Recent climatological trends and potential influences on forest phenology around western Lake Superior, USA.

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Key Points:

- Two measures of growing season duration increased around western Lake Superior during 1984-2013.
- Changes in lake temperature and ice cover regime strongly affect area *T* and *P* patterns and trends.
- Changed regional *T* and *P* patterns and trends may be significant enough to modify forest phenology.

Index Terms: Biosphere/atmosphere interactions (0426); Hydroclimatology (1833); Lakes (0746); Plant ecology (0476); Regional climate change (1637)

Keywords: climate variability; forest phenology; land/atmosphere interactions; precipitation; regional warming; Upper Great Lakes

Submitted on 5 April 2016 to the AGU *Journal of Geophysical Research–Atmospheres*

Revision 1 submitted on 21 August 2016; Revision 2 submitted on 4 November 2016

This article has been accepted for publication and undergone full peer review but has not been through the copyediting, typesetting, pagination and proofreading process which may lead to differences between this version and the Version of Record. Please cite this article as doi: 10.1002/2016JD025190

ABSTRACT

We assess long-term climatological means, trends, and interannual variability around the western end of Lake Superior during 1984-2013 using available weather station data. Our results focus on changes in basic and derived climate indicators from seasonal and annual temperature and precipitation, to the traditionally defined frost-free season, to a novel definition of the climatological growing season. We describe seasonal and year-to-year climate variability that influences forest phenology, using an alternative growing season metric is based on the warm-season plateau in accumulated chilling days as an indicator of environmental triggers for vegetation growth and senescence. Our results indicate +0.56°C regional warming during our 30-year study period, with cooler springs (–1.26°C) and significant autumn warming (+1.54°C). The duration of the climatological growing season has increased +0.27 days/y, extending primarily into autumn. Summer precipitation in our study area has declined by an average –0.34 cm/y, potentially leading to moisture stress that impairs vegetation carbon uptake rates and can render the forest more vulnerable to disturbance. Many changes in temperature, precipitation, and climatological growing season are most prominent in locations where Lake Superior exerts a strong hydroclimatological influence, especially the Minnesota shoreline and in forest areas downwind (southeast) of the lake. Observed trends in lake temperature and ice phenology have also changed, coincident with a large-scale climatological regime shift around 1998. A number of factors are likely altering forest phenology and the role of the forest in the climate system of this ecologically important and highly varied forest-and-lake region.

1. Introduction

Changes in the climatology and related weather patterns of the Great Lakes region [Sousounis and Grover, 2002; Hayhoe et al., 2010; Huff and Thomas, 2014] are expected to affect both vegetation phenology [Jolly et al., 2005; Schwartz et al., 2006; Groffman et al., 2012; Gunderson et al., 2012; Jeong et al., 2013; Richardson et al., 2013] and forest disturbance regimes [Hufkens et al., 2012; Filewood and Thomas, 2014]. Forests exhibit both direct and long-term indirect feedback responses to changes in climate [Heide, 1993; Pope et al., 2013; Marchin et al., 2015], complicating our ability to predict forest phenology, estimate carbon sequestration, and represent clearly the numerous land–atmosphere interactions within the climate system [Peñuelas et al., 2009; Richardson et al., 2013]. Proximity to large water bodies such as Lake Superior can affect local and regional temperature and precipitation patterns [Changnon and Jones, 1972; Scott and Huff, 1996; Hinkel and Nelson, 2012]. The interactions of land–lake processes with climate change will be complex; spatiotemporal warming patterns and consequential changes in forest phenology may vary considerably across the Great Lakes region [Reyer et al., 2013]. In this paper we assess long-term climatological means, trends, and interannual variability in large forested areas around the western end of Lake Superior. We focus on describing changes in basic and derived climate indicators, from seasonal and annual temperature and precipitation to a novel definition of the climatological growing season.

Vegetation and forest phenology refer here to the seasonal timing of events in a tree's annual physiological cycle, including leaf budburst and growth, senescence, and (for deciduous species) abscission. Phenology also includes flowering and seed production, the generation of annual growth rings in stemwood, winter hardening and spring sap flush, and other less visible processes. From the tree's perspective, the timing of phenological events involves complex and dynamic trade-offs among numerous processes that are both driven and constrained by environmental conditions: photosynthesis balances carbon assimilation with moisture loss and nutrient transport; carbon must be allocated among leaf and stem growth, wood production [Delpierre et al., 2016], chemical defenses, and reproductive processes; the tree must protect itself against temperature-related stresses such as freezing during spring and autumn transitions [Kim et al., 2014] and moisture-related stresses in hot and/or dry periods [Arend and Fromm, 2007; Anderegg et al., 2012]. The pace of photosynthesis is acutely sensitive to environmental conditions, especially temperature [Ali et al., 2015], and varies over time with the phenological phases of the leaf and tree. A tree must constantly adjust its allocation of resources (carbon, nutrients, moisture, sunlight) according to changing biological strategies, subject to biological and abiotic limitations, in a competitive environment.

Forest phenology in our midlatitude temperate study area is driven largely by the annual temperature cycle, and phenological events can typically be ordered in “thermal time” by the accounting of cold season chilling days (CD) and warm season growing degree-days (GDD) [Baskerville and Emin, 1969; Cannell and Smith, 1983; Heide, 1993; Trudgill et al., 2005]. Meteorological and climatological factors thus significantly influence vegetation seasonal phenology at the land surface [Jolly et al., 2005; Ceccherini et al., 2014; Hwang et al., 2014; Koster et al., 2014; Xie et al., 2015] and explain a large fraction of observed year-to-year variability in temperate forest phenology [Fisher et al., 2007; Marchin et al., 2015], specifically as drivers of the annual growing season start, intensity, and duration.

Climate change in the Great Lakes region is expected to proceed at different rates for different seasons, with greater average warming expected in winter than in summer [Hayhoe et al., 2007, 2010]. Precipitation regimes are expected to change, with a greater frequency of heavy precipitation events [Groisman et al., 2012] and a diminishing proportion of winter precipitation falling as snow [Feng and Hu, 2007; Mishra and Cherkauer, 2011]. Warming winters may interfere with dormancy periods for species with leaf bud differentiation and development requirements [Rohde and Bhalerao, 2007; Morin et al., 2009; Viherä-Aarnio et al., 2014; Williams et al., 2014], which can then affect the timing of spring budburst [Cannell and Smith, 1983; Murray et al., 1989]. Young trees and undergrowth can be affected by changes in winter precipitation regimes and seasonal snow cover, especially where snowpack often insulates seedlings and soil from hard freezing [Drescher and Thomas, 2013].

A regional trend toward earlier spring green-up [Schwartz et al., 2006; Morin et al., 2009; Jeong et al., 2011] can have substantial consequences for the ecosystem: overall seasonal carbon uptake and sequestration may increase [Saxe et al., 2001; Millard et al., 2007] but new growth is also exposed to an increased likelihood of spring frost events [Hänninen, 1991]. Some species may readily adapt to changing early-season freezing regimes [Saxe et al., 2001] depending on the magnitude of interannual variability driving such changes. Winter warming along with greater variability in spring meteorological conditions may lead to more “false spring” and frost events that can damage leaves and severely hinder phenological processes through the remainder of the growing season [Rigby and Porporato, 2008; Augspurger, 2013; Peterson and Abatzoglou, 2014].

Uncertainty regarding climate change impacts on forests is even greater for autumn transitions, generally because the senescence process and its triggers remain poorly understood [Estiarte and Peñuelas, 2015; Gallinat et al., 2015]. Autumn senescence and deciduous leaf abscission occur with photosynthetic downregulation [Hörtensteiner, 2006; Guo, 2013] and nutrient conservation as defenses against freezing injury [Killingbeck, 1996; Niinemets and Tamm, 2005]. A regional trend toward later leaf senescence [Jeong et al., 2011] can lead to overall longer growing seasons [Jeong et al., 2011; Gunderson et al., 2012] and possibly to increased total primary productivity [Nemani et al., 2003; Twine and Kucharik, 2009]. However, individuals and species for which phenological triggers adjust more slowly to environmental changes may remain susceptible to both frost and drought stress later in the growing season [Saxe et al., 2001; Parida and Buermann, 2014]. Under drought conditions, a nutrient conservation process similar to winter preparation may drive leaf dormancy and senescence [Munné-Bosch and Alegre, 2004; Marchin et al., 2010], and chronic moisture stress can compromise the long-term capacity for carbon assimilation in these forests [Noormets et al., 2008; Brzostek et al., 2014; Anderegg et al., 2012, 2013, 2015].

Understanding climate change impacts on forest phenology requires both the comprehensive characterization of recent climatological variability, so that we can more accurately assess important trends, and an improved understanding of forest responses to that variability, which can differ across species and landscapes. In this paper we analyze spatiotemporal climate variability and trends during 1984-2013 in forests of the Upper Great Lakes, focusing primarily on proximity to Lake Superior and long term climatological changes to the cold and warm seasons that may have had profound, observable effects on forest phenology in the region. We employ several threshold-based metrics in common use to characterize seasonal climatology: chilling days (CD) and freezing days (FD) from autumn through spring, and growing degree-days (GDD) from spring through autumn. We introduce an alternative definition of the climatological growing season based on the accumulation of CD through the year, similar but not equivalent to the more traditional frost-free growing season. Using these and several derivative metrics, we assess recent changes in regional climatology for the 1984-2013 period and examine spatiotemporal variability in growing season influences, identifying specific areas where changes and trends have differed markedly from others during our study period. This study presents an examination of the *climatological* growing season; our ongoing work seeks to associate that influence with the *observed* forest vegetation growing season across our study area. Follow-on work will apply our findings in this paper to an observational study of forest phenology over the same area and period using remote sensing methods.

2. Study Area and Background

Our study area covers ~202,000 km^2 of sub-boreal evergreen and midlatitude mixed forest around the western end of Lake Superior in the North American Upper Great Lakes (Fig. 1). This region hosts diverse forest and wildlife species, numerous protected and managed areas including state and national forests, widespread forest-related industry, and extensive tourism and recreational opportunities. A “tension zone” [Curtis and McIntosh, 1951] that traverses our study area is defined by a combination of geographic transitions, from warm and dry continental interior to cool and wet lake-influenced landscapes. This tension zone marks the approximate southern extent of the Laurentian glaciation with resulting gradients in soil types [Schaetzl et al., 2005; Danz et al., 2013] and encompasses a gradient in natural vegetation types [Wheeler et al., 1992; Bockheim and Schlieman, 2014], from prairie and hardwood forests in the southwest (now mixed with agriculture) to sub-boreal evergreen and temperate mixed forests closer to Lake Superior. This transition is clearly visible in land cover maps of the region based on USGS National Land Cover Database products [Jin et al., 2013; Homer et al., 2015] (Fig. S1) and can be identified using USEPA [2011] ecoregion maps of our study area [Omernik et al., 2000; Omernik, 2004] (Fig. S2).

Our choice of study area reflects a particular challenge that arises from the complex geography of the region. Situated at the prairie–forest ecotone, the region borders the North American boreal forest but also has strong agricultural influences, is subject to land–lake interactions and climatological influences of the Great Lakes. Forests in this region display a wide variety in observed phenology and disturbance events, with further variety expected as a consequence of climate change impacts on all of these elements. The forests of the Upper Great Lakes are increasingly vulnerable to disturbance factors and tree mortality due to seasonal moisture stress and fire risk under changing climatic conditions [Irland et al., 2001]. This region is also sensitive to changing forest ecology and management practices, increasing recreational use and logging pressure, anthropogenic warming and climate change, and evolving conditions in the Great Lakes themselves including rapid recent warming of Lake Superior. While all of these factors may affect the forest and its role in the carbon cycle, in this paper we concentrate on the dominant climatological factors that may affect regional forest phenology and its interannual variability.

Upper Great Lakes climatology is influenced by the polar jet stream that frequently traverses the region and generally marks a continental-scale boundary between cold/dry polar air and warm/humid subtropical air along the primary midlatitude storm track. The jet stream over the region is directed by interactions between global circulations, synoptic dynamics, and climatological teleconnections [Rohli et al., 1999; Grise et al., 2013]. Both the upper-level jet stream and surface land–lake interactions drive surface temperature gradients, frontal positions, and storm meteorology [Payer et al., 2011]. Seasonality is a key factor in meteorological patterns at temperate latitudes, with synoptic variability dominating the spring and autumn transition seasons [Grover and Sousounis, 2002; Small and Islam, 2009; Small et al., 2010] and the Great Lakes providing a strong regional influence on temperature and precipitation patterns throughout the year.

Given the position of the Great Lakes near the middle of the North American continent, jet stream and storm track patterns across the region are driven by numerous teleconnections including the Pacific–North America (PNA) pattern [Rodionov and Assel, 2001], the Arctic and North Atlantic Oscillations (AO and NAO, respectively) [Nie et al., 2008; Luo and Cha, 2012], and often the combination of these with the Pacific Ocean El Niño–Southern

Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) [Bond and Harrison, 2000; Grise et al., 2013]. Climatological teleconnections have been correlated most strongly with Great Lakes regional winter conditions [Rodionov and Assel, 2000, 2003; Wise et al., 2015] and seasonal lake ice cover [Assel and Rodionov, 1998; Assel et al., 2003; Bai et al., 2011, 2012; Benson et al., 2012; Bai and Wang, 2012; Wang et al., 2012]. The warm phase of the long-period Atlantic Multidecadal Oscillation (AMO) has been associated with cold winters in eastern North America [Peings and Magnusdottir, 2014]. With a 4- to 7-year cycle, the ENSO cycle is often associated with warm/dry winters in the Upper Midwest US during El Niño (warm) years and cool/wet winters during La Niña (cold) episodes [Trenberth et al., 1998; McPhaden et al., 2006].

