Forest Fire-Conducive Drought Variability in the Southern Canadian Boreal Forest and Associated Climatology Inferred from Tree Rings

Martin P. Girardin, Jacques C. Tardif, Mike D. Flannigan and Yves Bergeron

Abstract: Forest fires in Canada are directly influenced by the state of the climate system. The strong connection between climate and fire, along with the dynamic nature of the climate system, causes the extent, severity and frequency of fires to change over time. For instance, many reconstructions of the history of forest fires across boreal Canada report a general decrease in fire activity since ~1850 which could, in part, result from changes in climate. This paper describes progress in characterizing the variability in fire-conducive droughts in the central and eastern Canadian boreal forests during the past three centuries. An extensive network of drought-sensitive tree-ring records from Manitoba, Ontario and Quebec was used to develop five multi-century reconstructions of the mean July Canadian Drought Code and one reconstruction of mean July and August temperatures. Correlation analyses with regional fire statistics (common period 1959–1998) showed that drought estimates are accurate enough to approximate fire activity and, hence, the estimates are relevant for the study of climate change impacts on Canadian forests. Spatial correlation analysis over the period 1768–1998 revealed that variability between the west and east has increased since the mid-19th century, specifically the decade-to-decade variability and the frequency of extreme events. Based on the synoptic characteristics of recent droughts, we interpret this change in variability as a response to an increasing frequency of upper level ridging and troughing over western and eastern Canada, respectively. The increasing horizontal movement of humid air masses over eastern Canada since ~1850 could have contributed to the creation of moister conditions that are less suitable for fire.

Résumé : Les feux de forêt sont fortement influencés par l'état du système climatique. Le lien étroit qui existe entre le climat et les feux ainsi que la nature dynamique de l'état du système climatique conduisent à des variations temporelles dans l'étendu, la sévérité et la fréquence des feux. Par exemple, plusieurs reconstitutions historiques des feux de forêt à travers le Canada boréal rapportent une diminution de l'activité des feux depuis ~1850 qui pourrait en partie être due à des changements du climat. Dans la présente étude, nous décrivons les progrès réalisés dans la caractérisation de la variabilité des sécheresses

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propices aux feux de forêt au cours des trois derniers siècle du centre à l'est de la forêt boréale canadienne. Un réseau étendu de séries d'accroissement radial d'espèces arborescentes sensibles à la sécheresse en provenance du Manitoba, du Québec et de l'Ontario, a été utilisé pour développer cinq reconstitutions multi-centenaires de l'indice de sécheresse canadien (CDC) de juillet et une reconstitution des températures moyennes de juillet et août. Des analyses de corrélation effectuées sur des données régionales de l'activité des feux (période commune 1959–1998) ont démontré que les estimés de sécheresses étaient suffisamment fiables pour inférer la variabilité temporelle de l'activité des feux. Ces estimés sont donc pertinents pour l'étude de l'impact des changements climatiques sur la forêt canadienne. Des analyses de corrélation spatiale sur la période 1768–1998 ont démontré que la variabilité entre l'ouest et l'est s'est accentuée depuis le milieu du 19e siècle, en particulier pour la variabilité interdécennale et la fréquence d'événements extrêmes. À partir des caractéristiques synoptiques des sécheresses récentes, nous interprétons ce changement dans la variabilité comme une réponse à une augmentation de la fréquence des crêtes et creux barométriques au-dessus de l'ouest et de l'est du Canada, respectivement. L'accroissement du mouvement horizontal d'air humide sur l'est du Canada depuis ~1850 pourrait avoir contribué à la création de conditions plus humides qui sont moins propices aux feux.

Introduction

The American Meteorological Society (2000) defines drought as "a period of abnormally dry weather sufficiently long enough to cause a serious hydrological imbalance". However, they also note that "drought is a relative term; therefore, any discussion in terms of precipitation deficit must refer to the particular precipitation-related activity that is under discussion". In the boreal forest of central and eastern Canada, drought is particularly important as a control on the occurrence of forest fires. Drought influences fire characteristics because of its impact on fuel moisture and the effects of precipitation (particularly its frequency), relative humidity, air temperature, wind speed, and lightning (Flannigan and Harrington, 1988; Flannigan and Van Wagner, 1991; Agee, 1997; Harrington *et al.,* 1991; Johnson, 1992; Weber and Flannigan, 1997; Bergeron *et al.,* 2001). Despite the increasing importance of human activity as a source of fire ignition over the last few decades (Stocks *et al.,* 2003), dry forest fuels and wind remain major contributors to large stand-destroying fires (Johnson *et al.,* 1990; Masters, 1990; Johnson, 1992).

In the boreal forest of central and eastern Canada, forest fires are a primary natural process organizing the physical and biological attributes of the forest, shaping landscape diversity and influencing biogeochemical cycles (Bergeron, 2000; Bourgeau-Chavez *et al.,* 2000; Burton *et al.,* 2003). Forest communities are essentially a successional mosaic of broadleaf deciduous and needleleaf evergreen trees that reflect recovery from recurring fires (Bonan, 2002). The mosaics of different vegetation types are to a large extent an expression of their respective fire regimes and many boreal tree species show adaptation to fire. In addition, watershed fire disturbances have profound effects on water quality in boreal surface waters. Trees take up considerable amounts of water through the process of transpiration and their removal can lead to an excess of soil water. Consequently, enhanced erosion and leaching from soils of disturbed watersheds can lead to increases in particles and nutrients (e.g., phosphorus and nitrogen) in streams after snowmelt and rainstorms (Prepas *et al.,* 2003). Prepas *et al.* (2003) state that changes in nutrient fluxes after fires can lead to a complex series of biological changes that can affect the structure and function of aquatic ecosystems.

Because of the strong linkage between climate and forest fires, changes in climate trigger variations in historical observations of fire activity and in vegetation composition and structure (Flannigan and Harrington, 1988; Johnson, 1992; Swetnam, 1993; Bourgeau-Chavez *et al.,* 2000). A number of forest stand age distributions across boreal Canada (expressed in percentage of the study area per age class) reconstructed from living trees, snags, and downed woody material show a distinct decrease in the area originating from fire since ca. 1850 (Masters, 1990; Johnson and Larsen, 1991; Larsen 1997; Bergeron *et al.,* 2001; 2004a; b; Tardif, 2004). According to

Bergeron and Archambault (1993), the occurrence of lower fire activity in northwestern Quebec matched a long-term increase in mean ring width of *Thuja occidentalis* L. The authors suggested that the decrease in fire activity could reflect a decrease in the frequency of drought periods. The mechanism behind this climate change remained to be elucidated, though Hofgaard *et al.* (1999) provided evidence that it may have been triggered by an inhibition of dry arctic air outflows around 1850. Larsen (1997) suggested that the post-1860 decline in fire frequency in northern Alberta was also due to an increase of precipitation, likely driven by decreases in the frequency of high pressure anomaly over southern Yukon. Additionally, fire suppression during the past 50 to 80 years has contributed to changes in the recurrence of fire in many regions of Canada (Bourgeau-Chavez *et al.,* 2000).

In the past decade or so, significant progress has been made in characterizing the spatial and temporal variability of fire and fire-conducive weather conditions over the central and eastern Canadian boreal forest (Skinner *et al.,* 1999; 2002; Amiro *et al.,* 2004; Girardin *et al.,* 2004a). These studies, however, are based on observational data and are limited to periods not exceeding a few decades. This short time frame is problematic for fire-climate relationship analysis. Modern vegetation reflects the influence of past disturbances that occurred prior to the initiation of instrumental monitoring of climate. This is the case in the northwestern Quebec boreal forests, where over 80 percent of forest stands originated from forest fires that took place prior to 1920 and over 47 percent prior to 1850 (Bergeron *et al.,* 2004b). Furthermore, the period of largest area burned, the 1910-20s (approximately 22 percent of total forested area), has no equivalent in the present day in terms of extent (Bergeron *et al.,* 2004b). The climatic/synoptic causes for such extreme fire events are hence poorly understood. There is an increasing interest in incorporating natural disturbance regimes such as fire into boreal forest management. Management that favours the development of stand and landscape compositions and structures similar to those characterizing natural ecosystems should favour the maintenance of biological diversity and essential ecological functions (Hunter, 1999; Prepas *et al.,* 2003; Bergeron, 2004). To develop this type of management, there is a need to understand the past history of forests and the influence of climate on this history.

