2010 21st International Lightning Detection Conference
19 - 20 April • Orlando, Florida, USA
3rd International Lightning Meteorology Conference
21 - 22 April • Orlando, Florida, USA

CLOUD TO GROUND LIGHTNING IN CANADA 1999:2008

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1. INTRODUCTION

Cloud-to-ground (CG) lightning is a major cause of severe weather-related fatalities and injuries (Mills et al. 2009; Curran et al. 2000), property and infrastructure damage (Holle et al. 1996; Mills et al., 2008), forest fires (Stocks et al., 2002), and interruptions in or damage to electric power transmission and distribution systems (Cummins et al. 1998a). Several short-duration, limited-area studies have previously been conducted in Canada using provincial, territorial or private-sector lightning detection networks. Shortly after the Canadian Lightning Detection Network (CLDN) was established in 1998, the first national picture of lightning activity for the period 1989-1999 appeared in a Canadian Geographic article (Lanken, 2000). Lightning hotspots were shown to occur over the Prairie Provinces and southern Ontario. The first analysis of lightning characteristics across Canada was published by Burrows et al. (2002). Although only three years' data were available for that study, a complex pattern emerged, showing strong regional, diurnal and seasonal features. This paper presents an updated climatology of cloud-ground lightning over Canada for 1999-2008 at a resolution of 20 km as detected by the CLDN. We highlight some of the main findings for flash density, occurrence, polarity, multiplicity, and first-stroke peak current. Due to the length of the material we cannot show the entire analysis. This is the subject of two papers in the peer review process at the time of this writing. For the sake of brevity we do not have a separate section summarizing conclusions here.

2. DATA and ANALYSIS

The CLDN provides continuous lightning detection across most of Canada and offshore to about 300 km. The network is fully integrated with the United States National Lightning Detection Network (Orville et al., 2002).Both networks are operated by Vaisala Inc. The data we analyzed were the quality-controlled flashes sent after each month by Vaisala, and not the real time flash stream used by operational users and clients of the CLDN.

Figure 1 shows the locations of the 83 CLDN sensors as of November 2008. The CLDN is not a static entity. Technology improvements and available funds allowed sensors to be upgraded as time passed since 1998. Discrimination between CG and cloud lightning discharges was added in February 1999. Two new sensors were added in the northern Yukon in 2003. Beginning in 2005 some of the older LPATS-IV sensors were upgraded incrementally each year to LS7000 sensors as available funds permitted, mainly along the eastern and northern boundaries of the CLDN. Thirty-one detectors in Figure 1 are still the older LPATS-IV sensors, where lightning flashes are located by the time of arrival of radio pulses generated when a flash occurs. A flash must be detected by 3 of these sensors to determine its location. Fifty-two sensors are IMPACT/ES and LS7000 varieties in which magnetic direction finding technology and time-of-arrival technology are combined, thus a flash location can be determined if it is detected by 2 sensors. Details on how these sensors detect lightning are available in Cummins and Murphy (2009).

We analyze cloud-ground (CG) lightning flashes here. A CG lightning flash may have as many as 20 strokes (Cummins and Murphy, 2009), although the CLDN data is still truncated at 15 strokes. In any case the fraction of flashes with a multiplicity of 15 is exceedingly small (Kochtubajda and Burrows 2010). In our data the location of a flash is assigned as that of the first stroke, which is nearly always the strongest stroke. Median location accuracy is about 500 m or better according to Cummins and Murphy (2009). Vaisala has determined that detection efficiency of CG flashes is 80-90% or higher inside the CLDN, decreasing to about 70% just beyond the periphery and to about 30% at 300 km beyond the network. We have not corrected the data for detection efficiency. We show the approximate 70% detection efficiency contour in several figures in this paper where the scope is Canada-wide. Cloud flashes are also detected by the CLDN. Separate strokes are not reported by the CLDN for these. Burrows et al. (2002) reported detection efficiency of cloud flashes is about 1-4%, based on a current threshold of 5 kA to report a flash. However, beginning in 2006 the threshold was decreased to 3 kA, which raised the cloud flash detection efficiency. For this reason we do not show results for the combined total lightning as was done in Burrows et al. (2002). Thus while the large scale patterns of cloud-ground lightning occurrence identified here are very similar to those identified in Burrows et al. (2002), there are differences in the some of the details.

Certain details of the lightning event classification schemes changed during the ten year period. Raw data reported by the CLDN for all the 10 years were first scanned and positive flashes were reclassified according to guidelines discussed in Cummins and Murphy (2009) as a result of field campaigns described in Biagi et al. (2007). They state that positive events with current strength below 15 kA that were classified as CG previous to April 2006 would henceforth be classified as cloud discharges, and positive flashes with current strength greater than 20 kA that had been classified as cloud discharges were likely to be CG discharges. We corrected all our data since 1999 to these specifications. Cummins and Murphy (2009) mention that a new event classification scheme incorporating these changes is in the final stages of validation, but we had no information on it at the time of writing, and we do not expect it to differ substantially from what we know now.

Our analysis was divided into west and east sections due to the vast detection area. The west section includes the north area which is specially analyzed in some discussions to follow. Lightning flash locations were sorted into 20x20 km grid squares. In a latititude-longitude sense these are irregular polygons whose longitude span increases northward. Polygon vertices were determined with the 1980 Geodetic Reference System (GRS80) ellipsoid earth model (the MathWorks, 2008). This grid resolution approximates the upper bound of the 8-20 km region where thunder can be heard by an observer, as mentioned in (Huffines and Orville 1999).

3. RESULTS

Figure 2a shows the number of cloud-ground flashes detected over Canada in each of the years 1999-2008. We assumed all the lightning detected for January 1999 was CG. The impact on results is negligible since lightning activity in January is small over Canada. These numbers ranged from a low of 1.9835 million flashes in 2008 to a high of 2.9631 million flashes in 2005. The average for the ten years was 2.3549 million flashes per year. Figure 2b shows the ten-year average distribution of flashes by month. Lightning occurred in all months, but the vast majority of flashes occurred in the warm season months from April to October, with July being the month of greatest lightning occurrence.

3.1 FLASH DENSITY

Figures 3a and 3b show the ten-year average CG flash density (flash km^{-2} yr^{-1}) for western and eastern Canada, respectively. Average flash densities greater than 3 flash km^{-2} yr^{-1} occurred only in the United States and were truncated at 3 in order to highlight the patterns over Canada. The irregular light blue dashed lines around the perimeter of Canada show the location of 70% lightning detection efficiency. Outside of these lines only relatively stronger flashes were detected. The northernmost flash detected over Canada was over the eastern Beaufort Sea at 70.8811° N, 130.5910° W.

Lightning activity is highly influenced by length of season, proximity to cold water bodies, and elevation. The shape and locations of large scale patterns of lightning occurrence remained nearly the same as first reported in Burrows et al. (2002) where only three years of data were used, although some details were different. Flash density maxima were found in the same locations: the Swan Hills and foothills of Alberta, southeast Saskatchewan-southwest Manitoba, and southwest Ontario. A region of greater lightning occurrence but relatively low flash density south of Nova Scotia was found in the same location as previously reported. New areas of higher flash density were found along the international border in northwestern Ontario and southern Quebec, and appear to be northward extensions from the USA of higher flash density seen in the previous study. The majority of the country saw average flash densities of between .5 and 1.5 flash km^{-2} yr^{-1}.The greatest average CG flash density of 2.789 flash km^{-2} yr^{-1} occurred in southwest Ontario, as did the greatest single-year flash density (10.335 flash km^{-2} yr^{-1}). Prominent flash density minima were found east of the Rocky Mountain great divide in Alberta and the Niagara Escarpment in southern Ontario.

3.2 FLASH OCCURRENCE

The beginning and ending dates of the *lightning season* were estimated from the distribution of observed flashes in each 20x20 km grid square as the dates of the 0.5 percentile and the 99.5 percentile, respectively. Figures 4a, 4b, 5a and 5b show the result. The average length of the lightning season can be approximately estimated at each location from the span of days between the beginning and ending dates in Figures 4 and 5. As one would expect there is a gradual decrease in the length of the lightning season from southern Canada to northern Canada. The lightning season extends from about mid March to early November over much of southern Ontario. In the north the lightning season extends from mid to late May until late August to mid September. However there are significant differences from east to west as well. Lightning occurred virtually year-round in the Pacific coastal region and over southern Nova Scotia and offshore. Winter lightning is common in these regions as Arctic airmasses pass over much warmer water. Lewis (2000) noted this for the Maritime Provinces. Over a large portion of interior British Columbia the lightning season is much longer than surrounding mountainous regions, running from late March to late October. For most of the prairie province region the season extends from early-mid May to early-mid September, however southeast Manitoba has a longer lightning season from late March to late October. The interior of the country from southeast Manitoba across Ontario to southwestern Quebec sees a significantly longer lightning season than regions to the west and north, due to the influence of warm unstable airmasses originating in the Gulf of Mexico. Much of southern and western Quebec has a longer lightning season than eastern Quebec. This was also noted by Morissette and Gauthier (2008). The lightning season is much longer over the southern coast region of Newfoundland than elsewhere in Newfoundland. The length of the lightning season over southern Nova Scotia and southern New Brunswick is comparable to southern Ontario's. The streaky nature of early beginning dates over the Great Lakes is likely associated with pockets of intense lightning associated with passing winter low pressure systems rather than lake-effect snowsqualls since the banding is in the wrong orientation and the streaks are very long (Sills, private communication).

Figures 6a and 6b show the average number of days per year of CG lightning occurrence for western and eastern Canada, respectively. Values greater than the maximum value in the colorbars were set to that value so higher values over the United States would not overwhelm Canadian values in the plots. It should be noted that this field is dependent on the area of the grid boxes used (20x20 km^2) since lightning can occur in one section of a box but not on another on any given day. Lightning activity is seen to be highly influenced by length of season, proximity to cold water bodies, and elevation. The number of lightning days decreases from south to north in response to the decreasing length of the season where convection is possible. Lightning activity also decreases from the interior of the country with its warmer continental climate to the east and west coasts with their cooler maritime climate. CG lightning occurred on only 1-10 days per year approximately over coastal British Columbia, the far north, and the coastal Atlantic region, and on approximately 15-30 days per year over much of the interior of the country. A pronounced maximum exceeding 30 days per year is seen over the foothills and Swan Hills of Alberta to the east of a long prominent minimum of about 10 days per year or less that occurs at high elevation along the Rocky Mountain continental divide. An area of relatively greater lightning occurrence extends north from the United States plains into southern Saskatchewan and Manitoba, and northwest Ontario. Southwest Ontario saw on average the most days of lightning in Canada. The greatest ten-year average number of days with lightning in western Canada per year (32.9) occurred near

Rocky Mountain House, and in eastern Canada (35.9) in extreme southwest Ontario near Harrow. In the north, the southern Northwest Territories experienced the greatest number of days with lightning (10-15) anywhere in Canada north of 60°N. Lightning was diminished over, and for large distances to the lee of, ocean water bodies (James Bay, Hudson Bay, Pacific Ocean and Atlantic Ocean) and large inland lakes (Great Slave, Athabasca, Reindeer, Winnipeg, Superior, Huron-Georgian Bay, and Ontario), where the cold surface water dampens convection. During the ten years the greatest annual number of days with lightning in western Canada occurred in the foothills near Brazeau., Alberta (47), and in all of Canada, inland of the north shore of Lake Erie near Highgate, Ontario (50). The latter location lies in a well known lake-breeze convergence zone (King et al., 1996; Sills 1998).

The traditional view of thunderstorm activity holds that daytime solar heating causes thunderstorms to build during the day and dissipate in the late evening, resulting in a low fraction of nocturnal lightning. To see how well this view holds for Canada, the time of occurrence of each CG flash was calculated in local solar time to find the fraction of lightning that occurred between the late evening early morning hours, which we will refer to as *nocturnal* lightning. Figures 7a and 7b show the fraction of CG lightning that occurred between 22:30 and 10:30 local solar time where at least 10 flashes were detected over the ten years. In Figs. 7a and 7b we see a predominance of low fractions of nocturnal lightning (10% to 20% or less) in the Yukon Territory, interior of British Columbia, the western halves of Alberta and Northwest Territories, much of the eastern portion of Quebec, Labrador, and interior Newfoundland. Over the oceans and most of the very large lakes and oceans and to their lee (e.g. Lake Superior) we see close to 50% nocturnal lightning, as expected since their water surface is not affected by solar heating on a diurnal time scale. However over certain areas of the Atlantic Ocean, Gulf of St. Lawrence, Pacific Ocean west of Vancouver Island, Hudson Bay, and James Bay there is more than 60% nocturnal lightning occurrence. The reason for this is not evident. Over the interior portion of the country from eastern Alberta and eastern Northwest Territories to western Quebec there is a predominance of higher fractions of nocturnal lightning (30% to 50% or more). Thunderstorms in this region owe their existence to additional factors besides daytime heating. In fact over portions of east-central Alberta and the southern halves of Saskatchewan and Manitoba more than 50% of lightning is nocturnal, with an extreme of 65.7% near Quill Lake, Saskatchewan. In the extreme southern parts of Saskatchewan and Manitoba this can be attributed to long-lived thunderstorm complexes that form over the United States Great Plains during the day and drift northward into Canada. However in the central parts the reason is not obvious. Most of the local maxima can be found over local elevated terrain features, and these may cause enough low-level convergence to maintain thunderstorm complexes for a long time or to re-develop thunderstorms during the night as air destabilized aloft during the day moves east. There may also be an influence here by the low level jet, which can initiate and even strengthen convection over the plains at night (Bluestein 1993).