Several studies have associated a climatological regime shift in the Upper Great Lakes region around 1998 associated with influential long- and short-period oscillating teleconnections: the AMO entered a primarily warm phase around 1995, and the PDO entered a persistently cold phase in 1998 coincident with an anomalously strong El Niño episode during the 1997-1998 winter [Bond et al., 2003; Peterson and Schwing, 2003; Jo et al., 2014]. Changes in Lake Superior ice cover, evaporation, water level, and water temperature regimes and trends have been identified around the same time [Van Cleave et al., 2014; Watras et al., 2014]. Great Lakes surface water temperatures have warmed faster than surrounding land surface temperatures [Austin and Colman, 2007, 2008; Van Cleave et al., 2014] and Lake Superior ranks among the fastest-warming freshwater lakes in the world [O'Reilly et al., 2015]. This warming has driven changes in the lake ice regime: peak areal coverage has decreased [Howk, 2009; Wang et al., 2012], and observations indicate later ice onset in winter and break-up in spring [Assel, 2003; Howk, 2009; Assel et al., 2013]. Reduced lake ice cover allows an overall increase in lake surface evaporation through the early winter

[Brown and Duguay, 2010], resulting in greater winter lake-effect precipitation in downwind areas [Scott and Huff, 1996; Wright et al., 2013].

3. Data and Methods

We developed climatological maps and statistics using daily minimum and maximum temperature (T_{min} and T_{max}) and precipitation (P) observations from the merged Global Historical Climate Network–Daily (GHCND) dataset [Durre et al., 2010; Menne et al., 2012] for 410 weather stations in the vicinity of western Lake Superior over the period from 1 Jan 1983 through 31 Dec 2013 (Fig. 1). A number of these stations, including 83 stations in Canada and others outside of our analyzed portion of the US, are used here only for complete coverage of our study area. Our GHCND dataset and the Python code used for our analyses are available online at https://megarcia.github.io/WxCD. We relied on GHCND quality assurance flags and applied all valid station data for each date to generate gridded daily T_{min}, T_{max}, and P fields at 480-m spatial resolution using radial basis functions [Akkala et al., 2010]. This method in Python/SciPy is an exact multiquadric interpolator [Hardy, 1971, 1990; Franke et al., 1994] with results that compare favorably with other spatial interpolators [Garcia et al., 2008].

There is no explicit temporal interpolation of station values to cover periods of missing data at individual stations. As shown in Fig. 1, some weather stations cover almost the entire analysis period, while many were active for only portions of that period. We sought to generate the best possible maps on a daily basis using all available and valid station information for each day. All subsequent operations, including aggregation to seasonal averages or totals and statistical analyses for trends, are performed using this series of gridded daily observations. We evaluated spatial interpolation error characteristics by a jackknife procedure (Fig. S3) and found that our results are comparable to the Wisconsinoriented climatological study by Serbin and Kucharik [2009]. Our spatial interpolation methods produced a daily T_{min} bias (mean error) of +0.022°C and mean absolute errors (MAE) of 1.78°C, daily T_{max} bias of +0.038°C and MAE of 1.60°C, and daily P bias of +0.001 cm and MAE of 0.152 cm. These bias values are less than the reporting precision of the meteorological stations in the GHCND dataset (0.1°C and 0.025 cm for T and P, respectively) and our MAE values fall generally within expected interpolation accuracy [Garcia et al., 2008].

We calculated the daily average temperature [Cannell and Smith, 1983] from the gridded fields as

$$T_{avg} = \frac{T_{min} + T_{max}}{2} \tag{1}$$

We then accumulated chilling days (CD), chilling degree-days (CDD), and growing degree-days (GDD) [de Reaumur, 1735; Baskerville and Emin, 1969; Thompson and Moncrieff, 1982; Lechowicz, 1984] on a daily basis as:

$$CD = \sum_{i=0 \text{ at 1 Jul}}^{\text{current date}} d_i \quad \text{with} \quad d_i = \begin{cases} 1 \text{ if } T_{avg,\,i} < T_{base} \\ 0 \text{ otherwise} \end{cases} \tag{2}$$

$$CDD = \sum_{i=0 \text{ at 1 Jul}}^{\text{current date}} dd_i \quad \text{with} \quad dd_i = \max[0, (\, T_{base} - \, T_{avg,\,i})] \tag{3}$$

$$GDD = \sum_{i=0 \text{ at 1 Jan}}^{\text{current date}} dd_i \quad \text{with} \quad dd_i = \max[0, (\, T_{avg,\,i} - \, T_{base})] \tag{4}$$

We used T_{base} = 5°C following numerous empirical studies of tree physiology and spring phenology [Cannell and Smith, 1983; Murray et al., 1989; Hunter and Lechowicz, 1992;

Fisher et al., 2007; Schenker et al., 2014; Viherä-Aarnio et al., 2014; Körner, 2015]. Accepted values of T_{base} vary strongly with species and setting, and those used for analysis of forest areas can differ greatly from values used in agricultural application [*e.g.* Skaggs and Irmak, 2012]. In general, there is an inverse relationship between the value of T_{base} and the accumulation of GDD required to attain certain phenological phases such as budburst or flowering [Trudgill et al., 2005]. We also recognize that T_{base} and the biophysical efficiency of GDD and photoperiod use in early phenophases can vary widely among species and even within species across climatological settings. In the observational phase of our work, we may be able to evaluate such variety using remote sensing indicators of phenological phases. For simplicity in this climatological analysis, and based on an example discussed by Trudgill et al. [2005], we have selected a single value of T_{base} in common use for temperate tree species to be applied across our study area. On this common basis, we can use calculated GDD accumulations with vegetation greenness observations that will be incorporated in follow-on work to differentiate among fast- and slow-developing, or more and less cold-adapted, phenological types.

We also calculated the accumulation of freezing days (FD, using T_{base} = 0°C) through the cold season using both T_{avg} and T_{min} similar to eq. (2). Cold season variables (CD, CDD, and FD) are accumulated from 1 July through the following spring, and warm season variables (GDD) are accumulated from 1 January through the calendar year. We calculated seasonal temperature statistics as the 90-day mean and variance of T_{min}, T_{max}, and T_{avg}, total P over 90-day and longer periods up to a full year, and the accumulated days within a 90-day period with any ($P > 0$), moderate (1 cm $< P \leq$ 2.5 cm), and heavy ($P >$ 2.5 cm) precipitation. In total we examined 66 climatological indicators (excluding teleconnection indices; see Appendix A) including several that have been used previously to assess changing climate extremes [Frich et al., 2002; Alexander et al., 2006].

We defined the vegetation growing season in two ways. The first is a traditional definition based on the time between last spring frost and first autumn frost dates ($T_{min} < 0°C$) [Kunkel et al., 2004; Skaggs and Irmak, 2012; Yu et al., 2014; McCabe et al., 2015]. Frost dates and the duration of the frost-free season are useful for assessing long-term seasonal changes as well as acute indicators of possible vegetation freeze damage, especially in the transition from cold to warm seasons [Augspurger, 2013]. To evaluate this danger we examined GDD accumulation at the date of the last spring frost. With more accumulated GDD near the beginning of the growing season, there is a greater chance that the opening of flowers and leaves on many species makes them vulnerable to a freezing event that can adversely affect tree productivity through the remainder of the growing season. In some cases, severe late frost events have followed a "false spring" period brought on by shifts in synoptic influences over several weeks, from cold to warm and then to cold again. Two such events within our study area and period, in 2007 and 2010, will be discussed below.

Alternatively, we also defined the growing season as a function of chilling day (CD) accumulation. We observed that CD accumulation reaches a warm-season plateau soon after GDD accumulation begins in the spring. The CD accumulation then departs from that plateau in the autumn just before GDD accumulation ceases. Between these dates, $T_{avg} \geq T_{base}$ and vegetation defenses against freezing are limited while net primary productivity is generally dedicated to growth and reproduction [Schenker et al., 2014; Vitasse et al., 2014; Körner, 2015; Pagter et al., 2015]. Our resulting plateau-based growing season is several days longer than the frost-free season for the same year. The differences between these season starting and ending dates are of great interest to us: in these periods, many forest species are vulnerable to environmental conditions that could affect vegetation carbon uptake over the entire growing season, as in the case of a late spring frost, or bring the growing season to an early close with an autumn frost event that triggers leaf senescence and the tree's winter preparations.

Using our alternative definition of the growing season, we evaluated two aggregate seasonal measures. We defined the "cold season intensity" as

$$CSI = CDD_{plateau} / CD_{plateau} \tag{5}$$

Cold season duration is represented well by CD, but that measure does not indicate the severity of the season: the winter could be particularly cold or relatively mild, but CD will continue to accumulate as long as $T_{avg} < T_{base}$. Conversely, accumulated CDD alone has little phenological meaning unless it is related to a calendar duration of some importance. Together, CD and CSI provide a composite indication of both the duration and severity of the cold season. Two winters with similar CD accumulations may be differentiated by their CSI values, providing a distinction between cold and mild seasons. Although trees may not recognize a difference between one sub-freezing temperature and another, with regional warming we expect both CD and CSI (as calculated by our definitions) to diminish over time. Warmer winters can interfere with endodormancy ("winter chill") requirements for many forest species [Morin et al., 2009], leading to altered phenological cycles and reduced primary productivity in subsequent growing seasons.

Our aggregate measure for the warm season is the "growing season intensity" that we defined as

$$GSI = \frac{GDD_{plateau\ end} - GDD_{plateau\ begin}}{DOY_{plateau\ end} - DOY_{plateau\ begin}} \tag{6}$$

with DOY as the day-of-year and referring to the CD plateau as discussed above. In this case, both GDD and growing season duration remain relevant: phenological stages (*e.g.* budburst, leaf size thresholds, maturity) are frequently associated with GDD accumulation [Cannell and

Smith, 1983; Trudgill et al., 2005], and there is evidence that senescence timing depends on both inherent and external limits to leaf longevity, such as photoperiod [Kikuzawa et al., 2013]. Our GSI metric is separately useful as an indicator of temperature-related influences throughout the growing season, such as transpiration moisture demand and, if precipitation is inadequate through the season, the likelihood of vegetation moisture stress [Koster et al., 2014]. Tracking GSI through the growing season, especially using its rate of change on a short-term basis, enables sub-seasonal monitoring of vegetation status that can lead to reduced capacity for carbon uptake and growth [Teskey et al., 1987], leaf wilting [Munné-Bosch and Alegre, 2004; Marchin et al., 2010], and litter drying with evolving conditions conducive to forest fires [Yebra et al., 2013] and other disturbance agents.

To examine the possible effects of Lake Superior seasonal ice phenology on nearby land areas, we included observations of ice onset and break-up at Bayfield, Wisconsin, from two sources. We obtained records for 1984-2012 from the NOAA National Snow and Ice Data Center (NSIDC) [Howk, 2009] and supplemented those with operations records from the nearby Madeline Island Ferry Line (MIFL) for 2011-2014 [Mary Ross, 2015, personal communication]. Both of these sources indicate in their overlapping period that the area of Lake Superior near Bayfield did not completely freeze during the 2011-2012 winter season, and ice-on/ice-off dates otherwise differ between these sources by only 1-2 days, thus we considered MIFL observations a reasonable proxy for NSIDC records that were not yet available for this work.

Finally, we evaluated the possible influences of global teleconnections on our regional climatology using monthly indices reported by the NOAA National Centers for Environmental Prediction (NCEP) Climate Prediction Center (CPC) and the NOAA Earth Systems Research Laboratory (ESRL) Physical Sciences Division (PSD). From these records we calculated three-month averages for each season (DJF, MAM, JJA, and SON) for each climate oscillation and pattern described in Section 2 except ENSO. The ENSO signal is traditionally summarized over an extended winter (DJFM) period, for which we calculated the average index value in each of the reported equatorial Pacific Ocean regions in order to identify any differences in their influence: Niño-3 (eastern), Niño-4 (central), and Niño-3.4 (overlapping parts of -3 and -4).

There are a number of caveats to be noted regarding the time series analyses that we perform here. Although the 30-year period that we have examined generally meets the customary duration criterion for climatological analyses, it is a relatively short period for trend analysis. Within our study period, we have assumed and analyzed trends by linear regression instead of higher-order functions that may fit the examined time series better. We ignore possible breaks and shifts in reporting long-term trends for our 30-year analysis. Although we do examine any changes in (linear) trends across the 1997-1998 winter season that can be extracted from these time series, there is a necessary trade-off where each of those trends covers a much shorter period and is less likely to represent a long-term climatological trend. We recognize that trends important to the physical system may not demonstrate statistical significance, and that not all statistically significant trends indicate physical processes that can be justified in a conceptual model of the land–lake–atmosphere system examined here. We are particularly sensitive to the likelihood that derived climate indicators, such as plateau-based growing season duration, incorporate more basic variables that may demonstrate their own significant trends, even though a trend in the derived indicator may not itself end up being significant. Such cases can sometimes be attributed to opposing trends in the constituent indicators.

4. Results

Area-averaged 1984-2013 mean daily temperature observations (T_{min}, T_{max}, and T_{avg}) through the year are shown in Fig. 2, which also provides a conceptual overview of key temperature-related climate variables that we analyze: CD, FD and GDD accumulations, mean last spring and first autumn frost dates, and mean CD plateau beginning and ending dates. The area-averaged mean annual accumulations of CD and GDD are shown in Fig. 3 and are also marked with key dates related to the CD plateau. Area-averaged seasonal and annual precipitation (P) totals for 1984-2013 are shown in Fig. 4. Though we refer here to seasonal T and P means and trends over our study period, maps of individual seasonal indicators are included in the Supporting Information that accompanies this paper. The area-averaged climatological mean, variance, trend, and extreme values for a number of both seasonal and annual climatological indicators are summarized in Table 1. Maps of study-period mean and trend values over 1984-2013 for several key climatological indicators are shown in Figs. 5-10 with areas of statistically significant trends ($p < 0.05$) marked.