Tree rings can serve as proxies of past fire-conducive drought variability and thereby provide information that enhances the limited temporal coverage of observational records. The rationale for studying tree rings is that trees in temperate regions produce annual radial increments, where changes in ring width from one year to the next reflect changes in precipitation, temperature and drought, as well as other factors (Fritts, 2001). Changes in ring width may also reflect forest disturbance history caused by insect defoliation, forest fires, tree senescence or individual tree mortality. From this pattern of year-to-year changes in ring width, one can infer fire-conducive drought variability at times during which there was no observational record (Sheppard and Cook, 1988; Westerling and Swetnam, 2003; Girardin *et al.,* 2004b; 2006a; b; c).

This paper investigates patterns of drought that influence forest fire activity (hereafter referred to as fire-conducive drought) in the central and eastern Canadian boreal forest, primarily using six tree-ring based drought reconstructions from across the region (Figure 1). The objective is to demonstrate the relevance of these reconstructions for the study of climate change impacts on forest dynamics in the eastern and central Canadian boreal forest. First, the correlation between the inferred drought variability and observed fire activity records over the period 1959 to 1998 is shown. Second, the potential of the six reconstructions to approximate large-scale atmospheric circulation conducive to fire is demonstrated. Finally, the long-term history of fire-conducive drought variability and the value of this information are discussed.

Climate and Tree-Ring Data

Tree-Ring Records in the Central and Eastern Boreal Forest

Long-term variability in fire-conducive drought conditions over the central and eastern Canadian boreal forest was addressed by Girardin *et al.* (2004b; 2006a) with the development of tree-ring based summer drought reconstructions from six climate regions (Figure 2; Table 1). From west to east these were the Boreal Plains (BP), Lake Seul Upland and Lake of the Woods (LS), Lake Nipigon (LN), Abitibi

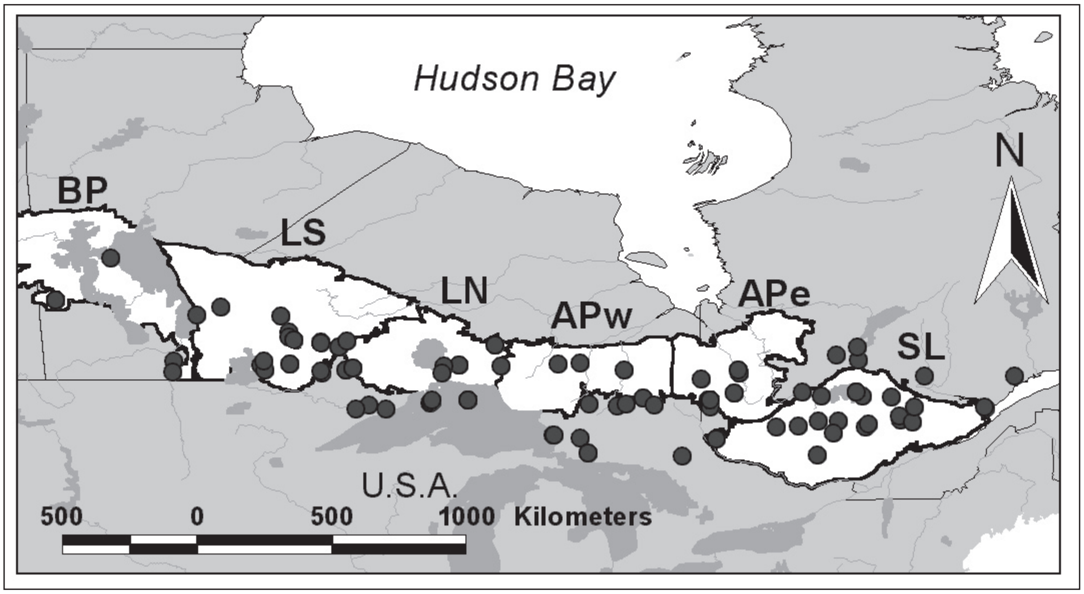

Figure 1. Six regions were selected for the climate reconstruction: Boreal Plains (BP), Lake Seul Upland and Lake of the Woods (LS), Lake Nipigon (LN), Abitibi Plains *west* (APw), Abitibi Plains *east* (APe), and Southern Laurentian (SL). Filled circles show the location of the 120 tree-ring width chronologies used for the climate reconstruction.

Plains west and east (APw and APe), and Southern Laurentian (SL) regions (Figure 1).

The reconstructions were developed from a network of 120 well-replicated tree-ring width sites distributed mainly on the Boreal Shield. The development of these tree-ring records represents the achievement of over a decade of research, sampling efforts and laboratory work by a number of researchers. Tree-ring width chronologies covering the eastern half of the Canadian boreal forest were obtained from the International Tree Ring Data Bank Library, Archambault and Bergeron (1992), Hofgaard *et al.,* (1999), Tardif and Bergeron (1997), Jardon *et al.* (2003), and several others (Girardin *et al.*, 2006a provide a complete listing). In 2002 and 2003, sampling campaigns were conducted in boreal Ontario to fill gaps within the network.

In total, over 4,400 ring width measurement series from 13 species and extending back to at least 1866 AD were gathered. Most sampled trees belonged to the genus *Pinus* (45 percent of all sites); the genus *Picea* (27 percent of all sites) and *Thuja occidentalis* L. (15 percent of all sites) also contributed to a large proportion of the total network. The drought reconstructions were considered reliable as proxies of past year-to-year and decade-to-decade variability in drought for the period covering the late 1760s to present. Because age/size-related trends were removed from each tree-ring measurement series, the drought reconstructions contain no information relative to century-long climate changes (Cook *et al.,* 1995). The tree-ring width series for the drought reconstructions were standardized with cubic splines such that mainly annual to decadal scale variance was retained in the chronologies, and hence, in the drought reconstructions. On average, 99 percent of the variance contained in frequencies lower than 19 years and 50 percent of the variance in frequencies lower than 60 years was preserved in the processed tree-ring time series. The mean length of measurement series (mean age of 137 years) was considered insufficiently long to allow robust reconstruction of low frequency climatic variations.

The Canadian Drought Code

Five of the six tree-ring reconstructions were developed for the mean daily Canadian Drought Code (CDC) in July (Figure 2a-e). The CDC is a component of the Canadian Forest Fire Weather Index (FWI) System. Beginning in the 1920s, researchers in Canada began tracking day-to-day susceptibility of forest fires using