Figures 8 and 9 show the hourly CG flash occurrence in a 50 km radius around the airport locations for sixteen Canadian cities. While comparison with thunderstorm statistics by human observers at each airport could be made, that is not our intention. In each panel the solid line shows the 1999-2008 mean, and the dashed line shows the maximum of the ten values at each hour. Vertical scales vary so patterns can be compared. The diurnal heating/cooling cycle is seen to exert the greatest control over lightning activity at all locations shown in Figs. 8 and 91 with the possible exception of St. John’s on the eastern extremity. The period of least lightning activity at nearly all locations was in the morning hours between approximately 6 LST and 9 LST, a time when the effect of overnight cooling is greatest. All locations had increased lightning activity between 13 LST and 17 LST, when the effect of solar heating is greatest. Lightning production at Whitehorse and Kelowna in the Rocky Mountain interior is primarily controlled by the diurnal cycle. This is true also at Vancouver on the west coast, although there is evidence of frontal influences there as well. The time of enhanced lightning activity is more complicated by translation and passages of transient dynamic features east of the Rocky Mountains. Lightning at Yellowknife in the north, and at Edmonton and Saskatoon in the western prairies, continued into the evening and early morning as thunderstorms that formed in the afternoon over the foothills drift south and east. The period of enhanced lightning activity on the eastern prairies and northwest Ontario continued past dawn, as seen at Estevan, Winnipeg, and Thunder Bay. Far from the mountains there are more directions from which thunderstorms can drift over a location, and transient dynamic processes such as the low level jet can initiate and even strengthen convection over the plains at night (Bluestein 1993). In the southern part of eastern Canada, while diurnal heating/cooling appears to be the dominant process affecting lightning, thunderstorm translation and transient feature passages have significant effects at many locations, as we see complicated diurnal patterns for Windsor, Toronto, Montreal, Fredericton, and Halifax. Lightning activity in the northern part of eastern Canada is primarily controlled by the diurnal cycle, as seen at Goose Bay.

3.3 POLARITY

The spatial distribution of positive flash percentage during the decade is plotted in Figure 10. The distinct areas and patterns of enhanced positive lightning across Canada are consistent with the initial findings reported in Burrows et al. (2002). In western Canada, positive flash percentages of more than 40% are detected along the Pacific coast and on Vancouver Island. During the winter months, relatively warm low-level temperatures (due to the warm sea surface) combined with cold upper-level temperatures (i.e in upper troughs and lows) result in vertical instability leading to winter convection and associated positive lightning. An isolated region of 25% to 30% positive flash percentage is also identified in southern Manitoba. This is likely due to the influence of intense convective systems that migrate into the area from the central United States during the summer (Lyons et al. 1998a). In northern Canada, a significant area of positive flash percentage of more than 35% extends over the forested regions of Yukon and northern British Columbia. Previous studies have suggested that thunderstorms entraining smoke from forest fires may exhibit enhanced positive CG lightning activity (Lyons et al. 1998b; Kochtubajda et al. 2002; Fernandes et al. 2006). An examination of the forest fire records in Yukon between 1999 and 2008 reveals that a record number of largely lightning-initiated forest fires occurred during the summer of 2004 (Government of Yukon, 2009). This evidence, while circumstantial, is compelling and further analysis of the influence of forest fire smoke on CG flash characteristics in Yukon is shown in this conference.

Several regions of 25% to 30% positive polarity in eastern Canada are observed including an area east of James Bay extending southward into the Rivière Rupert Plateau, the Minganie area north of the Gulf of St. Lawrence in Québec, the plateau area extending east of the Long Range Mountains in southern Newfoundland and off the coast of Nova Scotia in the Atlantic Ocean. These regions can experience freezing rain, wet snow and ice pellets during winter (Stuart and Isaac, 1999; Cortinas et al. 2004) and some positive lightning has been associated with these types of events (Hunter et al. 2001; Henson et al. 2007). The Maritimes have also been identified as an area of significant winter lightning occurrence (Lewis, 2000). The advection of smoke from forest fires in western Canada in the summer may influence the lightning polarity in Northern Québec. The areas of enhanced positive polarity in the Rivière Rupert Plateau region of Québec have also been observed with the Hydro-Québec lightning detection network (Morissette and Gauthier, 2008).

The mean monthly percentage of *positive* CG flashes recorded in each Canadian region from 1999-2008 is illustrated in Figure11. Geographic regions discussed below are defined in Figure 12. The mean monthly percentages of positive CG flashes reflect a strong seasonality in all regions, varying between 30% and 60% in winter compared to values between 7% and 15% in summer. It should be noted that some lightning flashes were detected between November and March in Northern Canada, however, due to the small sample size gathered during the decade (< 10 flashes), meaningful analyses could not be carried out; consequently these statistics were not included in any monthly summaries. This annual variation has been reported elsewhere in the world including Japan (Brook et al. 1982), the United States (Orville and Huffines, 2001), Austria (Schulz et al. 2005), Spain (Rivas Soriano et al. 2005) and Sweden (Sonnadara et al. 2006). An increase in positive flashes has previously been reported in low-level, wintertime thunderstorms (Takeuti et al. 1978). It has been proposed that cold season thunderstorms take place in sheared environments in which the horizontal displacement between the upper and lower negative charge centres facilitates the discharge of positive flashes (Brook et al. 1982). Thus the nature of convection during these colder months could be a factor. Several regions of 25% to 30% positive polarity in eastern Canada are observed including an area east of James Bay extending southward into the Rivière Rupert Plateau, the Minganie area north of the Gulf of St. Lawrence in Québec, the plateau area extending east of the Long Range Mountains in southern Newfoundland and off the coast of Nova Scotia in the Atlantic Ocean. These regions can experience freezing rain, wet snow and ice pellets during winter (Stuart and Isaac, 1999; Cortinas et al. 2004) and some positive lightning has been associated with these types of events (Hunter et al. 2001; Henson et al. 2007). The Maritimes have also been identified as an area of significant winter lightning occurrence (Lewis, 2000). The advection of smoke from forest fires in western Canada in the summer may influence the lightning polarity in Northern Québec. The areas of enhanced positive polarity in the Rivière Rupert Plateau region of Québec have also been observed with the Hydro-Québec lightning detection network (Morissette and Gauthier, 2008).

3.4 MULTIPLICITY

The geographical distribution of mean negative CG flash multiplicity is illustrated in Figure 13. In western Canada, areas of maximum negative multiplicity, exceeding 3.2 strokes, occur throughout the western Prairies and extend into southern Northwest Territories. In eastern Canada, an area of maximum negative multiplicity, exceeding 3.0 strokes, occurs off Anticosti Island in the Gulf of St. Lawrence. Areas exceeding 2.8 strokes occur in northern Ontario and extend into western Quebec. These patterns and areas of high multiplicity are consistent with the initial findings reported in Burrows et al. (2002), however the average multiplicity grid values are somewhat lower than earlier reported. This is likely due to a combination of the incremental network upgrades and the grid criteria used in the present analyses requiring a minimum detection of 10 CG flashes per grid (reduced from 25 CG flashes per grid in Burrows et al. 2002).

The monthly distributions of multiplicity stratified by region are shown in Figure 14. Positive flash multiplicity varies little throughout the year in all regions with values slightly over 1 stroke per positive flash. However, negative flash multiplicities exhibit regional and seasonal differences. These multiplicities show strong peaks of between 2.0 and 2.4 strokes in the late summer, and a minimum of about 1.2 strokes in the late fall and winter in northern and western Canada, respectively. In eastern Canada, a peak is observed in the fall, and the multiplicity values vary from about 1.8 to 2.3 strokes per negative flash throughout the year. It is possible that the variations in negative flash multiplicity are related to the seasonal variation of convective cloud depth as suggested by Orville and Huffines (1999).

3.5 FIRST-STROKE PEAK CURRENT

The geographical distributions of median positive and negative CG peak current flashes in Figures 15 and 16, respectively, show distinct patterns. In western and northern Canada the strongest peak positive currents are observed near Vancouver Island, in southern Manitoba, Yukon and in an area between Great Bear and Great Slave Lakes in the Northwest Territories. In eastern Canada, median peak positive currents are strongest at the extreme edge of the network. In general, the peak negative current on land increases from lower to higher latitudes. Peak currents over the southern Prairies, Ontario and Quebec from 12-15 kA increase to 20-24 kA in northern Canada. The high peak currents observed at the extreme edges of the network reflect the sensor bias as weaker flashes are not detected due to the distance or range from the nearest sensor. A transition from lower to higher median peak negative currents is evident off the eastern coast. Lyons et al. (1998a) also reported higher median peak currents over salt water than over land. Recent studies indicate that the lightning attachment process may explain these observations. Measurements of field risetime (one of the lightning waveform parameters) for both first and subsequent strokes suggest that the enhanced current appears to be associated with downwards-propagating negative stepped leaders attaching to a smooth, highly-conducting (ocean) surface (Cummins et al. 2005). Another possible mechanism is proposed by Murray et al. (2005). An analysis of 131 electric field waveforms produced by first return strokes in cloud-to-ocean lightning revealed that about 37% of the waveforms contain multiple pulses in dE/dt (the time derivative of the electric field) that are large and that occur within 1 μs of the dominant peak. They suggest that these multiple pulses could be due to upward leaders, but questions remain whether multiple pulses in dE/dt are present when lightning strikes land surfaces.

The relation between first-stroke peak current and flash multiplicity was also examined. Orville et al (2002) suggest "*that the key factors that determine the charge in the lower portion of the lightning channel are also related to the total charge available for producing a flash*". Figure 17 shows a general increase in the median *negative f*lash peak current with increasing multiplicity in all regions. Previous studies in the United States (Orville and Huffines, 2001) and Austria (Schulz, et al. 2005) have reported similar trends between first-stroke negative peak current and multiplicity. The high negative peak currents in the North reflect the sensor bias toward the stronger flashes Schulz et al. (2005) quantified the dependence on multiplicity by calculating the ratio of the median peak current for negative flashes with multiplicity of 10 strokes per flash to the median peak current for a negative single-stroke flash. They reported that this ratio in Austria was 2.3. These ratios for the North, West and East were found to be 1.4, 1.7 and 1.8, respectively, suggesting that the dependence of negative flash peak current on multiplicity in Canada is not as strong as compared to the values in Austria.

Figure 18 shows regional differences in the relationships between the median first-stroke peak current and flash multiplicity for *positive* flashes. All regions show an increase in median peak current for flashes with multiplicity up to 2. A decrease thereafter in the median peak current with increasing multiplicity is noted for eastern Canada, but similar decreases are not evident in the West or North. We also note that there are few positive flashes in the North with multiplicities greater than 10 strokes per flash.

CG flashes with peak currents ≥ 100 kA account for nearly 0.9 % of the annual CG flash activity, and significant fractions are positive flashes. While LCLFs are detected throughout the year in Canada, the majority occur in the summer months.

The fraction of negative polarity to positive polarity LCLFs stratified by peak current strength is illustrated in Figure 19. We find that negative LCLFs are more frequent than positive LCLFs in the 100-200 kA range (ratio of 1.2). This value is lower than the values calculated from observations by Lyons et al. (1998a) in the USA and Pinto et al. (2009) in Brazil, 5.4 and 2.8, respectively. Negative LCLFs in Canada are less numerous for peak currents > 200 kA (average ratio of 0.6), as compared to the findings by Lyons et al. (1998a) and Pinto et al. (2009), 4.1 and 1.1 respectively.

5. ACKNOWLEDGEMENTS

Vaisala Inc. of Tucson, Arizona, USA processes the CLDN data and provides telecommunication services to Environment Canada. The data are the property of Environment Canada.

