 1981). Our results also showed a spatial asymmetry with an observed northwest to southeast tilt in the composite 300-hPa wind speed pattern, which can be attributed to orographic forcing over the North American Rockies as noted by Brayshaw et al. (2009).

The composite analysis given here (Figs. 6 and 7) also shows that the relationship between synoptic fields and fire spread events has important regional and seasonal differences. The 300-hPa meridional wind patterns were displaced to the west when compositing events in eastern North America, which is consistent with the observation of a semipermanent atmospheric ridge over western North America (Bolin 1950;

Broccoli and Manabe 1992). Skinner et al. (2002) also identified this circulation feature from the summer climatology of 500-hPa heights over Canada, specifically: a weak trough to the west of the North American coast; a ridge extending over western North America to about 110°W; and a weak trough over eastern Canada called the Canadian polar trough, which is stronger in the winter, weakening and moving north in the summer (Maxwell 1986). Skinner et al. (2002) further noted there was considerable variability in these patterns, particularly in the summer months, with large fires occurring predominantly under an anomalously high western ridge. In this paper, the 300-hPa meridional wind patterns and 500-hPa geopotential height anomalies were also found (Fig. 8) to be weaker but to persist for longer in May–August when compared to April, September, and October. It may be reasonable to surmise that larger anomalies were necessary in the cooler months (i.e., shoulder fire season) to realize the persistent anomalous patterns conducive to large wildfire events. Composite patterns also persisted for longer in zones 1 and 2 compared with zone 3, also suggestive of the preferential position of an atmospheric ridge over the west of the continent. Regional and seasonal differences in surface weather composites for fire spread events were also apparent. Surface temperature anomalies largely reflected the same patterns as the 500-hPa geopotential height anomaly composites, as stratified by month (Fig. 6) and zone (Fig. 7). However, positive VPD anomalies were greatest for zone 1 (west) followed by zone 3 (east) and then zone 2 (central), in contrast to the temperature anomalies for each zone, which were similar in magnitude. Moreover, in contrast to the upper-level and temperature results, VPD anomalies for fire spread events increased throughout the fire season from April to October. The regional variations may reflect differences in regional climatology. Moreover, seasonal variations in VPD may also depend on phenological changes, which are not fully explained by the upper-level or temperature composite patterns. In general, the daily precipitation composites did not show a clear variation for individual months. However, zone 3 (east) did exhibit much larger negative precipitation anomalies for fire spread events than the other two zones, perhaps in part to the larger summertime precipitation that occurs in eastern North America.

Distributions of scalar positional metrics of the jet stream relative to fire spread events (Fig. 10) further confirm that fire spread events occur on average under an upper-level ridge under low split flow configurations. In contrast, fire spread events occur on average between jet stream core peaks for high split flow configurations, consistent with split flow blocking regimes. Our definition of split flow encompasses a wide range of jet stream configurations, which may include omega and Rex type blocks, the latter being often cited in the literature as a specific example of split flow (Barriopedro et al. 2006). In general, the role of split flow of the jet stream in predicting persistent anomalous patterns, and therefore fire spread events, is not clear. The climatology of the SFI (Fig. 11) shows that the majority of fire spread events occur during the period when the climatological SFI is a minimum. However, the SFI anomalies over North America during extreme fires days (i.e., days with more than 5 fire spread events) show a distribution

shifted to small positive values compared with other days) (i.e., 5 or fewer fire spread events). Moreover, recent studies that suggest quasi-resonant amplification is a mechanism that can lead to boreal summer weather extremes, depending on the formation of a double jet configuration (Coumou et al. 2014). Similarly we only found small positive anomalies in meridional flow (specifically the absolute value of the MCI) for extreme fire days in July and August. We also examined subseasonal variations in storm track positions and their relationship with wildfires. Specifically, anomalies of the meridional mean of the zonal 300-hPa winds for extreme fires days show little difference with nonextreme fire days, indicating that seasonal variations in the PJS northward displacement may be more important than subseasonal variability for determining fire conducive weather patterns. This finding should be contrasted with that of Wahl et al. (2019), who used paleoclimate reconstructions to infer the latitudinal profile of the zonal wind speed for both high and low fire years in California; they showed that high fire years are associated with small northward displacement of the jet compared with low fire years. However, their analysis differs from our own in a number of respects, most notably including the use of an annual mean for the zonal wind speeds, the use of paleoclimatic methodologies, and the focus on Californian fire regimes, which may have different fire weather drivers (i.e., the Santa Ana winds) than the study area considered here. In general we should emphasize that our study area was confined to 40° and 70°N latitude and the synoptic and surface weather patterns identified here may not correspond to drivers of fire activity outside of this area.