4.1 Temperature and Precipitation Indicators and Lake Superior Influences

Seasonal and annual summaries of area-averaged T and P indicators are listed in the upper half of Table 1. Maps of annual average T values and interannual trends are shown in Figure 5. Cooler temperatures in spring and summer are concentrated on the Minnesota (northern) shore of Lake Superior, and the coldest temperatures in all seasons are found in the land areas northwest of the lake (Fig. 5a, c, and e). Warmer temperatures in autumn and winter are generally concentrated on the Wisconsin/Michigan (southern) shore of Lake Superior, although some warm locations along the Minnesota shore are also apparent. The warmest temperatures in all seasons are found in the southwestern portion of our study area in primarily agricultural regions (Fig. S1). The resulting spatial gradient in T_{avg} is strongest in a general north–south orientation across the far western end of Lake Superior and more generally across the tension zone that traverses our study area. Long-term positive trends in T_{min} (Fig. 5b) are generally more widespread and stronger than those in T_{max} (Fig. 5d), which actually demonstrates some areas of long-term cooling both northwest and southeast of Lake Superior over the study period.

From the area-average trends listed in Table 1 we calculated a 30-year net warming of +0.56°C in annual mean T_{avg} for our study area. Long-term cooling in spring (net –1.26°C / 30y) has been offset by warming in all other seasons, especially autumn (net +1.54°C / 30y), with the remainder of overall 30-year net warming divided almost evenly between the winter and summer seasons. Though none of these overall temperature trends demonstrate statistical significance at the $p < 0.05$ level using area-averaged values, significant trends can be found in more localized portions of our study area. In particular, long-term warming is evident in T_{min} (Fig. 5b) and T_{avg} (Fig. 5f) on the Minnesota shore of Lake Superior and following the approximate location of the tension zone across our study area.

Maps of annual average *P* values and interannual trends are shown in Figure 6. Mean seasonal precipitation (Fig. S7) is lowest in winter and highest in summer, with generally ambiguous area-wide spatial gradients in those seasons, while spatial gradients in mean seasonal *P* across our study area are generally north-to-south in spring and west-to-east in autumn, with comparable total *P* in those seasons. These patterns result in a southeastward gradient in mean annual *P* that is most prominent across Lake Superior (Fig. 6a). We note here the lake-effect influence on spatial distributions of *P* in winter, with the areas of heaviest winter precipitation on the southern shore and immediately downwind (southeast) of Lake Superior. Winter precipitation is increasing in some areas, especially in the immediate vicinity of Lake Superior, but summer precipitation exhibits an area-average trend of –0.34 cm/y ($p < 0.01$), with areas of stronger trends along almost the entire lakeshore and extending in a number of directions from Lake Superior, especially along the tension zone in Wisconsin. Area-average reductions in summer precipitation are consistent with an overall trend in summer precipitation days ($P > 0$) of –0.25 days/y ($p < 0.001$). The resulting trends in total annual precipitation (Fig. 6b) are relatively neutral across much of our study area, with isolated locations of significant negative trends in areas where the summer trends dominate, especially south and southeast of Lake Superior.

A number of our results are consistent with the thermal influence of Lake Superior on regional climatological indicators. We found higher winter and lower summer temperatures along the northern lakeshore than in the forest region farther to the northwest, with a longer frost-free season along the lakeshore by more than 4 days. Growing season transitions, as indicated by the beginning and end of the CD plateau, are also delayed 2-3 days along the lakeshore in comparison with the forest areas to the northwest. The accumulation of area-wide winter precipitation days is positively correlated with the date of Lake Superior ice-on conditions ($p < 0.05$) and negatively correlated with ice cover duration ($p < 0.05$). The number of spring moderate precipitation days (1 cm $< P \leq$ 2.5 cm) in areas downwind (south and southeast) of Lake Superior is negatively correlated with ice duration ($p < 0.05$), a result consistent with ice cover inhibiting lake evaporation and thus lake-effect precipitation in those areas.

Our results suggest that the seasonal ice cover on western Lake Superior plays an important role in temperature and precipitation patterns on adjacent land areas. Our analyses indicated a trend in the date of Lake Superior ice-on conditions of +1.15 days/y ($p < 0.01$), though we recognize that a linear trend cannot adequately account for the large interannual variability in lake ice phenology. Notably, during autumn and winter the area of greatest precipitation in our study area is concentrated southeast of Lake Superior where prevailing northwesterly winds carry lake-evaporated moisture onshore. Conversely, spring and summer

P appears relatively suppressed in the same area. Cold-season ice cover from mid-winter through early spring inhibits lake evaporation and lake effect precipitation; during the summer the lake is generally cooler than the surrounding land areas, inducing a stable atmospheric boundary layer and inhibiting the development of storms that could then move onshore [Changnon and Jones, 1972]. Even the small trends in lake ice phenology that we found are enough to allow statistically significant increases in winter precipitation (Fig. S7) southeast of Lake Superior. However, spring precipitation seems also to increase in those same areas, which we cannot attribute to the slightly later lake ice breakup over time. In that case, our analyses may require a finer temporal division of seasonal precipitation in areas downwind of the lake so that portions of the spring season before and after the observed ice-off date in each year are treated separately. Such a division could help demonstrate the transition from winter (cold) to spring (warm) weather patterns and phenological processes in the area of lake influence.

4.2 Cold and Warm Season Indicators and Interseasonal Variability

As expected, the greatest accumulation of CD in our study area occurs in the forest areas northwest of Lake Superior (Fig. 7a). The 30-year trend in CD accumulation (Fig. 7b) is generally mixed across the study area, with only small areas of statistically significant reduction in CD accumulation that are generally coincident with locations of winter T_{min} and annual T_{avg} warming. There is slightly greater spatial variability evident in the detail of 30-year mean CSI (Fig. 7c) which is lowest in the immediate vicinity of Lake Superior and in downwind regions. This result, and decreasing CSI over time in many areas on the lakeshore (Fig. 7d), is consistent with the warming surface temperatures and changing ice phenology of Lake Superior.

The duration of the frost-free season (Fig. 8f), our traditional measure of the climatological growing season, shows an area-average trend of +0.32 days/y ($p < 0.05$) over our 30-year study period (Table 1) that is consistent with previous studies [Easterling, 2002; Kunkel et al., 2004; Yu et al., 2014; McCabe et al., 2015]. There is much spatial detail in maps of trends in the last spring and first autumn frost dates (Figs. 8b and 8d, respectively), with some areas experiencing statistically significant changes in both indicators consistent with lengthening of the frost-free season. Most of these areas occur in the forested areas northwest of Lake Superior and along the tension zone south and southeast of the lake.

We found quite different patterns for the CD plateau (Fig. 9), our alternative measure of the climatological growing season. The beginning of the CD plateau changed little over time in the area average but shows a large area downwind of Lake Superior with significant trends toward a later start to the climatological growing season (Fig. 9b). The end of the CD plateau showed an area-average trend of +0.27 days/y ($p < 0.05$) over our 30-year study period with large areas of significant trends (Fig. 9d) that are consistent with significant autumn warming. These include areas on the Minnesota (northern) shore of Lake Superior and across agricultural areas in the southwest portion of our study area, but notably not in the region immediately southeast of the lake. Trends in the duration of the CD plateau are strongly mixed across our study area (Fig. 9f), with an overall longer plateau evident in many areas where the end of the plateau now extends further into autumn, but a shorter plateau in areas directly upwind and downwind of Lake Superior. Our maps of plateau GDD accumulation, GSI, and their trends (Fig. 10) are consistent with slightly warmer climatological growing seasons over time, especially for isolated areas in immediate proximity to the lake.

We found throughout the year that seasonal mean average temperature is inversely correlated with the number of seasonal precipitation days ($P > 0$) but not with accumulated seasonal P, demonstrating that cloudy seasons are cooler seasons overall. We also found several significant (p < 0.05) interseasonal correlations among *T* and *P* indicators (Fig. 11), especially between winter and spring conditions. Winter *T* and *P* indicators are strongly tied to *T* and the number of precipitation days in the following spring. However, by these methods we found no statistically significant interseasonal correlations that might be used to extend *T* and *P* predictability beyond spring and through the remainder of the growing season.

4.3 Teleconnection Influences and a Climatological Regime Shift

Two large-scale teleconnections examined here, the AMO and PDO, demonstrated regime transitions around the middle of our study period in the course of their long-period oscillations. NOAA observations indicated that the AMO index increased during 1984-2013 and around 1995 shifted to a warm (AMO+) phase that is generally associated with drier conditions in the Upper Midwest U.S. [Enfield et al., 2001]. The PDO index generally decreased through our 30-year study period, with observations of distinct PDO+ periods (indicating warmer northeastern Pacific coastal waters) in the mid-1980s and mid-1990s followed by PDO– anomalies (cool northeastern Pacific coastal waters) beginning around 1998. Our results indicate that the AMO (PDO) index is positively (negatively) correlated with study area *T*, winter *P*, and the date of Lake Superior ice onset. The AMO (PDO) index is negatively (positively) correlated with summer *P* and lake ice duration. Local correlations with Pacific teleconnections (PDO, ENSO, and PNA) appear strongest in winter and spring, with Pacific indices overall positively correlated with *T* and negatively correlated with *P* in the Upper Great Lakes region. A long-term increase in autumn PNA index values (p < 0.05), and positive values of the PNA index in general, has been associated with relatively dry conditions over continental North America [Leathers et al., 1991]. The often-related AO and NAO indices showed no significant correlation with study area surface climatology except in autumn, when the AO index is negatively correlated with area-averaged *P*.

A distinct shift in climatological regimes across several global and regional indicators was thus observed around 1998. Our analyses indicated changes to, and even reversals of, statistically significant ($p < 0.05$) temporal trends for several climatological indicators between the 1984-1998 and 1998-2013 periods. Specifically, paired maps of trends in spring and autumn T_{avg} and annual P are shown in Fig. 12. Many locations with strong spring cooling during 1984-1998 (Fig 12a) shifted to near-neutral trends in the latter half of our study period (Fig. 12b), which can be attributed primarily to a shift in the trend of spring T_{max} with very little change in spring T_{min} trends across much of our study area. Locations with weak autumn warming during the earlier period (Fig. 12c) shifted to strong cooling trends in the 1998-2013 period (Fig. 12d), which we attribute to sharp reversals in both autumn T_{max} and autumn T_{min} trends across our study area. Despite this reversal to statistically significant cooling in many locations, our strongest 30-year area-averaged seasonal warming still appears in autumn (Fig. S6 and Table 1). Annual total P shifted from increasing trends in isolated locations during the 1984-1998 period (Fig. 12e) to widespread decreasing trends across much of our study area in the 1998-2013 period (Fig. 12f). We attribute this shift primarily to an area-average negative trend in summer P ($p < 0.01$, Fig. S7 and Table 1) that is also strongest in the latter half of our study period, and to a lesser extent diagnosed shifts to decreasing seasonal precipitation trends that we also found for winter and spring.

5. Discussion

Using available weather station data, we have examined the mean seasonal and annual climatology, temporal trends, teleconnection correlations, and the potential influences of Lake Superior on surrounding land areas during 1984-2013. Our study area is characterized by an extensive and ecologically important forest-and-lake landscape at the Upper Midwest U.S. prairie–forest ecotone, where we anticipate observable sensitivity to recent and ongoing climate change. Our results indicate regional warming of +0.56°C through our 30-year study period supported by mixed seasonal changes, with spring area-average cooling offset primarily by autumn warming. Long-term trends include warmer winters, wetter winter and spring seasons, a diminishing duration of Lake Superior ice cover, and a strong decrease in summer *P* since the 1998 PDO regime shift and El Niño event.

Using both traditional and new definitions of the growing season, we have identified an overall extension of the climatological growing season into autumn. Our new definition, based on a warm season plateau in CD accumulation, provides additional information for diagnosing early-season freezing risks and later periods of potential vegetation moisture stress, both of which can have a strong influence on seasonal and interannual variability in vegetation phenology. However, some risks such as late spring frost and “false spring” events may remain location-specific, and are not characterized well with area-averaged metrics.

Several prominent global and hemispherical climatic teleconnections correlate in varying degrees with seasonal and interannual variability in regional hydroclimatology and with lake ice phenology, all of the factors affecting seasonal *T* and *P* patterns in our study area. Changes in Lake Superior ice cover regimes have contributed to greater winter and spring lake-effect *P* in portions of our study area south and southeast of the lake. Lake Superior clearly affects *T* and *P* patterns in nearby land areas through both proximity and an interseasonal lag due to thermal inertia, with effects that are likely strong enough to alter forest phenology in those areas.

5.1 Climatological and Growing Season Trends

The overall area-average temperature increase that we estimated by linear regression is consistent with prior estimates of warming in this region over similar periods [Li et al., 2010; Groisman et al., 2012]. Our results show a consistent continuation of trends reported for

1951-1980 climatic changes in the Great Lakes region reported by Scott and Huff [1996]: increased minimum temperatures in all seasons, decreased spring and summer maximum temperatures, slight decreases in summer rainfall, and large winter precipitation increases for lake-effect areas primarily southeast of Lake Superior and as winter ice cover diminished. Choi et al. [2014] analyzed Serbin and Kucharik's [2009] climatological dataset for Wisconsin and showed statistically significant decreases in the frequency of heavy precipitation events during 1950-2006, especially southeast of Lake Superior in the same lake-effect areas where we also found slight decreases for heavy precipitation days ($P > 2.5$ cm) in summer and autumn. However, our analyses indicated slight increases in winter and spring heavy precipitation days in those areas. These differences with Choi et al. [2014] might be attributed to differing study periods but are likely related to technical differences (*e.g.* study area boundaries, station data selection, interpolation methods, grid resolution) in the examined precipitation fields that can demonstrate large spatial variability.