6. REFERENCES

BIAGI, C.J.; K.L. CUMMINS, K.E. KEHOE and E.P. KRIDER. 2007. National Lightning Detection Network (NLDN) performance in southern Arizona, Texas and Oklahoma in 2003-2004. *J. Geophys. Res.* **112**, D05208, doi:10.1029/2006JD007341.

BLUESTEIN, H. B., 1993. *Observations and Theory of Weather Systems. Vol. II, Synoptic-Dynamic Meteorology in Midlatitudes,* Oxford University Press, 594 pp.

BROOK, M.; M. NAKANO, P. KREHBIEL and T. TAKEUTI. 1982. The electrical structure of Hokuriku winter thunderstorms. *J. Geophys. Res.,* **87**, 1207-1215.

BURROWS, W.R., P. KING, P.J. LEWIS, B. KOCHTUBAJDA, B. SNYDER, and V. TURCOTTE. 2002. Lightning occurrence patterns over Canada and adjacent United States from lightning detection network observations. *Atmosphere-Ocean,* **40 (1):** 59-81.

_______________ and B. KOCHTUBAJDA, 2010. A decade of Cloud-to-Ground Lightning in Canada: 1999-2008 Part 1: Flash Density and Occurrence*. Submitted to Atmosphere-Ocean.*

CLODMAN, S. and W. CHISHOLM. 1996. Lightning flash climatology in the Southern Great Lakes Region. *Atmosphere-Ocean* **34**: 345–377.

CORTINAS, J.V.JR.; B.C. BERNSTEIN, C.C. ROBBINS and J.W.STRAPP. 2004. An analysis of freezing rain, freezing drizzle, and ice pellets across the United States and Canada: 1976-90.

CUMMINS, K.L., E. PHILIP KRIDER AND M.D., MALONE, 1998a. The U.S. National Lightning Detection Network and applications of cloud-to-ground lightning data by electric power utilities. *IEEE Trans. Electromagn. Compat.* **40**: 465-480.

______, K.L.; J.A. CRAMER, W.A. BROOKS and E.P. KRIDER. 2005. On the effect of land:sea and other earth surface discontinuities on LLS-inferred lightning parameters. 8th International Symposium on Lightning Protection. 21-25 Nov. 2005 Sao Paulo, Brazil.

__________, K. L. and M.J. MURPHY. 2009. An overview of lightning locating systems: History, techniques, and data uses, with an in-depth look at the U.S. NLDN. *IEEE Trans. Electromagn. Compat.* **51(3):** 499-518.

CURRAN, E.B., R.L. HOLLE, AND R.E. LÓPEZ, 2000. Lightning casualties and damages in the United States from 1954 to 1994. *J. Climate* **13**: 3448-3464.

FERNANDES, W.; I.R.C.A. PINTO, O.PINTO JR., K.M. LONGO and S.R.FREITAS. 2006. New findings about the influence of smoke from fires on the cloud-to-ground lightning characteristics in the Amazon region. *Geophys. Res. Lett*., **33**, L20810, doi:10.1029/2006GL027744.

GOVERNMENT OF THE YUKON. 2009. 2004 Yukon wildfire bulletins. *http://www.community.gov.yk.ca/firemanagement/1145.html*. (accessed June 15, 2009).

HENSON, W.; R. STEWART and B. KOCHTUBAJDA. 2007. On the precipitation and related features of the 1998 Ice Storm in the Montréal area. *Atmos. Res.,* **83**, 36-54. doi:10.1016/j.atmosres.2006.03.006

HOLLE, R.L., R.E. LÓPEZ, L.J. ARNOLD, AND J. ENDRES. 1996. Insured lightning-caused property damage in three western states. *J. Appl. Meteor.* **35**: 1344-1351.

HUFFINES, G. R. and R. E. ORVILLE. 1999. Lightning ground flash density and thunderstorm duration in the continental United States: 1989-96. *J. Appl. Meteorol.* **38:** 1013-1019.

KOCHTUBAJDA, B.; R.E. STEWART, J.R. GYAKUM and M.D. FLANNIGAN. 2002. Summer convection and lightning over the Mackenzie River Basin and their impacts during 1994 and 1995. *Atmosphere-Ocean,* **40(2):** 199-220.

__________________ and W.R. BURROWS. 2010. A decade of cloud-to-ground lightning in Canada: 1999-2008 Part 2: Polarity, multiplicity and first-stroke peak current. *Submitted to Atmosphere-Ocean.*

KING, P.W.S., M.J. LEDUC, D.M.L. SILLS, N.R. DONALDSON, D.R. HUDAK, P. JOE, and B.P. MURPHY. 1996. Lake breezes in southern Ontario and their relation to tornado climatology. *Wea. And Forecasting* **18:** 795-807.

LANKEN, D. 2000. Struck by lightning. *Canadian Geographic,* July/Aug, **120**, 20-32.

LEWIS, P.J., 2000: Winter lightning in the Maritime Provinces of Canada. *Proceedings 2000 Intl. Lightning Detection Conference, Tucson, AZ, 7-8 Nov 2000, Global Atmospherics Inc.*

LYONS, W.A.; M. ULIASZ and T.E. NELSON. 1998a. Large peak current cloud-to-ground lightning flashes during the summer months in the contiguous United States. *Mon. Wea. Rev.,* 126, 2217-2233.

LYONS, W.A.; T.E. NELSON, E.R. WILLIAMS, J.A.CRAMER and T.R. TURNER. 1998b. Enhanced positive cloud-to-ground lightning in thunderstorms ingesting smoke from fires. *Science*, **282**, 77-80.

MILLS, B.; D. UNRAU, C. PARKINSON, B. JONES, J. YESSIS and K. SPRING. 2008. Assessment of lightning-related fatality and injury risk in Canada. *Nat. Hazards*, 47, 157-183. doi:10.1007/s11069-007-9204-4

_______, B.; D. UNRAU, L. PENTELOW, and K. SPRING. 2009. Assessment of lightning-related damage and disruption in Canada. *Nat. Hazards,* doi:10.1007/s11069-009-9391-2.

MORISSETTE, J. and S. GAUTHIER. 2008. Study of cloud-to-ground lightning in Quebec: 1996-2005. *Atmosphere-Ocean*, **46**: 443-454.

MURRAY, N.D.; E.P. KRIDER and J.C.WILLET. 2005. Multiple pulses in dE/dt and the fine-structure of E during the onset of first return strokes in cloud-to-ocean lightning. Atmos. Res., 76, 455-480. doi:10.1016.j.atmosres.2004.11.038

ORVILLE, R.E. and G.R. HUFFINES. 1999. Lightning ground flash measurements over the contiguous united states: 1995-97. *Mon. Wea. Rev*. **127**: 2693-2703.

______________________________. 2001. Cloud-to-ground lightning in the United States: NLDN Results in the first decade, 1989-98. *Mon. Wea. Rev.*, **129**, 1179-1193.

ORVILLE, R.E., G.R. HUFFINES, W.R. BURROWS, R.L. HOLLE and K.L. CUMMINS. 2002. The North American Lightning Detection Network (NALDN) – First results: 1998-2000. *Mon. Wea. Rev.* **130**: 2098-2109.

PINTO, O.; I.R.C.A. PINTO, D.R. de CAMPOS and K.P. NACCARATO. 2009. Climatology of large peak current cloud-to-ground lightning flashes in southeastern Brazil. *J. Geophys. Res.*, **114**, D16105, doi:10.1029/2009JD012029.

SILLS, D. M. L., 1998: Lake and land breezes in southwestern Ontario: observations, analyses and numerical modelling. *PhD dissertation, CRESS, York University*, 338 pp.

SONNADARA, U.; V. COORAY and T. GÖTSCHL. 2006. Characteristics of cloud-to-ground lightning flashes over Sweden. *Phys. Scr.,* **74**: 541-548.

SORIANO, R., L.; F. DE PABLO and C. TOMAS. 2005. Ten-year study of cloud-to-ground lightning activity in the Iberian Peninsula. *J. Atmos, Solar-Terr. Phys.,* **67**, 1632-1639. doi:10.1016/j.jastp.2005.08.019

STOCKS, B.J., J.A., MASON, J.B., TODD, E.M., BOSCH, B.M.,WOTTON, B.D. AMIRO, M.D. FLANNIGAN, K.G., HIRSCH, K.A., LOGAN, D.L., MARTELL, AND W.R., SKINNER, 2002: Large forest fires in Canada, 1959–1997. *J. Geophy. Res.,* **108 (D1)**, 8149, doi:10.1029/2001JD000484.

STUART, R.A. and G.A. ISAAC. 1999. Freezing precipitation in Canada. *Atmosphere-Ocean*, **37**, 87-102.

TAKEUTI, T. M.; NAKANO, M. BROOK, D.J. RAYMOND and P. KREHBIEL. 1978. The anomalous winter thunderstorms of the Hokuriku coast. *J. Geophys. Res.* **83**: 2385–2394.

THE MATHWORKS. 2009. MATLAB Mapping Toolbox 3 Users Guide. The MathWorks. Inc., Natick, MA. 1710 pp.

7. FIGURES

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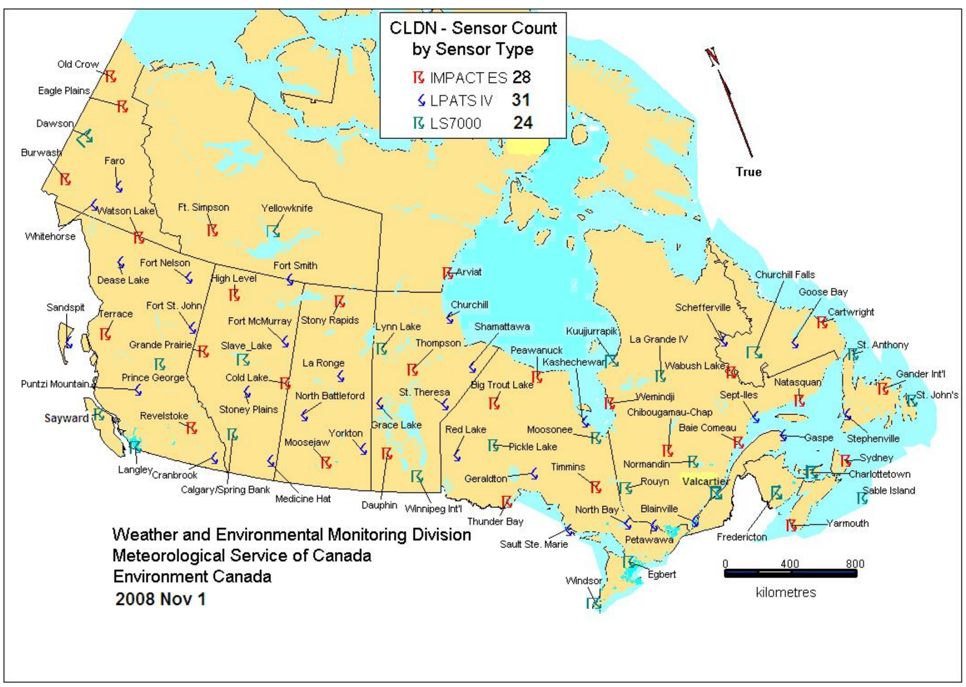
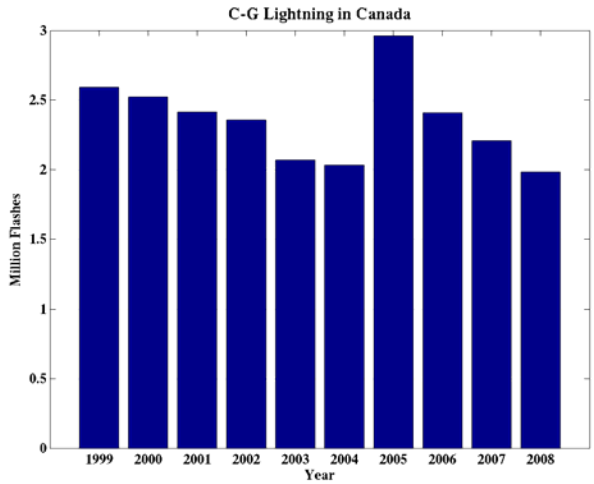


Figure 1. Location and type of CLDN sensors.

a)