Climate composites, such as used in this study, are a powerful tool for examining the relationship between the state of the atmosphere and phenomena of interest. However, caution should be exercised when interpreting composites, which represent the mean of a climate field conditioned on the occurrence of an event. As such, they may form a necessary but not sufficient condition for the event occurrence so that the relationship is nonreversible (i.e., in general, the climate state cannot predict the event of interest; Boschat et al. 2016). In the present case, for example, synoptic weather patterns conducive to fire may not lead to fire spread events in the absence of ignitions (i.e., human or lightning caused). Moreover, because each composite represents mean values, they may not correspond to the specific atmospheric pattern associated with an individual event. We note that the use of spatiotemporal composites used in this paper can partially address the limitation on predictability by providing information on atmospheric patterns preceding and succeeding fire events. The most common alternative to composite analysis is *weather typing* or *classification* that essentially clusters atmospheric patterns into different types, either using subjective methods (i.e., manual classification by domain expert) or objective methods (i.e., automatic classification) (Huth et al. 2008; Skific and Francis 2012). However, while these methods allow the estimation of probabilities of an event given different weather types, as with any clustering algorithm, such approaches are sensitive to the number of allowed clusters.

Our work highlights the role of both seasonal and subseasonal variability of the polar jet stream and surface

weather conditions and the relationship with extreme fire events. However, there are several other potential directions for future work, which include understanding better the drivers and (or) mechanisms underlying persistent patterns that include atmospheric blocking. This may further require resolving the issue of whether planetary wave configurations conducive to fire represent global (Petoukhov et al. 2013) or regional (Röthlisberger et al. 2016) patterns. Moreover, the role of Arctic amplification (AA) on extreme weather in the midlatitudes is still an active area of interest with conflicting views in the literature as to whether AA is leading to patterns with increased meridional flow (i.e., waviness) (Overland et al. 2015; Vavrus 2018; Blackport and Screen 2020). For example, the seasonal variation in the Fourier spectrum of the planetary wave patterns (300-hPa meridional wind speed) show that the dominant wavenumber changes from long wavelength ($k = 3$) during the boreal winter to shorter wavelength patterns ($k = 5$ – 7) in the boreal summer (see Fig. S3). Therefore, while the summertime planetary wave patterns appear weaker, they also correspond to shorter-wavelength patterns when compared with the winter, albeit with weaker amplitudes. A spectral analysis of the jet stream, as well as the global and regional wave properties, will be the subject of future work. Climate change may also affect midlatitude weather in other ways. For example, Ali et al. (2009) suggest that climate change may lead to less favorable conditions for wildfire in eastern Canada due to increased advection of humid air from Atlantic tropical zones with the northward displacement of the polar jet stream. It is therefore important to understand how midlatitude circulation patterns are changing and how these changes will influence surface weather conditions, including fire weather extremes.

Finally, fire management is challenging and there are increasing additional pressures such as climate change, expanding wildland urban interface, and declining forest health, as well as growing public awareness and expectations. Fire management agencies recognize that there are physical and economic limits to further manage fire and have come to realize that increasing fire suppression expenditures lead to decreasing marginal returns in terms of escaped fires or area burned (Flannigan et al. 2009). Fire management agencies operate with a narrow margin between success and failure; any advantage in dealing with fire spread events that are responsible for most of the area burned and impacts would be beneficial. Spatial and temporal information from the upper-level composite analyses (Figs. 5–8) could form the basis of an early warning system. An effective early warning system would allow the movement of resources to where the fire spread is expected and extinguish unwanted fires while they are still small and manageable.

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Data availability statement. NCEP Reanalysis 2 data are provided by the NOAA/OAR/ESRL PSL, Boulder, Colorado, USA, from their Web site at <https://psl.noaa.gov/>; CONUS fire data are provided by the Monitoring Trends in Burn Severity

(MTBS) from their Web site at <https://www.mtbs.gov/>; Alaska fire data are provided by the Alaska Interagency Coordination Center (AICC), from their Web site at <https://fire.ak.blm.gov/aicc.php>; Canada fire data provided by the Canadian National Fire Database–Agency Fire Data, Natural Resources Canada, Canadian Forest Service, are available from the Canadian Wildland Fire Information System (CWFIS; <http://cwfis.cfs.nrcan.gc.ca/ha/nfdb>). Hotspot data provided by MODIS Collection 6 NRT Hotspot / Active Fire Detections MCD14DL are available online (<https://earthdata.nasa.gov/firms>).

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