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A Meteorological Analysis of Important Contributors to the 1999–2005 Canadian Prairie Drought

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ABSTRACT

Drought is a complex natural hazard that is endemic to the Canadian prairies. The 1999–2005 Canadian prairie drought, which had great socioeconomic impacts, was meteorologically unique in that it did not conform to the traditional persistent positive Pacific–North American (PNA) pattern and west coast ridging paradigm normally associated with prairie drought. The purpose of this study is to diagnose the unique synoptic-scale mechanisms responsible for modulating subsidence during this drought. Using 30-day running means of the percent of normal precipitation from station data, key severe dry periods during 1999–2005 are identified. Analysis of the mean fields from reanalysis data shows that these dry events can be grouped into three upper-level flow categories: amplified warm, amplified cold, and zonal. Amplified warm cases match the traditional ridging paradigm, while amplified cold and zonal cases elucidate the fact that cold-air advection and downsloping flow, respectively, can also be important subsidence mechanisms during a Canadian prairie drought. In all, the 1999–2005 drought was more meteorologically complex on the synoptic scale than previous historic prairie droughts. Finally, a brief historical perspective shows that the drought was centered in 2001–02 and was not as severe as historical droughts, suggesting that societal vulnerability also played a substantial role in the impacts of the 1999–2005 drought.

1. Introduction

a. Motivation

In Canada, drought is endemic to the agriculture-dominated prairie provinces (Alberta, Saskatchewan, and Manitoba), where there have been no less than 40 drought occurrences in the last 200 years (Phillips 1990; Maybank et al. 1995; Nkemdirim and Weber 1999; Bonsal 2008). The Canadian Prairie drought of 1999–2005 has been called the worst drought for at least a hundred years in terms of its socioeconomic impact (Bonsal 2008), and prompted the creation of the first integrated drought network in Canada, the Drought Research Initiative (DRI; Hanesiak et al. 2011). Economic and agricultural impacts were staggering, amounting to a $5.8 billion (U.S. dollars) drop in gross domestic product (GDP) and a loss of 41 000 jobs in 2001–02 and, ultimately, led to the first occurrence of negative net farm incomes in over 25 years (Bonsal 2008; Wheaton et al. 2008). The impact of this drought reverberated across many economic and natural sectors, seen by a decrease in cattle numbers, increased forest fires, and reduced river and streamflow in Alberta and Saskatchewan, hindering the generation of hydroelectricity in Manitoba and British Columbia (Bonsal 2008; Wheaton et al. 2008).

The 1999–2005 drought was also notable in its meteorological conditions being distinct from earlier droughts. This drought did not conform to the circulation patterns normally associated with Canadian prairie drought, namely a persistently positive Pacific–North American (PNA; Wallace and Gutzler 1981) pattern (ridging and “blocking” over western Canada), and anomalously warm temperatures (Dey 1982; Knox and Lawford 1990; Bonsal et al. 1999; Bonsal and Wheaton 2005; Bonsal 2008; Wheaton et al. 2008; Bonsal et al. 2011a,b). Here, blocking refers to a quasi-stationary upper-tropospheric anticyclone that restricts westerly midlatitude propagation (positive PNA; Fig. 1). This challenge to the traditional drought paradigm was highlighted in Bonsal and Wheaton (2005), which showed distinct differences between the 500-hPa height patterns of 2001 and 2002 and two historical Canadian prairie droughts (1961 and
Bonsal and Wheaton (2005) and Bonsal et al. (2011a) show that the classic anomalous ridging in western Canada and North Pacific troughing set up an anomalous meridional flow pattern in western Canada in 1961 and 1988 but not in 2001 and 2002. Thus, it is not surprising that the spatial structures of surface temperature and precipitation anomalies were also distinct from previous Canadian prairie droughts. Bonsal and Wheaton (2005) showed that the 1961 and 1988 droughts were accompanied by strong positive temperatures anomalies (1.5°–2.5°C), as opposed to 0.5°–1°C in 2001 and 2002, with spring 2002 being the coldest spring in the prairies since 1948 (Bonsal and Wheaton 2005).

Another major aspect in which the 1999–2005 drought differed from previous Canadian prairie droughts was that there was no clear relationship to other teleconnection patterns, unlike other droughts in the world during this decade (Bonsal 2008; Bonsal et al. 2011a). There were persistently warm sea surface temperature (SST) anomalies in the western tropical Pacific and persistently cold SST anomalies in the eastern tropical Pacific (i.e., La Niña–like conditions) during 1999–2005 (Hoerling and Kumar 2003). However, while general circulation models (GCMs) forced by such SST signatures reproduced anomalous temperature and precipitation patterns similar to observations over the lower midlatitudes in the Northern Hemisphere (e.g., the United States), the GCMs lacked adequate representation of the observations in polar latitudes (e.g., Canada) during this time, particularly in upper-tropospheric height anomalies. Thus, it was suggested that the oceans may not be the main forcing for drought in higher latitudes (Hoerling and Kumar 2003). Moreover, it is generally El Niño that results in anomalously dry conditions across the prairies in the winter after onset (Shabbar et al. 1997; Shabbar and Skinner 2004), while La Niña is linked to wetter than normal conditions during the winter after onset. The dry (wet) conditions are explained by positive (negative) PNA-like patterns (Shabbar et al. 1997; Shabbar and Skinner 2004). Thus, the Canadian prairies should have been plagued by an abundance of rainfall during 1999–2005, not the worst drought in a century.