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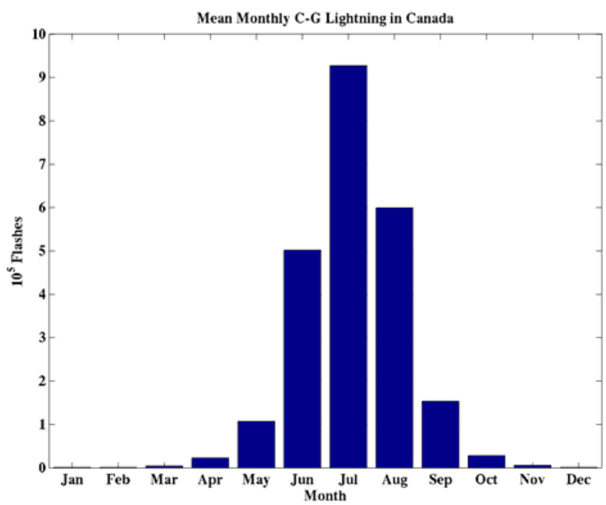


.**Figure 2.** a) number of CG flashes each year; b) number of CG flashes each month.

a)

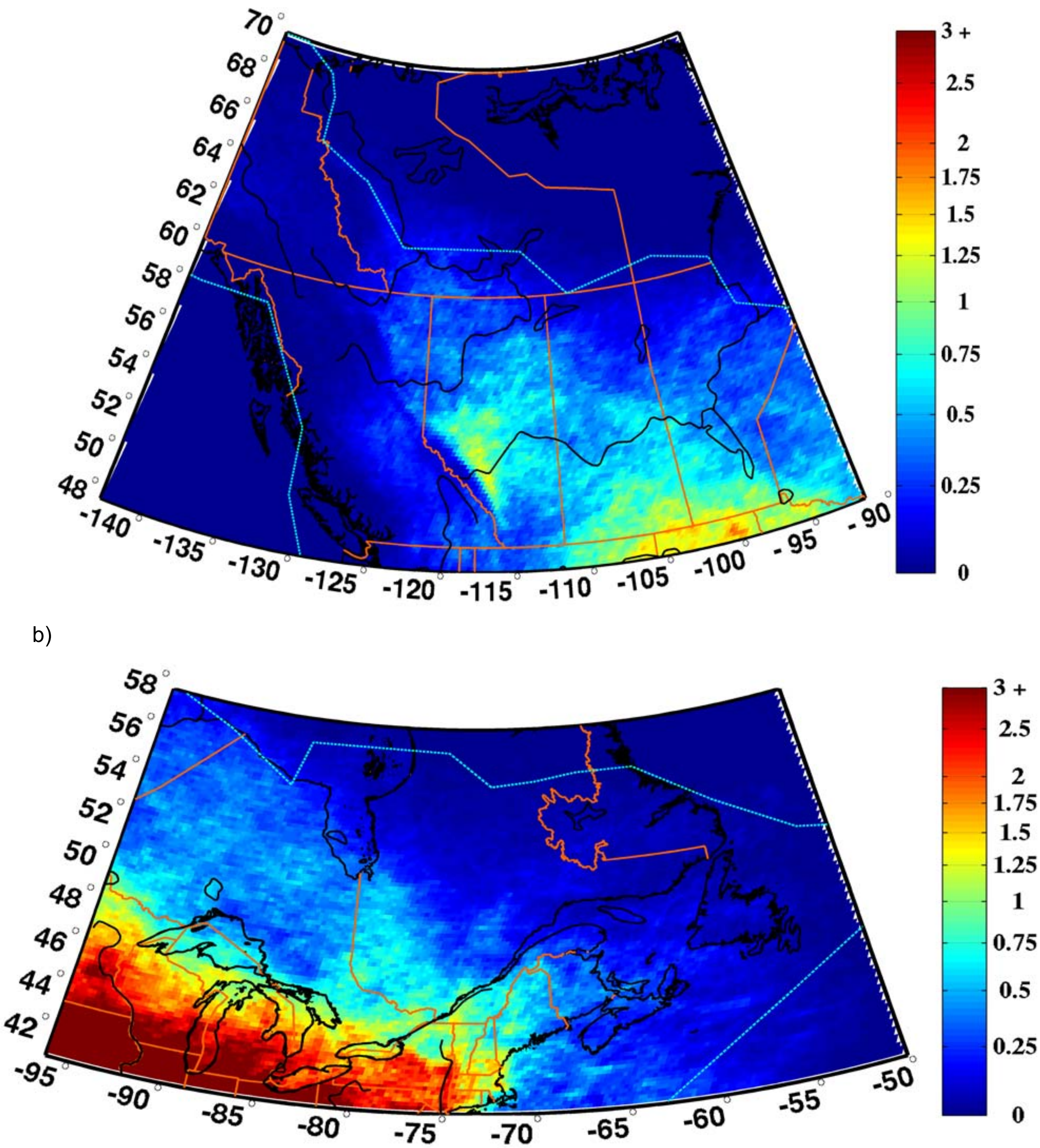
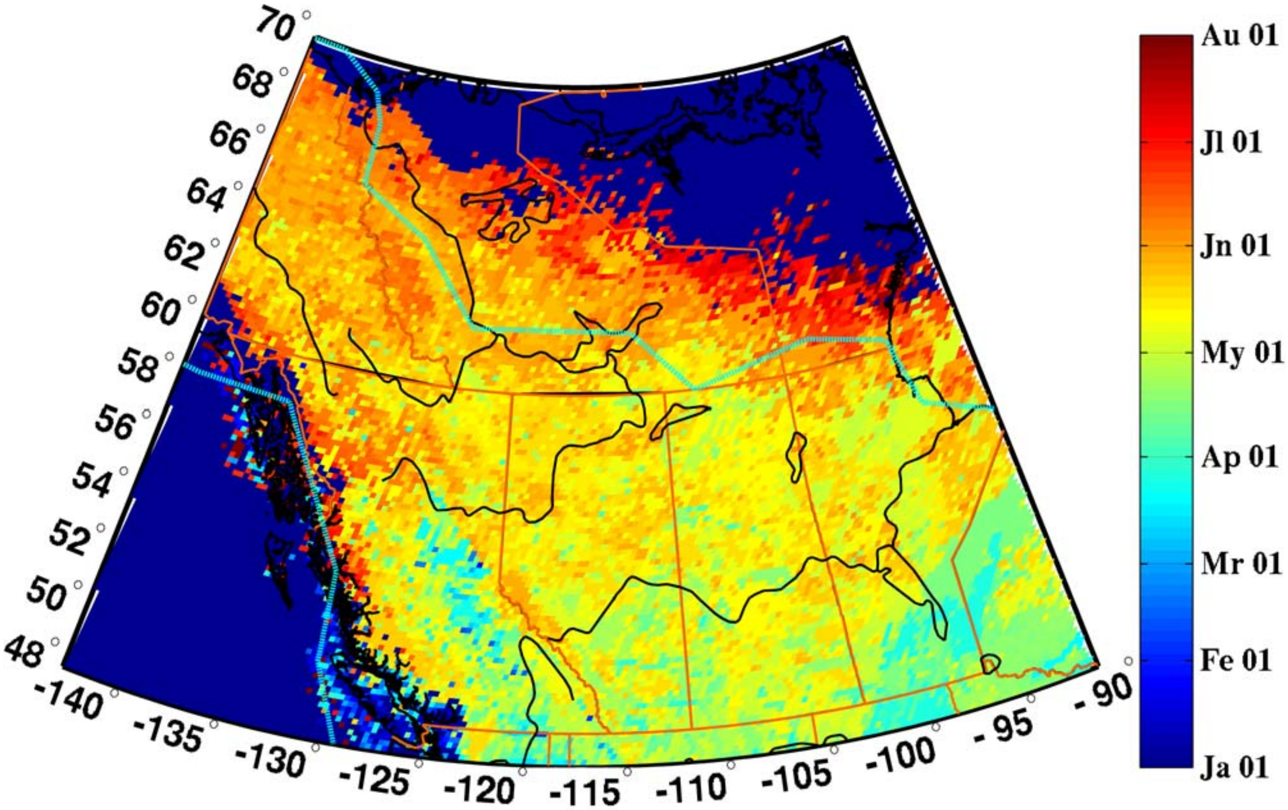


Figure 3. 1999-2008 average flash density (flash km^{-2} yr^{-1} for a) for western Canada; b) eastern Canada. Light blue irregular lines around the periphery are the approximate 70% detection efficiency as of 1 November 2008

a)



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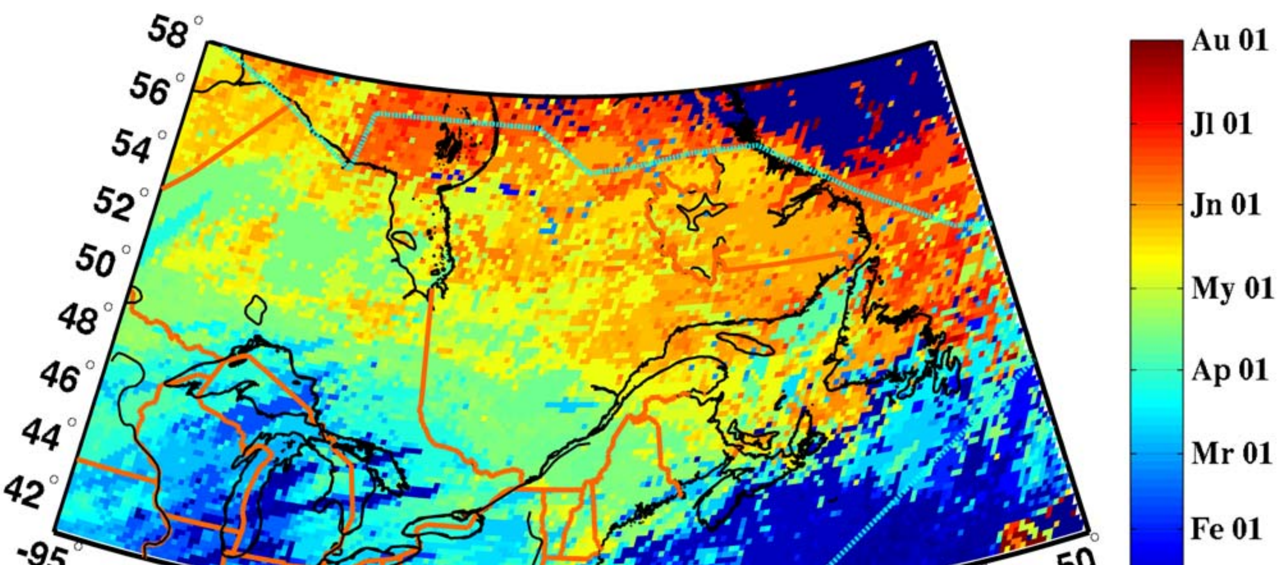
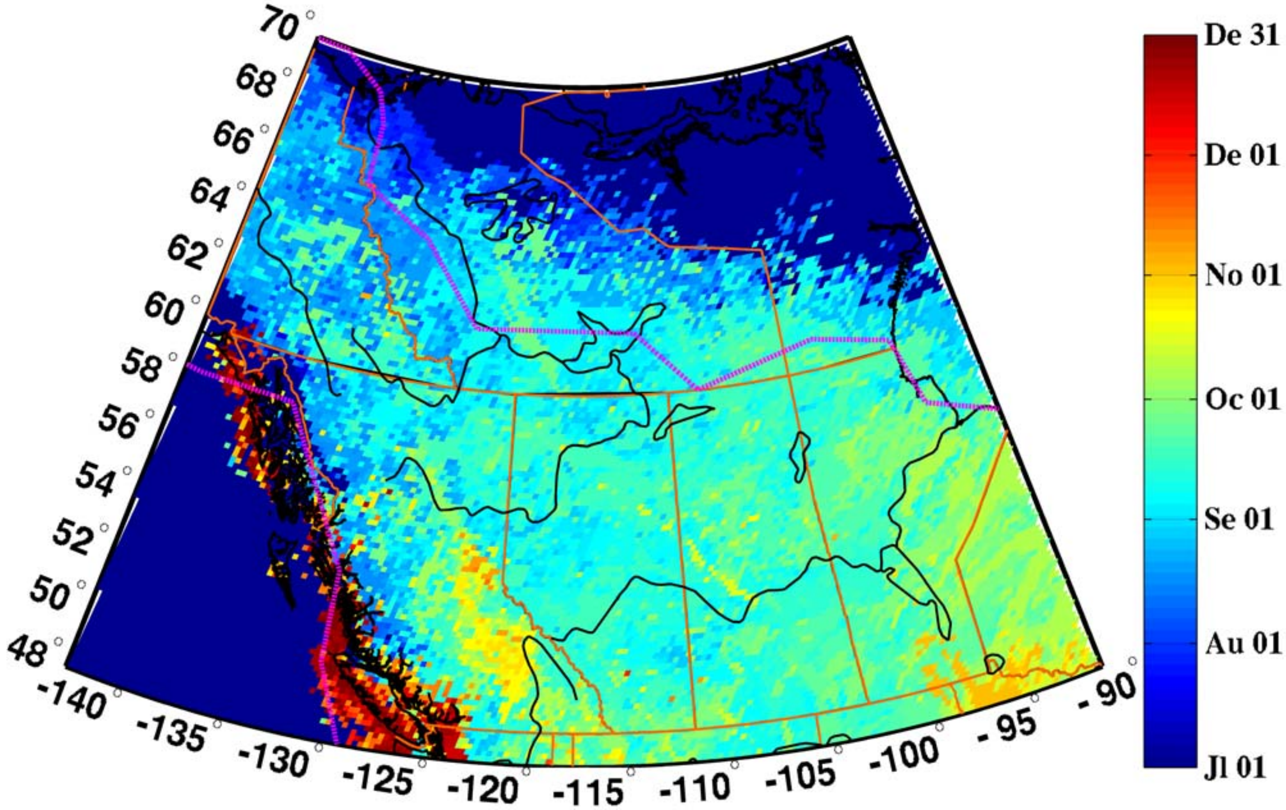