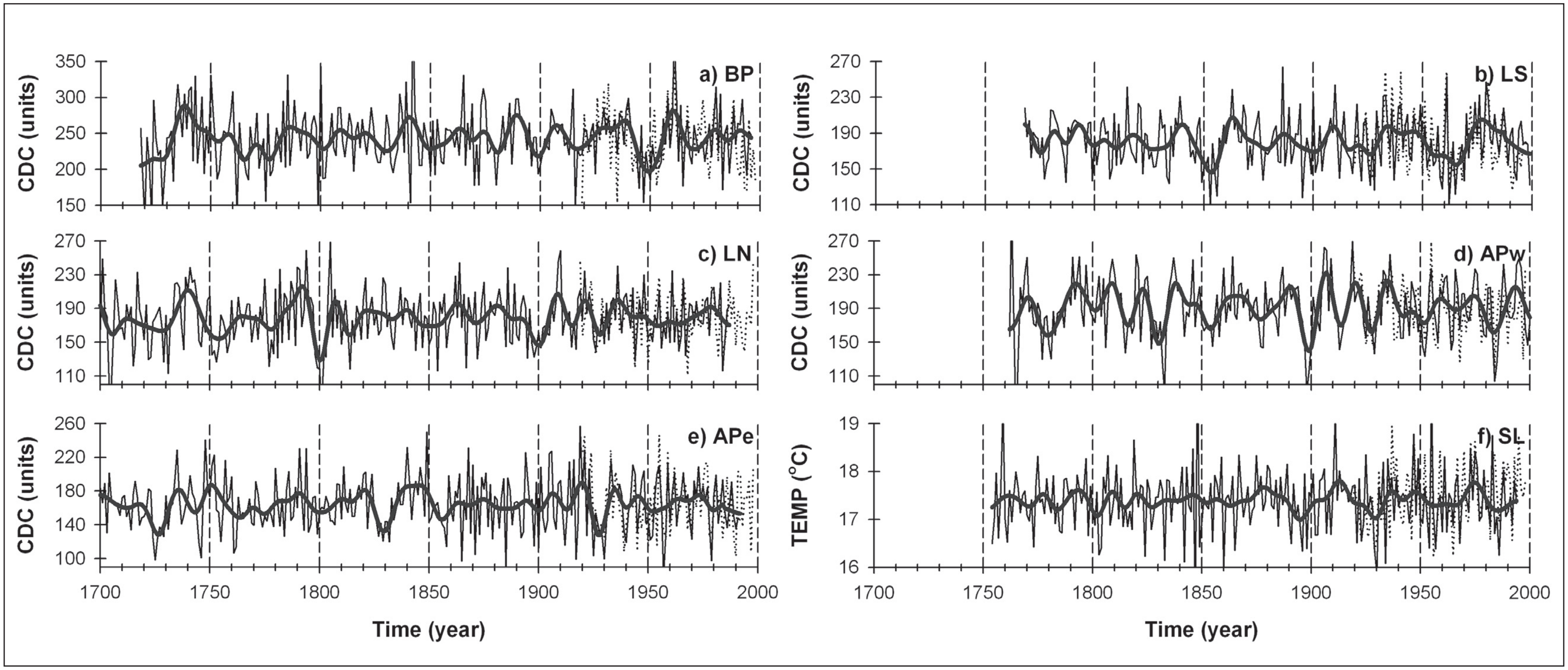

Figure 2. (a) to (e) Reconstructions of the July mean of the daily Canadian Drought Code (CDC) for the BP, LS, LN, APw, and APe regions (refer to Figure 1 for abbreviations). The CDC scale ranges from soil saturation (zero) to severe drought (>300). (f) Reconstruction of the SL mean July to August temperature (TEMP, °C). Smoothed lines show least-squares polynomial fittings (order 4) across a moving window of 10 years. Dotted lines show instrumental data (period 1913-1998); variance in instrumental data has been adjusted to correspond to reconstruction.

Table 1. Statistics of the six drought reconstructions developed by Girardin *et al.* (2006a). The period covered by each reconstruction, the number of chronologies used as predictors, and the maximum amount of variance R^2 in the instrumental data accounted for by each reconstruction (period 1919-1984) are indicated.

Reconstruction	Variable	Period Covered	Number of Chronologies	R^2
Boreal Plains	July CDC	1718-1996	13	39%
Lake Seul	July CDC	1768-1999	25	50%
Lake Nipigon	July CDC	1680-1987	24	36%
Abitibi Plains *west*	July CDC	1762-2001	13	34%
Abitibi Plains *east*	July CDC	1683-1992	20	46%
Southern Laurentian	Jul-Aug Temp	1754-1994	25	44%

rating systems for fire danger (Van Wagner, 1987). In subsequent decades, four different fire danger systems were developed for various regions of Canada, with each version based on field research on the fuel types of local importance. By the late 1960s, there was an increasing demand by forest fire control agencies for the development of a new fire danger-rating index; the result was the FWI System. This system retained a solid link with previous approaches by building on the best features and adding new components where necessary (Van Wagner, 1987). Today, the FWI System is used in the daily operation of Canadian fire management agencies to monitor forest fire danger across the country. Expansion of the monitoring system has begun in foreign countries. Forest fire researchers from Canada, Russia and Germany have recently developed methodologies for electronically gathering daily weather data and producing daily fire weather and fire behaviour potential maps for large portions of northern Europe and northern Asia (Fire Ecology Research Group, 2005).

As part of the FWI System, the CDC was developed to serve as an index of the water stored in the soil, on average about 20 cm deep, and to warn when lower layers of deep partly decomposed organic material may be drier than the upper ones (Appendix A). The CDC represents the net effect of daily changes in evapotranspiration and precipitation on cumulative moisture depletion. Real fire danger may be affected by the state of these layers, which are common in many parts of Canada (Turner, 1972; Van Wagner, 1987). At this soil depth, drought is a determining factor for forest fire severity, as dry conditions allow deep burning and smouldering. Therefore, via its influence on fire severity, drought also becomes an important controlling factor of postfire ecosystem structure and function through direct impacts on underground plant roots, reproductive tissues, and soil seed banks (Weber and Flannigan, 1997).

Canadian Drought Code Reconstructions

For the purpose of climate reconstruction, regional monthly CDC data files covering the period 1913 to 1998 were created by averaging daily CDC data computed from meteorological stations located in and near the climate regions (Girardin *et al.*, 2004a). In total, 62 meteorological stations were used for computation of regional monthly CDC indices. The CDC is cumulative and hence the five July CDC reconstructions presented in Figure 2 approximate the average moisture content of deep and compact organic layers for a season from May to July. Over 78 percent of the total area burned in Canada does so during this season (Stocks *et al.*, 2003). The rationale linking tree growth to the July CDC is that assimilation of carbohydrates and optimal tree growth occur only if soil moisture is sufficient to maintain foliage water potential and minimize vapour pressure deficits (Girardin and Tardif, 2005). By using a linear model relating tree growth to July CDC, past July CDC values (Figure 2a-e) can be inferred from multi-century tree ring chronologies (Girardin *et al.*, 2004b).

In some circumstances, the relationship between tree growth and the July CDC may be of little use in inferring past drought variability. This was the case with the SL region, where summertime drought did not appear to be an important limiting factor for tree growth (Girardin *et al.*, 2006a). The low relationship may also have been due to strong heterogeneity in precipitation in this region (Girardin, 2005). However, tree growth showed a significant negative correlation with summertime temperatures during the previous year (Girardin *et al.*, 2006a); this observation led to the development of a July to August mean temperature reconstruction for the SL region (Figure 2f; Girardin *et al.*, 2006a). Despite this difference, all reconstructions (Figure 2a-f) are referred to as "drought reconstructions" in the rest of the paper for simplicity. A list of persistent drought episodes inferred from the smoothed reconstructions is presented in Table 2. All reconstructions were updated to 1998 using instrumental data (Figure 2) prior to subsequent analysis.

Common Drought Signal

Over the central and eastern boreal forest, drought severity exhibits strong regional coherence; that is, drought in one region is related to drought in other regions, largely due to the common influence of large-scale features of atmospheric circulation (Girardin *et al.*, 2004a; b; 2006a). Therefore, it is appropriate to summarize the dominant information contained within the six fire-conducive drought reconstructions into a new set of uncorrelated variables. The variability within the reconstructions may be analyzed using Principal Component Analysis (PCA) (Legendre and Legendre, 1998; Barry and Carleton, 2001). In this procedure, the amount of variability is described using the same number of variables, but the first principal component (PC1) accounts for the maximum possible proportion of the total variance. Succeeding PCs, in turn, account for as much of the residual variance as possible. The loading of each reconstruction on each component (Figure 3), gives the spatial representation of the PCs. Hereafter PCs are referred to as "drought components". For the purpose of this paper, only the first and second drought components (Figure 3), accounting for 30.4 percent and 21.2 percent of the variance, respectively, were retained for subsequent discussions.

One may note from the smoothed drought components that decade-to-decade variability in drought severity over much of the domain under study

Table 2. Prolonged drought episodes across the Boreal Shield and eastern Boreal Plains inferred from the six reconstructions.