Precipitation in the Canadian prairies has high temporal and spatial variability (Maybank et al. 1995; Bonsal and Wheaton 2005). Up to two-thirds of the annual total precipitation in the prairies falls during the growing season (May–August) (Dey 1982; Bonsal et al. 1999). Thus, reductions in precipitation during the growing season have far more socioeconomic impacts than reductions in winter precipitation, although drought conditions can
be further aggravated by low snow cover. Additionally, the two principal moisture sources for the prairies are the Gulf of Mexico and the Pacific Ocean; the Gulf of Mexico is the most important moisture source for summer precipitation in the prairies, whereas the Pacific is dominant for winter precipitation (Liu et al. 2004). Thus, we would expect to find a high occurrence of flow that blocks moisture transport from the Gulf of Mexico during the growing season in Canadian prairie drought.

b. Objectives

The inconsistency with regards to teleconnection signals and the lack of conformity to the traditional drought paradigm leaves us in want of a dynamical explanation for the drought. Hanesiak et al. (2011) mentioned that very contrasting synoptic-scale processes were at play at different times during the drought, all of which led to anomalously dry conditions. Therefore, we are motivated to examine a finer time scale of circulation anomalies than what is typical of drought studies (i.e., seasonal averages), because the highly variable nature of precipitation, particularly in the prairies, is such that one or two events can bring about the end of a meteorological drought.

The primary objective of this study is to analyze the synoptic-scale dynamics during the 10 driest (most important) 30-day time periods of the 1999–2005 drought. As Garrity et al. (2010) and Bronnimann et al. (2009) have discussed, drought should be viewed three-dimensionally; our analysis thus includes time–height cross sections. The remainder of the paper is as follows: section 2 discusses the data and methods used, results are discussed in section 3, and a summary and conclusions constitute section 4.

2. Data and methods

a. Data

Corrected monthly station precipitation data were obtained from Environment Canada’s Climate Research Division Adjusted and Homogenized Canadian Climate Data (AHCCD; available online at http://ec.gc.ca/dchca-ahccd/Default.asp?lang=En&n=B1F8423A-1). Corrected daily station precipitation data for Alberta, Saskatchewan, and Manitoba have also been provided by the Climate Research Division. The monthly and daily precipitation data were used for 33 stations in the prairies (Fig. 2), which were chosen on the basis of record length (1948–2005), even spatial coverage representative of the study region, relatively few missing data, and the availability of both monthly and daily corrected precipitation data. These data have been corrected for known inhomogeneities and systematic biases, which include gauge errors arising from wind, evaporation, and wetting losses, in addition to errors derived from changes in instrumentation, measurement procedures, and the location of the stations (Mekis and Hogg 1999).

Daily PNA indices were obtained from the National Oceanic and Atmospheric Administration/National Weather Service/National Centers for Environmental
b. Diagnosis of key events

1) PERCENT OF NORMAL METHOD

Since one or two synoptic events can effectively end a meteorological drought, a temporal resolution of smaller than a month or season is beneficial in order to ensure that events on short temporal scales are captured. Three commonly used drought indices are the Palmer drought severity index (PDSI; Palmer 1965), the standardized precipitation index (SPI; McKee et al. 1993), and the percent of normal precipitation. While the PDSI is generally used to diagnose drought on longer time scales (e.g., Guttman 1998), the SPI and the percent of normal precipitation can be utilized over a multitude of different time scales (Guttman 1998; Wu et al. 2005). However, the 1999–2005 drought featured a consistent lack of precipitation, particularly in the northern part of our domain (Fig. 2). Dry periods were not interspersed with wet periods, thus rendering the choices for drought diagnostics (i.e., PDSI, SPI, percent normal) largely interchangeable. Since our focus is on the synoptic-scale dynamics of key dry periods during the drought, for the sake of simplicity we use percent of normal as the method for identifying key severe periods in this study.

There is an argument regarding spatial variance that goes against the usage of percent of normal (e.g., Quiring 2009). However, we justify our choice of percent normal in three ways. First, the spatial variance in normal growing-season precipitation in the Canadian prairies is not as large as that in a place such as Texas (Quiring 2009), and therefore there is less inherent error in using the percent normal diagnostic; in essence, the stations in the region are enough alike to be considered to occupy a similar location for synoptic-scale features. Second, the results of our percent normal diagnostics find the same qualitative results as studies that use other drought diagnostic methods (e.g., Bonsal and Regier 2007; Bonsal et al. 2011b, Hanesiak et al. 2011). For example, we show that the worst of the drought was clearly in 2001 and 2002 (Figs. 3 and 4), as do other studies that use SPI and PDSI (Bonsal and Regier 2007; Bonsal et al. 2011b; Hanesiak et al. 2011). Finally, our primary goal is to analyze the synoptic-scale characteristics and dynamics of key drought periods. Others (e.g., Bonsal and Regier 2007; Bonsal et al. 2011b; Hanesiak et al. 2011) have already defined the drought and its key periods, and our results concur.

The daily and monthly precipitation values were first averaged over the 33 stations (Fig. 2) from 1999 to 2005. We tried two different methodologies to calculate daily percent normal, the first being a more traditional method and the second being novel to this study. The first method utilized a daily precipitation climatology (e.g., averaging all 1 Januaries together, etc.) and a sliding 30-day running average, with the daily percentage of normal quantity being the simple ratio of the
actual accumulation to the expected value for that day from the climatology.

The second method derived a daily climatology of precipitation accumulation from the monthly precipitation climatology, as follows:

- The monthly values of accumulated precipitation (mm) are divided by the number of days in the respective month (January, 31 days; February, 28.25 days, etc.) to get a “daily” value. For example, if the climatology for May is 60 mm, the daily value would be 60 mm/31 days = 1.935 mm day$^{-1}$.

- Instead of assigning this one value to every day in the month, giving a climatology that only changes dramatically at the beginning and end of each month, this daily value is assigned only to the middle (15th day) of the month. Following the previous example, the value of 1.935 mm day$^{-1}$ is assigned to 15 May.