Figure 4. Beginning date of lightning season, defined as the date of the 0.5 percentile of the total number pf CG flashes that occurred 1999-2008 in each 20 km-sided square for a) western Canada; b) eastern Canada. Irregular lines around the periphery same as Fig. 3.

a)



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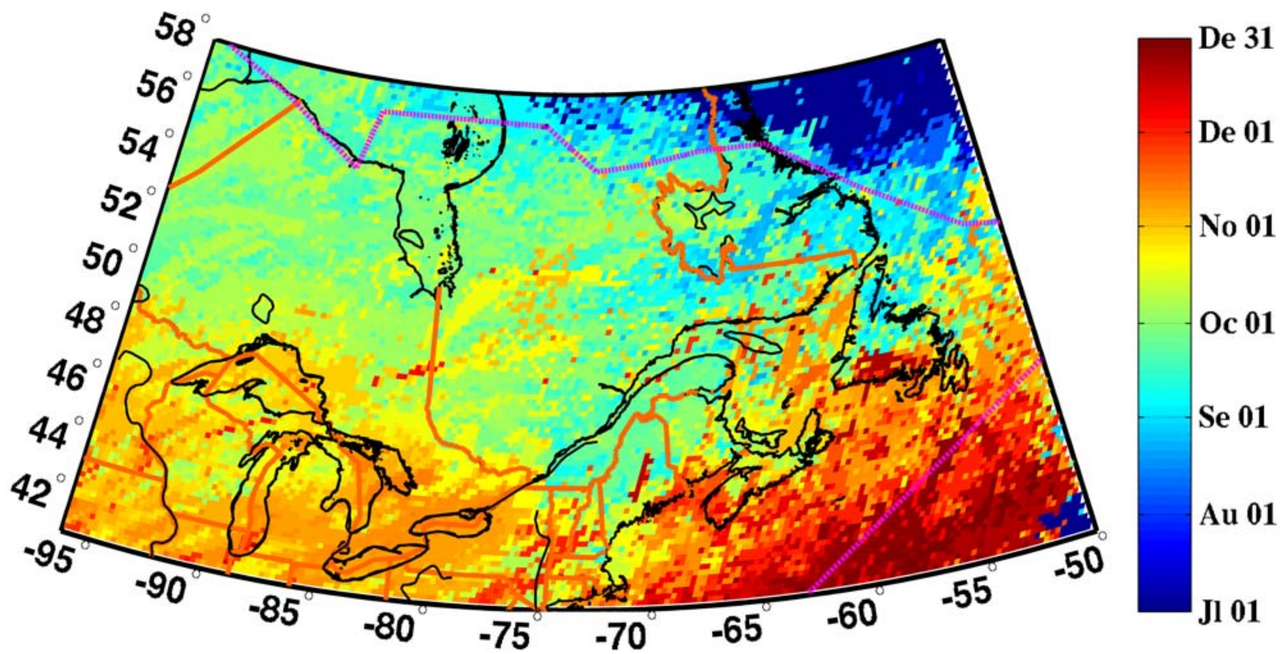


Figure 5. Ending date of lightning season, defined as the date of the 99.5 percentile of the total number pf CG flashes that occurred 1999-2008 in each 20 km-sided square for a) western Canada; b) eastern Canada. Irregular lines around the periphery same as Fig. 3.

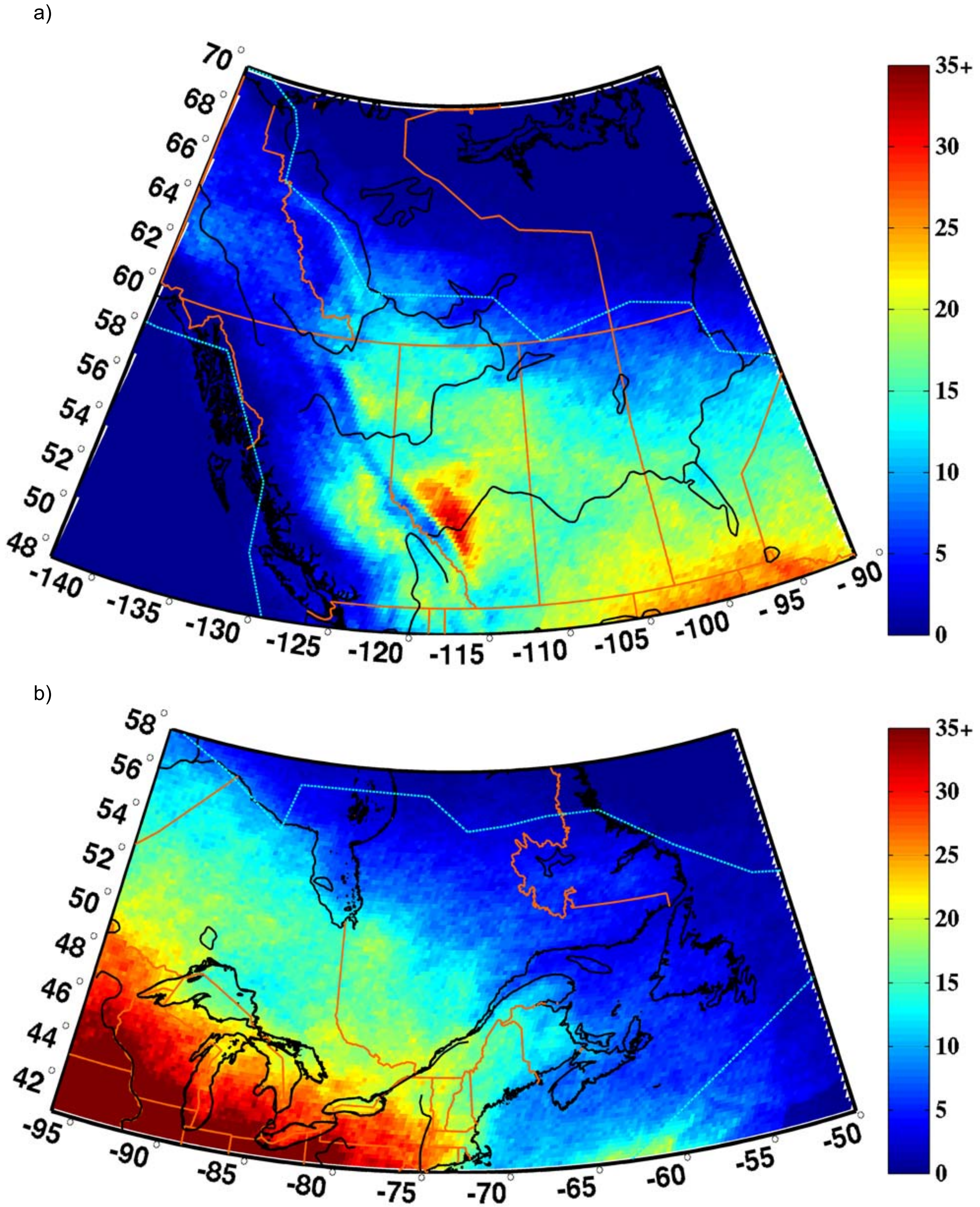
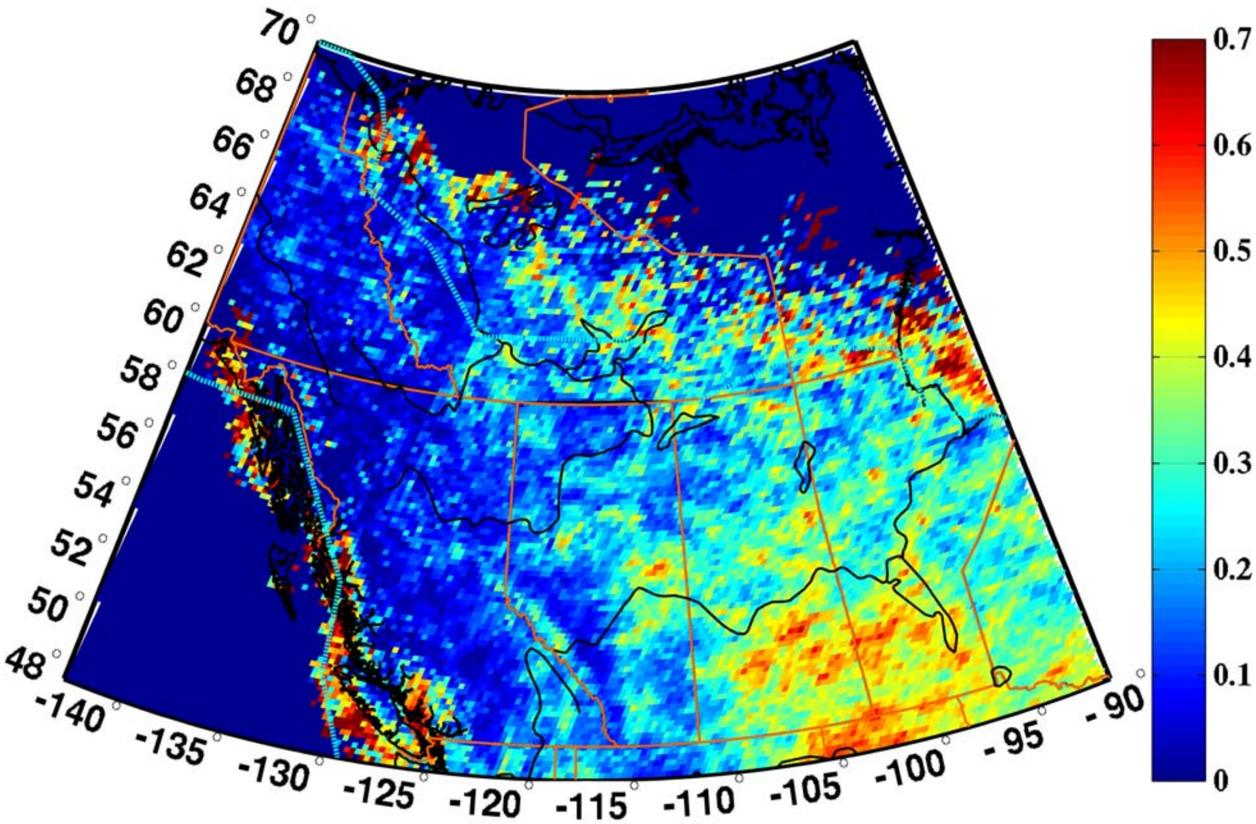


Figure 6. 1999-2009 average number of days per year that at least 1 CG flash was detected in a 20 km-sided square for a) western Canada; b) eastern Canada. Irregular lines around the periphery same as Fig. 3.

a)



b)

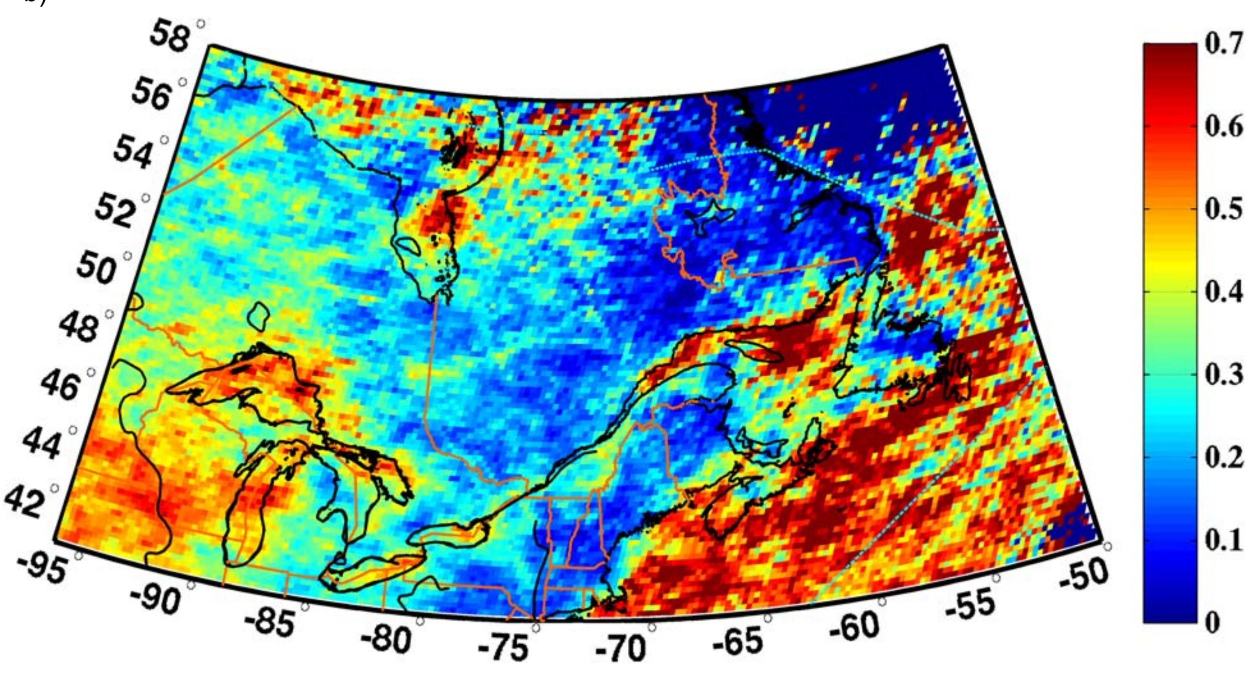


Figure 7. Fraction of CG lightning that occurred between 22:30 and 10:30 local solar time. Irregular lines around the periphery same as Fig. 3.

