# Trends in Canadian Precipitation Intensity

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ABSTRACT *Past research has unveiled important variations in total precipitation, often related to large-scale shifts in atmospheric circulation, and consistent with projected responses to enhanced greenhouse warming. More recently, however, it has been realized that important and influential changes in the variability of daily precipitation events have also occurred in the past, often unrelated to changes in total accumulation.*

*This study aims to uncover variations in daily precipitation intensity over Canada and to compare the observed variations with those in total accumulation and two dominant modes of atmospheric variability, namely the North Atlantic Oscillation (NAO) and the Pacific/North America teleconnection pattern (PNA). Results are examined on both annual and seasonal bases, and with regions defined by similarities in monthly variability.*

*Seasonally increasing trends in total precipitation that result from increases in all levels of event intensity during the 20$^{th}$ century are found in southern areas of Canada. During the latter half of the century increases are concentrated in heavy and intermediate events, with the largest changes occurring in Arctic areas. Variations in precipitation intensity can, however, be unrelated to variations in the total accumulation. Consistent with these differences, the precipitation responses to the NAO and PNA are often found to occur only at specific levels of event intensity. Precipitation responses to the NAO occur in northeastern regions in summer and winter with the intensity affected in both seasons. The PNA strongly influences precipitation in many regions of the country during autumn and winter. In particular, it strongly influences variations in southern British Columbia and the Prairies, affecting the intensity in only some areas. However, it only influences the frequency of heavier events in autumn and winter in Ontario and southern Quebec, where this response is actually more robust than the response in total accumulation. During these seasons a negative PNA generally leads to more extreme precipitation events.*

RÉSUMÉ *Des recherches antérieures ont révélé d'importantes variations dans les précipitations totales, souvent reliées aux changements à grande échelle de la circulation atmosphérique et consistantes avec les réactions prévues du réchauffement dû à l'effet de serre. Toutefois, plus récemment, on a constaté que des modifications importantes et déterminantes*

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*dans la variabilité des événements de précipitations quotidiennes ont aussi eu lieu dans le passé, mais souvent non reliées à des modifications dans les précipitations totales.*

*Cette étude a pour objet de révéler les variations dans l'intensité des précipitations quotidiennes sur le Canada et de comparer les variations observées avec celles de l'accumulation totale et de deux modes dominants de la variabilité atmosphérique, notamment l'Oscillation Nord-Atlantique (ONA) et celui de la configuration de la téléconnexion du Pacifique/Nord-Américain (PNA). Les résultats sont examinés sur des bases à la fois saisonnière et annuelle, avec des régions définies par des similarités dans la variabilité mensuelle.*

*Les tendances saisonnières croissantes dans les précipitations totales ont été observées sur des régions méridionales du Canada, dues aux augmentations dans tous les niveaux d'intensité d'événement au cours du vingtième siècle. Au cours de la dernière moitié du siècle, les augmentations sont concentrées sur les événements majeurs et intermédiaires, les modifications les plus grandes se produisant dans les régions de l'Arctique. Toutefois, les variations dans l'intensité des précipitations peuvent être non reliées aux variations dans l'accumulation totale. Consistentes avec ces différences, les réactions des précipitations à l'ONA et à la téléconnexion du PNA se produisent souvent à des niveaux spécifiques d'intensité d'événement. Les réactions des précipitations à l'ONA se produisent dans les régions du nord-est en été et en hiver, l'intensité étant affectée dans les deux saisons. La téléconnexion du PNA influence fortement les précipitations dans plusieurs régions du pays au cours de l'automne et de l'hiver. En particulier, ce dernier mode influence fortement les variations dans la province de la Colombie-Britannique méridionale et dans les Prairies, affectant l'intensité dans seulement quelques régions. Cependant, en Ontario et sur le Québec méridional, il influence seulement la fréquence d'événements majeurs en automne et en hiver, où cette réaction est plus vigoureuse que celle dans l'accumulation totale. Au cours de ces saisons, une configuration négative de la téléconnexion du PNA mène généralement à plus d'événements de précipitations extrêmes.*

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## 1 Introduction

Of all climate variables, precipitation probably impacts humanity most directly and significantly, with variations or changes often bringing enormous economic, environmental, social and political repercussions. This importance justifies the analysis of causes of past variations in its own right, as a means of understanding future changes. Moreover, in the context of global warming due to anthropogenic greenhouse gas emissions, such analysis serves the additional role of a method of monitoring such climate change.

In theory, enhanced greenhouse warming should force stronger large-scale convection, resulting in enhanced moisture transport from the tropics towards the poles (Gordon et al., 1992; Kattenberg et al., 1996; Hennessy et al., 1997). Furthermore, a warmer atmosphere should hold more precipitable water, resulting in a more dynamic hydrological cycle. Indeed, general circulation models (GCMs) consistently project an increase in global total precipitation when greenhouse gas concentrations are increased, with the largest relative changes projected to occur at high latitudes in winter, with frequent extension into middle latitudes (Kattenberg et al., 1996, and references therein; Hennessy et al., 1997; Hulme et al., 1998; Boer et al., 2000a; Boer et al., 2000b).

In most regions an alteration of the frequency or timing of extreme precipitation events would have a more pronounced economic impact than a change in total precipitation (Katz and Brown, 1992). As mentioned, a warmer and moister atmosphere should lead to a more dynamic hydrological cycle and thus greater atmospheric instability, resulting in more extreme events. Noda and Tokioka (1989), Gordon et al. (1992), Kattenberg et al. (1996) and Hennessy et al. (1997) found shifts to more convective and less non-convective precipitation in GCM-simulated climates under $CO_2$-doubling, implying more intense precipitation events. Zwiers and Kharin (1998) examined the daily precipitation simulated by the Canadian Centre for Climate Modelling and Analysis (CCCma) GCM, and found more intense precipitation globally under double $CO_2$. During a realistic transient climate change experiment Cubasch et al. (1995) found slight regional increases in both rainfall intensity and drought duration. Kharin and Zwiers (2000) obtained results comparable to Zwiers and Kharin (1998) from an ensemble of CCCma GCM transient integrations. Increases in extreme precipitation were broadly significant over land at the time of $CO_2$-tripling, and more robust than changes in total accumulations.

In general, global (Bradley et al., 1987; Diaz et al., 1989; Vinnikov et al., 1990; Eischeid et al., 1991; Dai et al., 1997) and regional (Groisman and Easterling, 1994; Nicholls et al., 1996, and references therein; Karl and Knight, 1998; Hennessy et al., 1999; Mekis and Hogg, 1999) studies of total precipitation have found an increasing trend, concentrated during the winter in high northern latitudes, consistent with enhanced greenhouse warming projections. However, a limited-number of regional studies (e.g., Iwashima and Yamamoto, 1993; Nicholls and Kariko, 1993; Karl et al., 1995; Karl and Knight, 1998; Hennessy et al., 1999) have detected no overall trend in extreme heavy precipitation events, finding instead a mixture of regional trends of varying sign and magnitude. Using a statistical model based on observed precipitation parameters, Groisman et al. (1999) deduced that over most of the extratropics the observed increase in total precipitation has resulted in a disproportionate increase in the frequency of heavy daily events.

Much of the observed variability in atmospheric pressure and circulation takes the form of persistent and recurring large-scale patterns of pressure anomalies (Wallace and Gutzler, 1981; Barnston and Livezey, 1987; Thompson and Wallace, 1998). Several investigations have revealed that variations in regional precipitation, including trends, relate to changes in these patterns (e.g., Nicholls and Kariko, 1993; Hurrell, 1995; Dai et al., 1997; Appenzeller et al., 1998; Cayan et al., 1998; Robertson and Ghil, 1999). The possibility exists that enhanced global warming could alter these atmospheric circulation modes (Corti et al., 1999; Fyfe et al., 1999).

Over southern Canada, Vinnikov et al. (1990) and Dai et al. (1997) found a linearly increasing trend over the last century. Groisman and Easterling (1994) found a similar increase, concentrated in the east, as well as a larger increase in the last four decades in northern Canada. Mekis and Hogg (1999) found corresponding smaller but significant changes using an improved dataset, and also noted distinct seasonal patterns in the trends. Groisman et al. (1999) concluded that this increase would be reflected in a disproportionate increase in heavy precipitation. In fact, Mekis and Hogg (1999) found a decrease in the heaviest decile of precipitation events in southern Canada during 1910–1995, but an increase in northern Canada during 1940–1995. Hogg et al. (1998) noticed seasonal and regional properties in these trends and in the observed long term variability of these extreme events. Akinremi et al. (1999) examined daily events of different classes of intensity over the Prairies, and found an increase in the frequency of lighter events over the 1921–1995 period. As the only analyses of temporal changes in Canadian precipitation extremes thus far, these results are significant; however, difficulties exist with the interpretation of the quantile-derived results in northern regions, where the top decile can correspond to a threshold as low as 0.5 mm. Also, considering the regional nature of variations detected in other regions of the world and the results of Hogg et al. (1998), these trends probably possess more complicated seasonal and spatial structures.

The purpose of this paper is to elucidate further variations in Canadian precipitation and its intensity, and to relate these to variations in the dominant large-scale modes of atmospheric variability. The Canadian precipitation record clearly suits such an analysis: reliable measurements exist at stations spanning the country for the latter half of the twentieth century, and for most of the century in southern regions. Furthermore, a new daily station dataset (Mekis and Hogg, 1999) improves upon their reliability. The Canadian record is also well situated geographically for this purpose: the Canadian landmass straddles both middle and high latitudes, where the projected precipitation responses to enhanced greenhouse warming should be strongest (Kattenberg et al., 1996).

The remainder of this paper proceeds as follows. Section 2 describes the data and methods used in this investigation, as well as the regional character of the observed precipitation variations. Section 3 consists of an analysis of these variations and trends according to changes in the intensity of daily precipitation events, and a comparison with variations in total accumulation. Section 4 contains a comparison of the variations in observed precipitation with variations in the NAO and the PNA teleconnection pattern, in order to elucidate relationships between them. Finally, Section 5 provides a discussion that assimilates these results and examines them in relation to previous findings.

## 2 The data and methods

Consistent, long term records of climate variables are difficult to obtain. The introduction of new, more advanced or efficient instruments or observing procedures, changes in the unit of measurement, and changes in local environment (e.g. relocation, changes in wind obstruction) all introduce discontinuities into the records. Due to its nature, records of precipitation are especially susceptible to these biases (e.g. Groisman and Easterling, 1994; Metcalfe et al., 1997; Mekis and Hogg, 1999).