Boreal Plains	Lake Seul	Lake Nipigon	Abitibi Plains *West*	Abitibi Plains *East*	Southern Laurentian
1735 to 1743		1736 to 1744		1734 to 1738	
				1748 to 1754	
	1791 to 1795	1787 to 1795	1790 to 1795	1789 to 1792	1791 to 1795
		1806 to 1809	1807 to 1811		
				1819 to 1822	
1838 to 1843	1838 to 1842		1837 to 1840	1837 to 1849	
	1860 to 1867				
					1876 to 1882
1887 to 1892			1889 to 1892		
	1908 to 1911	1907 to 1911	1905 to 1909		1909 to 1916
		1920 to 1923	1919 to 1922	1917 to 1922	
1936 to 1940	1932 to 1937	1934 to 1938	1933 to 1937		
1958 to 1963					
	1973 to 1983				1971 to 1978
		1991 to 1995	1992 to 1995		

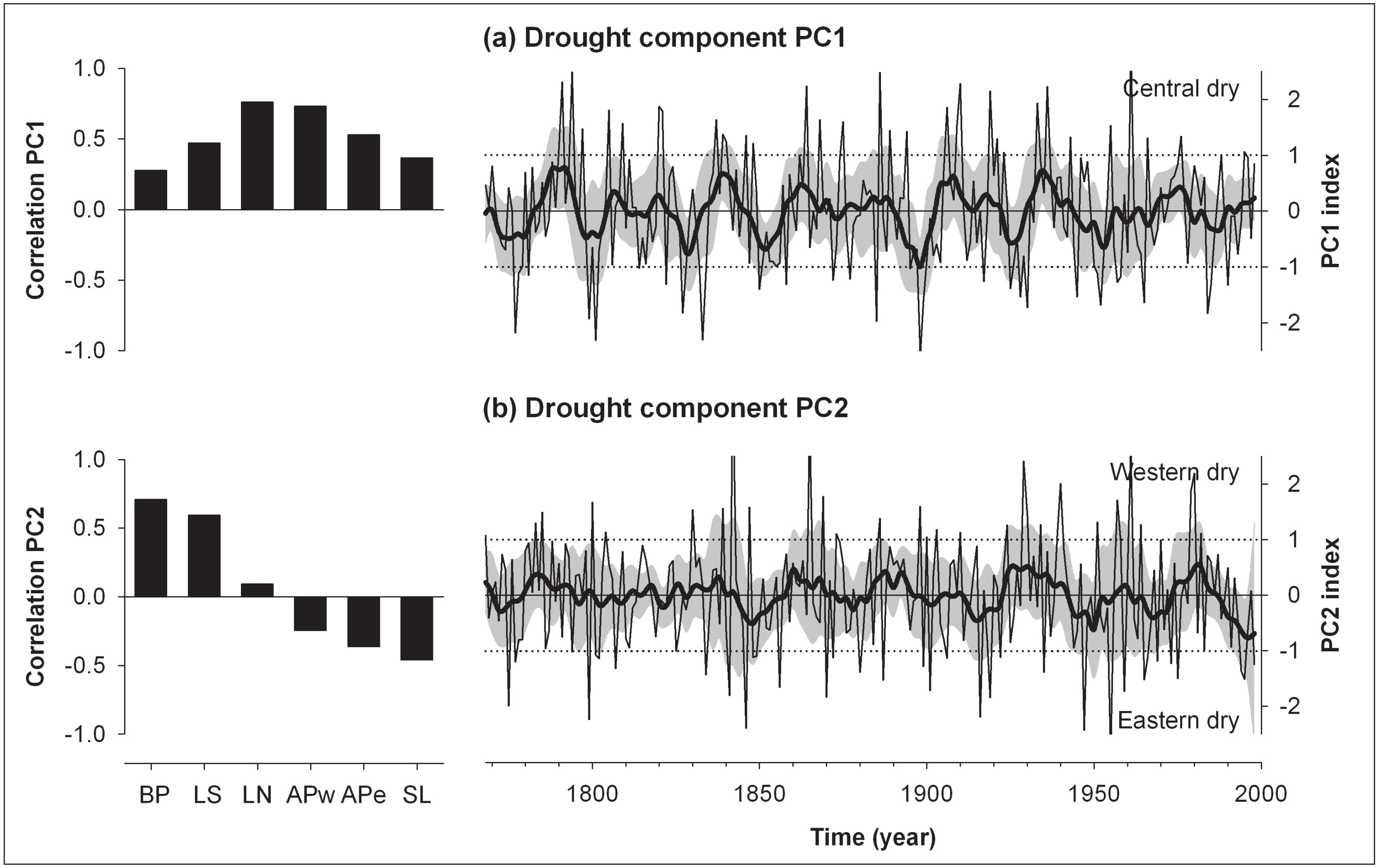

Figure 3. Reconstruction loadings (*vertical bars at left*) for the first and second drought components (PCs) (*indices at right* — bold lines show 10-year moving averages with 95 percent confidence bands). The reconstruction loadings show the correlation coefficients between the drought components and the six drought reconstructions of Figure 2 across the study area over their common period 1768-1998. From these correlations we deduce that in *(a)* the first drought component is associated with a domain-wide pattern of drought variance centred in the *'central'* portion of the study area, where a positive PC1 value indicates dry conditions; a year showing a negative PC1 value denotes low drought severity. The second drought component (*b*) is associated with a dipole between the western and eastern regions. A year showing a positive PC2 value indicates that drought severity in the *'western'* regions is higher than in the eastern regions; a year showing a negative PC2 value indicates higher drought severity in the *'eastern'* regions.

remained in the same order of magnitude during the past two centuries (Figure 3a). Conversely, decade-to-decade differences in drought severity between the western and eastern regions were less pronounced from ca. 1770 to 1850s (Figure 3b). This observation was corroborated with a wavelet analysis on PC2, showing greater variance in the nine to 32-year/cycle frequency band since ca. 1850s (Girardin *et al.*, 2006a). The increasing contrast is also apparent in changing correlations between the BP and APe regions (Figures 2a and 2e): from 1718 to 1840, the CDC reconstructions from these regions were significantly positively correlated (r = 0.24, p = 0.007), but were not significantly correlated after 1840. An analysis of spatio-temporal variability on 90 multi-century tree-ring chronologies covering the eastern half of Canada (101°W to 61°W and 41°N to 61°N) also support the observation of an increasing contrast in the relationship between western and eastern regions (Girardin *et al.*, 2006a).

Associations between Drought and Fire

The capability of drought components PC1 and PC2 to approximate fire activity was examined. In this procedure, correlation coefficients were computed between the drought components and records of forest fire observed in the last half of the 20th century (Figure 4). For this analysis, annual fire data were obtained from the Large Fire Database (LFDB; Stocks *et al.*, 2003), which is a compilation of forest fire data from Canadian fire management agencies, including provinces, territories, and Parks Canada for the period covering 1959 to 1998. The data set includes only fires greater than 200 ha in final size, which represent only a few percent of all fires but account for most of the area burned (usually more than 97 percent; Stocks *et al.*, 2003). Time series of annual area burned (ha) and number of large forest fires were compiled for the five ecoregions covering the area under study (Figure 4).

As shown by the correlation analysis, the drought components are roughly representative of the variability in regional fire activity over the period 1959 to 1998 (Figure 4). The first drought component, PC1, is significantly correlated with both area burned and the number of large fires in the LN and AP ecoregions (Figure 4a). The second drought component is significantly correlated with the fire variables in the BP and LS ecoregions, but did not correlate with fire activity in eastern ecoregions (Figure 4b). However, it was highly correlated with the Quebec provincial area burned record over the interval 1972 to 1998 (Figure 5) (fire data from Ministère des Ressources naturelles et de la Faune du Québec, 2005). It should be noted that analyses of fire activity data are often complicated by biases in the data. Many problems can occur with fire statistics, primarily because of inconsistent and expanding detection systems; the quality of forest-fire statistics varies over both time and space and tends to be better in recent years (Murphy *et al.*, 2000; Podur *et al.*, 2002). Analyses may further be complicated by other factors that influence fire statistics, including fuels, land use, topography, landscape fragmentation, fuel characteristics, fire site accessibility, simultaneous fires, and fire management policies. Large spatial units (provincial or ecozone levels) may also provide better results when linking fire activity to climate factors—fire is highly variable among years and the use of a large number of samples provides some statistical smoothing (Amiro *et al.*, 2004; Flannigan *et al.*, 2005; Girardin *et al.*, 2006b). Nonetheless, based on the analysis presented in Figures 4 and 5, for most of the study area, one can assume that the two components can provide meaningful information on fire conditions at the time during which there were no observational fire records (Girardin *et al.*, 2006b).

Connection between Drought and Atmospheric Circulation

Atmospheric circulation plays a key role in creating drought conditions that are conducive to fire and can also act as a determinant factor for fire ignition. This section will address the relationship between extreme large-scale droughts and atmospheric circulation.