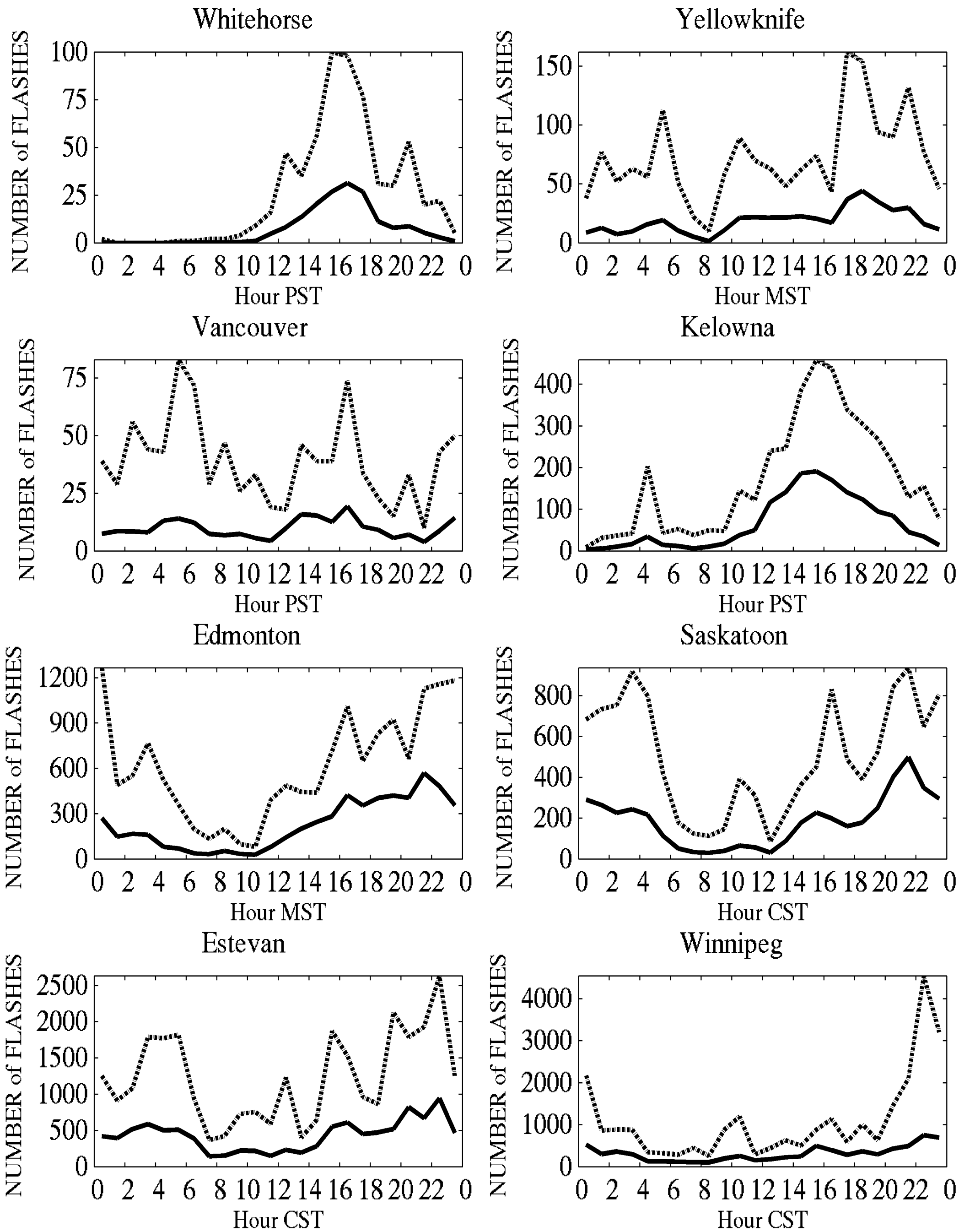


Figure 8. Number of CG flashes that occurred in each hour local standard time for at 8 cities in the west half of Canada. In each panel the solid line shows the 1999-2008 mean, and the dashed line shows the maximum of the ten values at each hour.

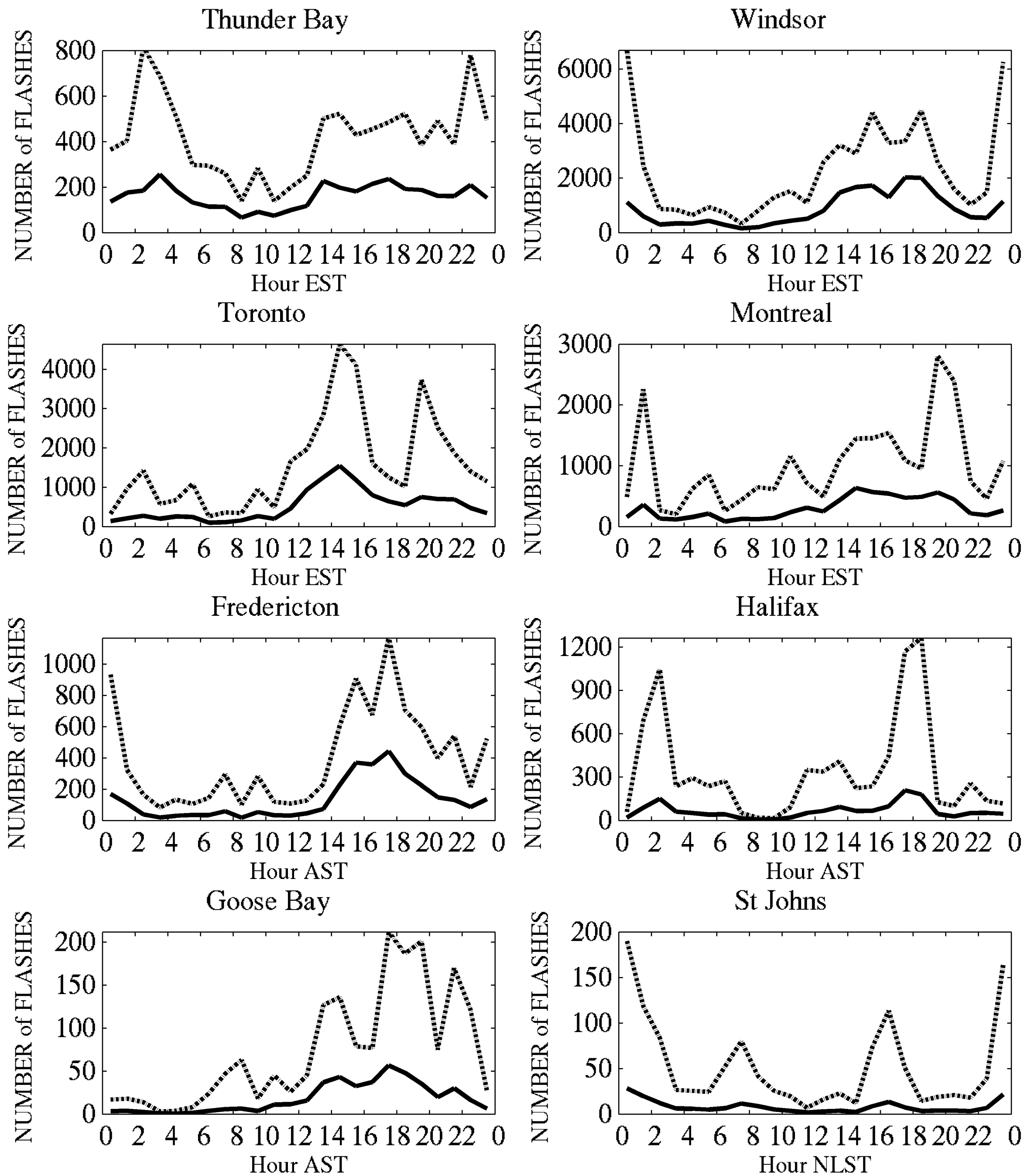


Figure 9. Number of CG flashes that occurred in each hour local standard time at 8 cities in the east half of Canada. In each panel the solid line shows the 1999-2008 mean, and the dashed line shows the maximum of the ten values at each hour.

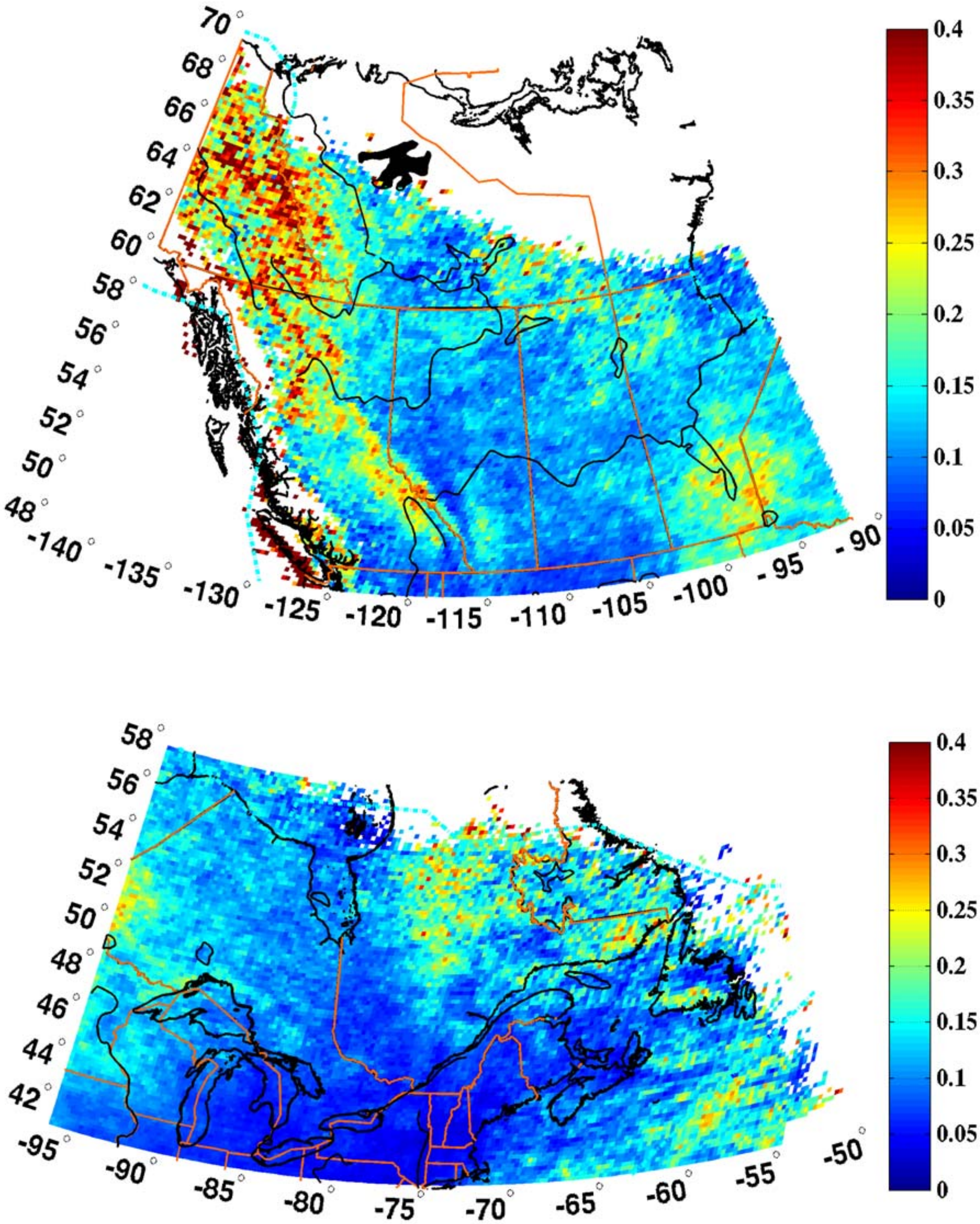


Figure 10. Spatial variation of the fraction of positive CG flashes (1999-2008). Irregular lines around the periphery same as Fig. 3.