In order to reduce the influence of observational biases, this analysis uses a corrected dataset provided by the Climate Monitoring and Data Interpretation Division (CCRM), Climate Research Branch, Meteorological Service of Canada, Environ-

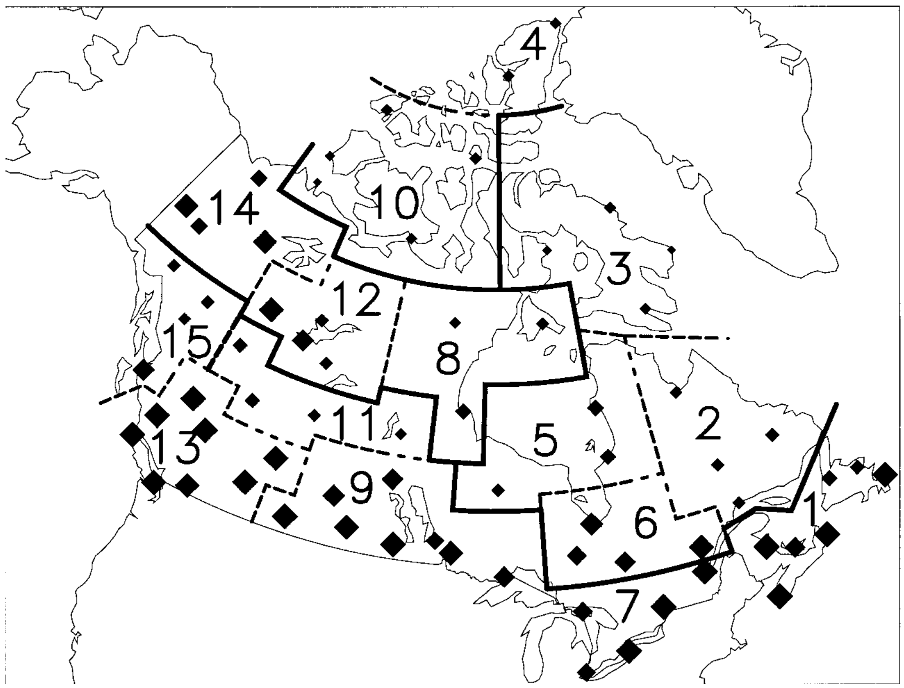


Fig. 1 Map of the stations in the precipitation dataset. Diamond size represents the length of the station record, ranging from 34 to 102 years. See Table 2 for a description of the regions and districts. Solid lines represent borders between districts (and regions), while dashed lines represent borders between regions only.

ment Canada (Mekis and Hogg, 1999). It contains the daily rain-equivalent precipitation records, for liquid and solid precipitation separately, for 69 stations scattered across Canada (Fig. 1). Record lengths range from 34 to 102 years within the 1895–1996 period covered. Adjustments made to the dataset include systematic corrections for evaporative, wetting and wind losses from gauges; adjustments for changes in measurement procedures over time; and quantitative allowance for trace measurements. Snowfall amounts, obtained from ruler measurements in order to minimize discontinuities, are adjusted to more accurate rain-equivalent values using station-dependent ratios obtained from a comparison with reliable present-day gauge measurements. The effects of station relocations are also accounted for by using an adjustment based on data obtained simultaneously at both sites.

In order to deal with missing data, months, seasons and years are defined as active at a given station provided they possess a specified critical number of observations (Table 1). Rather strict criteria are chosen for seasonal and annual data since precipitation possesses a strong annual cycle in many areas. However, the number of rejected years (or months or seasons) is rather insensitive to the exact criteria values.

TABLE 1. Restrictions on missing data.

| Data format | Minimum number of measurements |
|---|---|
| Monthly | 25 days in the month |
| Seasonal | All 3 months in the season |
| Annual | 355 days in the year |

TABLE 2. Properties of the geographical regions and districts. The listed active period, when station density (separation) is greater than $3 \times 10^{-6}$ km$^{-2}$ (580 km), is for monthly data (intervals for annual or seasonal data are sometimes shorter).

| ID | Region or district name | Area ($10^3$ km$^2$) | Station number | Station density ($10^{-6}$ km$^{-2}$) | Mean station separation (km) | Station density (separation) $\geq 3 \times 10^{-6}$ km$^{-2}$ (580 km) |
|---|---|---|---|---|---|---|
| SEC | Southeastern Canada | 708 | 12 | 17 | 240 | 1895–1996 |
| 1 | Atlantic | 327 | 7 | 21 | 220 | 1895–1996 |
| 7 | Great Lakes-St. Lawrence | 381 | 5 | 13 | 280 | 1895–1996 |
| NEC | Northeastern Canada | 2824 | 15 | 5 | 450 | 1943–1996 |
| 2 | Northern Quebec-Labrador | 866 | 4 | 5 | 450 | 1944–1996 |
| 3 | Baffin Island | 704 | 4 | 6 | 410 | 1957–1996 |
| 5 | Hudson Bay | 588 | 3 | 5 | 450 | 1926–1993 |
| 6 | Southern Quebec | 665 | 4 | 6 | 410 | 1895–1996 |
| AC | Arctic Canada | 1071 | 7 | 7 | 380 | 1948–1996 |
| 4 | Queen Elizabeth Islands | 257 | 2 | 8 | 350 | 1948–1996 |
| 10 | Kitikmeot | 814 | 5 | 6 | 410 | 1948–1995 |
| SWC | Southwestern Canada | 3124 | 24 | 8 | 350 | 1897–1995 |
| 9 | Southern Prairies | 1154 | 8 | 7 | 380 | 1899–1996 |
| 11 | Northern Prairies | 702 | 4 | 6 | 410 | 1944–1995 |
| 13 | Southern British Columbia | 818 | 8 | 10 | 320 | 1895–1995 |
| 15 | Northern British Columbia | 451 | 4 | 9 | 330 | 1939–1995 |
| NWC | Northwestern Canada | 2233 | 11 | 5 | 450 | 1932–1996 |
| 8 | Keewatin | 718 | 3 | 4 | 500 | 1949–1996 |
| 12 | Great Slave Lake | 734 | 4 | 5 | 450 | 1942–1996 |
| 14 | Yukon-MacKenzie | 781 | 4 | 5 | 450 | 1925–1995 |
|  | Total | 9959 | 69 | 7 | 380 |  |

According to these criteria, all stations are active over the 1960–1990 period, hereafter defined as the "climatological mean" period.

Pooling stations into regional groupings assists in understanding the spatial pattern of precipitation variations. Thus fifteen regions, described in Table 2, are defined on the condition that all stations within a region must be significantly correlated (one-sided 5% significance level) in their de-trended 1960–1990 monthly fractional precipitation anomalies, defined according to

$$P^f_{a,m} = \frac{P_{a,m} - \bar{P}_{1960\text{–}1990,m}}{\bar{P}_{1960\text{–}1990,m}}$$

where $P_{a,m}$ is the total accumulation for month $m$ of year $a$, and $\bar{P}_{1960\text{–}1990,m}$ is the 1960–1990 mean accumulation for month $m$. $P^f_{a,m}$ is then the fractional anomaly for month $m$ of year $a$. The fifteen regions defined in this manner are the minimum required for inclusion of all stations. They also largely conform to regions used in other studies defined according to mean climatology (e.g. Mekis and Hogg, 1999). Regional time series are constructed by computing the arithmetic mean of the station fractional anomalies.

Many similarities exist between the variations of these regions; thus, they are further grouped into five larger regions (Table 2), hereafter denoted “districts” for clarity. While it is not the case at the national level, within regions and districts station density remains fairly uniform, thus minimizing spatial biases. Both regions and districts are considered active for years when the station density is greater than $3.0 \times 10^{-6}$ stations/km$^2$, corresponding to an average station separation of at most 580 km. The spatial correlation between stations decreases to 0.5 at station separations of 200–300 km, but is still statistically significant at the 5% level for station separations of up to 500 km.

A single threshold cannot satisfactorily define extreme heavy daily events at all stations since the intensity of precipitation events varies considerably between stations. Station-dependent thresholds are therefore required. However, difficulties arise when using quantile-defined thresholds due to measurement resolution changes and adjustments accounting for observational biases. For example, prior to 1977 precipitation was recorded in units of 0.01 in. (0.25 mm), after which the units were switched to 0.20 mm (Metcalfe et al., 1997). An event, $p$, that actually produced between 0.015 in. (0.38 mm) and 0.025 in. (0.64 mm) of precipitation would have been recorded as 0.02 in. (0.51 mm) prior to 1977. After the switch to the metric system, this same event would have been recorded as either 0.40 mm (if 0.38 mm $< p <$ 0.50 mm) or 0.60 mm (if 0.50 mm $< p <$ 0.64 mm). Quantile-defined thresholds tend to separate these values, producing an artificial change in the frequency of events above or below the threshold in 1977. Thus different station-dependent thresholds must be used. Three intensity classes are therefore defined following the criteria of Table 3. The first threshold, 0.6 mm da$^{-1}$, separates the small trace values most susceptible to observational biases. The 2 mm da$^{-1}$ threshold appears a useful division at all stations, separating “light” and “intermediate” events. The final threshold value, separating “intermediate” and “heavy” events, is station dependent, but is always a multiple of 5 mm da$^{-1}$. The value of this threshold for each station is the highest possible value that still results in at least five events per year or three-month season (over the 1960–1990 record), depending on which period is being analysed. This ensures that heavy precipitation event frequencies are large enough to allow in their analysis the approximation of the binomial distribution of the count

TABLE 3. Definition of precipitation intensity classes. For each station, $n$ is the highest non-negative integer that still results in an average of at least five "heavy" precipitation events per year (over the 1960–1990 period). At stations where this criterion cannot be satisfied, the value $n = 0$ is used. Thresholds for seasonal data are obtained analogously.

| Intensity class | Definition criteria | |
|---|---|---|
| Light | $\geq$0.60 mm da$^{-1}$ | $<$2.00 mm da$^{-1}$ |
| Intermediate | $\geq$2.00 mm da$^{-1}$ | $<$(5.00 + 5.00 × $n$) mm da$^{-1}$ |
| Heavy | $\geq$(5.00 + 5.00 × $n$) mm da$^{-1}$ | — |

data by the Gaussian distribution. At some northern stations this criterion cannot be satisfied even for a 5 mm da$^{-1}$ threshold; however, this value can still be used when stations are grouped into districts or regions.

Time series are constructed for each intensity class and station using the above definitions, recording the number of daily events per year or season. These time series are then joined together as follows to produce regional and district time series. The frequency of daily events in an intensity class at each station is converted to a fractional anomaly value according to

$$P_a^f = \frac{P_a - \bar{P}_{1960\text{–}1990}}{\bar{P}_{1960\text{–}1990}}$$

where $P_a$ is the number of daily events in the intensity class occurring in year $a$, and $\bar{P}_{1960\text{–}1990}$ is the average of $P_a$ over the 1960–1990 period. Regional and district fractional values are the arithmetic average of the corresponding station values for each year. Fractional anomaly values for three-month seasons are obtained analogously.

Section 4 consists of an investigation of the effects of large-scale modes of atmospheric variability on both Canadian precipitation accumulation and the frequency of events in each intensity class. The atmospheric modes considered are the NAO and the PNA, which are persistent and recurring large-scale patterns of tropospheric pressure and circulation anomalies (Wallace and Gutzler, 1981; Barnston and Livezey, 1987; Thompson and Wallace, 1998). Time series of the amplitudes of these two patterns were obtained from the Climate Prediction Center, National Oceanic and Atmospheric Administration, U.S. Department of Commerce (http://nic.fb4.noaa.gov:80/data/teledoc/telecontents.html). Values were calculated in a manner similar to Barnston and Livezey (1987). The most influential patterns, or dominant modes of variability, were determined by applying a Rotated Principal Component Analysis to monthly 700-mb height anomalies between January 1964 and July 1994. Modes for each calendar month were calculated on a seasonal basis (i.e., January, February and March for the February modes), resulting in a larger sample size and greater continuity between months. Monthly amplitudes were then obtained through a least squares regression analysis onto the dominant modes. Data are available for all months since January 1950.