- Linear interpolation is then carried out from mid-month to mid-month to achieve values for every day in the year, showing the monotonic increase or decrease in accumulation between months. The values for the remaining days in each month are thus determined by the climatology of two months. For example, the

**FIG. 4.** Percent normal of GPCP precipitation, for the growing seasons of (a) 2001 and (b) 2002. The outline is the approximate extent of the study area, with the northern stations denoted in red.
values for the beginning half of May are dependent on April and May climatology, whereas the values after 15 May are a combination of both May and June climatology.

- The daily percentages of normal values are then smoothed by taking 30-day running averages, or 30-day moving windows, with the averaged value assigned to the middle of the period. For example, the average for 1–30 May 1999 is assigned to 15 May 1999, and the average for 2–31 May 1999 is assigned to 16 May 1999, etc.

The second methodology attempts to provide a longer time filter while keeping the daily/synoptic time-scale memory, thus amalgamating the long duration time scale of drought with the synoptic time scale. For the sake of simplicity, our results here (e.g., Fig. 3) are presented using the first, more traditional, method. However, we found the correlation between the two methods to be 0.995 for 1999–2005 for percent normal at the northern stations (e.g., Fig. 3a), and similarly high for the 33 stations overall. Individual stations (e.g., Saskatoon) had only slightly lower correlations (0.985). Thus, we believe the second method is novel and quite viable, particularly over areal averages, and may be quite useful in situations where daily precipitation climatology data are not available.

The spatial representation of the drought also needs to be taken into account. As noted by Bonsal and Wheaton (2005), the spatial pattern of the 1999–2005 drought was unique in terms of its northward extent, particularly during 2001 and 2002. Figure 4 shows the spatial representation of the drought period using version 2.1 of the Global Precipitation Climatology Project (GPCP) combined satellite–station precipitation dataset, available as monthly mean precipitation rates at 2.5° × 2.5° resolution (Alder et al. 2003). Monthly GPCP data have been shown to be reliable compared with gauge data over North America (e.g., Krajewski et al. 2000). Here, the percent normal of precipitation is shown for the most severe period of the drought, namely the growing seasons of 2001 (Fig. 4a) and 2002 (Fig. 4b). A vast area of below 80% of normal precipitation is present for the entire Alberta–Saskatchewan agricultural region (triangular outline) in 2001 (Fig. 4a). However, the spatial pattern of the drought is dramatically different during the growing season of 2002 (Fig. 4b), as a region of above-normal precipitation occurs in southern Saskatchewan. This is predominantly related to a single extreme precipitation event in June 2002 (fully detailed in Szeto et al. 2011). Its occurrence during the “worst drought in a hundred years” serves to highlight the complexity of this drought. Meteorological drought does not entail a complete absence of high-precipitation events, but a typical decrease in the number of such events (Evans 2008) and/or an increase in the number of consecutive dry days. This northern drought signature also appears in averaged PDSI results over the whole 1999–2005 period [Fig. 5,
using the monthly Dai PDSI dataset (Dai et al. 2004) based on historical global temperature and precipitation records over land areas, with a horizontal resolution of 2.5° × 2.5°], and is confirmed in previous studies (Bonsal and Wheaton 2005; Bonsal and Regier 2007; Bonsal et al. 2011b). We show the PDSI as confirmation of drought conditions, since PDSI is the most commonly used diagnostic for meteorological drought (Dai 2011).

To avoid the drought signature being masked by a single “wet” event, we selected a northern subset of 11 stations from which to complete the diagnosis of key severe 30-day periods. This northern subset is seen in Figs. 2 (red stations), 4b, and 5, with its southern extent at approximately 51.75°N. This latitude roughly corresponds to the boundary between the dry–wet feature in Fig. 4b.

We show the 30-day running mean time series for both the 33 stations and the 11 northern stations for comparison (Fig. 3c). The two time series have a high correlation of 0.80. There are only three main periods where the two curves deviate—autumn 2000, spring 2001, and the growing season of 2002—which can be explained by the enhanced northern drought conditions in 2000–01 and the relatively wet growing season in 2002 (Szeto et al. 2011).

The key dry periods are identified in Fig. 3a as the lowest 10 (in red) 30-day running mean values for the northern stations (1999–2005; Table 1). The frequency of the identified dry events is the greatest during the heart of the drought from autumn 2000 to autumn 2002. The identified dry events in Table 1 are slightly skewed to late autumn and winter periods, owing to the climatological paucity of precipitation in these seasons. We reiterate that the amount of winter precipitation (and associated spring snowmelt) can help to initiate or accentuate a drought, the lack of growing season precipitation is a more important factor in prairie drought severity. For example, a deficit of 30 mm of precipitation during December–February, which receives on the order of 60 mm climatologically, would appear more significant than a deficit of 30 mm of precipitation during June–August, which climatologically receives approximately 200 mm of precipitation, but the latter is when crops most require the precipitation. Eight out of the 10 dates identified as key dry events from the northern subset of stations were common to those from the full 33-station average, albeit in an altered order (Tables 1 and 2); the consistency of the identified extreme events from the northern station and 33-station curves gives robustness to our results.

2) Quasigeostrophic Dynamics

The 10 driest 30-day periods (Fig. 3b, Table 1) are synoptically analyzed using quasigeostrophic (QG) dynamics to investigate mechanisms for subsidence (descent). The adiabatic, frictionless form of the QG omega equation (Bluestein 1992) indicates that there are two dynamic mechanisms associated with subsidence: vorticity advection increasing with respect to pressure (anticyclonic vorticity advection, AVA) and localized cold-air advection (CAA):

\[
\left(\nabla^2 + \frac{f_0^2}{\sigma} \frac{\partial}{\partial p}\right) \omega = -\frac{f_0}{\sigma} \frac{\partial}{\partial p} \left[-\mathbf{v}_g \cdot \nabla_T (\zeta_g + f)\right] - \frac{R}{\sigma p} \nabla^2 (\mathbf{v}_g \cdot \nabla_T T),
\]

where \(f_0\) is the constant Coriolis parameter (s\(^{-1}\)), \(\sigma\) is the static stability parameter (m\(^2\)s\(^{-2}\)Pa\(^{-1}\)), \(\omega\) is the vertical velocity (Pa s\(^{-1}\)), \(\zeta_g\) is the geostrophic relative vorticity (s\(^{-1}\)), \(f\) is the latitude-dependent Coriolis parameter, \(p\) is pressure, \(R\) is the universal gas constant.