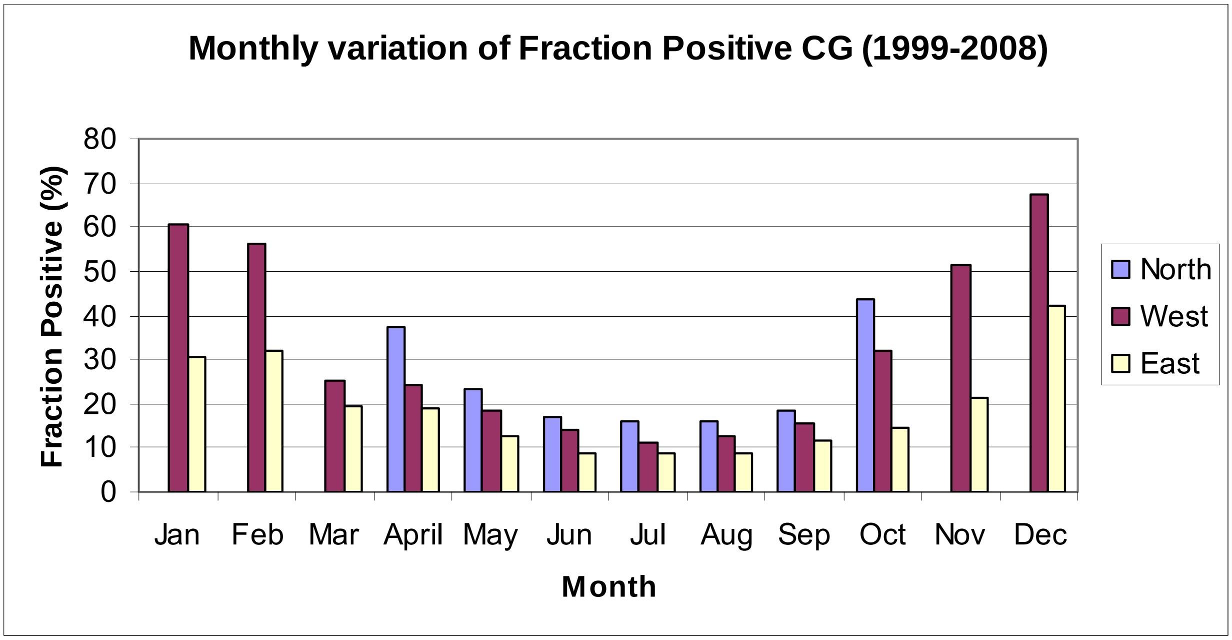


Figure 11. Monthly variation of the percentage of positive CG flashes detected in each region (1999-2008).

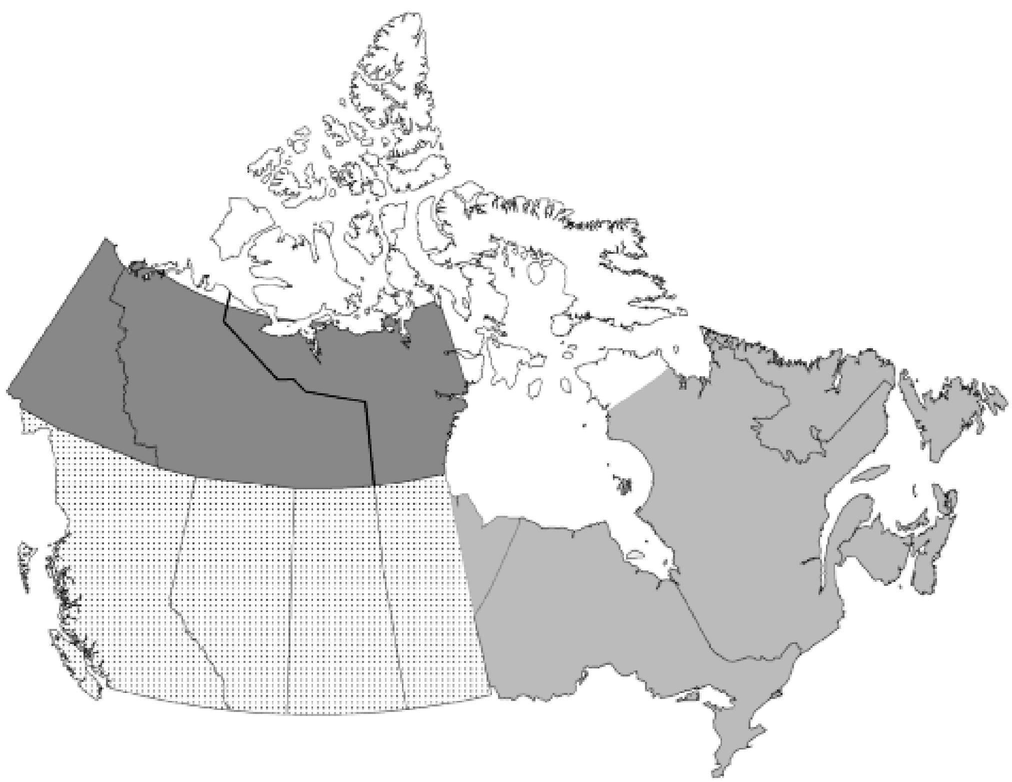


Figure 12. Map identifying the eastern, western and northern regions of Canada

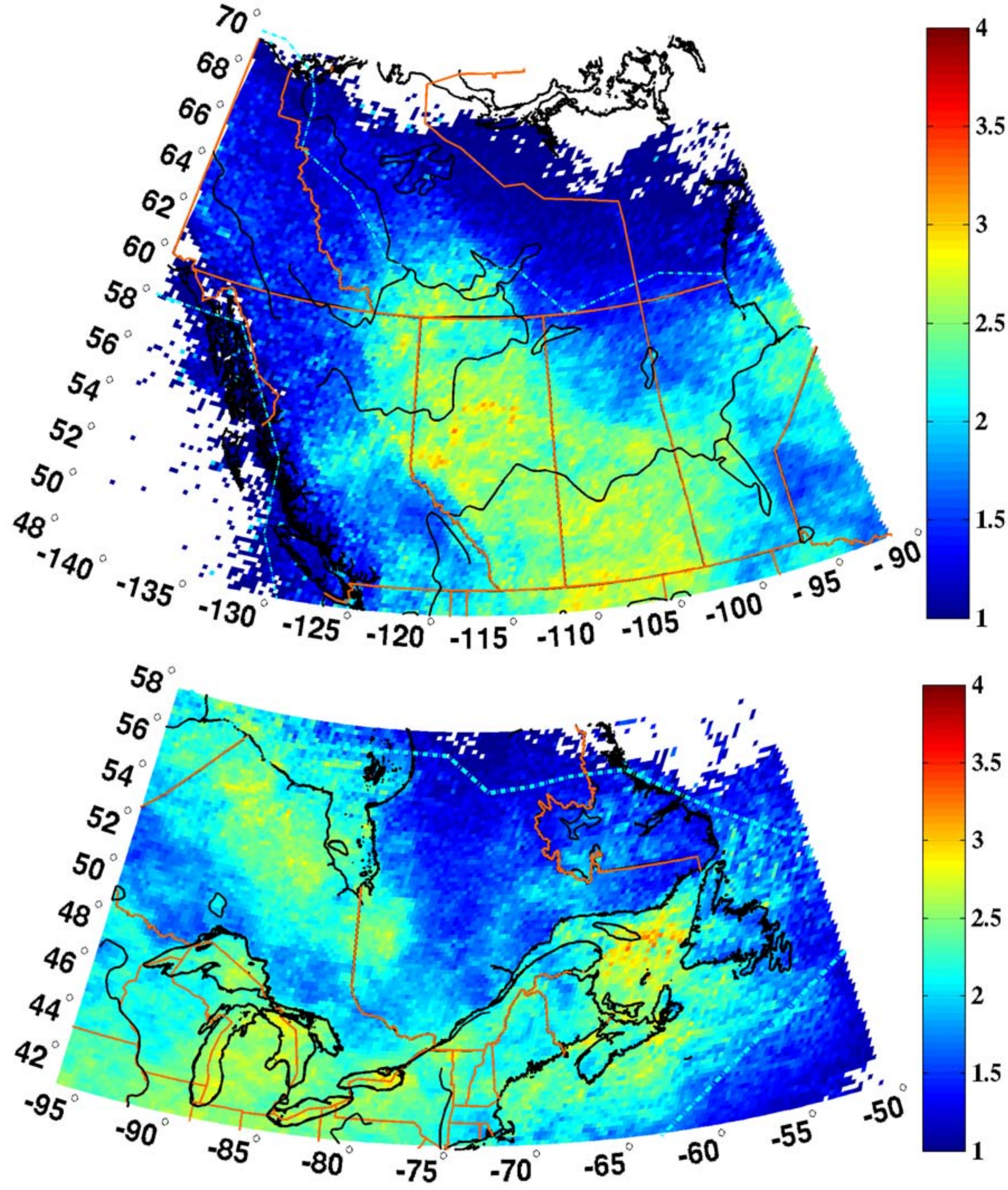


Figure 13. Spatial variation of the mean negative CG flash multiplicity (1999-2008). Irregular lines around the periphery same as Fig. 3.

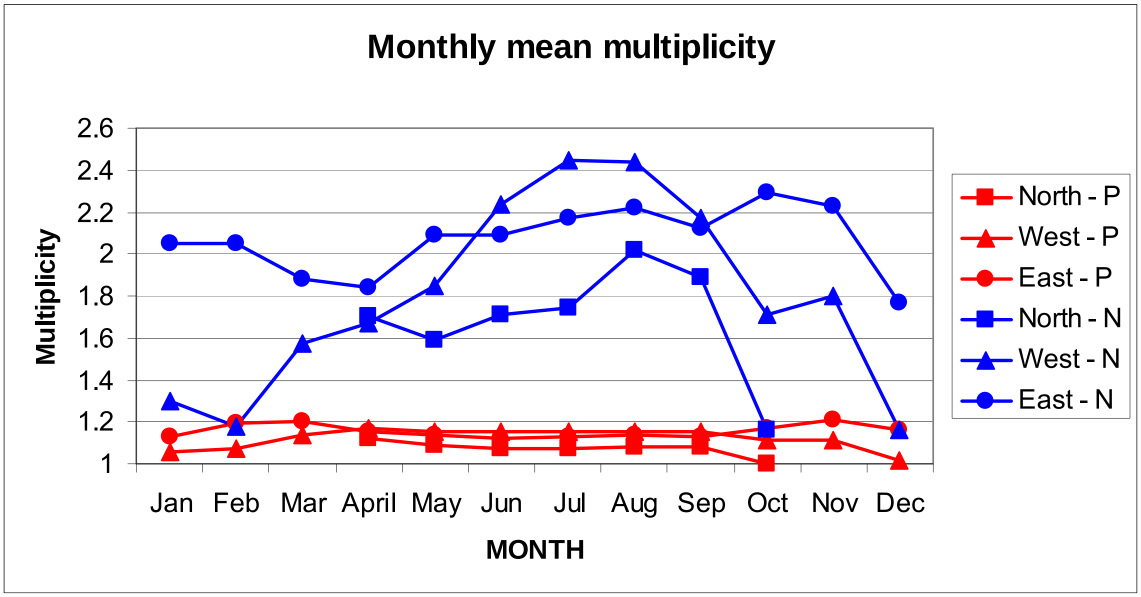


Figure 14. Monthly variation of the mean multiplicity for negative (in blue) and positive (in red) CG flashes in each region (1999-2008). The solid squares, triangles and circles denote the northern, western and eastern regions, respectively.