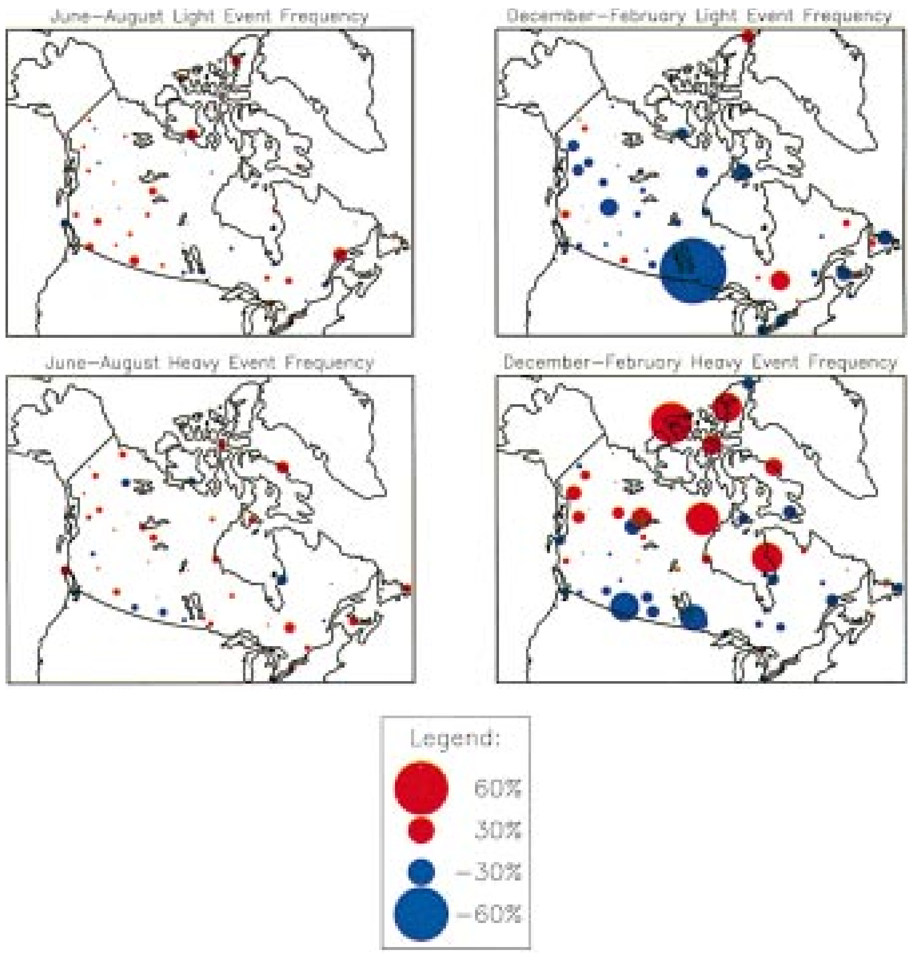


Fig. 2 Map of station trends in the frequency of summer and winter light and heavy daily precipitation events over the 1950–1990 period, when data are available at most stations. Trends are in percent of the 1960–1990 mean. Red disks denote positive trends, while blue disks denote negative trends. Trend values are in percent of mean per decade. See Table 3 for intensity class definitions.

## 3 Observed trends and variations

A district-by-district analysis of seasonal variations and trends in the frequency of events in the different intensity classes is performed. Seasons are denoted by their initials (e.g. JFM for January–March). Figure 2 shows a map of the trends at each station for summer and winter light and heavy events for the 1950–1990 period. Data are available at most stations during this period. The most notable feature of these four maps is the large increasing trend in the frequency of heavy daily winter precipitation at northern stations. There are no comparable increases in the frequency of light daily winter precipitation. Figures 3, 5, 7, 9 and 11 display the annual and seasonal variations in both total accumulation and intensity class event frequency for each of the five districts, while Figs 4, 6, 8, 10 and 12 display the 1950–1995 linear trends in each of these time series.

**a** *Southeastern Canada*

Light, intermediate, and heavy events all become progressively more frequent from 1920 to 1970, after which the rate of intermediate and heavy events remains fairly constant (Fig. 3). The frequency of light events, however, begins to decrease in about 1985. This pattern generally holds in all four seasons for intermediate and heavy events. The large increase prior to 1940 in winter light events could be related to observational biases due to the greater difficulties in measuring small snowfall accumulations (Metcalfe et al., 1994), however, Mekis and Hogg (1999) found no evidence of any changes in observing procedures at this time (W.D. Hogg, personal communication). Figure 4 displays the seasonal intensity class trends in this district for the 1950–1995 period. While light events decrease in winter, intermediate and heavy events increase in autumn and summer respectively.

**b** *Northeastern Canada*

This district experiences notable semi-periodic variations in precipitation, generally visible in all seasons and in heavy and intermediate intensity classes (Fig. 5). The frequencies of events in summer classes vary largely in tandem with each other, with the same occurring with the winter classes. On the other hand, light events are visibly different from the other two classes in the spring. In autumn intermediate events do not reflect the low number of light and heavy events prior to 1965. In 1985 all three classes in this season start increasing. Figure 6 displays the 1950–1995 seasonal trends in each of the intensity classes in this district. Increases statistically significant at the 5% level occur in all three classes in autumn, with an increase in heavy events also occurring during the spring.

**c** *Arctic Canada*

Figure 8 displays the 1950–1995 linear trends in event intensity for Arctic Canada. Most strikingly, while heavy events generally increase significantly at the 5% level in all seasons but the summer, and intermediate events increase in all seasons, light events show no statistically significant seasonal trends. Overall, the magnitudes of the increasing trends in intermediate and heavy events are comparable, with the largest increases occurring in winter and spring. In most cases, the bulk of the increase occurs from 1950 to 1970 (Fig. 7), with levels generally constant afterwards.

**d** *Southwestern Canada*

Intermediate and heavy events in winter, and for a shorter period in autumn, reach some of their lowest levels of the century during the years around 1930, correspond-

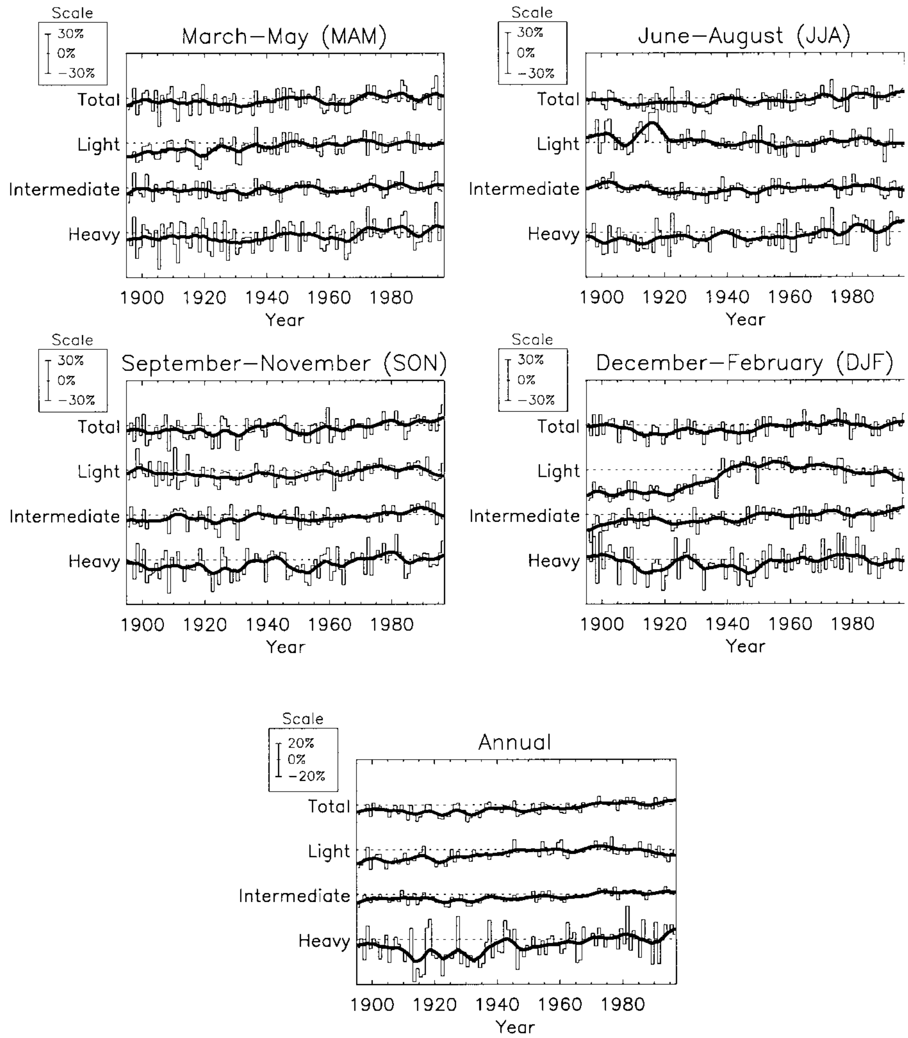


Fig. 3 Annual and seasonal anomalies in Southeastern Canada daily precipitation intensity classes. The vertical scale is the frequency of the daily events in percent of the 1960–1990 mean. Note that the scale, given in the top left corner, differs between plots. The top time series in each plot is the total accumulation. The dotted line is the 1960–1990 mean; the thick line is the 9-year running mean using a triangle filter. See Table 3 for intensity class definitions.

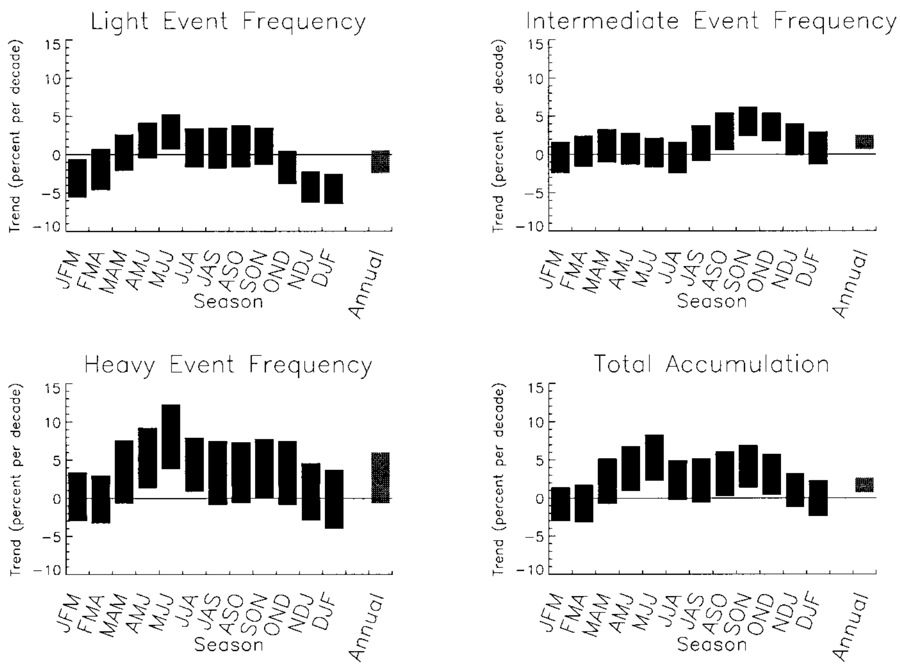


Fig. 4 1950–1995 trends in Southeastern Canada precipitation intensity classes. Bars denote the 95% confidence interval for each trend value. Seasons are denoted by their initials. See Table 3 for intensity class definitions.

ing to the "Dust Bowl" years (Fig. 9). As in Southeastern Canada, the large increases in light events prior to 1940 in autumn, winter, and spring light events may result from observational biases affecting small snowfall accumulations (Metcalfe et al., 1994). Intermediate events during these seasons also show smaller increases over this period. Later, only light and intermediate events decrease in spring and summer in the late 1960s. Meanwhile, all three classes record unusually high numbers in winter from 1960 to 1975. At this time all three winter classes drop substantially, not returning to normal levels until at least 1990. This is partially mirrored in all three autumn classes, as well as spring light events. These changes coincide with major shifts in atmospheric circulation associated with the PNA, discussed in Section 4. Figure 10 displays the seasonal intensity class trends in this district for the 1950–1995 period. Decreases significant at the 5% level in light events occur in winter and spring, accompanied by intermediate events in FMA only. Heavy events increase significantly in MJJ during this interval.