Extreme Large-Scale Droughts

Droughts that affect a large fraction of the central and eastern boreal forest are often linked to specific patterns in the atmospheric circulation over Canada. To identify circulation patterns that lead to widespread drought, the methodology proposed by Mudelsee (2006) was adapted to detect the occurrence of extreme drought years on the Boreal Shield. This approach identifies

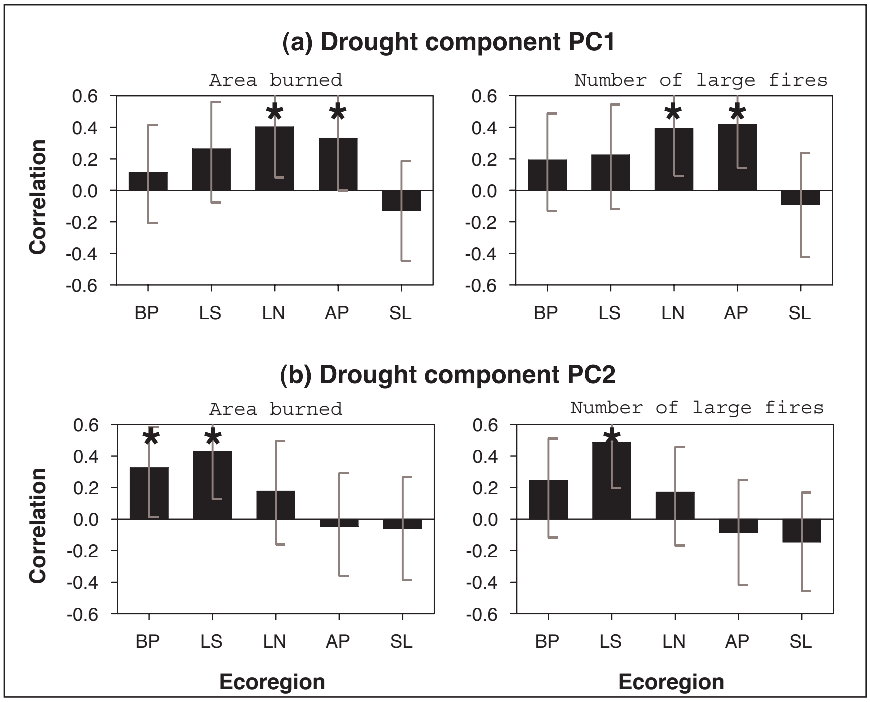

Figure 4. Spearman correlation coefficients *(black bars)* of the relation between the drought components (PCs) a) PC1 and b) PC2 and seasonal area burned *(left)* and number of large forest fires by years *(right)* for five ecoregions. *Error bars* show bootstrapped [1000] 95 percent confidence intervals; significant correlations are marked with an asterisk. Fire data were extracted from the large forest fire database, which includes observed fires larger than 200 ha (Stocks *et al.,* 2003). The period of analysis is 1959 to 1998.

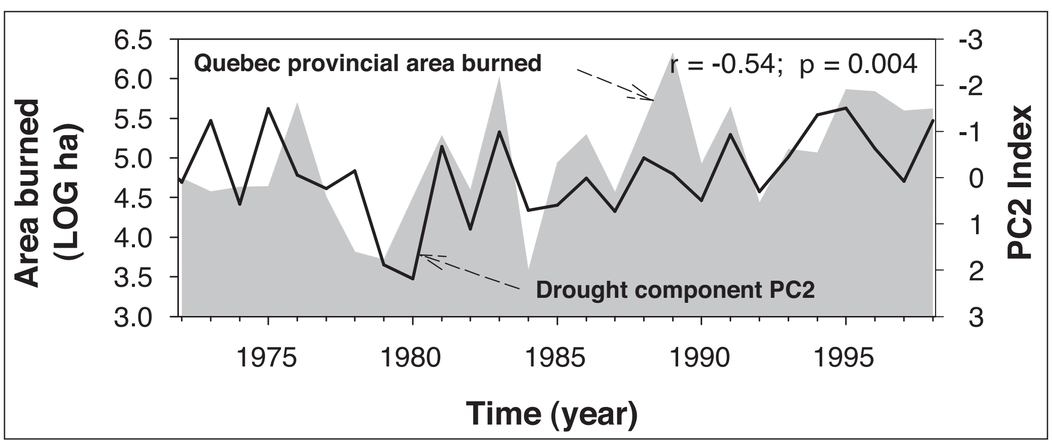

Figure 5. Relationship between the drought component PC2 *(solid line)* and Quebec provincial area burned *(shaded area)* during the period 1972 to 1998. Note the inverted *y*-scale for PC2 at right. The Pearson correlation coefficient is shown along with the corresponding *p*-value obtained from 10,000 permutations (if both series are first detrended: $r = 0.48$, $p = 0.011$). Area burned data were obtained from Ministère des Ressources naturelles et de la Faune du Québec (2005) and transformed using the logarithmic function.

events as ‘extreme’ when they exceed a detection threshold computed from a running median smoothing (Mudelsee, 2006). This approach is most useful when both the background state and interannual climate variability change through time, as is common in many types of records and climate archives (Mudelsee, 2006). A similar methodology (Mudelsee, 2002) was employed for detection of occurrence rates of extreme fire years on the Boreal Shield (Girardin *et al.,* 2006b).

The records of extreme large-scale droughts (expressed as the number of extreme droughts per decade) were developed for the central (PC1), western (+PC2) and eastern (-PC2) regions (Figure 6). The exceedance threshold used to identify extreme droughts was chosen subjectively to approximately correspond to the 85th percentile. The analysis showed that the frequency of extreme large-scale drought events over the central regions of the study area (Figure 6a) has remained relatively constant during the past two centuries. Conversely, contrasting trends appeared with the analysis of drought differences between the western and eastern regions (Figures 6b and 6c). The frequency of extreme large-scale drought events over western regions increased after ca. 1850 (Figure 6b). The mean frequency of extreme droughts per decade detected during the interval 1850 to 1980 was twice that of the interval 1770 to 1850 (1.8-event decade^{-1} versus 0.9-event decade^{-1}, respectively; *t*-test = 2.499, 100.000 permutations $p = 0.030$). There was no apparent trend for eastern Canada, although high frequencies of extreme events were experienced prior to ca. 1850 and during the last 50 years (Figure 6c).

Circulation during Extreme Droughts

Large forest fires in boreal Canada are associated with prolonged blocking high-pressure systems in the upper atmosphere over or upstream from the affected regions (Skinner *et al.,* 1999; 2002). These systems are defined by the American Meteorological Society (2000) as ‘anomalous’ circulation patterns that typically remain nearly stationary or move slowly westward, and persist for a week or more. The prolonged blocking high-pressure systems cause obstruction, on a large scale, of the normal west-to-east progress of migratory storms. These systems also cause air subsidence in the upper atmosphere, resulting in typically sunny, warm days that create dry fuel conditions that can extend over several hundreds of kilometres (Newark, 1975; Johnson and Wowchuk, 1993; Bessie and Johnson, 1995; Skinner *et al.,* 1999; 2002). When the high-pressure systems have significant moisture or begin to break down, convective activity leads to numerous lightning strikes and forest fire ignition (Nash and Johnson, 1996).

The reconstructed drought components suggest that the frequency of extreme droughts has changed in the past (Figure 6). This change could involve changes in the frequency of circulation patterns. The drought components’ capability to reflect atmospheric circulation conducive to large fire activity is presented below. The analyses were conducted using 500 hPa NCEP reanalysis geopotential height composite charts (Kalnay *et al.,* 1996; http://www.cdc.noaa.gov/cgi-bin/Composites/printpage.pl). Composite charts are often used in synoptic climatology to reveal ‘average’ circulation patterns during signature years (in this case, years of widespread drought). These patterns can be used to investigate patterns of upper-atmospheric wind flow and the movement of weather systems. Geopotential height at the 500 hPa level approximates the height above sea level of the 500 hPa pressure surface (roughly 5500 m above sea level). Winds at the 500 hPa level tend to flow parallel to the isohypses (height contours). The winds are called meridional when the isohypses form a strong wave-like pattern, and zonal when the isohypses are nearly parallel to the lines of latitude (Ahrens, 2003). Within these waves, an elongated area of high heights is known as a ridge and an elongated area of low heights is a trough. Geopotential height data exist for the study area for the period from 1948 to the present.