### Table 1. Lowest 10 values from the northern stations time series (Fig. 3a). The northern station dates are denoted in red in Fig. 3.

<table>
<thead>
<tr>
<th>Central date</th>
<th>30-day percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>29 Dec 2001</td>
<td>12.99</td>
</tr>
<tr>
<td>30 Nov 2002</td>
<td>17.74</td>
</tr>
<tr>
<td>18 Oct 2000</td>
<td>19.41</td>
</tr>
<tr>
<td>20 May 2002*</td>
<td>21.41</td>
</tr>
<tr>
<td>21 Jan 2001</td>
<td>21.73</td>
</tr>
<tr>
<td>12 Dec 2003</td>
<td>22.98</td>
</tr>
<tr>
<td>28 Apr 2001</td>
<td>23.01</td>
</tr>
<tr>
<td>8 Nov 2004</td>
<td>23.24</td>
</tr>
<tr>
<td>15 Oct 1999*</td>
<td>26.64</td>
</tr>
<tr>
<td>30 Apr 2005</td>
<td>29.52</td>
</tr>
</tbody>
</table>

* Key period that is unique to the northern stations.

### Table 2. Dry events: lowest 10 values from the 33-station time series (Fig. 3c).

<table>
<thead>
<tr>
<th>Central date</th>
<th>30-day percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>20 Aug 2001</td>
<td>19.26</td>
</tr>
<tr>
<td>07 Oct 2001</td>
<td>19.66</td>
</tr>
<tr>
<td>10 Dec 2003</td>
<td>29.34</td>
</tr>
<tr>
<td>16 Jan 2001</td>
<td>29.78</td>
</tr>
<tr>
<td>14 Nov 2004</td>
<td>30.23</td>
</tr>
<tr>
<td>1 Dec 2002</td>
<td>30.95</td>
</tr>
<tr>
<td>7 Oct 2003</td>
<td>33.92</td>
</tr>
<tr>
<td>29 Apr 2001</td>
<td>34.08</td>
</tr>
<tr>
<td>27 Dec 2001</td>
<td>34.81</td>
</tr>
<tr>
<td>30 Apr 2005</td>
<td>36.66</td>
</tr>
</tbody>
</table>
(287 J kg$^{-1}$ K$^{-1}$), $\mathbf{v}_g$ is the geostrophic wind vector (m s$^{-1}$), and $T$ is temperature (K).

In the absence of dynamic mechanisms, subsidence can be associated with downsloping flow over terrain, which must be considered since the prairies lie in the rain-shadow region of the Rocky Mountains. Downslope flow is represented by the homogeneous form of the QG omega equation as expressed at the surface (Bluestein 1993):

$$
\nabla^2 \omega_o = -\frac{f^2}{\sigma} \frac{\partial^2 \omega_o}{\partial p^2},
$$

(2)

where $\omega_o$ is the orographic omega. During downslope flow, $\omega_o > 0$. Using the statement of continuity, under statically stable conditions, this leads to divergence $\delta$ decreasing with pressure,

$$
-\frac{\partial \delta}{\partial p} > 0,
$$

(3)

and thus convergence and increasing cyclonic vorticity at the surface. Consequently, flow over terrain results in surface cyclone development on the leeward (downslope flow) side (Bluestein 1993). Adiabatic warming and drying precludes the development of precipitation in the downsloping region.

We also utilize the $\mathbf{Q}$-vector form of the inviscid adiabatic QG omega equation (Hoskins et al. 1978):

$$
\left(\nabla^2 + \frac{f^2}{\sigma} \frac{\partial}{\partial p}\right) \omega = -2\nabla_p \cdot \mathbf{Q}.
$$

(4)

Here, the sense of vertical motion is related to the divergence of the $\mathbf{Q}$ vector, supported by Hoskins et al. (1978), who state that “in quasi-geostrophic theory ... vertical velocity is forced solely by the divergence of $\mathbf{Q}$.” In other words, areas of $\mathbf{Q}$-vector divergence are associated with subsidence.

3. Results

There is inherent difficulty in studying a long timescale feature (drought) using synoptic-scale fields. This is addressed by taking a 30-day mean of the synoptic fields, as a compromise between the two temporal scales. However, these fields can be smeared if the signal is weak or if there is large variability in the pattern throughout the 30-day period. The smearing issue is addressed using time–height cross sections, which show the variability during the 30-day period.

Table 3 shows a manual typing of the 10 most severe dry cases identified in Table 1 based on characteristics of the mean flow regime. While the typing is done with strong consideration paid to subsidence mechanisms (i.e., AVA, CAA, and downsloping), the mechanisms are not mutually exclusive. For instance, AVA and CAA often occur concomitantly. However, in persistent ridging regimes, adiabatic warming associated with subsidence can overwhelm CAA, leading to highly positive monthly temperature anomalies. When simply looking on monthly or seasonal time scales, this may lead one to incorrectly conclude that CAA is not an important subsidence mechanism during positive PNA events.