**e** *Northwestern Canada*

Heavy events generally occur less frequently in all seasons prior to 1960, as do winter intermediate events and both autumn light and intermediate events (Fig. 11).

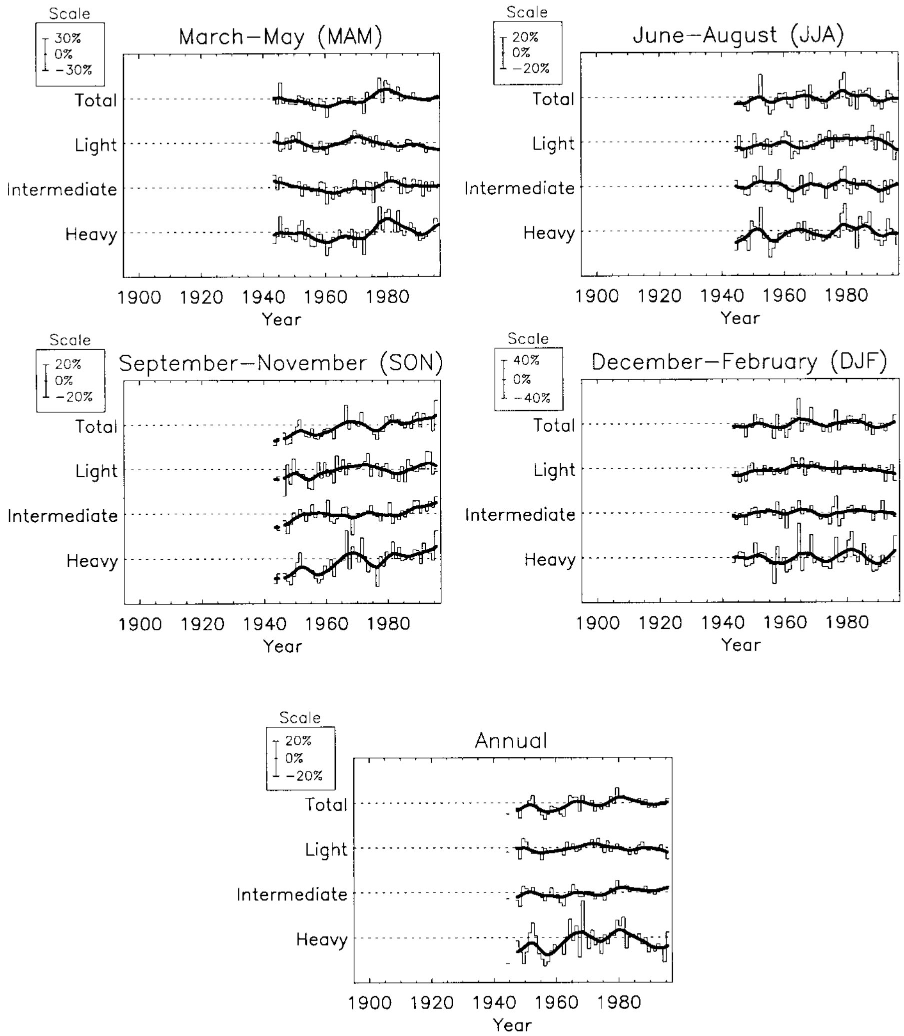


Fig. 5 As in Fig. 3 but for Northeastern Canada.

Otherwise, all three classes largely vary together. For instance, all three, except for heavy events in spring and autumn, record lower numbers during the early 1980s in autumn, winter and spring. Figure 12 displays the 1950–1995 seasonal trends in each of the intensity classes in this district. Heavy events become significantly more frequent in the winter and spring during this period, accompanied by intermediate events in DJF only. Light events, on the other hand, decrease at this time.

In all five regions variations in the frequency of heavy and intermediate daily events are generally reflected in the total accumulation; this relationship does not

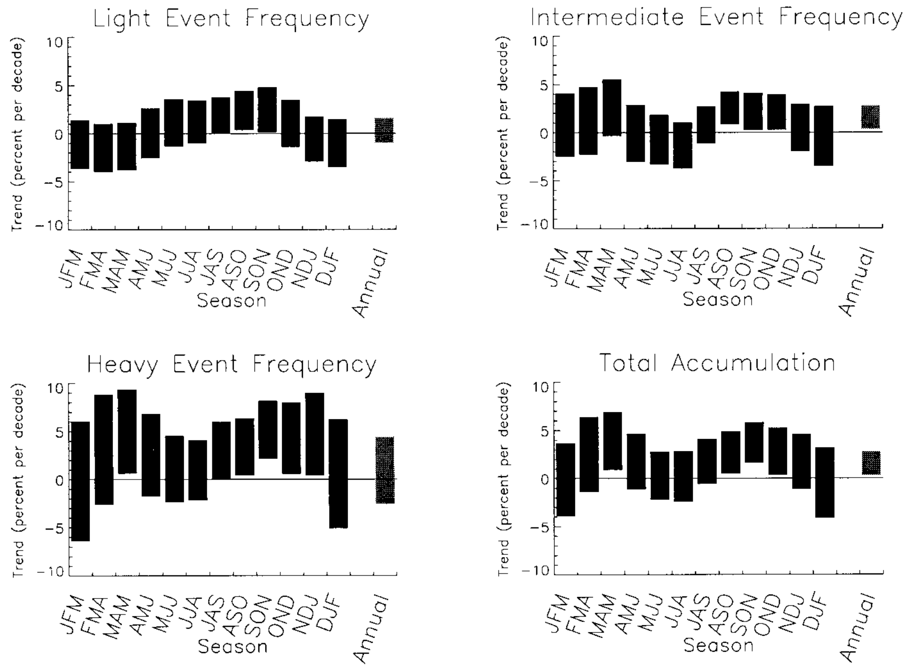


Fig.6 As in Fig. 4 but for Northeastern Canada.

tend to hold with light daily events. This arises naturally from the higher weighting of heavier events in their contribution to the total accumulation. Overall, variations in precipitation over Canada since 1950 are dominated by an increasing trend, concentrated in northern regions. This trend results mainly from an increase in the frequency of heavy and intermediate daily events, with little change in the frequency of light daily events.

## 4 Relationships with modes of atmospheric variability

Recent analyses demonstrate the importance of certain persistent and recurring atmospheric circulation anomalies (Wallace and Gutzler, 1981; Barnston and Livezey, 1987; Thompson and Wallace, 1998) in determining long term variations in regional precipitation (Hurrell, 1995; Dai et al., 1997; Appenzeller et al., 1998; Cayan et al., 1998; Robertson and Ghil, 1999). Further indirect evidence (Gershunov and Barnett, 1998) suggests that important responses occur not only in the total accumulation, but also in the intensity. This section examines the role of two well known and dominant modes of atmospheric variability, specifically the NAO and the PNA, in determining long-term regional precipitation variations. They are the two strongest and most robust modes of variability in atmospheric pressure and circulation in the extratropical Northern Hemisphere on timescales ranging from

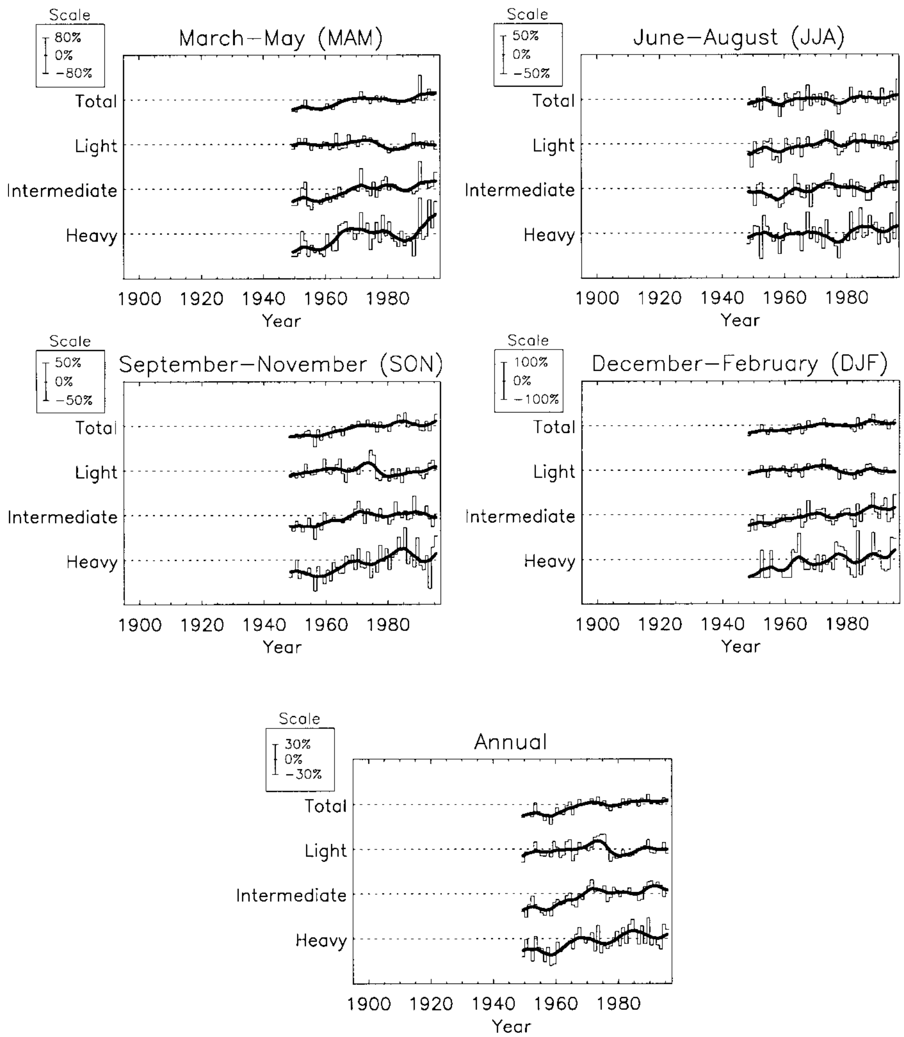


Fig. 7 As in Fig. 3 but for Arctic Canada.

weeks to years. Furthermore, they are both known to exert a noticeable influence on Canadian climate (e.g. Hurrell, 1996; Dai et al., 1997; Shabbar et al., 1997; Cayan et al., 1998).