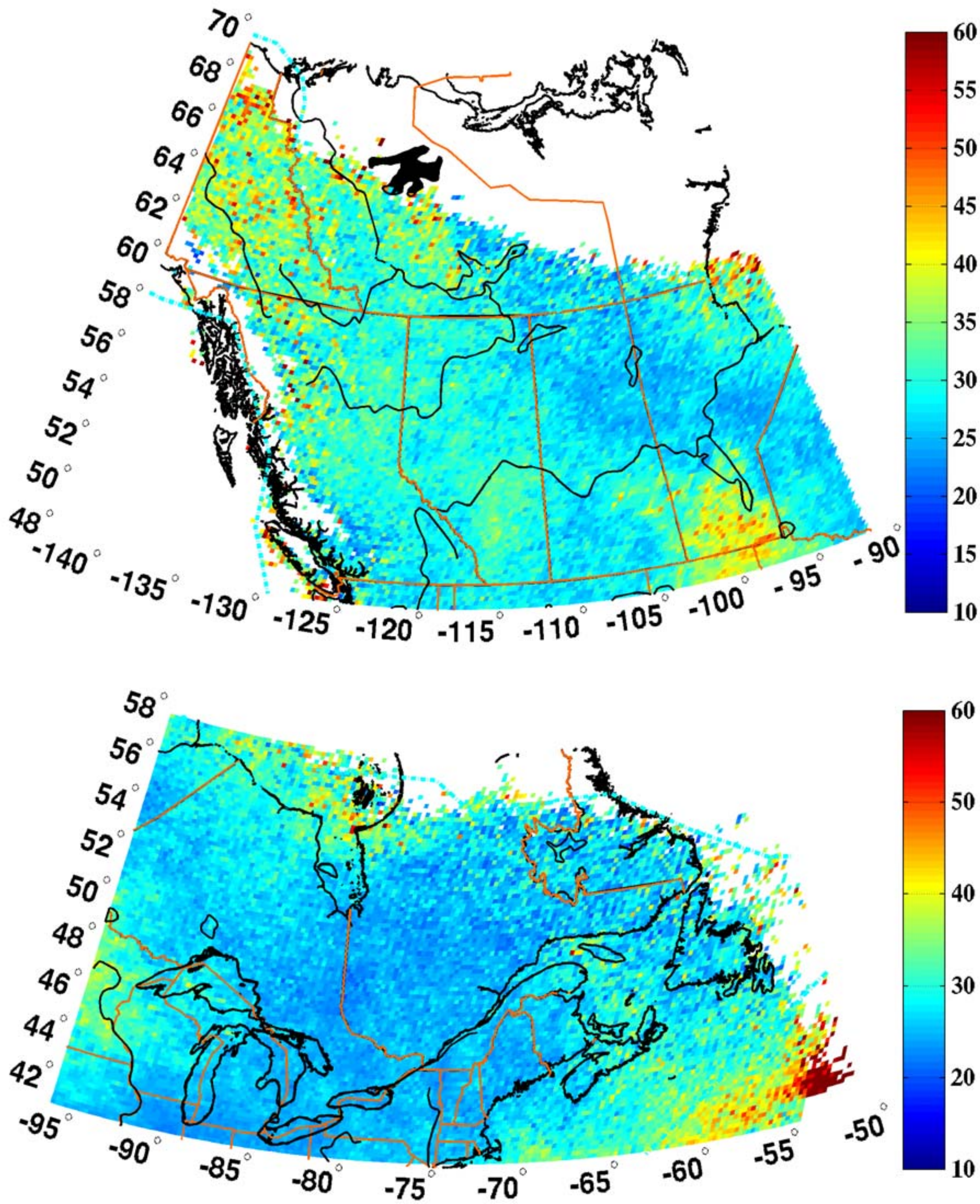


Figure 15.. Spatial variation of median positive CG peak current flashes (1999-2008). Irregular lines around the periphery same as Fig. 3.

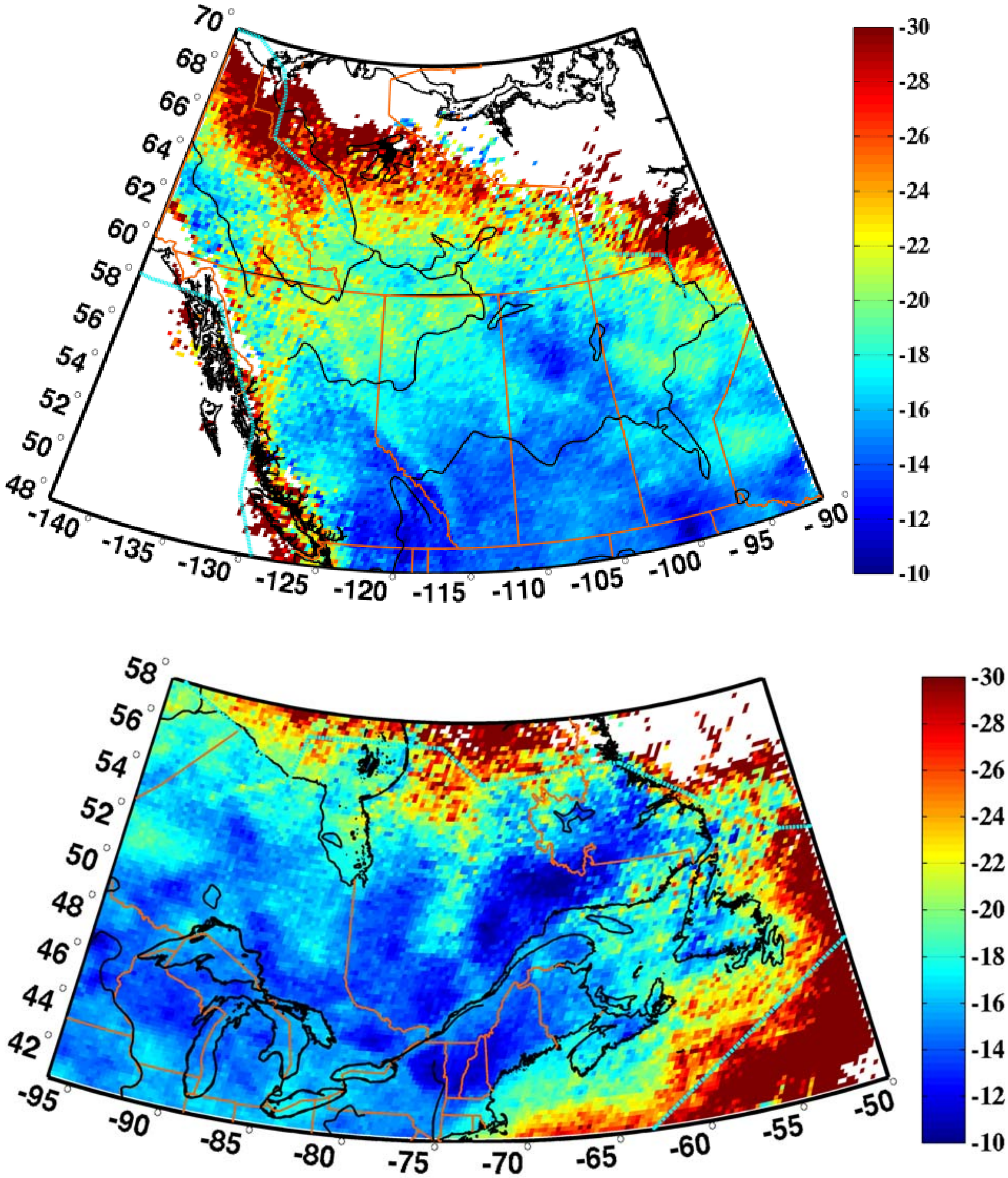


Figure 16. Spatial variation of median negative CG peak current flashes (1999-2008). Irregular lines around the periphery same as Fig. 3.

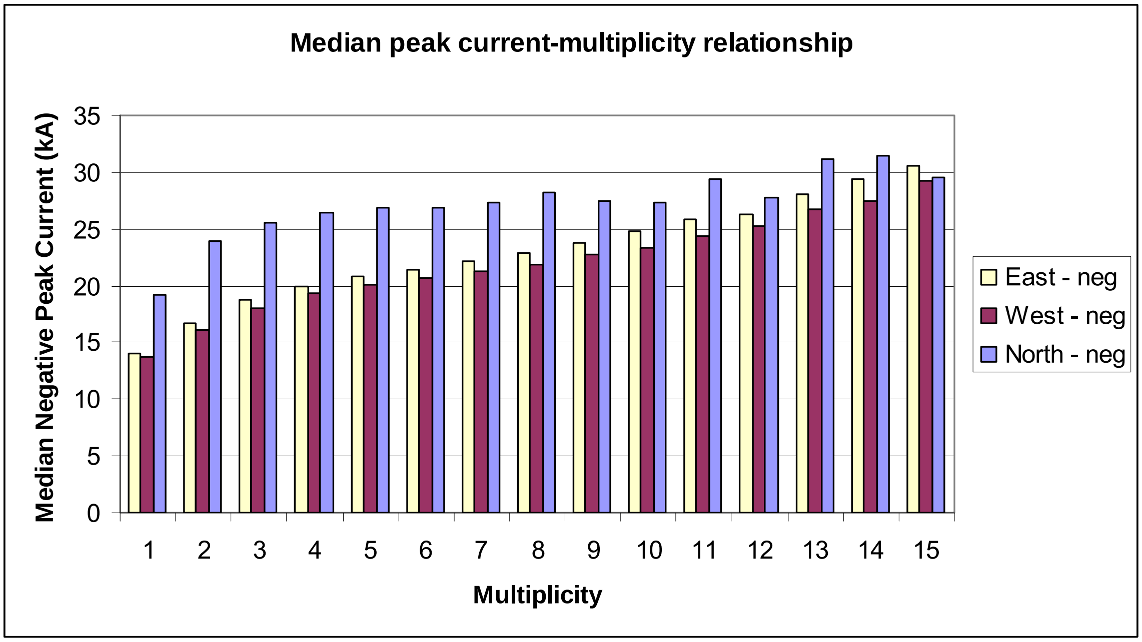


Figure 17. Median peak current as a function of multiplicity for negative CG flashes

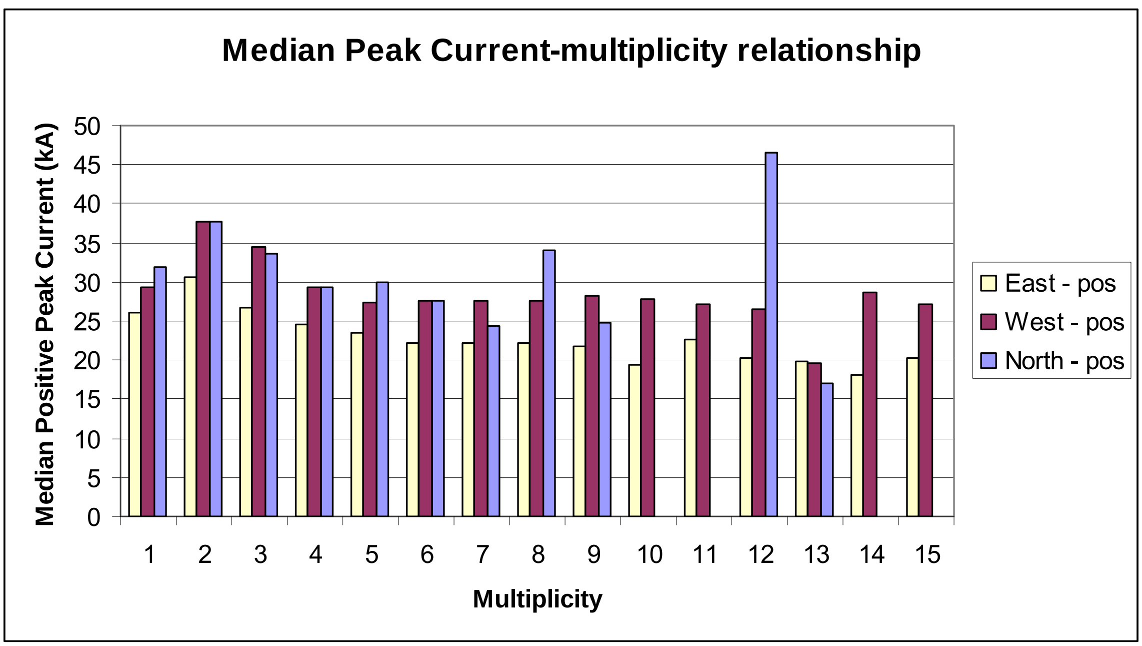


Figure 18. Median peak current as a function of multiplicity for positive CG flashes.