The mean features of the upper atmospheric circulation over Canada include the presence of troughs located over the North Pacific (West Coast Trough, WCT) and northeastern Canada (Canadian Polar Trough, CPT) and a ridge across the northeastern Pacific to the West Coast (Continental Ridge, CR) (Figure 7). It is during strong meridionality (i.e., an increased wave-like pattern) that obstruction of the normal west-to-east progress of migratory storms occurs (Ahrens, 2003). Fire-conducive drought events over the study area are generally linked with such conditions (Figure 8). Extreme large-scale drought events over the central regions (Figure 8a) were, on average, associated with positive height anomalies (ridging) over western Hudson Bay and negative

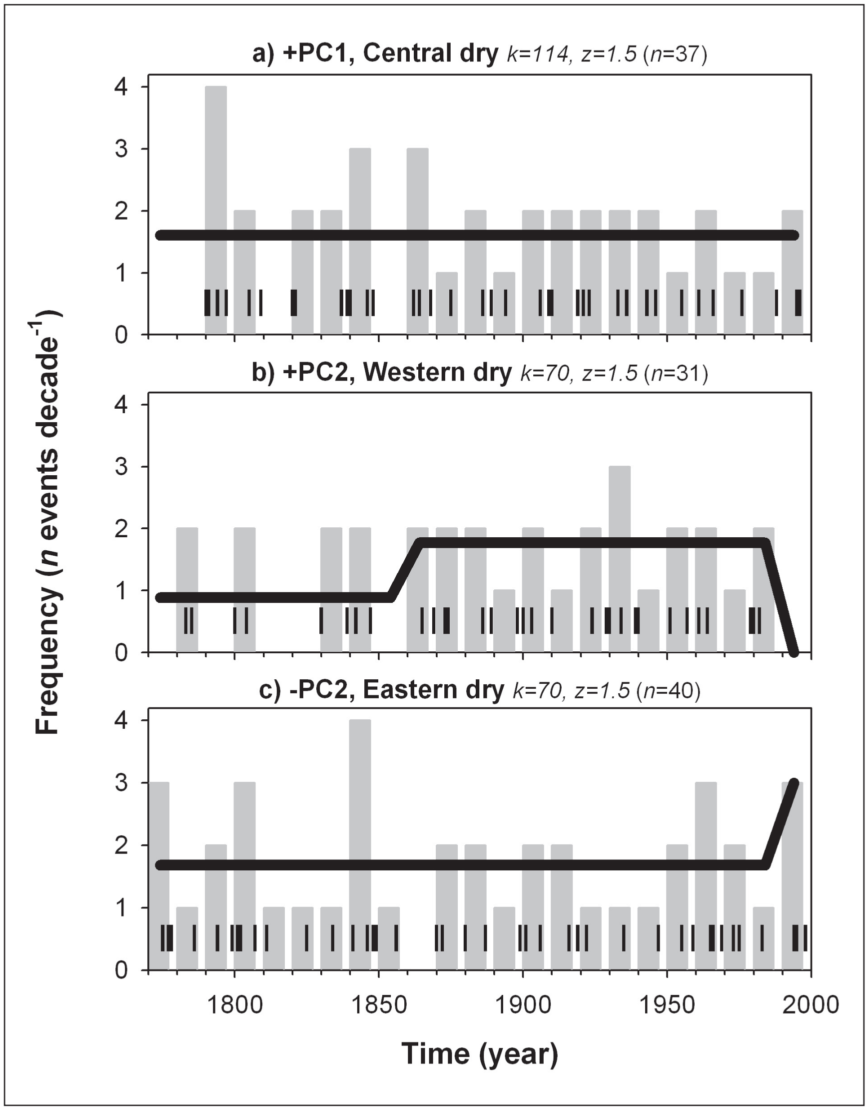

Figure 6. Frequency of extreme large-scale drought events per decade inferred from a) +PC1 scores, b) +PC2 scores and c) –PC2 scores *(grey-vertical bars)* between 1770 and 1998. The extreme events *(vertical lines)* in the drought components (Figure 3a and 3b) were detected using median smoothing (see Mudelsee, 2006 for details). Parameter values *k* (window width) and *z* (threshold selection) used for the median smoothing are indicated for each analysis; selection of *k* was guided by cross-validation. Changes in the mean frequency of events per decade *(bold line)* as investigated using regime shift detection (significance level was set at σ = 0.10; cutoff length = 10 decades) (Rodionov, 2004).

height anomalies (troughing) over the eastern Pacific and Iceland. Similarly, extreme large-scale drought events over the western regions (Figure 8b) and low drought severity over eastern regions were, on average, characterized by positive height anomalies (ridging and drying) over the Gulf of Alaska and Greenland and negative height anomalies (troughing) over eastern Canada. This configuration indicates that the wave-like pattern present in the mean climatology was accentuated. Horizontal transport of moist and unstable air from the subtropical North Atlantic basin could have contributed to the moister conditions (and low fire activity) over eastern Canada during these signature years. Finally, drought conditions over eastern regions (Figure 8c) are generally associated with positive height anomalies (ridging) over eastern

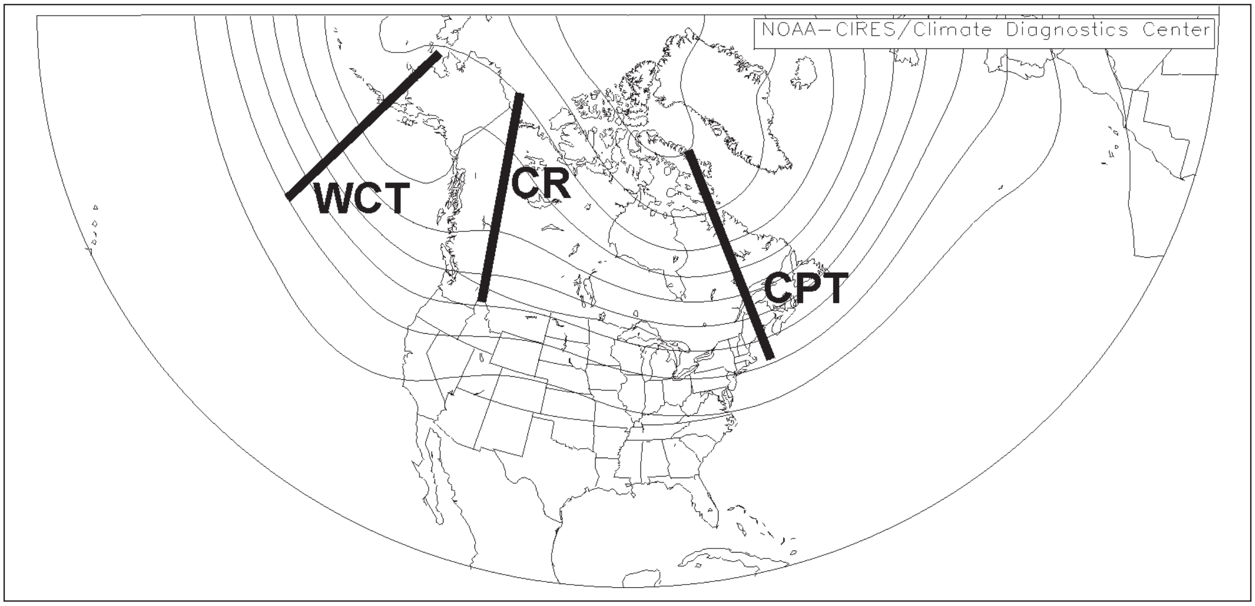

Figure 7. Mean June and July 500 hPa geopotential height composite chart (50 m contour intervals; *thin lines)* for the reference period 1968–1996. *Thick vertical lines* indicate the location of the main features affecting North American climate. From west to east these are the West Coast Trough (WCT), the Continental Ridge (CR), and the Canadian Polar Trough (CPT). Analyses were made using the NCEP/NCAR reanalysis data (Kalnay *et al.,* 1996); images were provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado (http://www.cdc.noaa.gov/). Composite anomalies (Figures 8 and 9) should be compared with this climatology.