**a** *The North Atlantic Oscillation*

The NAO is one of the top three modes of variability in atmospheric pressure and circulation in most months, although it is most influential during the winter (Barnston and Livezey, 1987). It consists of a north-south dipole of pressure anomalies,

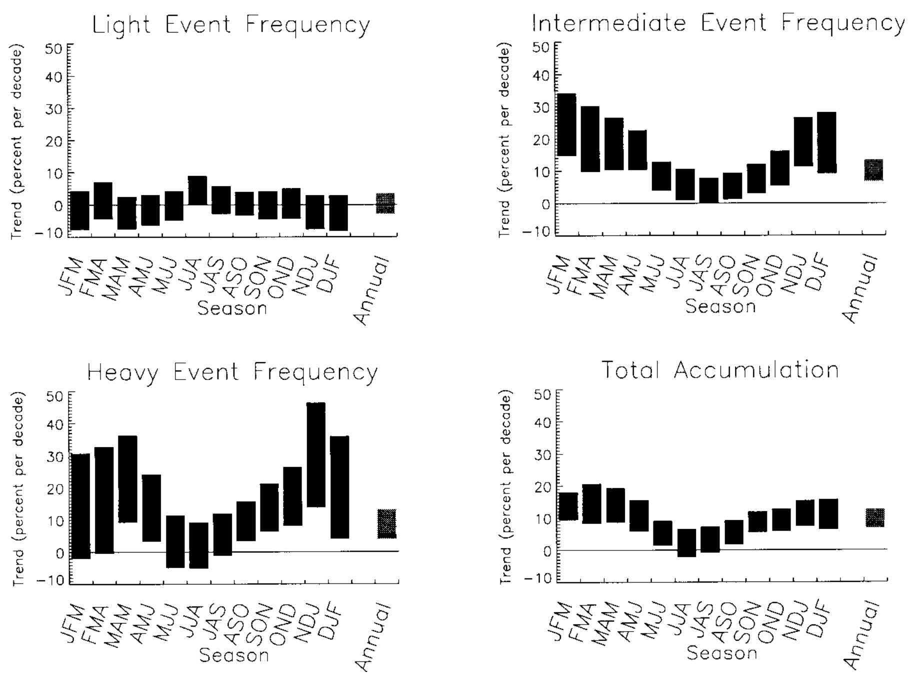


Fig. 8 As in Fig. 4 but for Arctic Canada.

with one centre over Greenland and another of opposite sign spanning the North Atlantic between 35°N and 40°N. By convention, the phase is defined by the sign of the North Atlantic band.

Table 4 lists the correlation of the NAO index with the regional frequency of daily events in the three intensity classes, as well as the total accumulation, for the three-month seasons during which a statistically significant number of regions and intensity classes are affected. One would expect about 5% of correlations to be significant at the 5% level by random change when there is no relationship with the NAO index. About 8.1% of the correlations shown in Table 4 (including the part of the table that is not displayed, but excluding total accumulation) is significant. However, in the case of the NAO, one would expect to find physical relationships restricted to the east since the NAO only affects atmospheric circulation in this area. Examining the eight easternmost regions along with Queen Elizabeth Islands (Region 10), all three levels of event intensity, and all twelve three-month seasons, there are 324 correlations, of which 35, or 10.8%, are significant at the 5% level. Calculation of the binomial distribution with probabilities 0.05 and 0.95 finds that at least 60 degrees of freedom are required to accept the ensemble of values as significant at the 5% level (Livezey and Chen, 1983). Even supposing that there are in fact only four degrees of freedom in seasons and two in event intensity, this requirement

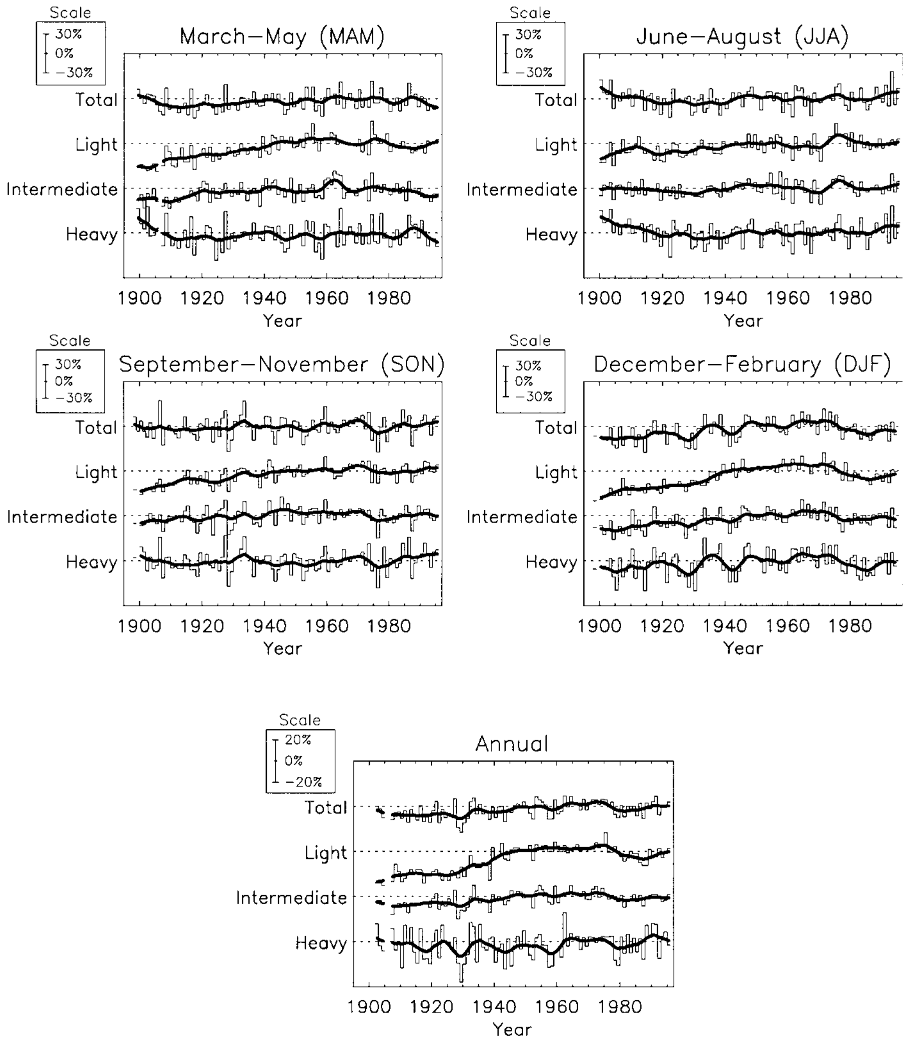


Fig. 9 As in Fig. 3 but for Southwestern Canada.

is satisfied. Thus we can conclude that the NAO does indeed have a statistically significant influence on the precipitation intensity classes in eastern Canada. However, these effects are concentrated in only a few seasons, and only these are displayed in Table 4. Figure 13 displays maps of the correlation between light and heavy daily station precipitation and the NAO phase during the JFM season. A description of the responses to the positive phase of the NAO follows (responses to the negative phase are opposite).

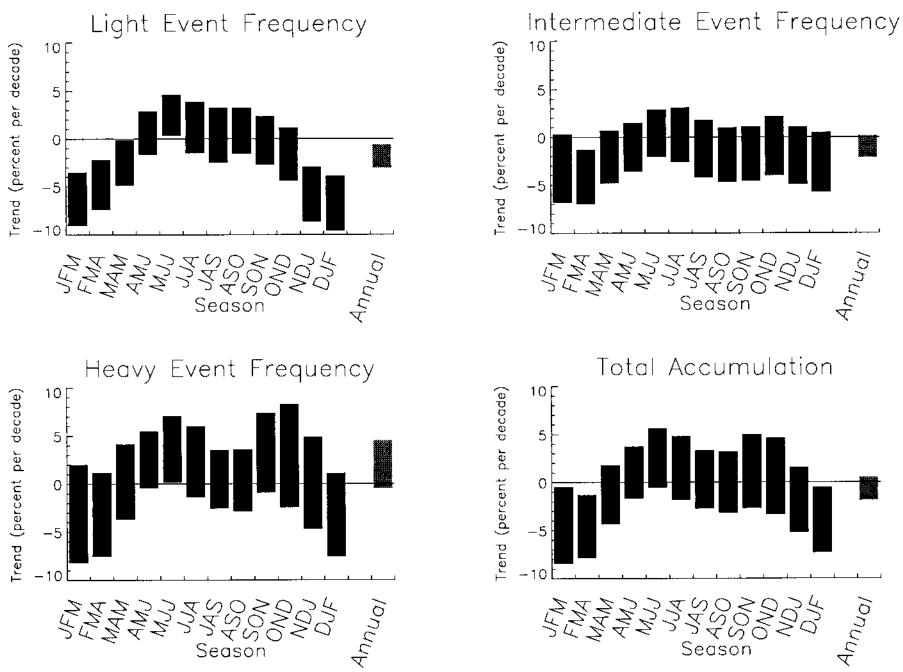


Fig. 10 As in Fig. 4 but for Southwestern Canada. Values for NDJ and DJF are for the 1950–1994 period.

Correlations significant at the 5% level affecting a large number of regions and intensity classes occur in four of the twelve three-month seasons: NDJ, JFM, MJJ and JJA. NDJ and JFM (Fig. 13) share many features; for instance, in both seasons an unusually low number of heavy events in Northern Quebec-Labrador (Region 2) and of both heavy and intermediate events in Baffin Island (Region 3) accompany the positive NAO phase. This spreads to heavy events in Hudson Bay (Region 5) and light events in Northern Quebec-Labrador in JFM (Fig. 13). However, these relationships do not exist in either MJJ or JJA. Instead, both of these seasons receive fewer events in the intermediate and light classes in Hudson Bay and Southern Quebec (Region 6). Furthermore, Keewatin (Region 8) receives fewer intermediate events in both seasons, extending to light events in MJJ and intermediate events in Great Slave Lake (Region 12) in JJA. Meanwhile, Queen Elizabeth Islands (Region 4) and Kitikmeot (Region 10) experience increases in some classes, but only during JJA.