Our estimated +0.56°C area-average temperature change, along with the accelerated warming of Lake Superior surface waters (+2.5°C) [Austin and Colman, 2007, 2008; Van Cleave et al., 2014], clearly indicates regional warming over the 30-year study period. Though the beginning of the CD plateau in spring has changed little, we have noted the extension of the CD plateau later into autumn consistent with long-term temperature increases concentrated primarily in that season. Overall these results point to a longer climatological growing season due more to warm-season extension later into autumn than earlier into spring, consistent with findings by Jeong et al. [2011]. Annual total precipitation demonstrated a slight negative area-average trend over the study period, and a particularly strong negative trend since 1998 (Fig. 12) driven by a sharp decrease in summer precipitation (Fig. S7). Our analyses of temperature trends support the small negative trends that we found for cold season CD and FD and a consequent increase in the duration of the frost-free season.

Along with observed influences of large-scale teleconnections (AMO+, PDO–, and increasing PNA), these changes suggest an overall drying trend [Parida and Buermann, 2014] within our study period for the region.

Area-wide climatological warming and drying during 1984-2013 means that forests in our study area may have experienced increasing moisture stress in that period, especially in summer. It is important here to distinguish between changes affecting moisture availability, based primarily on *P*, and those that drive vegetation moisture demand, based primarily on *T*. Vegetation moisture stress occurs with the combination of these, when moisture availability is insufficient to meet moisture demand. Conditions promoting moisture stress can inhibit transpiration and reduce growth [Teskey et al., 1987], promote leaf wilt and early senescence [Munné-Bosch and Alegre, 2004; Marchin et al., 2010], enhance tree mortality [Anderegg et al., 2012, 2013], and reduce overall forest carbon uptake [Brzostek et al., 2014; Koster et al., 2014].

For tree species that adapt quickly to changing climate conditions and phenological cues, a longer climatological growing season may drive changes in growing season net primary productivity [Nemani et al., 2003; Twine and Kucharik, 2009]. A warmer spring is typically associated with an earlier start to the vegetation growing season, provided adjustment to the timing and speed of leaf growth (subject to freezing risks) to these warmer conditions. Higher autumn temperatures may support longer vegetation growing seasons for those forest species that can adjust their leaf longevity and senescence triggers to warmer autumn conditions. However, we are still learning to identify and understand the many cues for leaf senescence that influence leaf phenology, including photoperiod and temperature [Kikuzawa et al., 2013; Ali et al., 2015] and biochemical limits on leaf longevity [Keenan and Richardson, 2015; Seki et al., 2015].

5.2 Seasonal Transitions

Climate during the shoulder seasons can be critical to vegetation phenology. A lack of statistically significant interseasonal correlations between summer and the shoulder seasons (Fig. 11) leads to uncertainty regarding aspects of forest phenology such as the progression of green-up, seasonal peak greenness in early summer, and deciduous autumn senescence with follow-on effects for the related winter feeding patterns of overwintering herbivores. Summer and autumn conditions strongly influence seed production in many forest species, with consequences for wildlife feeding and reproduction patterns [Yang et al., 2010]. These growing season conditions affect biochemical processes that control leaf bud set and hardening prior to late autumn freezing, with consequences for optimum productivity in the next growing season [Vitasse et al., 2014; Estiarte and Peñuelas, 2015]. Seasonal climatological conditions can thus affect forest phenology and overall primary productivity in the same season and well into the forest life cycle [Noormets et al., 2008; Anderegg et al., 2012, 2013, 2015; Brzostek et al., 2014].

Regarding late frost events and false spring occurrences, we found an area-average trend of –2.1 degree-days/y ($p < 0.05$) in the accumulation of GDD before the last spring frost. This trend, along with a slight trend toward earlier last spring freezing nights (Table 1), suggests a slowly diminishing risk of vegetation-damaging spring frost events over our study period. This result is not necessarily consistent with the frequency-based examination of frost-based spring vegetation damage [Augspurger, 2013] that suggests an increasing frequency of occurrences in the Midwest US. We take particular note of “false spring” and late frost events that were observed in the US Upper Midwest in 2007 [Augspurger, 2009; Gu et al., 2008] and 2010 [Fereday et al., 2012; Hufkens et al., 2012; Filewood and Thomas, 2014; Ning and Bradley, 2014]. For these years the area-average accumulations of GDD before the last spring frost were 147 and 156 degree-days, respectively, both very close to the

30-year average for that metric, and with earlier-than-normal last spring frost dates in both years. By contrast, a spring heatwave in 2012 [Ellwood et al., 2013; Peterson et al., 2013] led to an area-average accumulation of 200 GDD by the time of the last spring frost, which also occurred earlier than the 30-year normal date, yet no particular frost-related damage was reported across the region. Hypothetically, we expect higher-than-normal GDD accumulations (suggesting greater likelihood of early leaf growth) before a possibly later-than-normal last spring frost date for these events. However, false springs in ENSO-neutral 2007 and El Niño-influenced 2010, but not in La Niña-influenced 2012, ran contrary to that expectation. Under that hypothesis, we further suspect that "false spring" events occurred for portions of our study area in 1986 and 1992, as indicated in Table 1. We anticipate that our continuing observational study of regional forest phenology will provide us with greater insight into these occurrences, given the potential importance of spring late frost events to seasonal phenological patterns and their impact on tree growth.

The seasonal and inter-seasonal climatological relationships that we have found may promote some interesting phenological patterns. Within a season, the inverse relationship between average temperature and precipitation days (as a proxy for cloudy days) corresponds to days with greater water availability (from precipitation) also having lower evaporation and transpiration moisture demand. A cycle may be established in which forest soil and canopy moisture are allowed to build up during periods with cloudy/wet days and then deplete during periods with clear/dry days. Lower leaf primary productivity in cloudy periods may then be offset by enhanced productivity during clear periods when supported by greater soil moisture availability. Increased soil moisture is also conducive to seed germination and seedling growth, especially for shade-tolerant species, and can thus have potential impacts on mixed forest understory structure and species distributions [Nowacki and Abrams, 2008].

Conversely, dry soils and forest litter can lead to inhibited soil respiration, slower litter decay, and potential interference with seed germination and seedling growth, also altering forest structure and composition over time [Gustafson and Sturtevant, 2013; Peters et al., 2015]. Reduced moisture availability can eventually lead to a shift of the surface energy balance away from latent heating (evaporation, transpiration) in the growing season to greater sensible heating at the surface and in the forest canopy, a positive feedback cycle that can enhance local warming and exacerbate forest canopy moisture stress [Anderegg et al., 2012]. A warmer spring and earlier start to the growing season may therefore compensate for a cooler summer in some years in terms of total growing season primary productivity, but the reverse is not necessarily true: a warm summer may not compensate for a cool spring and a late start to the growing season, but instead exacerbate temperature and moisture stresses that reduce forest productivity throughout the growing season. Numerous influences generate complex interactions around the Upper Great Lakes, and the lakes themselves contribute to spatiotemporal variability in the system through their internal mixing regimes, providing some “memory” of conditions across seasons (via thermal inertia) and potentially over several years [Bennett, 1978; Gerten and Adrian, 2001; Piccolroaz et al., 2015].

5.3 Teleconnection Influences

We are concerned not only with the mean climatology but also its year-to-year variability and extremes, as these also influence forest phenology and growth [Bouriaud et al., 2005; Voelker et al., 2012]. We may generalize for the Upper Midwest U.S. that El Niño (ENSO+) events foster warm/dry years and La Niña (ENSO–) events support cool/wet years: Table 1 lists extremes in many of our climatological indicators that can be associated with ENSO events within our study period. El Niño conditions that persisted through 1987 and 1988 produced the driest winter and warmest growing season in our study period in conjunction with regional drought conditions [Trenberth et al., 1988; Weaver et al., 2009]. On the other hand, Lake Superior did not fully freeze during both the El Niño winter of 1997-1998 [Changnon, 1999] and the La Niña winter of 2011-2012 [Peterson et al., 2013; Dole et al., 2014], the latter event leading into one of the most spatially and temporally variable growing seasons during our study period [Hoerling et al., 2014]. Teleconnection variability may have influenced the wettest year of our study period in 1985 (ENSO–, AMO–, and PDO+) and the driest year in 2006 (ENSO–, AMO+ and PDO–), both La Niña years. It remains to be seen whether our observed teleconnection influences are part of a persistent long-term trend or might change, such as with eventual regime shifts in the long-period AMO and PDO.

Understanding interannual climatological variability, combined with interactions of the land–atmosphere system, in our study area is clearly more complex than reliance on any single element such as teleconnections [Wise et al., 2015] or another external indicator will allow. As another example, Weaver et al. [2009] analyzed the 1993 Upper Midwest summer flood event, which was supported by both AO– and NAO– conditions, and found that moisture from the distant Gulf of Mexico contributed significantly to large rainfall totals in the Great Lakes and Upper Mississippi River regions that year. An event or season can reach climatological extremes by the interaction and reinforcement of influences: the NAO phase, which can drive synoptic organization over the Great Lakes but may not necessarily generate a significant event on its own, interacted with the alignment of AO-driven meridional flows to drive the northward transport of Gulf moisture along the entire length of the Mississippi River valley.

5.4 Lake Effects

Climatological characteristics are distinguished across the western end of Lake Superior and reflect the influence of the lake on *T* (northwestern shore) and *P* (to the southeast), both of which can be linked to the thermal inertia of Lake Superior surface waters during seasonal transition periods. Warm lake surface waters in autumn and winter enhance the land–lake temperature contrast, contributing also to strong *P* gradients across the lake from lower accumulations in the northwest to higher totals to the southeast. Delayed ice formation in warm winters allows a longer period of surface evaporation, feeding lake-effect *P* maxima to the southeast through both autumn and winter. Conversely, Lake Superior ice cover that extends well into spring in some years may contribute to diminished spring *P* as well as delayed last spring frost dates for that area. Because of the thermal and moisture effects of the lake, we anticipate that vegetation phenological transitions in the spring season for areas south and east of Lake Superior are typically delayed several days, possibly weeks, compared with those transitions west and north of the lake.

The influence of Lake Superior and its own changes over time on the climatology of our study area is substantial, including both recent lake warming [Austin and Colman, 2007, 2008; Van Cleave et al., 2014; O'Reilly et al., 2015] and changing ice phenology [Assel, 2003; Howk, 2009; Wang et al., 2012; Assel et al., 2013]. Observational studies regarding the effects of the Great Lakes on their surrounding land areas [*e.g.* Li et al., 2010] remain essential to our growing understanding of land–atmosphere processes in the surrounding forest areas. Given their prevalence on the study area landscape (Fig. S1), the roles of smaller lakes in regional climatology and forest phenology are also of interest to our work [Johnson and Stefan, 2006; Plank and Shuman, 2009; Mishra and Cherkauer, 2011; Mishra et al., 2011a, 2011b]. Overall, while we remain interested in phenological events in the autumn season and the potential effects of Lake Superior thermal inertia on delayed timing of those events in the vicinity of the lake, the spring transition and its complexity because of that proximity is vital for understanding phenological transitions at the start of the vegetation growing season and will be an interesting point of focus in future work.

5.5 Potential Applications

Our analyses indicate that climatological influences are strongly heterogeneous across our study region but may be relatively homogeneous, or at least similar, over smaller distances on the order of 100 km. Analysis of climate trends in this area requires detailed examination of patterns at the sub-regional scale. We cannot assume that climatology and its influences on forest areas southeast of Lake Superior are similar to those in areas northwest of the lake, or even that climatology is consistent throughout the nearby ecoclimatic tension zone. Along the northern shore of Lake Superior in Minnesota, an analysis of climatological influences on forest phenology and health must include localized details. Different regimes of land–atmosphere interaction among these regions may require differing approaches to the identification of dominant influences on forest phenology, the causes of forest disturbances, and the processes involved in post-disturbance forest recovery. Recent and continued climatic changes may promote altered trajectories of forest health (*e.g.* due to moisture stress and other disturbance factors) and forest species composition (favoring more temperate and drought-resistant species) [Duveneck et al., 2014a, 2014b], complicating options for regional forest management toward long-term goals [Rittenhouse and Rissman, 2015]. Trophic interactions between forest vegetation, insects, and wildlife become more uncertain with altered timing of phenological events [Foster et al., 2013; Roberts et al., 2015]. Forest managers with responsibility for planning winter and dry-season harvest territories, or for undertaking intervention activities to mitigate forest disturbances due to insect pests and fire, can benefit from any potential improvements in seasonal predictability.

We seek a better understanding of recent climate and forest phenological variability to support the development and validation of land surface models that incorporate both permanent and transient land cover change as well as seasonal vegetation processes [White et al., 1997]. There is an increasing need for improved accuracy in representation of land surface states and processes as regional and global climate models progress from relatively coarse (Δx = 10-100 km [Prein et al., 2015]) to finer representative scales. As we develop a better understanding of this coupled system at a range of spatial and temporal scales, we will enable the capability to model feedbacks between forest phenology and climate change at local scales [Peñuelas et al., 2009; Richardson et al., 2013; Brzostek et al., 2014], essential for accurate assessments of forest carbon states and their spatiotemporal variability [Desai et al., 2008; Jeong et al., 2013; Schwartz et al., 2013; Lu et al., 2015].

In regions such as the Great Lakes, we must exercise caution: most land–atmosphere models do not yet incorporate even large lakes in the physical system, producing often significant errors in the simulation of temperature and moisture fluxes [Bryan et al., 2015]. A number of researchers have pursued better parameterizations of lake processes for use in energy- and water-balance models [Plank and Shuman, 2009; Mishra et al., 2010] and for proper representation of lakes in regional climate models [Gula and Peltier, 2012; Bennington et al., 2014; Mallard et al., 2014]. It is important that large and small lakes, their surface temperatures and winter ice phenology, and their influences on surrounding areas are properly included in analyses of climate change effects on temperature and precipitation patterns and trends, so that regional modeling efforts [Hayhoe et al., 2010; Gula and Peltier, 2012; Mallard et al., 2014; Harding and Snyder, 2015; Notaro et al., 2015] can provide more accurate insight and guidance to scientists and decision-makers.