Canada near 70°W. Moister conditions occur over the western United States. These three situations (Figures 8a, 8b, 8c) contrast with a situation in which moisture-bearing systems from the north Pacific Ocean are free to move across the continent (Figure 9). This last situation is generally associated with low fire activity over much of the Boreal Shield forests (Girardin *et al.*, 2006b).

The increasing frequency of extreme +PC2 events after 1850 (i.e., extreme drought in western regions combined with low drought severity in eastern regions) and changes in correlation between western and eastern regions, suggests that the second pattern of atmospheric circulation (Figure 8b) has become more common since the middle of the 19th century. The underlying causes for changes in atmospheric circulation and the apparent change at 1850 are not entirely understood. Analyses suggest that a coupling between the tropical ocean and the atmosphere ultimately affects drought variability over much of Canada (Girardin *et al.*, 2004a; 2006a; Shabbar and Skinner, 2004). Interactions between tropical sea surface temperature (SST) anomalies and the atmosphere can affect air mass circulation and climate over vast geographical areas (Minobe, 1997; Nigam *et al.*, 1999; Latif *et al.*, 2000; Barlow *et al.*, 2001; Bonsal *et al.*, 2001). There is evidence that warm tropical Pacific SST anomalies have increased in frequency since ca. 1850 (D'Arrigo *et al.*, 2005) but the link with the observed change in drought variability in Canada remains to be demonstrated. Summertime 500 hPa wave-like atmospheric circulation anomalies have also been associated with anomalies in the polar region's lower stratosphere (Girardin and Tardif, 2005). In the wintertime, anomalies in the stratosphere can affect the momentum of tropospheric polar longitudinal winds and can induce meridional circulations that extend to the earth's surface, causing amplification of the wave-like pattern (Thompson *et al.*, 2002). The relationship linking summertime stratospheric circulation to boreal drought variability still needs to be explored (Girardin and Tardif, 2005).

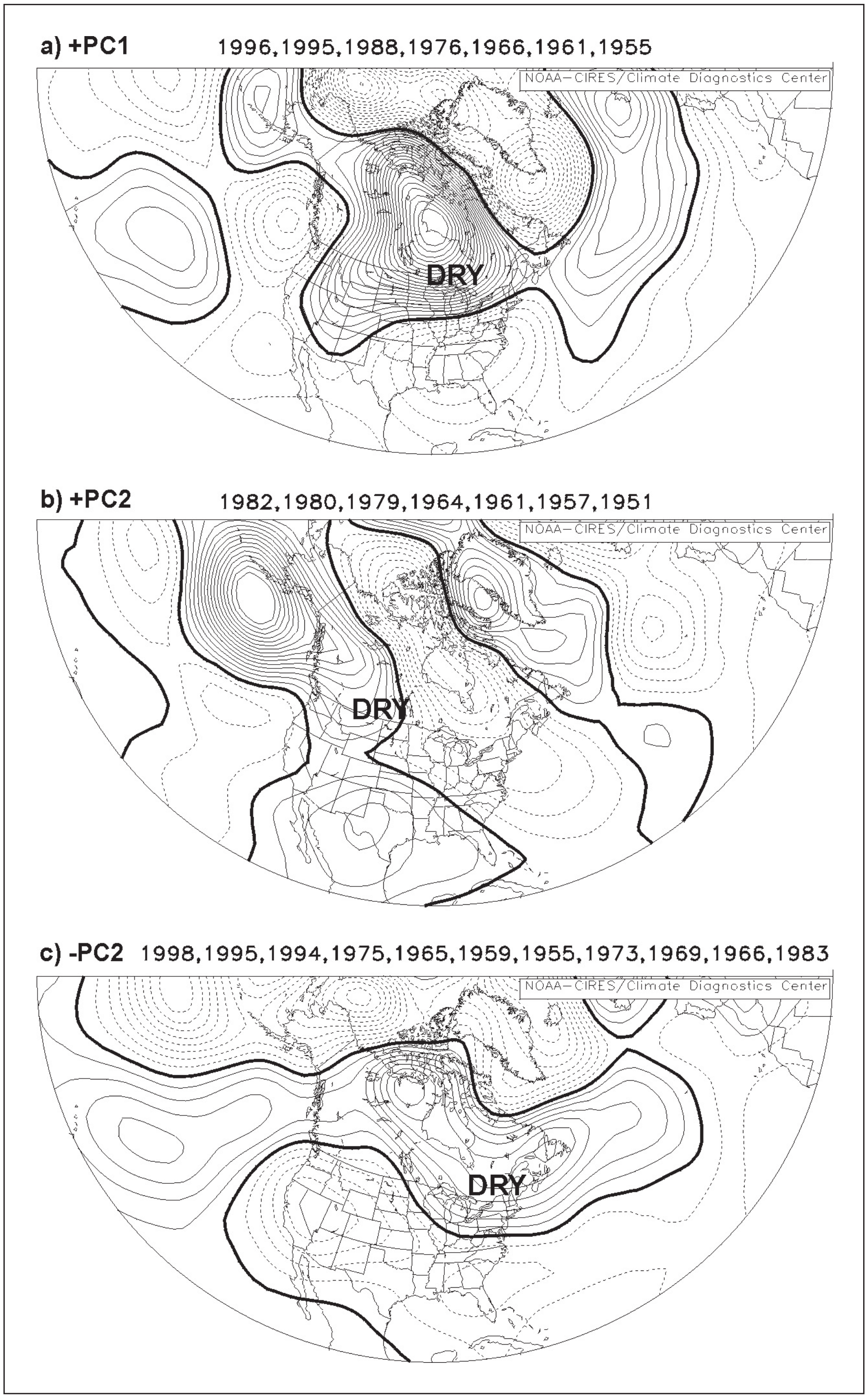

Figure 8. June to July 500 hPa geopotential height anomaly composite charts (2 m contour intervals) for years of extreme (a) +PC1 scores, (b) +PC2 scores, and (c) –PC2 scores (years obtained from the detection of extreme events, Figure 6). Regions of high drought severity (DRY) are indicated. *Solid contours* indicate positive height anomalies (ridging); *dashed contours* indicate negative height anomalies (troughing). Zero interval is bold. The anomalies were calculated from the mean of the reference period 1968-1996 (Figure 7). Years used for the construction of the composite charts are listed above each chart.

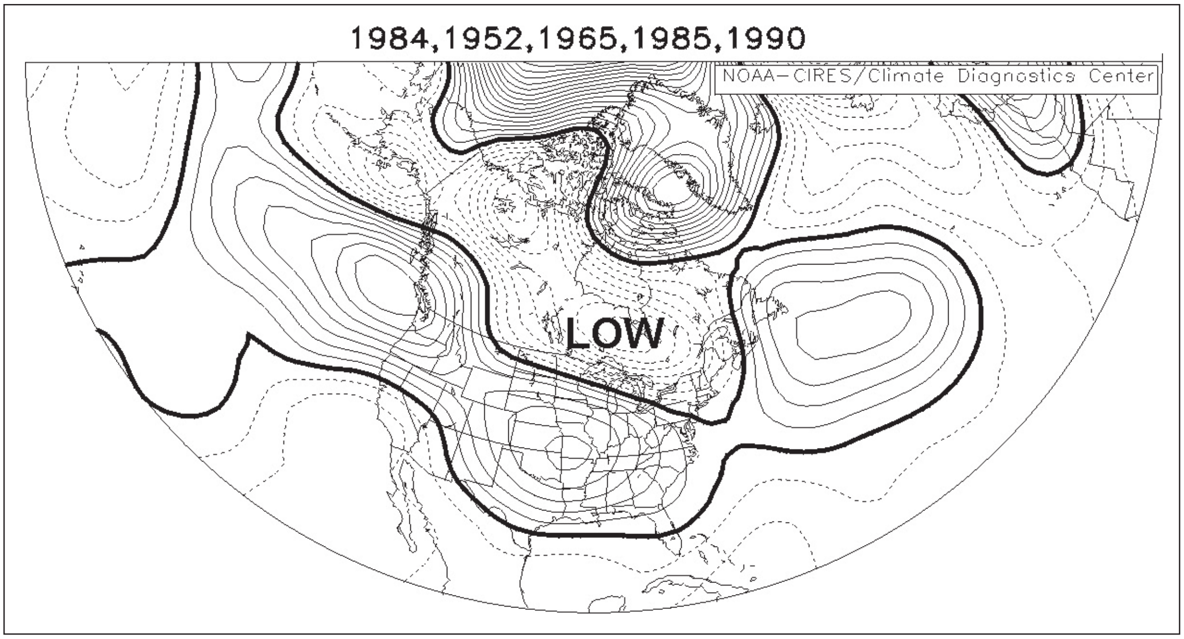

Figure 9. June to July 500 hPa geopotential height anomaly composite charts (2 m contour intervals) for the five years of lowest PC1 scores. Region of low drought severity (LOW) is indicated.