**b** *The Pacific/North America Teleconnection Pattern*

The PNA is the second-most important mode of inter-annual variability in the extratropical Northern Hemisphere after the NAO, being the most influential during the

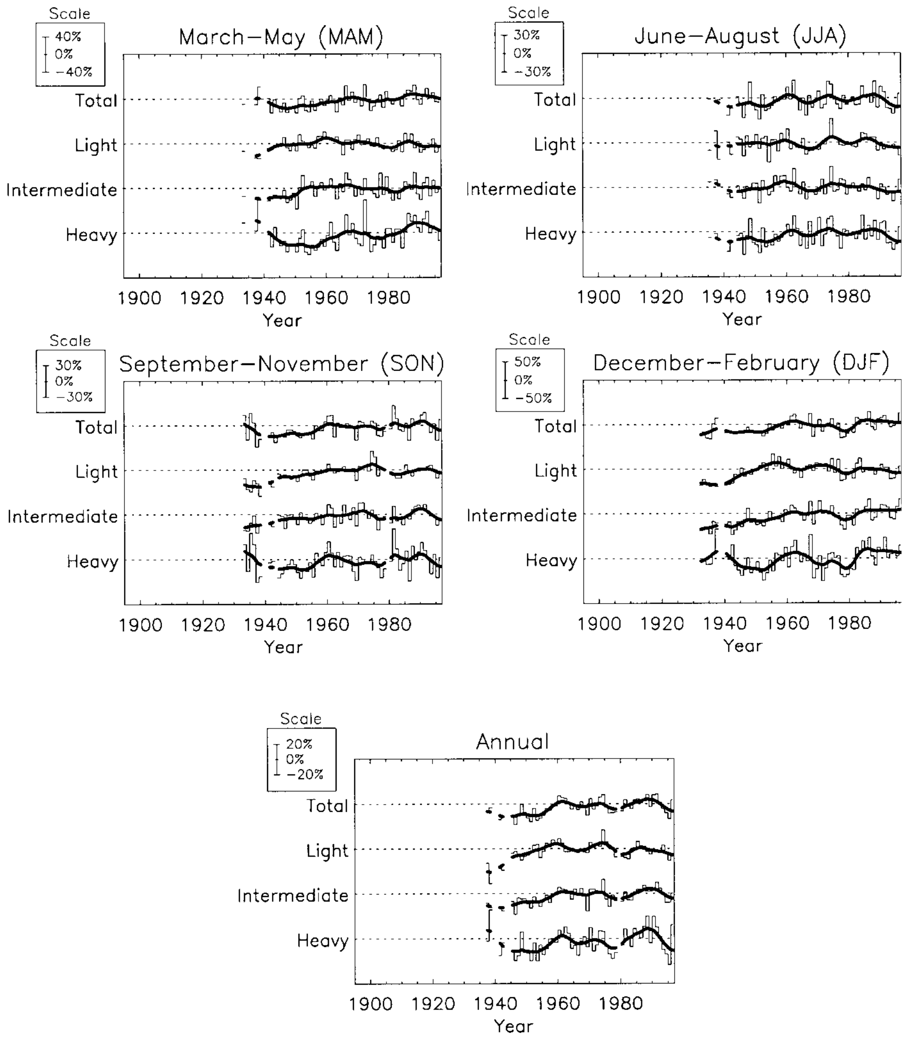


Fig. 11  As in Fig. 3 but for Northwestern Canada.

winter (Barnston and Livezey, 1987). Unlike the NAO, the PNA consists of a quadripole pattern of pressure anomalies. Two anomalies of like sign are located south of the Aleutian Islands and over the southeastern United States, while another two of the opposite sign are centred near Hawaii and over the Canadian Rocky Mountains. The sign of the latter two anomalies corresponds to the sign of the PNA.

Table 5 lists the regional correlation of the PNA with the three classes of event intensity, as well as the total accumulation, during the three-month seasons in which a large number of regions and intensity classes are affected. The correlations listed

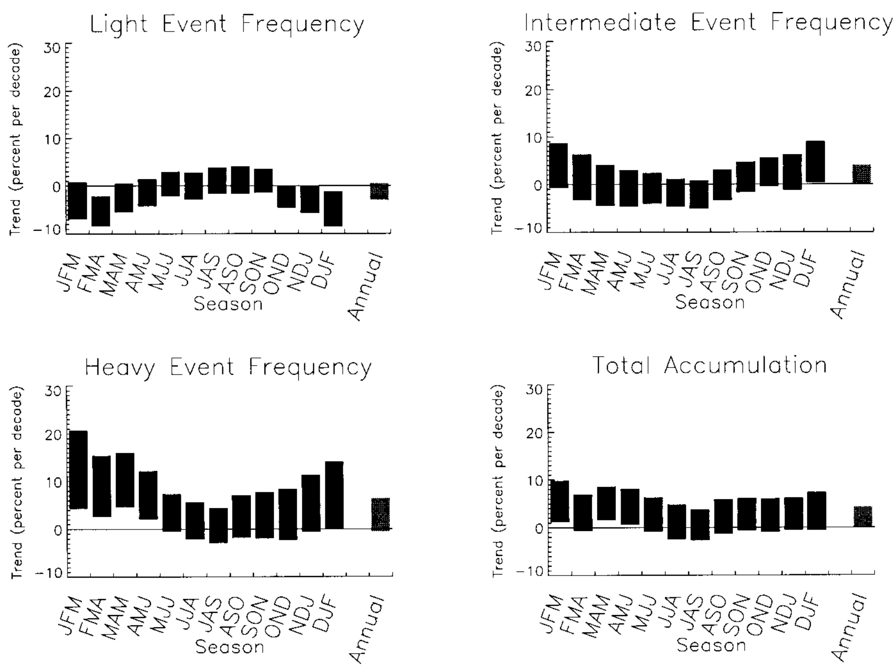


Fig. 12 As in Fig. 4 but for Northwestern Canada.

in Table 5, together with those for the seasons not listed, are collectively “field significant” at the 5% level when evaluated in a manner similar to that described above for the NAO. Figure 14 displays maps of the correlation between light and heavy daily station precipitation and the PNA phase during the JFM season. A brief description of the seasonal progression of these relationships follows for the positive PNA phase (responses to the negative phase are opposite).

The strongest relationship with the PNA occurs in Southern Prairies (Region 9), accounting for 35% of the variance in total accumulation during NDJ. In ASO to JFM (Fig. 14) a drop in the frequency of heavy and intermediate events accompanies the positive PNA phase, extending to light events in NDJ and DJF. In both Northern Prairies (Region 11) and Southern British Columbia (Region 13) an initial dearth of light events shifts later to intermediate events. Meanwhile, Northern British Columbia (Region 15) receives fewer light events from OND to JFM (Fig. 14). To the north, Yukon-McKenzie (Region 14) experiences a decrease in intermediate events in ASO and SON, shifting to light events for OND. Keewatin (Region 8) is unusual in receiving significantly more events, specifically light events from OND to JFM, extending to Great Slave Lake (Region 12) in JFM only (Fig. 14). In the east, Hudson Bay (Region 5) receives fewer intermediate events in ASO, shifting to heavy events in SON and OND. Similarly, fewer heavy events occur in Great

TABLE 4. Seasonal correlation coefficients between the NAO index and the frequency of daily precipitation events in the regional intensity classes. Only seasons with a large number of statistically significant correlations are listed. Values are for the 1950–1995 period, with linear trends removed. Bold values are significant at the 5% level (two-sided test). See the text for an evaluation of the table's statistical significance. District and region abbreviations and identification numbers are listed in Table 2. Seasons are designated by the initials of their months (e.g. JFM = January-March). Intensity class legend: H = Heavy events, I = Intermediate events, L = Light events, T = Total accumulation.

| | | Region | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | SEC | | NEC | | | | AC | | SWC | | | | NWC | | |
| Season | Class | 1 | 7 | 2 | 3 | 5 | 6 | 4 | 10 | 9 | 11 | 13 | 15 | 8 | 12 | 14 |
| MJJ | H | –13 | 0 | –9 | –22 | –14 | –6 | 12 | **32** | 25 | 15 | 3 | 5 | –7 | –12 | **32** |
| | I | –1 | –7 | 3 | 4 | **–47** | –27 | –2 | 23 | –3 | –9 | 8 | –7 | **–40** | –7 | –5 |
| | L | –12 | –13 | –21 | –3 | –3 | **–35** | 28 | –2 | –2 | –16 | 7 | 8 | **–30** | –22 | 7 |
| | T | –17 | 9 | 2 | –9 | –26 | –12 | 15 | 24 | 18 | 11 | 5 | –3 | –23 | –17 | 23 |
| JJA | H | –2 | –9 | –24 | –16 | 1 | –18 | 13 | 21 | –10 | 16 | 4 | 22 | –8 | –9 | 22 |
| | I | –17 | –22 | –9 | 17 | **–34** | **–39** | **39** | **32** | –27 | –21 | 6 | –1 | **–35** | **–36** | 0 |
| | L | 19 | –12 | –26 | –2 | **–31** | –18 | **42** | 6 | **–37** | –23 | 18 | 12 | –13 | –17 | –4 |
| | T | –6 | –4 | –20 | –6 | –15 | –26 | **38** | 22 | –11 | 8 | 7 | 13 | –13 | –21 | 10 |
| NDJ | H | –6 | –11 | **–30** | **–49** | 9 | –22 | –24 | –3 | –28 | –6 | –8 | 15 | 10 | 3 | 4 |
| | I | –1 | **37** | 22 | **–48** | –25 | 5 | –3 | –6 | –3 | 1 | –17 | 10 | –2 | 10 | 6 |
| | L | –7 | –10 | –16 | –1 | –6 | –28 | 7 | –4 | –3 | **–30** | –18 | –11 | –29 | 11 | 21 |
| | T | –8 | 8 | **–35** | **–43** | 3 | –19 | –14 | –4 | –23 | –10 | –15 | 12 | –5 | –1 | 6 |
| JFM | H | 5 | –4 | **–36** | **–51** | **–36** | 28 | –11 | 4 | 5 | –1 | 27 | 6 | –11 | 15 | –12 |
| | I | 17 | 25 | –19 | **–40** | –25 | 8 | 11 | **31** | 4 | –4 | 23 | –2 | 7 | **–39** | 9 |
| | L | –13 | 7 | **–29** | –21 | 12 | –25 | 20 | –3 | –8 | –15 | –17 | –3 | –8 | –10 | 28 |
| | T | 16 | 10 | **–35** | **–51** | **–38** | 20 | 14 | 24 | 4 | 1 | 26 | 5 | –11 | –5 | 2 |

Lakes-St. Lawrence (Region 7) from OND to JFM (Fig. 14), expanding to Southern Quebec (Region 6) during NDJ and DJF, and accounting for up to 20% of the variance. Meanwhile, Atlantic (Region 1) receives fewer intermediate events in DJF and JFM.