6. Conclusions

A wide variety of environmental factors contribute to seasonal and year-to-year variability in forest phenology around the Upper Great Lakes. In this work we have identified several trends showing recent and likely ongoing changes to the climatological growing season and the availability of moisture to regional forests. Our results also show large spatial variability in trends such as autumn and winter warming, diminishing summer precipitation, and the extension of the climatological growing season into autumn. Dry summers can lead to moisture stress that impairs carbon uptake rates and may render the forest more vulnerable to disturbance factors.

Our new definition of the climatological growing season, based on a warm-season plateau in accumulated chilling days, provides information that can be combined with observations of the traditional frost-free growing season to show the speed, and potential dangers to vegetation, of the transition seasons. The timing and rates of spring green-up and autumn senescence can have large variability, and have significant impacts on the seasonal carbon assimilation capacity of affected forest areas. The extension of warm periods into autumn can alter the environmental cues for leaf senescence and the preparation of trees for winter conditions, potentially impacting phenology in the following growing season as well. It is important to note that, while meteorology changes with the season, and climate factors change from year to year, trees in the forest maintain memory of these conditions through the impacts on phenology and growth in subsequent seasons and thus over a lifetime.

Adding to the complexity of the study area, Lake Superior exerts an identifiable influence on surrounding and especially downwind land areas. The warming lake, with its changing winter ice phenology, contributes to winter warming in the immediate vicinity of the lakeshore. One of the greatest potential influences of Lake Superior on regional phenology is the contribution of longer ice-free periods in autumn and winter to lake-effect precipitation in large forest areas south and southeast of the lake. The resulting extension of cold periods later into spring may delay spring green-up in those areas at the time of year when seasonal leaf expansion and wood production are vital to the relative competitive advantages of different species in mixed forests, affecting both carbon sequestration and resource optimization to sustain optimum growth rates through the remainder of the warm season. Forest management in the context of climate change must therefore consider the range of factors affecting the critical green-up and senescence periods bounding the growing season.

Finally, our work provides valuable support for an observational effort to understand long-term changes in vegetation phenology including seasonal environmental cues and the impacts of climatological variability on forest phenology, disturbances, and post-disturbance forest recovery. All of these processes inform our developing understanding of forest responses to recent and ongoing climate change. Feedbacks from surface conditions (including lakes, land cover, land use, snow cover, vegetation phenology, and forest disturbances) to the climate system [*e.g.* Sobolowski et al., 2010; Richardson et al., 2013; Rydzik and Desai, 2014] are integral to the co-evolution of the land–atmosphere system and thus important to any examination of regional phenoclimatology.

ACKNOWLEDGEMENTS

We acknowledge Prof. Ankur Desai’s comments regarding an early draft of this work. Comments and suggestions by two anonymous reviewers significantly improved our manuscript. Support for this work was provided by the USDA Forest Service–Northern Research Station, USDA NIFA McIntire–Stennis funding to the University of Wisconsin–Madison (project WIS01621), UW–Madison College of Agricultural and Life Sciences, UW–Madison Department of Forest & Wildlife Ecology, NASA Biodiversity and Ecological

Forecasting Project NNX14AC36G, and the first author’s 2015-2016 Wisconsin Space Grant Consortium Graduate Research Fellowship. This work employed the computing resources and assistance of the UW–Madison Center for High Throughput Computing (CHTC) in the Department of Computer Sciences. The CHTC is supported by UW–Madison, the Advanced Computing Initiative, the Wisconsin Alumni Research Foundation, the Wisconsin Institutes for Discovery, and the National Science Foundation, and is an active member of the Open Science Grid supported by the National Science Foundation and the US Department of Energy's Office of Science. Our GHCND station dataset and Python analysis code are described in the Supporting Information (Supplemental Appendix B) and available online at https://megarcia.github.io/WxCD.

APPENDIX A

Seasonal and annual climatological indicators examined in this work. Traditional seasons are defined as 90-day periods ending on the equinox and solstice dates.

Temperature indicators (from spatially interpolated GHCND station data records)

- Winter mean and temporal standard deviation of T_{min}, T_{max}, and T_{avg} (°C)
- Spring mean and temporal standard deviation of T_{min}, T_{max}, and T_{avg} (°C)
- Summer mean and temporal standard deviation of T_{min}, T_{max}, and T_{avg} (°C)
- Autumn mean and temporal standard deviation of T_{min}, T_{max}, and T_{avg} (°C)
- Annual (365-day) mean and temporal standard deviation of T_{min}, T_{max}, and T_{avg} (°C)

Precipitation indicators (from spatially interpolated GHCND station data records)

- Winter total precipitation (cm) and precipitation days ($P > 0$)
- Winter moderate (1 cm < $P \leq$ 2.5 cm) and heavy ($P >$ 2.5 cm) precipitation days
- Spring total precipitation (cm) and precipitation days ($P > 0$)
- Spring moderate (1 cm < $P \leq$ 2.5 cm) and heavy ($P >$ 2.5 cm) precipitation days
- Summer total precipitation (cm) and precipitation days ($P > 0$)
- Summer moderate (1 cm < $P \leq$ 2.5 cm) and heavy ($P >$ 2.5 cm) precipitation days
- Autumn total precipitation (cm) and precipitation days ($P > 0$)
- Autumn moderate (1 cm < $P \leq$ 2.5 cm) and heavy ($P >$ 2.5 cm) precipitation days
- Annual (365-day) total precipitation (cm) (evaluated on 31 December)

Lake Superior (from NSIDC/MIFL records)

- Ice-on and ice-off dates (day-of-year)
- Ice duration (days)

Cold season indicators (derived from daily gridded dataset)

- CD and CDD at beginning of plateau (accumulated from 1 Jul using T_{avg})

- FD at beginning of plateau (accumulated from 1 Jul using T_{min})
- Cold season intensity (CSI, calculated using plateau CDD and CD, in °C)
- Date (day-of-year) of last spring freeze/frost (using T_{min})
- Date (day-of-year) of first autumn freeze/frost (using T_{min})

<u>Warm season indicators (derived from daily gridded dataset)</u>

- Duration of frost-free season (days)
- Date (day-of-year) at beginning of CD plateau (using T_{avg})
- Date (day-of-year) at end of CD plateau (using T_{avg})
- Duration of CD plateau (days)
- GDD at date of beginning of CD plateau (accumulated from 1 Jan using T_{avg})
- GDD at date of last spring freeze/frost (accumulated from 1 Jan using T_{avg})
- GDD accumulated during spring and summer (using T_{avg})
- GDD at date of end of CD plateau (accumulated from 1 Jan using T_{avg})
- GDD accumulated during CD plateau (using T_{avg})
- Growing season intensity (GSI, calculated from plateau GDD and duration, in °C)

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Table 1: 1984-2013 climatological indicator statistics averaged over the US portion of our study area. Trend significance is marked as * at $p < 0.05$ and ** at $p < 0.01$. Extreme years affected by El Niño (ENSO+) and La Niña (ENSO–) events are marked with # and %, respectively.

Temperature Indicators	**Mean**	**Std. Dev.**	**Trend (units/y)**	**Minimum (Year)**	**Maximum (Year)**
Winter Average Temperature (°C)	–9.6	2.3	+0.04	–13.7 (1994)	–4.8 (2012$^{\%}$)
Spring Average Temperature (°C)	9.2	1.4	–0.04	6.7 (1996$^{\%}$)	11.6 (1987$^{\#}$)
Summer Average Temperature (°C)	17.7	0.9	+0.03	15.1 (1992$^{\#}$)	19.4 (2002)
Autumn Average Temperature (°C)	0.7	1.6	+0.05	–2.8 (1985$^{\%}$)	3.9 (2001$^{\%}$)
Annual Average Temperature (°C)	4.5	1.0	+0.02	2.8 (1996$^{\%}$)	6.8 (1998$^{\#}$)
Precipitation Indicators					
Winter Total Precipitation (cm)	9.2	1.9	+0.03	4.3 (1987$^{\#}$)	12.8 (1997)
Spring Total Precipitation (cm)	23.4	5.1	+0.13	14.1 (1988$^{\#}$)	36.4 (2001$^{\%}$)
Summer Total Precipitation (cm)	28.9	5.6	–0.34**	18.4 (2012$^{\%}$)	39.6 (1999$^{\%}$)
Autumn Total Precipitation (cm)	17.0	4.9	–0.02	9.7 (1989$^{\%}$)	26.6 (1996$^{\%}$)
Annual Total Precipitation (cm)	80.0	8.7	–0.19	63.9 (2006$^{\%}$)	96.5 (1985)
Lake Superior Indicators					
Lake Superior Ice-On Date (DOY)	15.9	18.7	+1.15**	–13 (1986/1990)	69 (2002)
Lake Superior Ice-Off Date (DOY)	88.3	12.2	+0.69	63 (2000$^{\%}$)	115 (1996$^{\%}$)
Lake Superior Ice Duration (days)	72.4	26.9	–0.46	0 (1998$^{\#}$/2012$^{\%}$)	109 (1996$^{\%}$)
Cold Season Indicators					
Freezing Days (using T_{min})	183.5	10.3	–0.28	167.0 (2010$^{\#}$)	200.6 (1996$^{\%}$)
Chilling Days (using T_{avg})	176.9	10.5	–0.14	158.6 (2012$^{\%}$)	196.3 (1996$^{\%}$)
CSI (degrees)	11.0	1.3	–0.03	8.7 (2012$^{\%}$)	13.3 (1996$^{\%}$)
Last Spring Freezing Night (DOY)	137.8	5.7	–0.12	127.2 (2001$^{\%}$)	148.8 (1992$^{\#}$)
First Autumn Freezing Night (DOY)	269.8	5.1	+0.20	261.5 (1986)	279.1 (2013)
Warm Season Indicators					
Last Spring Freezing Night (GDD)	155.6	47.0	–2.12*	79.7 (1996$^{\%}$)	244.4 (1992$^{\#}$)
Frost-free Season (days)	132.0	6.8	+0.32*	116.9 (1992$^{\#}$)	144.5 (2005)
Beginning of CD Plateau (DOY)	127.4	6.8	+0.16	112.7 (1985$^{\%}$)	139.7 (1992$^{\#}$)
End of CD Plateau (DOY)	275.1	6.1	+0.27*	262.9 (1991)	286.6 (2013)
Plateau Duration (days)	147.7	7.3	+0.11	134.4 (1991)	165.5 (1998$^{\#}$)
Plateau GDD (degree-days)	1659.6	134.5	+1.88	1316.4 (1992$^{\#}$)	1896.8 (1988$^{\#}$)
GSI (degrees)	11.2	0.8	+0.00	9.5 (1992$^{\#}$)	13.0 (1991)

Figure 1: Study area (dashed box) in regional context with collected GHCND surface weather stations, each indicated with its measurement type(s) and observation record duration.

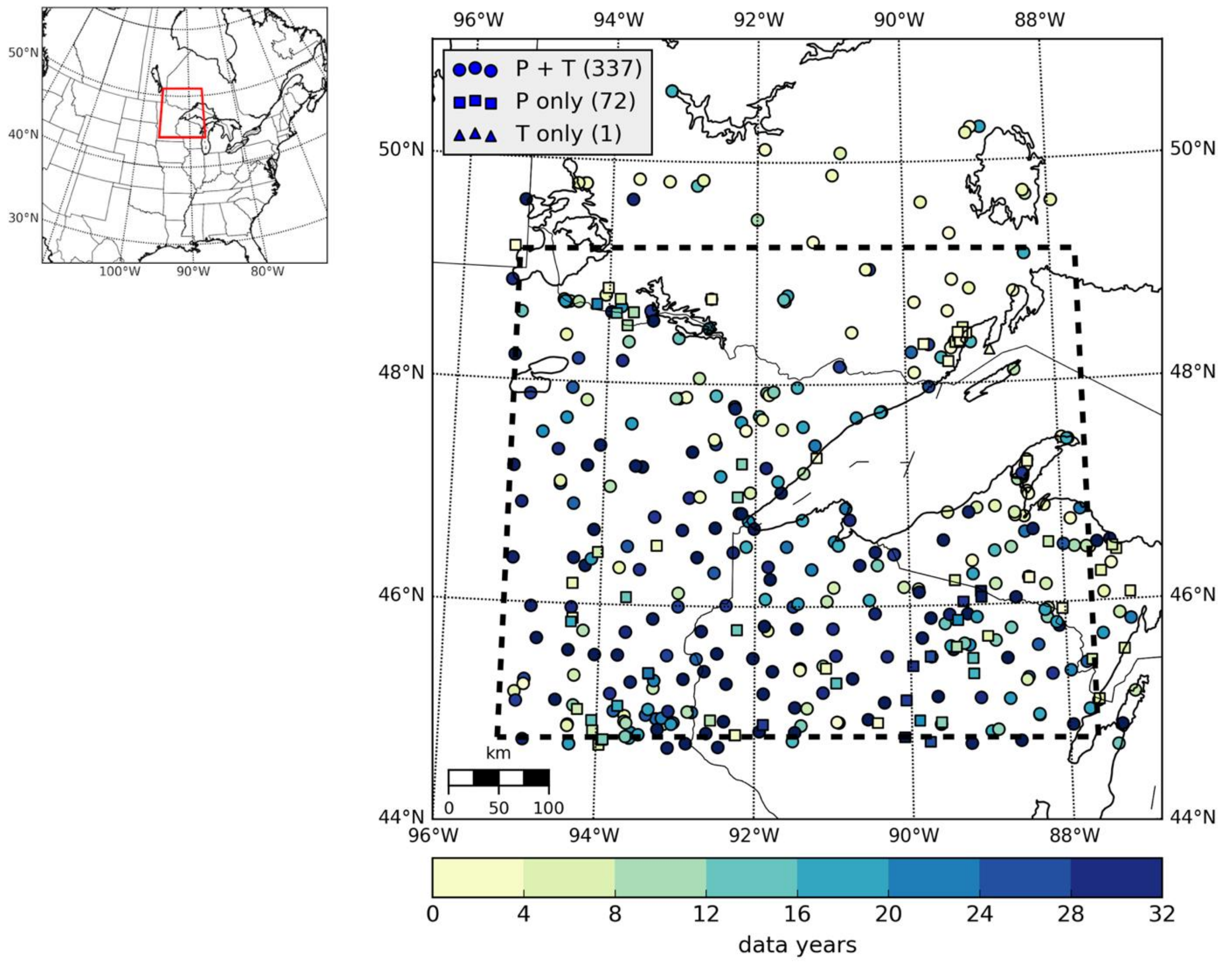


Figure 2: 1984-2013 mean daily temperatures averaged over the US portion of our study area, with observed seasonal indicators marked. See Table 1 for indicator values and additional details.