Challenges to Understanding Relationships between Fire and Climate in the Boreal Shield Region

The current network of drought reconstructions, although providing valuable information, has several limitations. Uncertainties within the calculation of the observational July CDC arise from the assumption that forest floors in Canada generally receive enough winter precipitation to saturate the CDC layer in spring (Turner and Lawson, 1978; Alexander, 1982; Girardin *et al.,* 2004a). However, exceptional years may occur in regions west of Lake Nipigon and in the Great Plains. In some situations, winter precipitation may be insufficient to prevent a seasonal drought carryover that may affect the fire season of the following year (Girardin *et al.,* 2006a; c). The current July CDC reconstructions may thus underestimate fire-conducive drought conditions originating from multi-seasonal droughts. This issue could be solved by integrating the seasonal drought carryover in the computation of the observational CDC (Alexander, 1982) and hence in its reconstruction.

Other problems may be encountered because seasonal changes in vegetation, soil type, drainage, slope, watershed, atmospheric turbidity, cloudiness, as well as many other meteorological, geological, and ecological parameters that can affect fire conditions are not accounted for by the reconstructions. Therefore, some of the variability in fire activity is unlikely to be explained by the CDC reconstructions. Also, the network does not allow comprehension of the climatic events that led to decreased fire activity over western regions (Johnson and Larsen, 1991; Larsen, 1997; Bergeron *et al.,* 2004a). Most of the area burned in northwestern Canada is due to extreme ridging over and upstream from western Canada (Skinner *et al.,* 1999) and these areas are not covered by the presented network of drought reconstructions. These issues could be addressed with the expansion of the tree-ring network. Finally, the drought reconstructions contain no information relative to century-long climate change (Wang *et al.,* 1994; Beltrami *et al.,* 2003; Laird *et al.,* 2003). The tree-ring width chronologies used for the drought reconstructions were detrended to remove growth trends (Girardin *et al.,* 2004b; 2006a), such that mainly annual to decadal scale variance was retained. The contribution of other types of proxies could be valuable in attempting to superimpose low-frequency changes on those reported in this work.

Conclusions

Climatological and ecological records obtained from tree rings provide insights on interannual to decadal scale climate variability for periods during which there are no direct observations. For the forest sector, these records provide a quantitative means for measuring the influence of climate on fire activity (Bergeron and Archambault, 1993; Carcaillet *et al.,* 2001; Westerling and Swetnam, 2003; Girardin *et al.,* 2006b). Furthermore, as opposed to observational fire records, the information provided by tree rings is less influenced by human intervention such as changes in management policies and expansion of fire detection practices. These reconstructions also: (1) provide valuable insights into the behaviour of the 'natural fire regimes'; (2) identify changes in the occurrence of extreme but infrequent events (Girardin *et al.,* 2006a); and (3) detect the presence of periodic components in the climate system that may drive forest disturbance regimes (Girardin *et al.,* 2004b; 2006c).

The six tree-ring drought reconstructions presented in the current paper provide empirical evidence of a change in spatio-temporal drought variability over the boreal forest at ca. 1850 (Girardin *et al.,* 2004b; 2006a). Based on the synoptic characteristics of recent droughts, we interpret this change in variability as a response to an increasing frequency of upper level ridging and troughing over western and eastern Canada, respectively. The increasing horizontal movement of humid air masses over eastern Canada since ~1850 favoured by increased cyclonic activity could have contributed to the creation of moister conditions that are less suitable for fire. The changes in fire activity at ca. 1850, as indicated by age distributions of forest stands reconstructed from living trees, snags, and downed woody material (Bergeron *et al.,* 2001; 2004a; b), could reflect a detectable influence of this climatic variability. That said, atmospheric circulation is highly variable and this variability likely gave rise to some historical years of extreme fire activity in the eastern boreal forest during the past 80 years or so (Girardin *et al.,* 2006b).

Atmospheric circulation associated with prolonged blocking high-pressure systems has been identified as one of the leading causes of large forest fires in Canada (Skinner *et al.,* 1999). Research has suggested that the frequency and persistence of blocking ridges in the upper atmosphere will increase in an enhanced CO_2 climate, particularly over west-central North America (Lupo *et al.,* 1997; Meehl and Tebaldi, 2004). It is likely that such variations will also influence forest fires (Bergeron *et al.,* 2004a; Flannigan *et al.,* 2005) and vegetation dynamics during the next century. Paleo-climatological and paleo-ecological studies may significantly contribute to addressing the impacts of climate change by allowing a comparison of current and future disturbance conditions with historical reconstructions (Flannigan *et al.,* 2001; Bergeron *et al.,* 2004a). Such studies will be building blocks for the development of a dynamic vegetation, climate and disturbance model, will help describe the potential impacts of climate change and allow adaptation strategies to be explored.

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Appendix A

Methods used to calculate the Canadian Drought Code have been previously described only in technical reports (Turner 1972; Van Wagner 1987). Because these sources are not commonly available, this appendix provides a detailed description of the methods used to obtain CDC records.

Moisture losses in the CDC are the result of daily evaporation and transpiration, while daily precipitation accounts for moisture gains. Evaporation and transpiration losses are first estimated as maximum potential evapotranspiration based on temperature and seasonal day length. Second, this maximum potential evapotranspiration value is scaled by the available soil moisture to reflect the fact that as soil moisture content is reduced, evaporation becomes increasingly difficult (Turner, 1972). The maximum water-holding capacity of the CDC is 100 mm for a layer with dry weight of about 25 kg/m^2, which amounts to approximately 400 percent of water per unit of mass. The minimum CDC value of zero represents soil saturation. A CDC rating of 200 indicates high drought and a rating above 300 indicates severe drought.

First, simulation of snowmelt is performed (refer to Girardin *et al.,* 2004a). At that time (day$_d$), the CDC is calculated by attributing a starting value of 15 to current drought (D)

$$\mathrm{CDC} = D + 0.5\mathrm{PET} \quad (1)$$

The calculation of the potential evapotranspiration (PET; from Thornthwaite and Mather, 1955) is given by

$$\mathrm{PET} = 0.36T + L \quad (2)$$

where T represents the maximum daily value of temperature at day$_d$ and L represents a seasonal day length adjustment. From April to October, the values of L are 0.9, 3.8, 5.8, 6.4, 5.0, 2.4, and 0.4 (Van Wagner, 1987).

The computation is carried to day$_{d+1}$ with the computation of the rainfall phase (RP)

$$\mathrm{RP} = [800/\exp(\mathrm{CDC}_{d-1}/400)] + 3.937\mathrm{ER} \quad (3)$$

The RP represents the moisture equivalent after rain and is expressed in percent of dry soil. The RP ranges from zero to 800, with 800 being saturation in water and zero the driest condition normally encountered. The RP never exceeds 800; excess water is considered as runoff and not accounted for by the CDC. CDC$_{d-1}$ represents the drought value of the previous day and ER (effective rainfall; this represents the amount of rainfall available for storage after interception by the canopy) is calculated from

$$\mathrm{ER} = 0.83P - 1.27 \quad (4)$$

where P represents the daily value of precipitation for day$_{d+1}$ above 2.80 mm (intercepted rainfall). The ER is not computed unless precipitation exceeds that amount of precipitation.

Current drought (D) for day$_{d+1}$ is computed using

$$D = 400 \ln(800/\mathrm{RP}) \quad (5)$$

where the constant 400 represents the maximum theoretical moisture content of the soil.

Using Equations (5) and (2), one may calculate the CDC (Equation (1)) for day$_{d+1}$. The procedure is carried over until October 31st. Van Wagner (1987) and Girardin *et al.* (2004a) provide more details on the CDC calculation procedure.