## 5 Discussion

With the rehabilitated dataset used in this analysis, Mekis and Hogg (1999) examined linear trends for eleven regions spanning the Canadian landmass. Their detection of a tendency towards increasing accumulations across the country, larger in northern regions during the winter, was consistent with previous analyses based on monthly data (e.g. Groisman and Easterling, 1994). Here it has been found that statistically significant trends have occurred in the frequency of daily precipitation events of different classes of intensity. For example, during the latter half of the century the increased precipitation in the summer in the Southeastern district is related to increasingly frequent heavy events, with the same occurring in the Arctic district

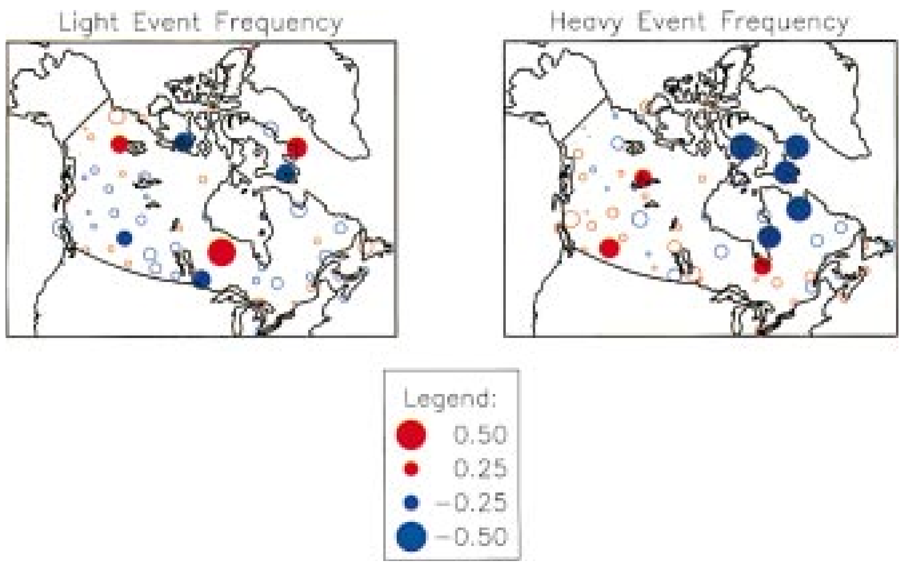


Fig. 13 Map of the correlation between the frequency of January–March light and heavy daily precipitation events at each station and the NAO over the 1950–1995 period. Red disks and circles denote positive correlation, while blue disks and circles denote negative correlation. Disks denote correlation statistically significant at the 5% level. See Table 3 for intensity class definitions.

in all seasons but the summer (Fig. 2). On the other hand, the decreasing frequency of light events in Southeastern Canada during the winter does not translate to changes in the total accumulation, which instead reflects the more constant frequency of heavy and intermediate events. Southwestern Canada is unusual in that it has experienced a significant decrease in winter precipitation since 1950. As noted by Akinremi et al. (1999), this is due mainly to a decrease in the frequency of light events. Notably, while significant trends over the 1950–1995 period in lighter intensity events tend to be negative, those in heavier intensity events are almost exclusively positive.

The largest variations tend to occur abruptly, often in widely separated regions, generally in the years around 1940, around 1960, and in the late 1970s. While these changes may reflect large-scale regime shifts in Canadian precipitation climates, they may also be the result of systematic changes in instrumentation or observational procedures not (fully) corrected for in the adjustments made to this dataset (Mekis and Hogg, 1999). The changes observed around 1940, other than those in light event frequency, in both Southeastern and Southwestern Canada probably reflect real climate changes. Firstly, these changes actually occur at different times, the late 1930s in the Southeast and the early 1940s in the Southwest. Secondly, they occur generally in both rain and snow (measured separately), and in varying seasons.

Mekis and Hogg (1999) note that abrupt increases occurring between 1955 and

TABLE 5. As in Table 4 but for the PNA index. See the text for an evaluation of the table's statistical significance.

| | | Region | | | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | SEC | | NEC | | | | AC | | SWC | | | | NWC | | |
| Season | Class | 1 | 7 | 2 | 3 | 5 | 6 | 4 | 10 | 9 | 11 | 13 | 15 | 8 | 12 | 14 |
| ASO | H | 27 | –11 | 24 | –5 | –8 | 6 | –7 | 29 | **–31** | –3 | 27 | –4 | –11 | –3 | –4 |
| | I | –1 | –14 | –12 | 10 | **–33** | –15 | –2 | –6 | **–36** | –17 | 7 | –4 | –13 | –22 | **–35** |
| | L | –12 | –19 | 26 | 5 | 0 | –16 | –11 | –17 | –20 | **–29** | 8 | –22 | –22 | **–29** | –6 |
| | T | **30** | –1 | 17 | 2 | –20 | 2 | 4 | 15 | –22 | 2 | 29 | –8 | –16 | –6 | –14 |
| SON | H | **37** | –22 | 13 | –5 | **–32** | 4 | –10 | –8 | **–44** | –13 | –17 | 5 | –10 | 1 | –4 |
| | I | –2 | 3 | –7 | –4 | –12 | –20 | 4 | –9 | –28 | **–48** | –26 | 2 | 4 | –21 | **–36** |
| | L | 7 | 18 | 8 | 19 | 7 | 4 | 9 | –20 | –7 | –12 | –16 | –14 | –14 | –26 | –9 |
| | T | **36** | –7 | 12 | –4 | **–41** | –5 | 1 | –11 | **–35** | –19 | –22 | –6 | –11 | 0 | –15 |
| OND | H | 13 | **–38** | –7 | –15 | **–38** | –23 | –12 | 6 | **–42** | –11 | –14 | 19 | –22 | –6 | –1 |
| | I | 0 | 0 | –5 | 9 | 12 | 5 | –3 | –4 | **–40** | **–48** | –28 | –4 | 10 | –29 | –25 |
| | L | 0 | 24 | 6 | 8 | 17 | –17 | 9 | 5 | –24 | –21 | **–35** | **–33** | **39** | –26 | **–32** |
| | T | 20 | –22 | –9 | –18 | –25 | –18 | –8 | 5 | **–44** | –25 | –25 | 12 | –15 | –11 | –23 |
| NDJ | H | 9 | **–45** | 9 | 15 | –10 | **–42** | 12 | –16 | **–52** | –9 | –26 | 12 | 9 | 22 | 25 |
| | I | –27 | –11 | –20 | **37** | 12 | –3 | –9 | 0 | **–42** | **–32** | **–51** | –7 | 1 | –3 | –7 |
| | L | –15 | 10 | –12 | 12 | –5 | –3 | –1 | –6 | **–36** | –27 | **–35** | **–37** | **31** | –19 | –6 |
| | T | –2 | **–39** | 4 | 29 | –2 | **–42** | 8 | –10 | **–59** | –21 | **–41** | 4 | 21 | 12 | 12 |
| DJF | H | 7 | **–30** | 15 | 22 | –11 | **–30** | 10 | –21 | **–43** | –9 | –17 | 2 | 18 | 29 | 13 |
| | I | **–42** | –23 | –8 | 32 | –1 | –6 | –10 | –6 | **–43** | –4 | **–52** | –23 | 18 | 26 | 3 |
| | L | –29 | –4 | 9 | 2 | 11 | –8 | –4 | –8 | **–34** | –21 | –29 | –22 | **31** | 11 | –1 |
| | T | –10 | –27 | 12 | 23 | –6 | **–32** | 2 | –16 | **–46** | –13 | **–33** | –16 | **36** | 29 | 8 |
| JFM | H | 6 | **–31** | 10 | –2 | –21 | –12 | 1 | –13 | **–41** | –12 | –20 | –26 | 37 | 25 | 15 |
| | I | **–42** | **–41** | –7 | –1 | –13 | –5 | 5 | –10 | **–49** | –3 | **–32** | –23 | 14 | **38** | –3 |
| | L | –23 | –11 | –1 | –4 | –7 | 5 | 10 | –24 | –19 | –14 | –10 | **–37** | **46** | **36** | 8 |
| | T | –4 | **–30** | 4 | –3 | –20 | –16 | 8 | –25 | **–42** | –14 | –26 | **–37** | **40** | **39** | 4 |

1965 in most northern areas result entirely from increased snowfall. Realizing that the use of Nipher shielded snow gauges commenced around this time, they remark that the new gauge measurements could have indirectly influenced the ruler measurements used in this dataset, although how this effect could arise is unclear. However, in Arctic Canada, and similarly in the other two northern districts, the increase arises entirely from changes in the frequency of intermediate and heavy events. Light events show no discernible change; however, this statistical analysis has indicated that lighter events are more susceptible to the effects of biases than are heavier events (e.g. Fig. 3). This suggests that indeed real increases in precipitation occured in northern regions around 1960.

Finally, the change in the late 1970s to the metric system should only measurably affect trace measurements, where indeed abrupt changes occur (not shown). The

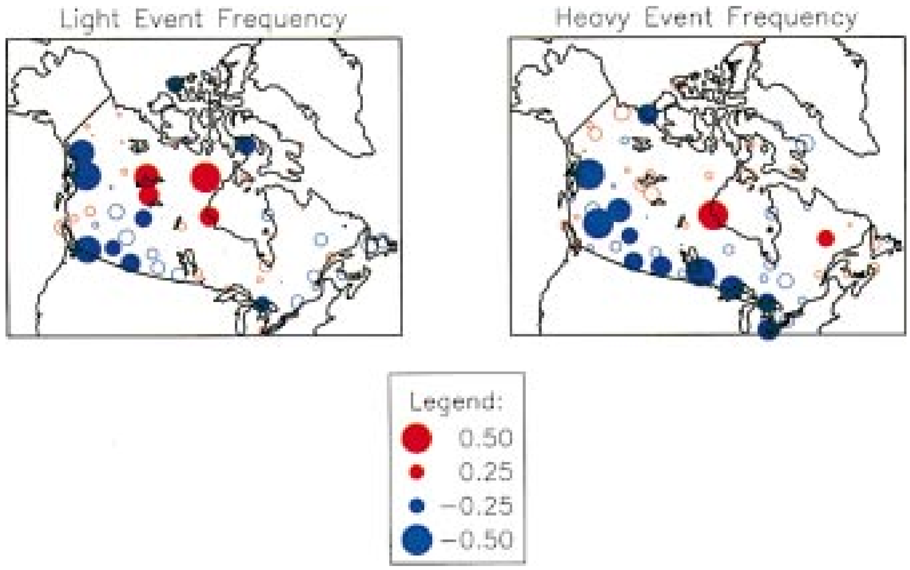


Fig. 14 As in Fig. 13 but for the PNA.

introduction of the Type-B rain gauge, also occurring at this time, was included in the corrections of Mekis and Hogg (1999). In fact, the only large change occurring at this time is a decrease in Southwestern Canada, suggesting that values over this time period are indeed related to coincident changes in the PNA pattern.

Of the two modes of atmospheric variability examined here, the PNA clearly dominates in its influence on precipitation variations. The strongest responses occur over Southwestern Canada (Table 5 and Fig. 14), reflected in the dryness of the 1976–1988 period when the PNA was unusually positive. Of particular note, in the Great Lakes-St. Lawrence and Southern Quebec regions only extreme heavy events are affected by the PNA. While variations in the PNA account for up to 20% of the variance of the frequency of heavy events in these two regions from late autumn to late winter, other levels of event intensity remain unaffected. In fact, this response in heavy events is more statistically robust than the response in total accumulation.

The relationship between the PNA and precipitation over the southwest is likely a result of variations in the intensity and location of the North Pacific jet stream. During the positive PNA phase the jet stream on average diverges southward before reaching the North American coast, which should result in less intense moisture advection into southern British Columbia and the Prairies. During the negative phase, on the other hand, the mean jet stream location follows a more zonal and concentrated path towards the southwest, which should result in enhanced moisture advection over this region. Responses over Ontario and southern Quebec likely result from increased advection of warm, moist air from the southeastern United

States during the negative phase, causing more frequent storms and thus more intense precipitation.