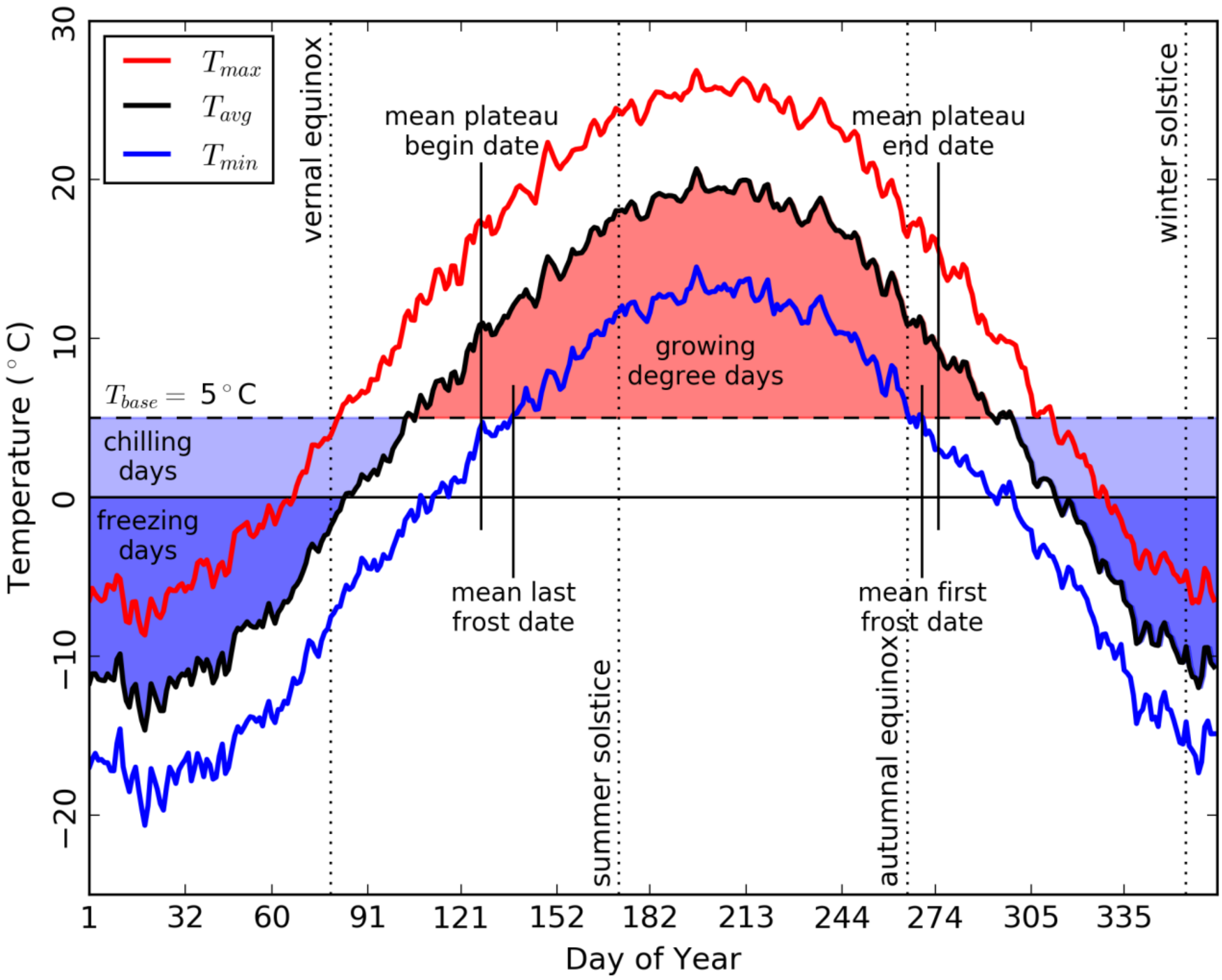


Figure 3: 1984-2013 mean CD and GDD accumulations averaged over the US portion of our study area, with seasons marked. Dashed lines represent the maximum and minimum accumulations within our study period. See Table 1 for indicator values and additional details.

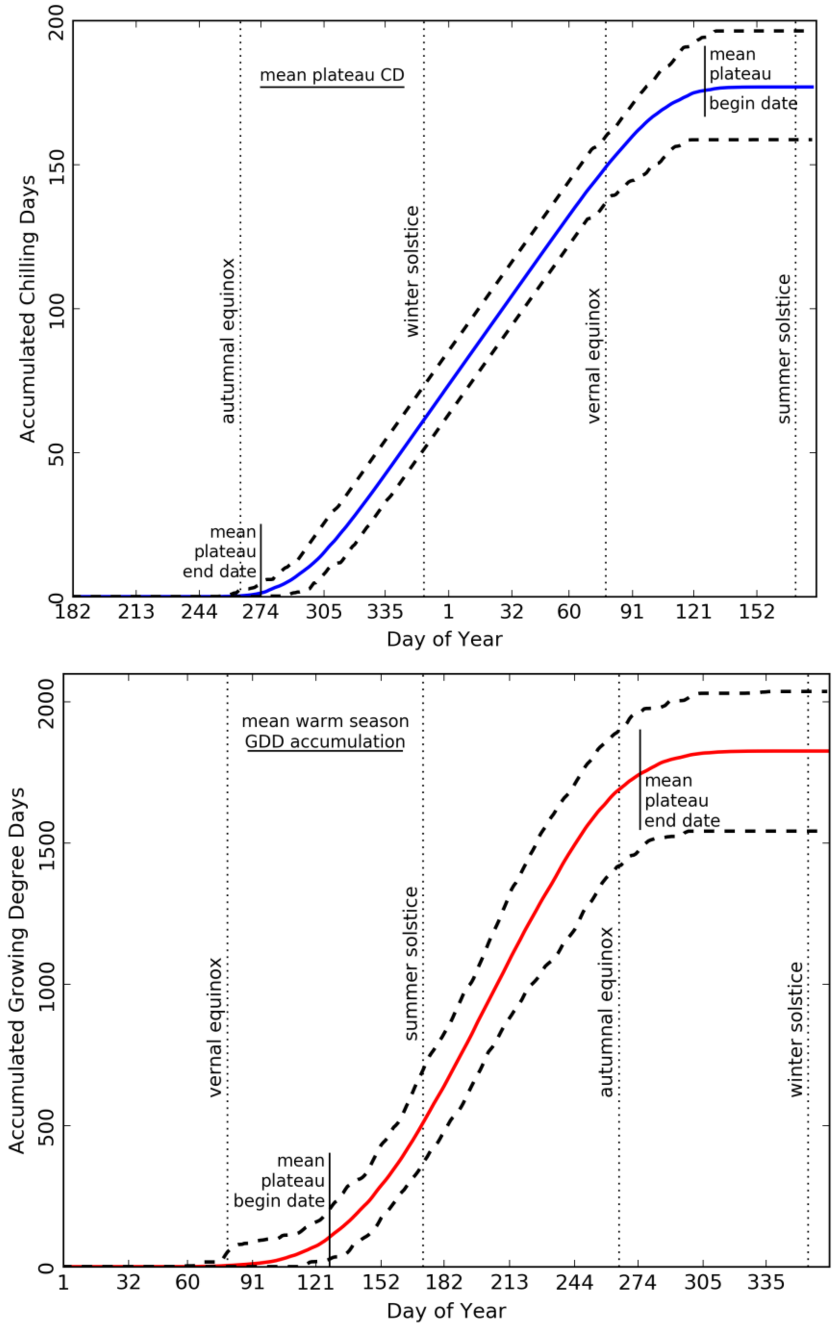


Figure 4: 1984-2013 seasonal and annual precipitation averaged over the US portion of our study area. Error bars indicate the spatial variability (±1σ) in annual total precipitation.

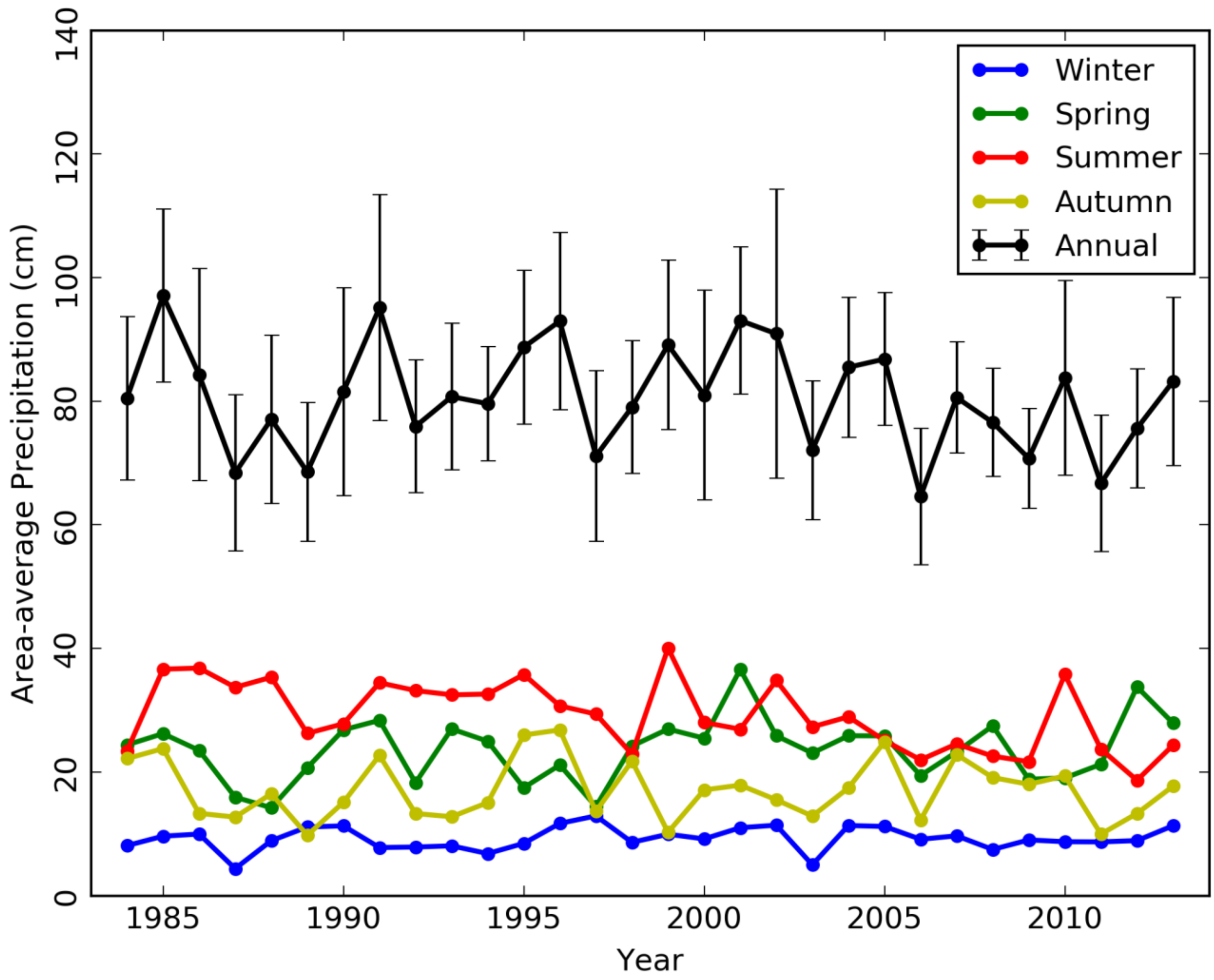


Figure 5: 1984-2013 mean annual T_{min}, T_{max}, and T_{avg}, with trends. Areas of trend significance at $p < 0.05$ in maps b, d, and f are stippled.