We divide our cases into amplified and zonal upper-level flow regimes, supported by Hanesiak et al. (2011), which recognized two different wave flow types (zonal and meridional, May 2001 and May 2002, respectively) that were both associated with anomalously low precipitation. We further subdivide our amplified flow regimes into warm and cold, dependent on the mean 1000–500-hPa thickness anomaly for the 30-day period. In general, an amplified warm event is associated with a persistent ridge situated over western Canada, typical of a positive PNA pattern. Conversely, an amplified cold event is associated with a meridional polar jet and a distinct and relatively robust subtropical jet, or can be a result of large changes in the flow regime. The separation of the two jet streams in this case precludes the incursion of relatively mild Pacific air masses into polar

<table>
<thead>
<tr>
<th>Type</th>
<th>Central date</th>
<th>Dry ranking</th>
<th>AVA</th>
<th>CAA</th>
<th>Downsloping</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amplified warm</td>
<td>Case A, 29 Dec 2001</td>
<td>1</td>
<td>Yes</td>
<td>(Yes)</td>
<td>No</td>
</tr>
<tr>
<td></td>
<td>Case B, 30 Nov 2002</td>
<td>2</td>
<td>Yes</td>
<td>(Yes)</td>
<td>No</td>
</tr>
<tr>
<td></td>
<td>Case C, 21 Jan 2001</td>
<td>5</td>
<td>Yes</td>
<td>(Yes)</td>
<td>No</td>
</tr>
<tr>
<td></td>
<td>Case D, 12 Dec 2003</td>
<td>6</td>
<td>Yes</td>
<td>Yes</td>
<td>(Yes)</td>
</tr>
<tr>
<td>Amplified cold</td>
<td>Case E, 30 Apr 2005</td>
<td>10</td>
<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td></td>
<td>Case F, 20 May 2002</td>
<td>4</td>
<td>Yes</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Zonal</td>
<td>Case G, 18 Oct 2000</td>
<td>3</td>
<td>Yes</td>
<td>No</td>
<td>Yes</td>
</tr>
<tr>
<td></td>
<td>Case H, 8 Nov 2004</td>
<td>8</td>
<td>Yes</td>
<td>(Yes)</td>
<td>Yes</td>
</tr>
<tr>
<td></td>
<td>Case I, 28 Apr 2001</td>
<td>7</td>
<td>(Yes)</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td></td>
<td>Case J, 15 Oct 1999</td>
<td>9</td>
<td>Yes</td>
<td>(Yes)</td>
<td>Yes</td>
</tr>
</tbody>
</table>
latitudes, leaving the prairies under the general influence of cold and dry polar air masses. Finally, the zonal flow regime is essentially dominated by downsloping. While synoptic-scale disturbances still pass through the region, their small amplitude and high speed of propagation in the absence of significant moisture limits their efficacy in producing precipitation.

An approximate Eady model (Eady 1949) representation (i.e., 300-hPa heights and mean sea level pressure, MSLP) for representative cases of each type are shown in Fig. 6. These cases were chosen based on both their severity rankings as well as their ability to distinctly show the different subsidence mechanisms. A time series of the daily PNA index is also included in Fig. 6 to orient the events on a synoptic scale, and to help offset possible smearing in the mean synoptic fields. The Eady representation is supplemented by a Q-vector and 500-hPa mean flow analysis (Fig. 7), as well as precipitable water and moisture transport averages at 700 hPa (Fig. 8). In addition, time–height cross sections are shown from Saskatoon, Saskatchewan (CYXE, No. 25 in Fig. 2). The fact that the cross section is taken at a single point does not hinder the results, as a time–height cross section approximates a snapshot of an east–west cross section for synoptic-scale features traveling eastward through the station. Saskatoon was chosen since it is located approximately in the middle of the study region and was particularly affected by the drought (Fig. 3b).

### a. Amplified warm cases

Cases A–D (Table 3), all winter examples, are dominated by persistent positive PNA ridging in western North America. Case B (16 November–15 December 2002) is chosen as a representative case (Fig. 6a). The
primary subsidence mechanism in this type agrees with the traditional ridging paradigm for Canadian prairie drought. In fact, three of the five driest cases (Table 1) are associated with ridging, indicating that this is still the

Fig. 7. Cases (a) B, (b) F, and (c) I averaged over the 30-day period. NCEP–NCAR global reanalysis composite 500-hPa Q-vector divergence (shaded, $\times 10^{-16}$ $\text{Km}^{-2}\text{s}^{-1}$), composite Q vectors ($\times 10^{-18}$ $\text{Km}^{-2}\text{s}^{-1}$, arrows), 500-hPa geopotential height (m, solid black), and 700–400-hPa thickness (m, green dashed). For reference, CYXE is marked with a black star and the approximate area of the northern stations is outlined with a blue box.

Fig. 8. The 30-day averages of 700-hPa geopotential heights (dam, solid, 6-hPa interval), PW anomalies (mm, shaded), and 1000–700-hPa moisture transport (kg m$^{-1}$s$^{-1}$, arrows) for cases (a) B, (b) F, and (c) I.
most effective regime for subsidence and drought in the Canadian prairies. However, cases A–D occur during winter, which is climatologically dry in the prairies, and thus may have less impact on drought severity. The prairies are in an expansive region of AVA in the northwesterly flow aloft downstream of the ridge. This sets up a broad region of descent and adiabatic warming [Eq. (1)]. The associated surface divergence leads to anticyclogenesis, as seen by the MSLP anticyclone in the northwestern United States, with an inverted ridge creeping into the southern prairies. Both the upper-level and MSLP ridges deflect storms poleward, preventing entry into the prairies. The MSLP trough extending from the cyclone in the Gulf of Alaska into northern Canada (Fig. 6a), which in a 30-day mean represents a storm track, shows this northward deflection of storms.