The influence of the NAO on Canadian precipitation is generally restricted to Northeastern Canada, where varying classes of intensity are affected depending upon the season. While the NAO influences the frequency of lighter events during the summer around Hudson Bay, it influences the frequency of heavier events during the winter over Baffin Island and Labrador (Table 4 and Fig. 13). During the winter a positive NAO phase allows more frequent outbreaks of cold, dry Arctic air into the Baffin Island region, resulting in less precipitable moisture over the region. During the summer the positive phase likely causes enhanced advection of dry air from the northwest into the Hudson Bay region, resulting in less precipitation there. During the negative NAO phase, less dry air than normal would be diverted into these regions, resulting in wetter conditions. It is not clear why similar atmospheric circulation patterns do not affect precipitation in this area during the transition seasons.

In summary, this analysis has uncovered statistically significant variations in daily precipitation intensity over Canada with important seasonal and regional characteristics. Trends in the frequency of three classes of precipitation intensity were examined for the latter half of the 20$^{th}$ century over five districts spanning the country. Heavy and intermediate daily precipitation events have become statistically significantly more frequent in over one third of all seasons in the five districts, indicating a shift to more intense precipitation, with increases concentrated in northern areas of the country. Moreover, the variability in precipitation intensity is at least partially related to dominant modes of atmospheric variability. For instance, autumn and winter precipitation intensity in Ontario and southern Quebec is significantly modulated by variations in the PNA, with important implications for seasonal forecasting. In this region the negative PNA phase generally leads to more extreme precipitation.

While the purpose of this study was not to address formally the attribution of the observed precipitation trends to anthropogenic effects, we nevertheless note that our results are consistent with climate model projections of climate change over the 20$^{th}$ century. Specifically, the strong trend towards increasing precipitation in northern areas of Canada, noted previously in other studies, and the evidence of increasingly frequent extreme precipitation events are quite striking in the observational record and both are projected by climate model simulations.

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## References

AKINREMI, O.O.; S.M. MCGINN and H.W. CUTFORTH. 1999. Precipitation trends in the Canadian Prairies. *J. Clim.* **12**: 2996–3003.

APPENZELLER, C.; J. SCHWANDER, S. SOMMER and T.F. STOCKER. 1998. The North Atlantic Oscillation and its imprint on precipitation and ice accumulation in Greenland. *Geophys. Res. Lett.* **25**: 1939–1942.

BARNSTON, A.G. and R.E. LIVEZEY. 1987. Classification, seasonality and persistence of low-frequency atmospheric circulation patterns. *Mon. Weather Rev.* **115**: 1083–1126.

BOER, G.J.; G. FLATO, M.C. READER and D. RAMSDEN. 2000a. A transient climate change simulation with greenhouse gas and aerosol forcing: experimental design and comparison with the instrumental record for the 20th century. *Clim. Dyn.* in press.

———; ——— and D. RAMSDEN. 2000b. A transient climate change simulation with greenhouse gas and aerosol forcing: projected climate for the 21st century. *Clim. Dyn.* in press.

BRADLEY, R.S.; H.F. DIAZ, J.K. EISCHEID, P.D. JONES, P.M. KELLY and C.M. GOODESS. 1987. Precipitation fluctuations over Northern Hemisphere land areas since the mid-19th century. *Science*, **237**: 171–175.

CAYAN, D.R.; M.D. DETTINGER, H.F. DIAZ and N.E. GRAHAM. 1998. Decadal variability of precipitation over western North America. *J. Clim.* **11**: 3148–3166.

CORTI, S.; F. MOLTENI and T.N. PALMER. 1999. Signature of recent climate change in frequencies of natural atmospheric circulation regimes. *Nature*, **398**: 799–802.

CUBASCH, U.; J. WASZKEWITZ, G. HEGERL and J. PERLWITZ. 1995. Regional climatge changes as simulated in time-slice experiments. *Clim. Change*, **31**: 273–304.

DAI, A.; I.Y. FUNG and A.D. DEL GENIO. 1997. Surface observed global land precipitation variations during 1900–88. *J. Clim.* **10**: 2943–2962.

DIAZ, H.F.; R.S. BRADLEY and J.K. EISCHEID. 1989. Precipitation fluctuations over global land areas since the late 1800s. *J. Geophys. Res.* **94**: 1195–1210.

EISCHEID, J.K.; R.S. BRADLEY and P.D. JONES. 1991. A comprehensive precipitation data set for global land areas. DOE Tech. Rep. TR051, DOE ER-69017T-H1, 81 pp. (Available from National Technical Information Service, U.S. Dept. of Commerce, 5285 Port Royal Road, Springfield, VA 22161.)

FYFE, J.C.; G.J. BOER and G.M. FLATO. 1999. The Arctic and Antarctic Oscillations and their projected changes under global warming. *Geophys. Res. Lett.* **26**: 1601–1604.

GERSHUNOV, A. and T.P. BARNETT. 1998. ENSO influence on intraseasonal extreme rainfall and temperature frequencies in the contiguous United States: Observations and model results. *J. Clim.* **11**: 1575–1586.

GORDON, H.B.; P.H. WHETTON, A.B. PITTOCK, A.M. FOWLER and M.R. HAYLOCK. 1992. Simulated changes in daily rainfall intensity due to the enhanced greenhouse effect: Implications for extreme rainfall events. *Clim. Dyn.* **8**: 83–102.

GROISMAN, P. YA. and D.R. EASTERLING. 1994. Variability and trends of precipitation and snowfall over the United States and Canada. *J. Clim.* **7**: 184–205.

———; T.R. KARL, D.R. EASTERLING, R.W. KNIGHT, P.F. JAMASON, K.J. HENNESSY, R. SUPPIAH, C.M. PAGE, J. WIBIG, K. FORTUNIAK, V.N. RAZUVAEV, A. DOUGLAS, E. FØRLAND and P.-M. ZHAI. 1999. Changes in the probability of heavy precipitation: important indicators of climatic change. *Clim. Change*, **42**: 243–283.

HENNESSY, K.J.; J.M. GREGORY and J.F.B. MITCHELL. 1997. Changes in daily precipitation under enhanced greenhouse conditions. *Clim. Dyn.* **13(9)**: 667–680.

———; R. SUPPIAH and C.M. PAGE. 1999. Australian rainfall changes, 1910–1995. *Aust. Met. Mag.* **48**: 1–13.

HOGG, W.D.; X. ZHANG and E. MEKIS. 1998. Spatial and temporal characteristics of heavy precipitation events in Canada. *In*: Proc. Ninth Am. Meteorol. Soc. Global Change Symposium, 11–16 January 1998, Phoenix, Arizona, USA. pp. 147–151.

HULME, M.; T.J. OSBORN and T.C. JOHNS. 1998. Precipitation sensitivity to global warming: Comparison of observations with HadCM2 simulations. *Geophys. Res. Lett.* **25**: 3379–3382.

HURRELL, J.W. 1995. Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science*, **269**: 676–679.

———. 1996. Influence of variations in extropical wintertime teleconnections on Northern Hemisphere temperature. *Geophys. Res. Lett.* **23**: 665–668.

IWASHIMA, T. and R. YAMAMOTO. 1993. A statistical analysis of the extreme events: Long-term trend of heavy daily precipitation. *J. Met. Soc. Jpn.* **71**: 637–640.

KARL, T.R.; R.W. KNIGHT and N. PLUMMER. 1995. Trends in high-frequency climate variability in the twentieth century. *Nature*, **377**: 217–220.

——— and ———. 1998. Secular trends of precipitation amount, frequency, and intensity in the U.S. *Bull. Am. Meteorol. Soc.* **79**: 231–241.

KATTENBERG, A.; F. GIORGI, H. GRASSL, G.A. MEEHL, J.F.B. MITCHELL, R.J. STOUFFER, T. TOKIOKA, A.J. WEAVER and T.M.L. WIGLEY. 1996. Climate models – Projections of future climate. In: *Climate Change 1995: The Science of Climate Change*, J.T. Houghton, L.G. Meira-Filho, B.A. Callander, N. Harris, A. Kattenberg, and K. Maskell (Eds), Cambridge University Press, Cambridge, U.K., pp. 285–357.

KATZ, A.W. and B.G. BROWN. 1992. Extreme events in a changing climate: Variability is more important that averages. *Clim. Change*, **21**: 289–302.

KHARIN, V.V. and F.W. ZWIERS. 2000. Changes in the extremes in an ensemble of transient climate simulations with a coupled atmosphere-ocean GCM. *J. Clim.* accepted.

LIVEZEY, R.E. and W.Y. CHEN. 1983. Statistical field significance and its determination by Monte Carlo techniques. *Mon. Weather Rev.* **111**: 46–59.

MEKIS, E. and W.D. HOGG. 1999. Rehabilitation and analysis of Canadian daily precipitation time series. ATMOSPHERE-OCEAN, **37(1)**: 53–85.

METCALFE, J.R.; S. ISHIDA and B.E. GOODISON. 1994. A corrected precipitation archive for the Northwest Territories. Environment Canada-Mackenzie Basin Impact Study, Interim Report #2, pp. 110–117.

———; B. ROUTLEDGE and K. DEVINE. 1997. Rainfall measurement in Canada: Changing observational methods and archive adjustment procedures. *J. Clim.* **10**: 92–101.

NICHOLLS, N. and A. KARIKO. 1993. East Australian rainfall events: Interannual variations, trends, and relationships with the Southern Oscillation. *J. Clim.* **6**: 1141–1152.

———; G.V. CRUZA, J. JOUZEL, T.R. KARL, L.A. OGALLO and D.E. PARKER. 1996. Observed climate variability and change. In: *Climate Change 1995: The Science of Climate Change*, J.T. Houghton, L.G. Meirra-Filho, B.A. Callander, N. Harris, A. Kattenberg, and K. Maskell (Eds), Cambridge University Press, Cambridge, U.K., pp. 133–192.

NODA, A. and T. TOKIOKA. 1989. The effect of doubling the $CO_2$ concentration on convective and non-convective precipitation in a general circulation model coupled with a simple mixed layer ocean model. *J. Met. Soc. Jpn.* **67**: 1057–1067.

ROBERTSON, A.W. and M. GHIL. 1999. Large-scale weather regimes and local climate over the western United States. *J. Clim.* **12**:1796–1813.

SHABBAR, A.; K. HIGUCHI, W. SKINNER and J.L. KNOX. 1997. The association between the BWA index and winter surface temperature variability over eastern Canada and west Greenland. *Int. J. Climatol.* **17**: 1195–1210.

THOMPSON, D.W.J. and J.M. WALLACE. 1998. The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.* **25**: 1297–1300.

VINNIKOV, K. YA.; P. YA. GROISMAN and K.M. LUGINA. 1990. Empirical data on contemporary global climate changes (temperature and precipitation). *J. Clim.* **3**: 662–677.

WALLACE, J.M. and D.S. GUTZLER. 1981. Teleconnections in the geopotential height field during the Northern Hemisphere winter. *Mon. Weather Rev.* **109**: 784–812.

ZWIERS, F.W. and V.V. KHARIN. 1998. Changes in the extremes of the climate simulated by CCC GCM2 under $CO_2$ doubling. *J. Clim.* **11**: 2200–2222.