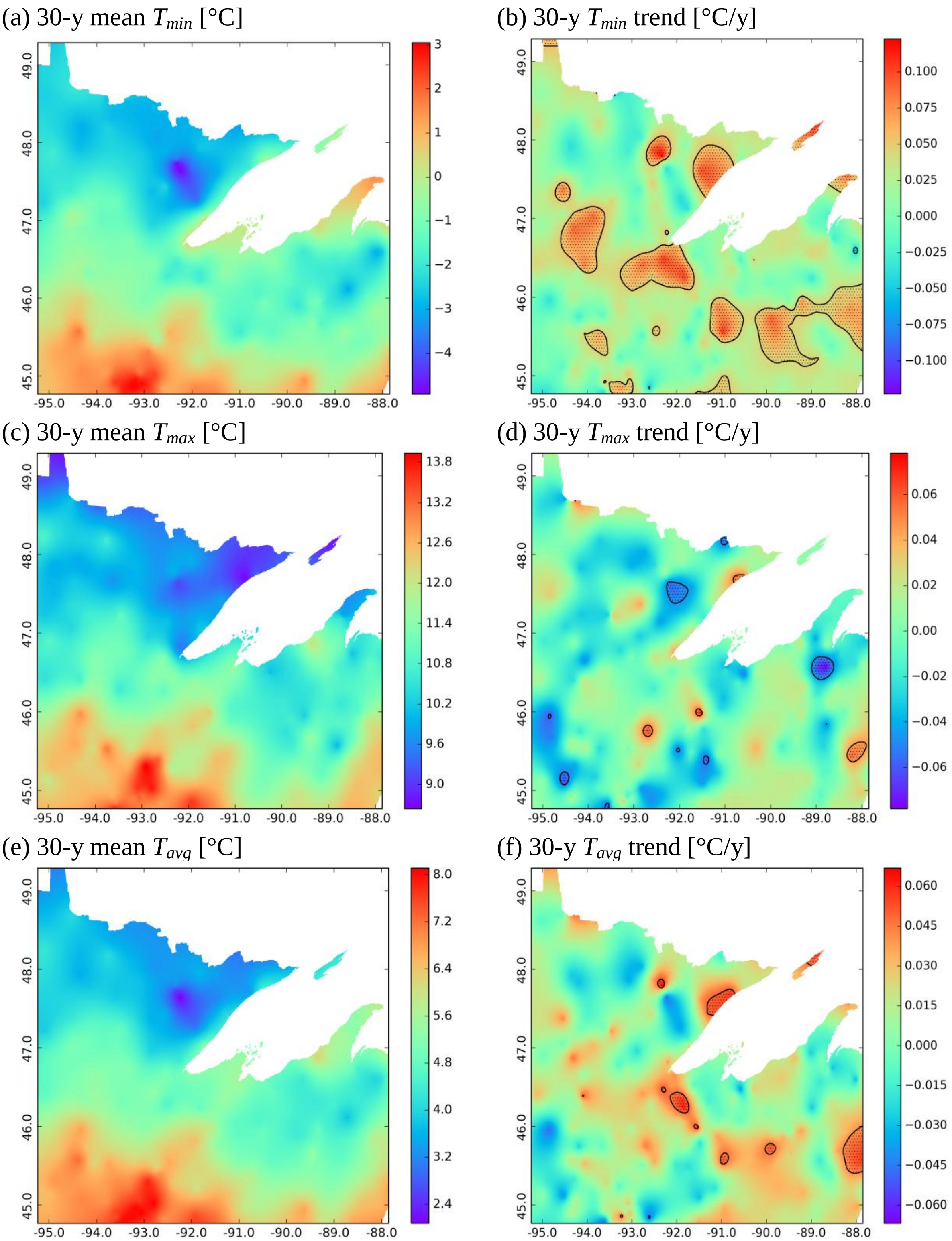
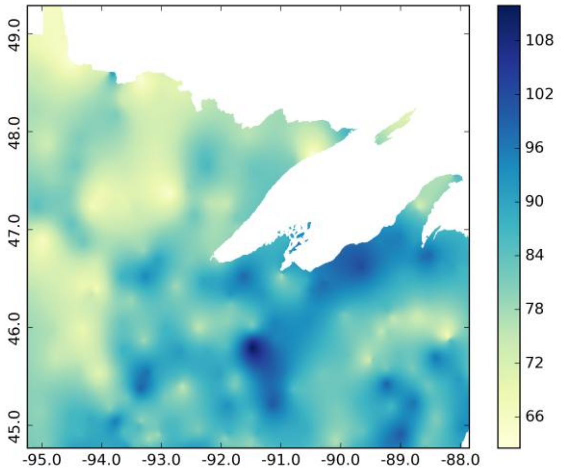


Figure 6: 1984-2013 mean annual P, with trends. Areas of trend significance at $p < 0.05$ in map b are stippled.

(a) 30-y mean P [cm]



(b) 30-y P trend [cm/y]

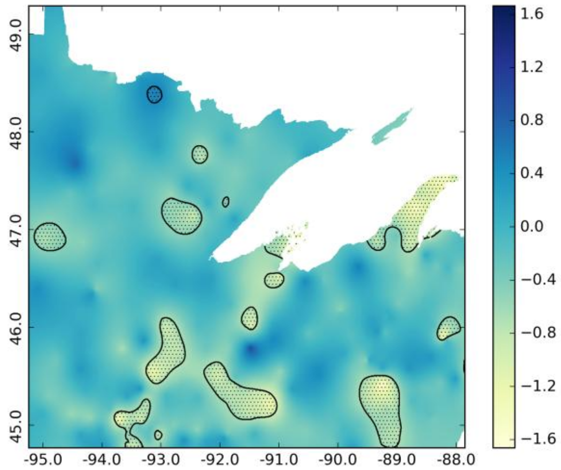


Figure 7: 1984-2013 mean derived cold season indicators, with trends. Areas of trend significance at $p < 0.05$ in maps b and d are stippled.

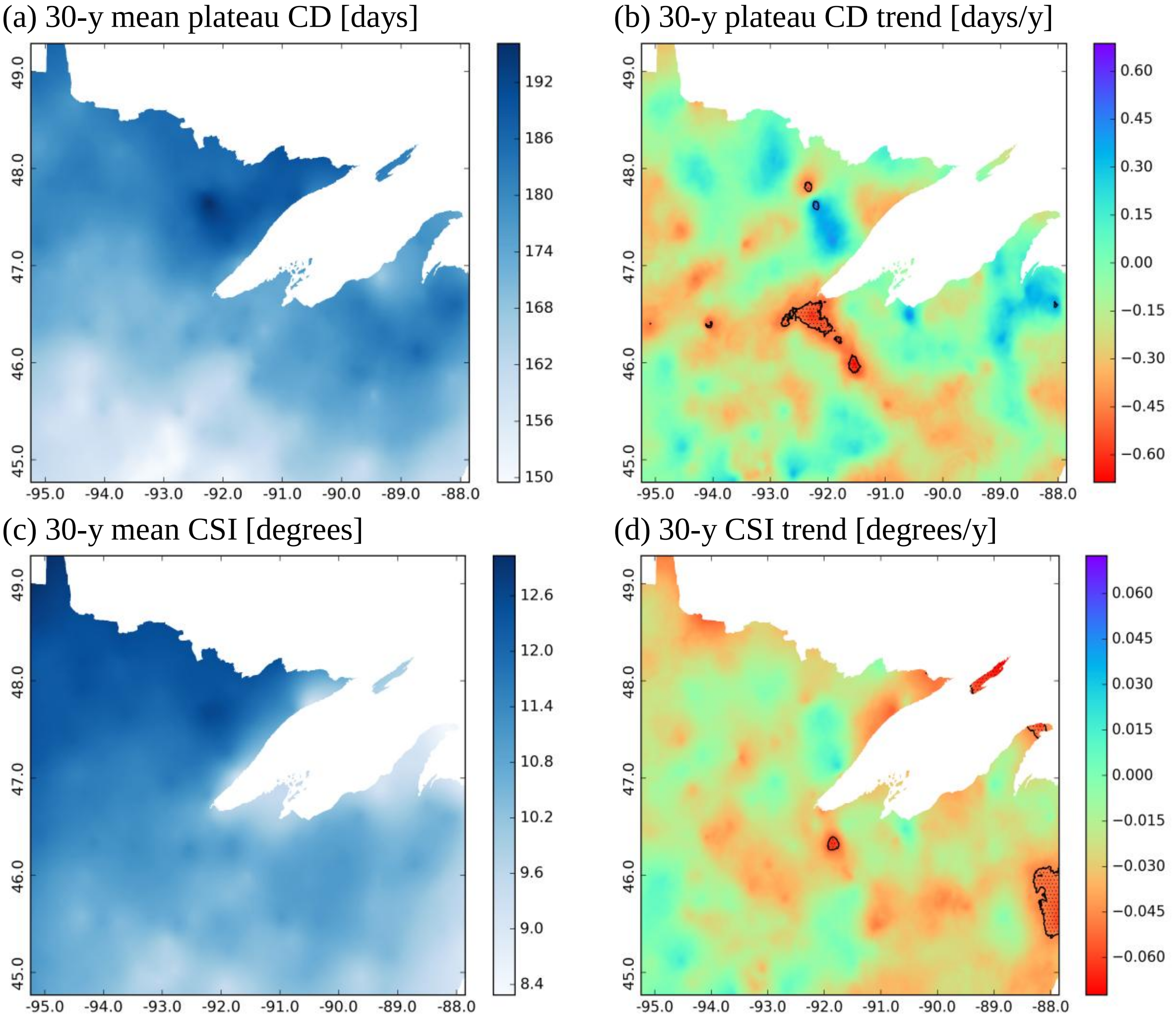


Figure 8: 1984-2013 mean frost-based growing season start, end, and duration, with trends.

Areas of trend significance at $p < 0.05$ in maps b, d, and f are stippled.

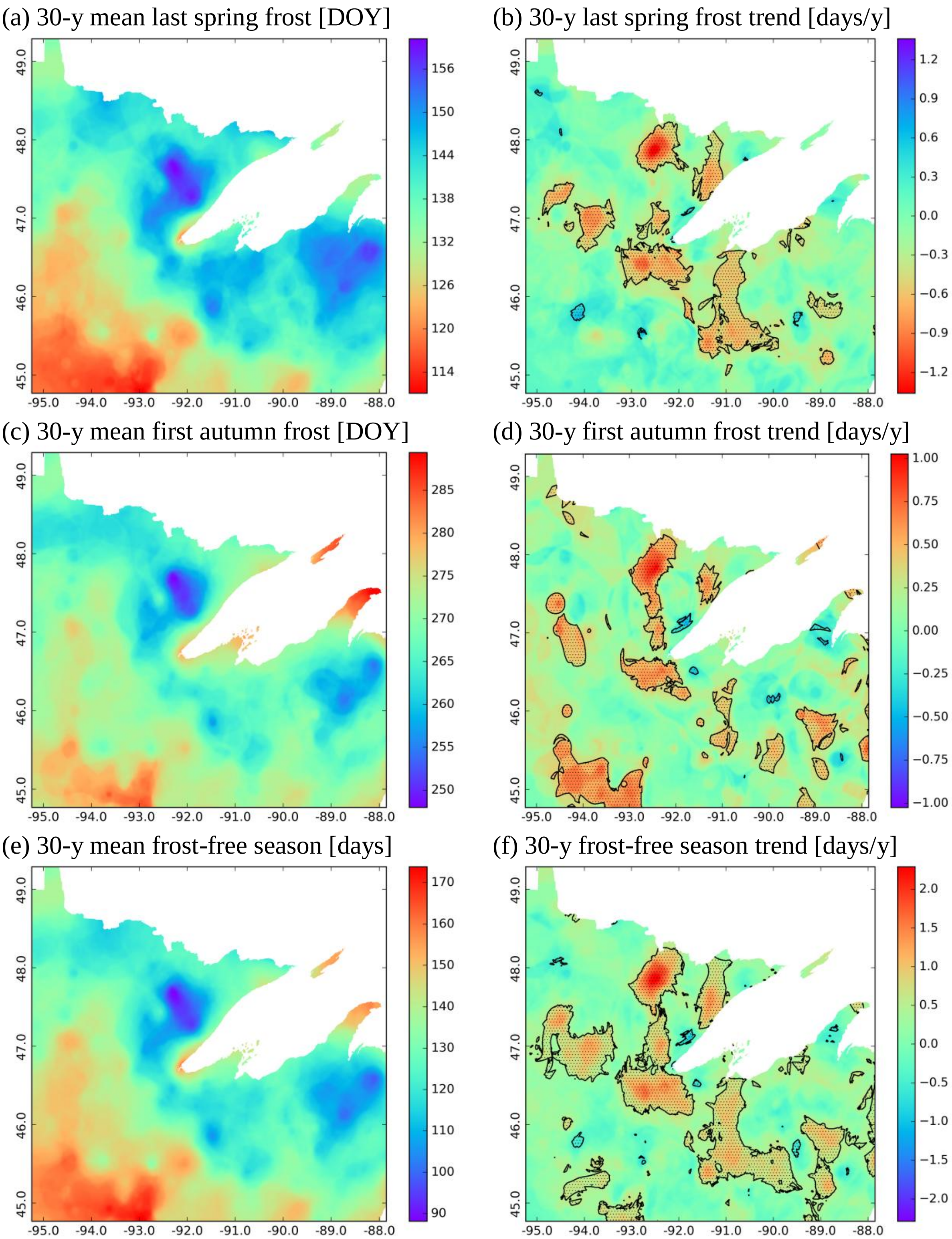


Figure 9: 1984-2013 mean plateau-based growing season start, end, and duration, with trends. Areas of trend significance at $p < 0.05$ in maps b, d, and f are stippled.