The strong cyclone in the Gulf of Alaska, which in conjunction with the continental anticyclone to the southeast produces very strong southerlies off the coast of British Columbia (Fig. 8a). Prior studies have shown that strong Gulf of Alaska cyclones and associated intense water vapor transport (IWVT) into the high latitudes of western North America ("pineapple express") during the cool season enhance downstream ridging in the prairies through warm-air advection (WAA) and diabatic effects due to latent heat release from precipitation (e.g., Roberge et al. 2009). Not only do the surface southerlies in the amplified warm cases transport moisture northward off the coast of British Columbia (Fig. 8a), but they are also responsible for large amounts of southerly WAA in the same area (Fig. 6a). The WAA also creates a thermal ridge in this area (see the 540-dam thickness contour in Fig. 6a). Using the QG height tendency equation (Bluestein 1992), the WAA and diabatic heating are associated with geopotential height rises at 500 hPa, thereby building the downstream ridge (Fig. 6a). In turn, these anomalously positive heights increase the AVA and adiabatic warming due to subsidence, leading to the creation of the anomalously warm thicknesses over western Canada (Fig. 6a). The fact that cases B and D correspond to the 25 November 2002 and 18 December 2003 IWVT events (Roberge et al. 2009) supports the conclusion that IWVT events in the Pacific can impact prairie drought.

Insight into the synoptic-scale variability associated with this 30-day period is provided by the time–height diagram shown in Fig. 9. Geostrophic winds are plotted so that the temperature advection can unambiguously be diagnosed through use of the thermal wind relationship. The dynamic tropopause (DT), as defined by the 2-PVU [potential vorticity unit (PVU), where 1 PVU = 10^{-2} \text{Km}^2\text{kg}^{-1}\text{s}^{-1}] surface (thick black line), is displayed as an indication of the number of synoptic-scale systems that travel through the area. Significant variations in the pressure of the DT can be thought of as transient troughs (higher pressures) and ridges (lower pressures) translating through the region.

A prominent feature of case B is the persistence of northwesterly geostrophic flow in the mid- to upper troposphere from 16 November to 10 December 2002 (Fig. 9a). This reinforces the notion that Saskatoon (CYXE) was situated downstream (upstream) of an upper-level ridge (trough), which is not conducive to synoptic-scale ascent. This period of northwesterly flow coincides with a relatively persistent period of negative anomalies in the specific humidity field (Fig. 9b). This is consistent
with the relatively dry northwesterly trajectory of air into the region. However, there is significant variation in the pressure level of the DT, indicative of a series of short waves that were traversing around the periphery of the persistent upper-level ridge. The two most prominent of these disturbances occur on approximately 23 November and 3 December 2002. In the case of 23 November 2002, relative humidity values fail to saturate throughout the column in spite of substantially deep-column ascent in the presence of WAA, as indicated by the veering geostrophic winds. The failure to reach saturation may be due to persistent descent in the week preceding the passage of the disturbance. While the disturbance on 3 December 2002 is associated with both ascent and a relatively saturated troposphere, the magnitude and extent of the vertical motion are mitigated by the presence of CAA, as indicated by the backing geostrophic winds in the column. The end result is a total of 0.2 mm of liquid equivalent precipitation, which fell in the form of snow on 3 December 2002 (Fig. 9b); this represents the only measurable precipitation during the entire 30-day period.

b. Amplified cold cases

Cases E and F (Table 3) differ from cases A–D in that cold and dry air masses tended to dominate the period. Case F (6 May–4 June 2002; Fig. 6b), which ranked as the fourth driest 30-day period, is chosen as the representative case. The rank of case F is impressive, as it occurred in the spring (the wet season for the prairies) during the 2002 growing season, when the southern prairies experienced greater than normal precipitation, while extreme drought conditions persisted in the northern prairies (Fig. 4b).

AVA still plays a role in case F, but not to the same extent as in cases A–D. In fact, the ridging over the prairies is more evident at the 700-hPa level (Fig. 8b) than the 300-hPa level (Fig. 6b). The ridge weakening with height additionally indicates relatively cold conditions. Split flow (i.e., divergent 300-hPa geostrophic winds on the ridge axis) characterizes this period, with the ridge over the prairies and a trough over the southwestern United States. The fact that the ridging is more confined to the northern jet is supported by the poleward displacement of mean Ω-vector divergence relative to case B (see Figs. 7a and 7b). The PNA in cases E (not shown) and F (Fig. 6b) can be divided into two sections wherein the PNA value approaches zero in the middle of the 30-day period (transition phase). The regimes shift in case F from strongly negative PNA to positive PNA raises the likelihood for smearing in the 30-day mean fields.

The upper-level flow in the Gulf of Alaska is strikingly different between cases B and F; the absence of 300-hPa geopotential height and MSLP contours in this region in case F is remarkable compared to the tightly packed contours in case B. This weak upper-level flow in the Gulf of Alaska in case F leads to a confluence of height contours east of the prairies, effectively deflecting storm tracks through the stronger southern stream and then northeastward, thus avoiding the prairies. The relatively weak surface flow over the prairies in case F is an indication of a relatively inactive atmosphere (lack of disturbances on the DT; Fig. 10a).

The most notable difference between the amplified warm and cold cases is the association of the latter events with strong CAA and anomalously cold thicknesses over the prairies. Despite some smearing, an inverted MSLP ridge and its associated northwesterlies in north-central Canada (Fig. 6b) are evident, which leads to a broad region of CAA in the Northwest Territories, Saskatchewan, Manitoba, and Ontario. This is indicative of cold polar air traveling southward. Using Eq. (1), CAA is associated with subsidence. This broad region of CAA leads to the extensive region of the cold thickness anomalies in central Canada (Fig. 6b), as well as anomalously low 300-hPa heights and negative precipitable water (PW) anomalies (Fig. 8b). This broad region of anomalously cold thickness can be linked to a massive cold-air outbreak at the beginning of this 30-day period (not shown). The fact that the thickness and PW anomaly fields over the prairies in cases B and F are almost
opposite directly challenges the traditional drought paradigm for the Canadian prairies; drought in this region can effectively occur in below-normal temperatures and in the absence of extensive ridging.

Compared to case B, there is significantly more variability in the mid- and upper-tropospheric geostrophic wind directions (especially prior to 22 May 2002), suggesting that persistent upper-level ridging alone is not the main culprit for the excessive dryness of the period. Instead, there seems to be a consistent dearth of saturated or near-saturated deep-tropospheric RH values. This lack of moisture is most evident in the large negative specific humidity anomalies that occur at the beginning of the period (Fig. 10b) in conjunction with the aforementioned cold-air outbreak. Perhaps the two periods most conducive to rain during the 30-day period are 21–22 May and 2–4 June 2002. As early as 15 May, geostrophic southerlies become evident around the back side of a lower-tropospheric anticyclone. However, the deep subsidence associated with the approaching upper-tropospheric ridge appears to preclude saturation of the lower troposphere until 22 May, at which point the sense of the temperature advection has already changed to CAA, signaling the end of the period of ascent. Consequently, it would appear that most of the ascent during this period produced clouds as opposed to producing precipitation.

The dynamics associated with the period of 2–4 June 2002 appear to be much more quiescent, as indicated by the relative lack of geostrophic flow. Perhaps the most notable aspect is the persistent westerly geostrophic flow in the mid- and upper troposphere, which would be consistent with downsloping flow from the Rockies, and precludes the presence of substantial moisture in the lower troposphere. This is supported by the observation that the positive specific humidity anomalies during this disturbance are most pronounced in the mid- and upper troposphere (Fig. 10b), and by measured rainfall amounts of <1 mm. Inspection of the surface observations (not shown) also indicates that 3 June is likely the only day that had any substantial cloud cover, as diurnal temperature variations on 2 and 4 June were 15° and 18°C, respectively. Since the geostrophic winds preclude any significant temperature advections, we can safely conclude that these diurnal temperature variations are associated with the presence or absence of cloud cover.

This above-normal precipitation in the southern prairies seems surprising when negative thickness (Fig. 6b) and PW anomalies (Fig. 8b) extend over this region. However, the 300-hPa trough in the southern branch of the jet generates CVA and associated ascent and surface convergence downstream near Texas, evident by the MSLP cyclone in that location. The accompanying southerly WAA and moisture transport (Fig. 8b) from the Gulf of Mexico northward into the Great Plains drives a thermal ridge south of the CAA-induced thermal trough and, thus, an area of frontogenesis just south of the prairies (Fig. 6f). The frontogenesis is also aided by the easterly (upslope) surface flow near the Canada–U.S. border, and results in conditions favorable for precipitation. The southern prairies thus capture the northern edge of the systems traveling in the southern stream.

c. Zonal cases

Cases G–J are markedly different from the amplified cases. Specifically, cases G–J (Fig. 6c, cases G, H, and J not shown) have an average zonal geostrophic wind of 50 kt (over 30 days; 1 kt = 0.5144 m s⁻¹) at 300 hPa over the prairies as opposed to 35 kt (over 30 days) in cases A–F (Fig. 6, cases A and C–E not shown), an increase of approximately 50%. Zonal flow over the Rocky Mountains leads to forced ascent and surface divergence on the windward side, and downslope flow (descent) and surface convergence on the lee side. Therefore, the increased zonal component of the wind in cases G–J suggests that downslope flow has a greater role in generating the subsidence in these cases. Less direct evidence for downsloping lies in the relative weakness of the Q-vector divergence during this period (Fig. 7c), especially compared to the amplified warm and cold cases (Figs. 7a,b). We would not expect downsloping to be associated with Q-vector divergence, as the latter [Eq. (4)] only explicitly represents the right-hand side of the traditional QG omega equation [Eq. (1)]. Additionally, cases G–J are characterized by a fluctuating low-amplitude PNA signal, in contrast to the persistence and strength of the signal in the amplified cases.

Case I (14 April–13 May 2001; Fig. 6c) ranks as the seventh-driest 30-day period and is chosen as the representative zonal case, since it occurred during the pivotal growing season of 2001. The most striking characteristics of this regime are indicated by the 700-hPa heights and 1000–700-hPa moisture transport (Fig. 8c). The strongest transport of moisture in a time mean sense from the Pacific Ocean is actually situated to the south of Canada. It is also evident that the flow is somewhat diffuent along the western seaboard with a weaker southern branch extending into the United States, while a stronger northern branch crosses southern Canada. As in all of the various drought regimes, a key component contributing to the dry conditions in the Canadian prairies is the lack of southerly moisture transport vectors originating from near the Gulf of Mexico (Fig. 8c). The stronger westerly flow of the northern branch and the implied downsloping results in a time-mean lee trough
situated over Alberta in the mean sea level pressure field (Fig. 6c). Furthermore, unlike the other types, the thickness anomalies suggest an enhanced equatorward-oriented temperature gradient and stronger upper-tropospheric westerlies.

Figure 11 shows a predominance of low-RH values over the 30-day period. This appears to be set up by the passage of a very deep trough at the beginning of the period as indicated by DT pressures approaching 425 hPa and the strongly backing northeasterly-to-northerly winds in the troposphere. Strong subsidence is occurring in conjunction with the backing winds, and RH values of less than 30% dominate the troposphere below 300 hPa until 19 April. At that point, a weak short wave moves through the area, as evidenced by the midtropospheric winds shifting west-southwesterly to west-northwesterly. However, both the magnitude of the wind shift as well as the relatively small change in DT pressure indicate that this is a small-amplitude feature. The most significant trough passage during this period (29 April–2 May) does in fact produce a positive specific humidity anomaly and a small amount of precipitation (Fig. 11b). However, in spite of the generally favorable synoptic-scale conditions, only 2 mm of rain fall at the station and little in the way of ascent is indicated. This is likely a result of the generally westerly winds in the troposphere resulting in orographically forced descent even in the presence of synoptic-scale forcing for ascent.