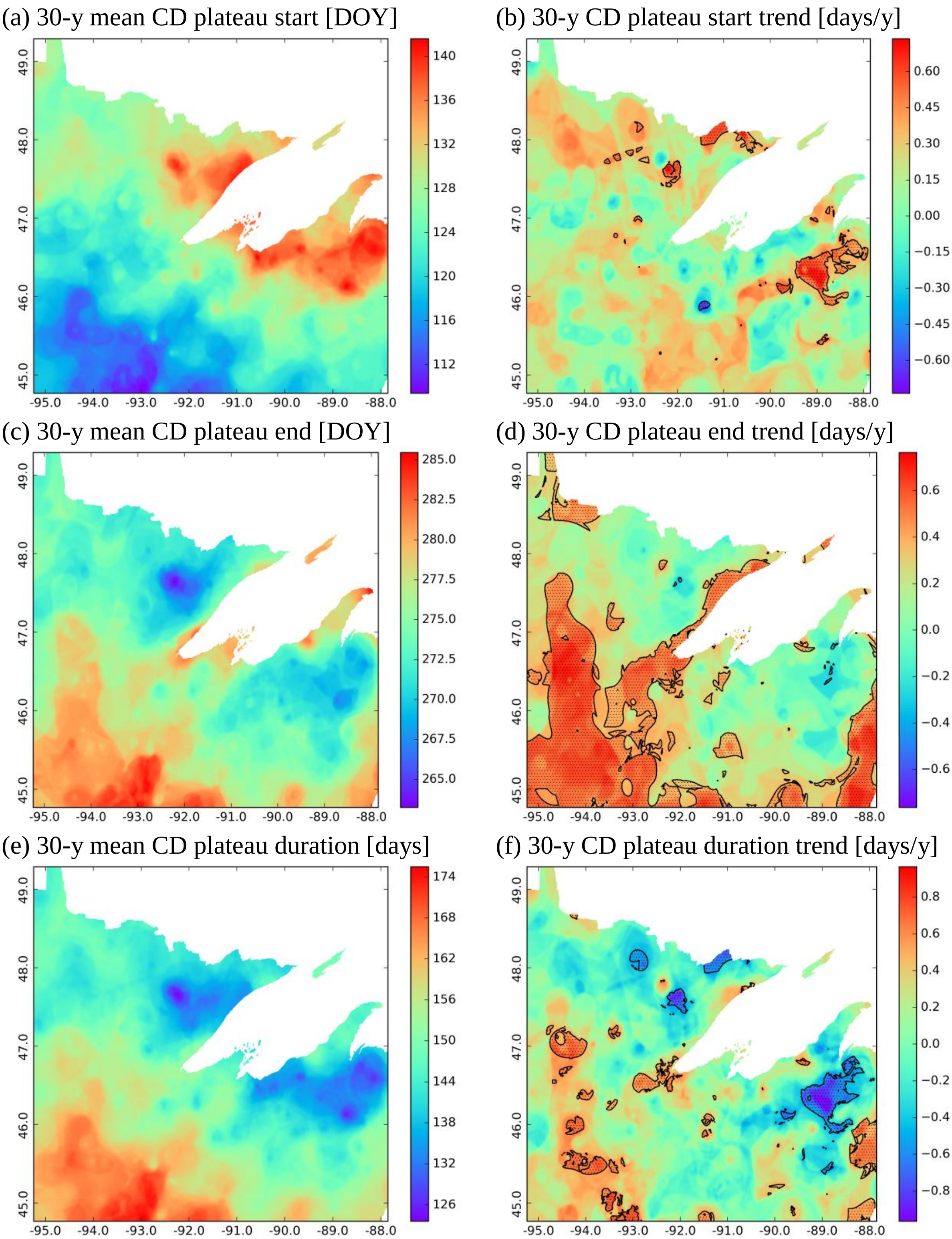


Figure 10: 1984-2013 mean plateau-based growing season indicators, with trends. Areas of trend significance at $p < 0.05$ in maps b and d are stippled.

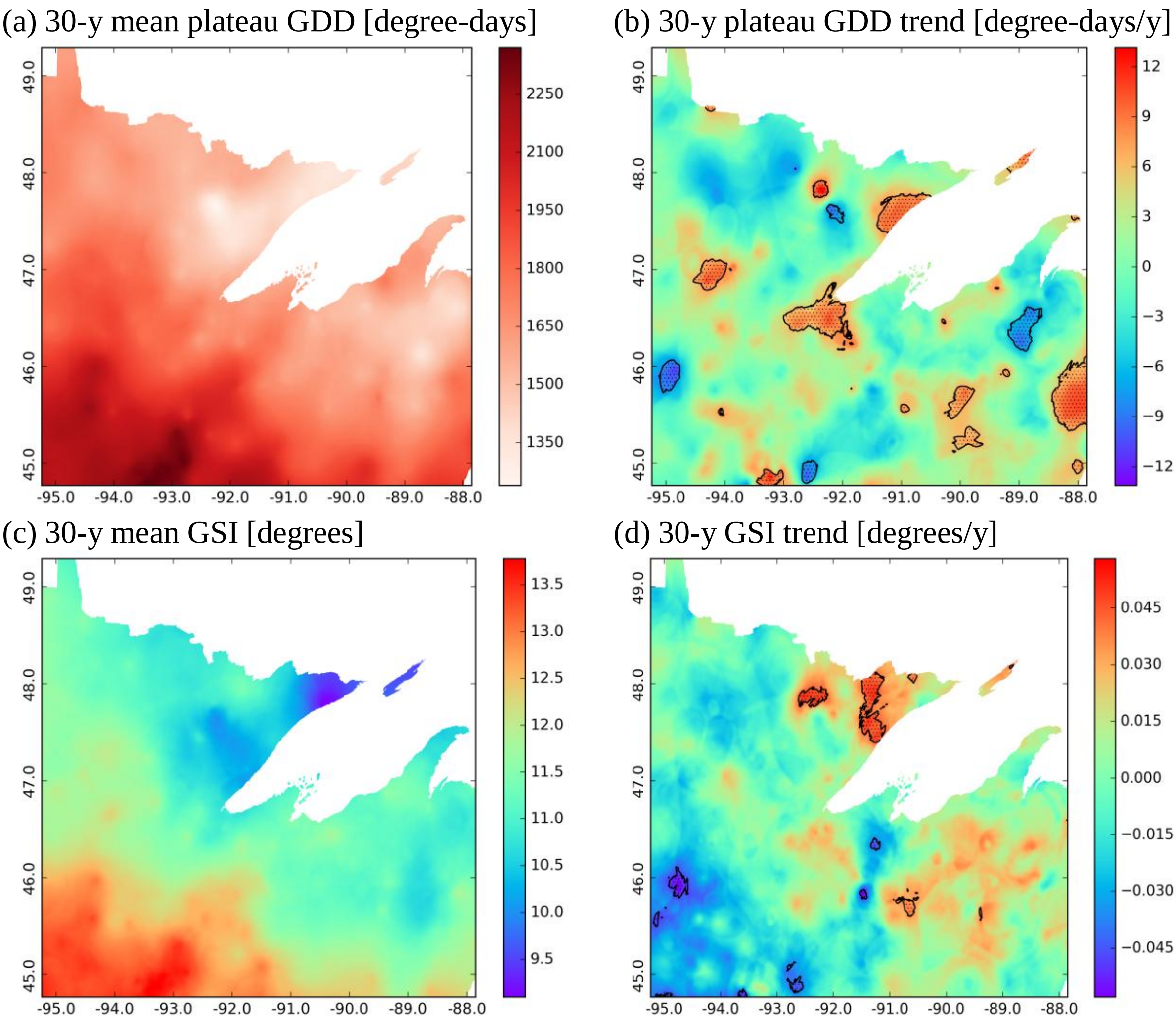


Figure 11: 1984-2013 intra- and interseasonal study area temperature and precipitation correlations. Only statistically significant correlations ($p < 0.05$, * at $p < 0.01$, ** at $p < 0.001$) are shown.

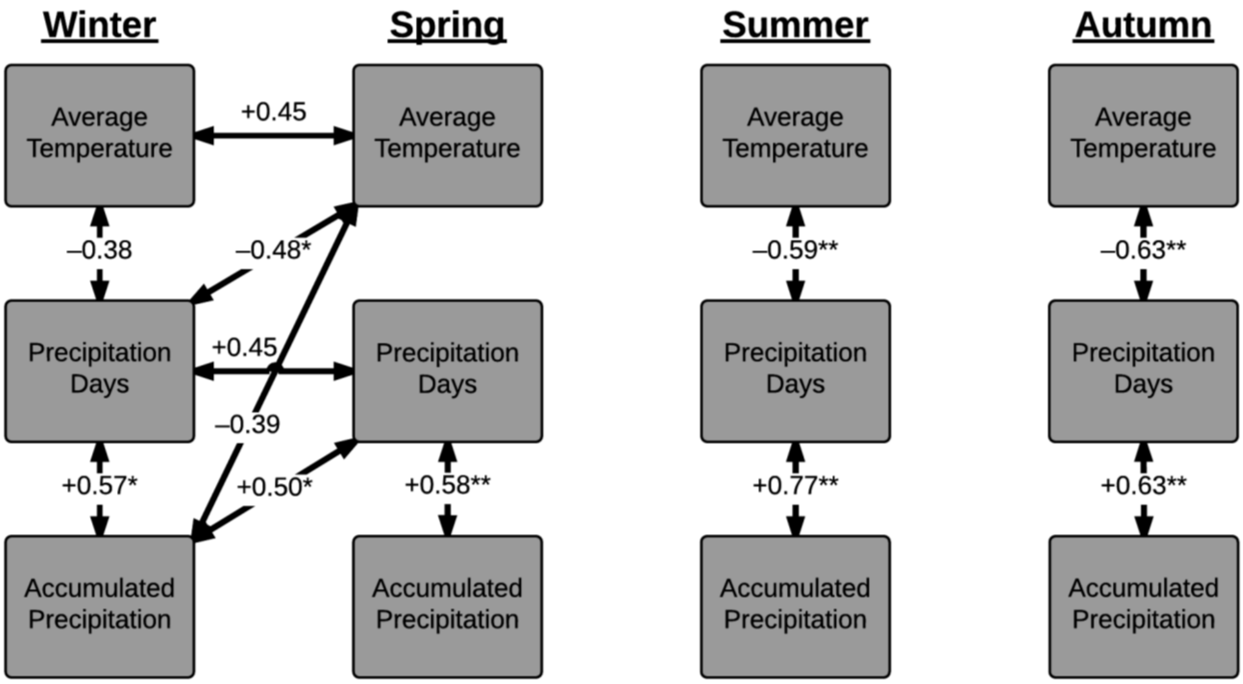
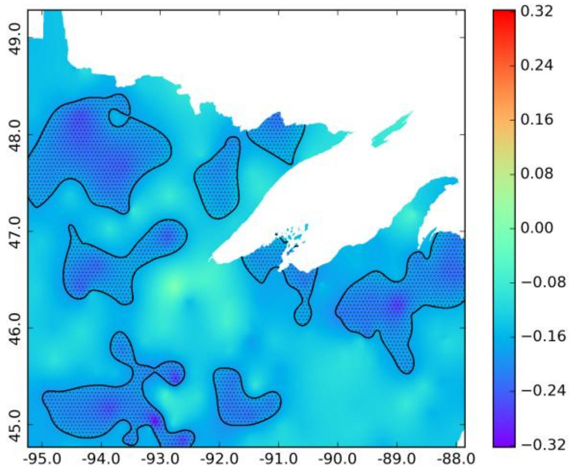
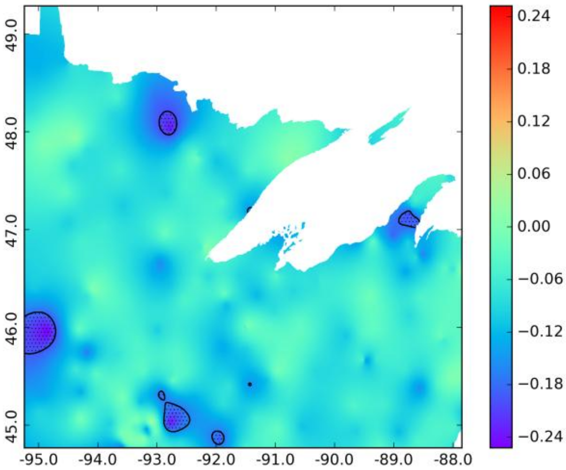


Figure 12: 1984-1998 and 1998-2013 trends in selected climatological indicators. Areas of trend significance at $p < 0.05$ in maps b, d, and f are stippled.

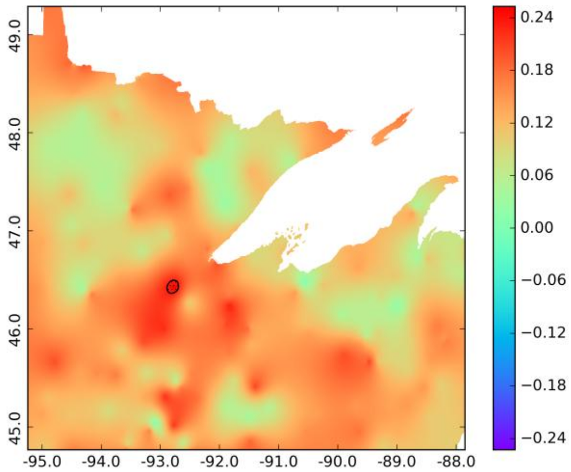
(a) 1984-1998 spring T_{avg} trend [°C/y]



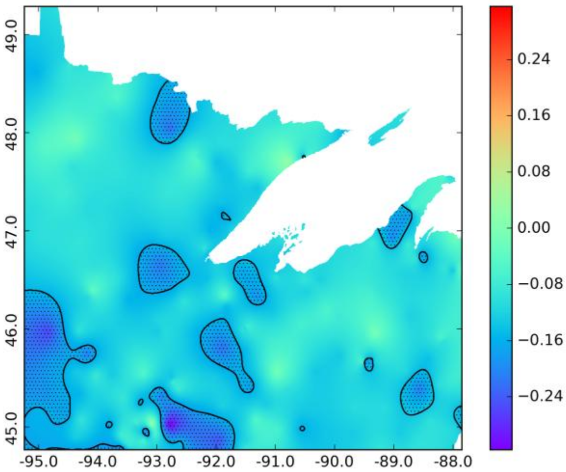
(b) 1998-2013 spring T_{avg} trend [°C/y]



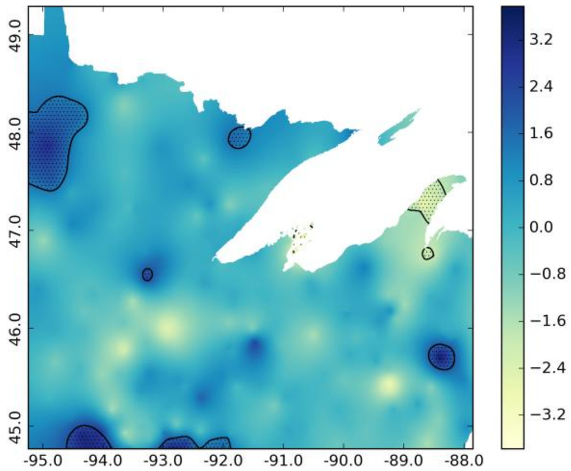
(c) 1984-1998 autumn T_{avg} trend [°C/y]



(d) 1998-2013 autumn T_{avg} trend [°C/y]



(e) 1984-1998 annual P trend [cm/y]



(f) 1998-2013 annual P trend [cm/y]