Table 4. Net 30-day 700–400-hPa omega ($\times 10^{-3}$ Pa s$^{-1}$) over Saskatoon for each of the three representative cases.

<table>
<thead>
<tr>
<th>Case</th>
<th>30-day net omega ($\times 10^{-3}$ Pa s$^{-1}$)</th>
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<tbody>
<tr>
<td>B</td>
<td>+30.89</td>
</tr>
<tr>
<td>F</td>
<td>−15.37</td>
</tr>
<tr>
<td>I</td>
<td>+23.05</td>
</tr>
</tbody>
</table>

As a summary, the net 700–400-hPa omega over Saskatoon for the 30-day period of each representative case has been calculated, and is shown in Table 4. This omega was calculated over a 3 × 3 gridpoint box centered over Saskatoon (7.5° × 7.5°). As expected, the amplified warm case B shows the greatest value of subsidence (+30.89 × 10$^{-3}$ Pa s$^{-1}$), followed closely by zonal case I at +23.05 × 10$^{-3}$ Pa s$^{-1}$. Surprisingly, however, the amplified cold case F shows net ascent (−15.37 × 10$^{-3}$ Pa s$^{-1}$). The reason that this net ascent did not produce precipitation has to do with the extreme anomalously dry state of the troposphere during this period, which is clearly evident from the specific humidity anomalies in the cross section for case F (Fig. 10b). The full-column negative humidity anomalies exist for almost 20 out of 30 days and are very impressive compared to those of cases B and I (Figs. 9b and 11b). Despite the ascent, case F ended up being the driest 30-day growing season period of the drought, adding to the complexity of the drought mechanisms in the Canadian prairies.

4. Conclusions

This study focuses on the synoptic-scale dynamics of the 1999–2005 drought in the Canadian prairies. The meteorology of this drought did not universally conform to traditional paradigms (i.e., positive PNA-like persistent ridging over western Canada) common to previous prairie droughts. It was also unique in its northward extent and lack of persistent teleconnection patterns (Bonsal and Wheaton 2005).

A method involving the percent normal of daily precipitation, with a 30-day running mean filter, was used to identify the 10 driest periods during 1999–2005. A subset of 11 northern stations was used because of the atypical northern extent of this drought, as particularly seen in the high–low precipitation couplet in the growing season of 2002 that helped to break the drought in the southern prairies (Szeto et al. 2011).

A meteorological analysis of the 10 driest cases revealed three event types based on flow regime. The first regime (amplified warm) agrees with the traditional paradigm for Canadian prairie drought: positive PNA ridging in western North America, resulting in a broad region of AVA in the prairies, and deflecting storms to
the north. In amplified warm cases, a strong Gulf of Alaska cyclone coupled with a continental anticyclone to the southeast drove extreme southerly WAA of the coast of British Columbia, enhancing the downstream ridge (Roberge et al. 2009). All four events in this regime occurred in winter, and three out of the top five most extreme dry events fell into this category. This suggests that large-scale ridging and associated AVA are still the most effective mechanisms for subsidence in the prairies, and that this regime is most prevalent in the winter.

The second regime (amplified cold) is characterized by cold and dry air masses, with AVA playing a less important role. Both cases that fell into this regime occurred during spring and had weak upper-level flow in the Gulf of Alaska. The third regime (zonal) has a much greater zonal component in the upper-level flow, making downsloping the dominant subsidence mechanism. In zonal cases, the PNA was variable and of small amplitude. Downslope flow was aided at the surface by the positioning of an anticyclone directly south of the Gulf of Alaska cyclone, enhancing surface westerlies. This downsloping regime is a very effective mechanism for drought, with one of these cases ranking just as high as amplified warm cases. Thus, during periods of weak and inconsistent PNA, downsloping appears to be an important subsidence mechanism for the prairies.

Overall, the meteorological analysis of the 10 driest periods showed that synoptic-scale flow regimes that bring about drought in the Canadian prairies are more complex than was originally proposed by Dey (1982). There was no single flow regime that was distinct to the 1999–2005 drought. It did not disprove the traditional paradigm for Canadian prairie drought; large-scale positive PNA ridging remains the most effective subsidence mechanism, particularly during the winter season. However, this analysis brought to light the importance of downslope flow as a smaller-scale subsidence mechanism that was just as effective at causing subsidence. This analysis shows that drought conditions in the Canadian prairies can be effectively maintained even without persistent large-amplitude positive PNA, since the amplified cold cases featured both positive and negative PNA patterns over a 30-day period, and the zonal cases were characterized by a near-neutral PNA signal.

Our study of the 1999–2005 drought reveals the extreme sensitivity of the Canadian prairies to drought; it does not require one persistent pattern in order for meteorological drought to occur. This leads to the idea that this drought peaked in meteorological severity in 2001 and 2002, but was not a particularly exceptional meteorological drought when compared to the entire 1948–2005 period (Bonsal and Regier 2007). This can be seen in Fig. 12, where the 1999–2005 drought is not more prominent than the other two droughts since 1948 (1961 and 1988). Overall, the great economic impacts of this drought did not match up to the degree of meteorological significance. This implies that the larger issue is that society is increasing its vulnerability to natural hazards such as drought. With the changing climate and the expected poleward expansion of arid areas (Cook et al. 2010), it is inevitable that drought will still occur often in the Canadian prairies. Whether these future droughts will be meteorologically unprecedented or not, perhaps the Canadian prairie society should direct its efforts toward drought mitigation and adaptation in order to decrease its drought vulnerability, bringing the impacts closer in line with their meteorological significance.

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