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Streamflow in the Mackenzie Basin, Canada

MING-KO WOO[1,2] and ROBIN THORNE[1]

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ABSTRACT. Rivers of the Mackenzie Basin exhibit several seasonal flow patterns that include the nival (snowmelt dominated), proglacial (influenced by glacier melt), wetland, prolacustrine (below large lakes), and regulated flow regimes. The Mackenzie amalgamates and moderates these regimes to deliver spring peak flows, followed by declining summer discharge and low winter flows, to the Arctic Ocean. The mountainous sub-basins in the west (Liard, Peace, and northern mountains) contribute about 60% of the Mackenzie flow, while the interior plains and eastern Canadian Shield contribute only about 25%, even though the two regions have similar total areas (each occupying about 40% of the total Mackenzie Basin). The mountain zone is the dominant flow contributor to the Mackenzie in both high-flow and low-flow years. A case study of the Great Slave system demonstrates the effects of natural runoff, regulated runoff, and lake storage on streamflow, as well as the large year-to-year variability of lake levels and discharge. Despite a warming trend in the past three decades, annual runoff of the Mackenzie Basin has not changed. Significant warming at most climatic stations in April (and at some, also in May or June) could have triggered earlier snowmelt. The first day of hydrograph rise for the main trunk of the Mackenzie (seen as a proxy for breakup) has advanced by about three days per decade, though the trend was not statistically significant for the mountain rivers. Peak flows do not reveal any trend, but the arrival of the spring peaks has become more variable. More evidence is needed to interpret these flow phenomena properly.

Key words: streamflow, regimes, Mackenzie River, climate change, streamflow variability, peakflow

RÉSUMÉ. Les rivières du bassin du Mackenzie manifestent plusieurs modèles d’écoulement qui comprennent les régimes d’écoulement nival (dominé par la fonte des neiges), proglaciaire (influencé par la fonte glaciaire), de marécages, prolascustre (en aval de grands lacs) et régularisé. Le Mackenzie combine et modère ces régimes pour donner des débits de pointe au printemps, suivis d’un débit à la baisse en été, puis de faibles débits en hiver, en direction de l’océan Arctique. Les sous-bassins montagneux occidentaux (Liard, Peace et montagnes du Nord) contribuent pour environ 60 % au débit du Mackenzie, tandis que les plaines intérieures et le Bouclier canadien oriental ne contribuent que pour environ 25 %, même si les deux régions ont une superficie globale semblable (chacune occupant environ 40 % de la superficie totale du bassin du Mackenzie). La zone montagneuse apporte la contribution majeure au régime du Mackenzie, dans les années à fort débit comme dans celles à faible débit. Une étude de cas du réseau du Grand lac des Esclaves révèle l’impact sur le débit fluvial de l’écoulement naturel, de l’écoulement régularisé et de la hauteur d’eau dans le lac, ainsi que la grande variabilité d’une année sur l’autre du niveau et du débit des lacs. Malgré la tendance au réchauffement des trois dernières décennies, l’écoulement annuel du bassin du Mackenzie n’a pas changé. Un réchauffement notable enregistré à la plupart des stations climatiques en avril (et à certaines aussi en mai ou juin) pourrait avoir provoqué une fonte nivale précoce. Le premier jour où se manifeste l’augmentation du régime hydrique pour l’artère principale du Mackenzie (considéré comme un indicateur de la débâcle) a avancé d’environ trois jours par décennie, bien que statistiquement cette tendance ne soit pas significative pour les rivières de montagne. Les débits de pointe ne révèlent aucune tendance, mais l’arrivée des pics printaniers est devenue plus variable. Il faudrait des preuves supplémentaires pour interpréter correctement ces phénomènes d’écoulement.

Mots clés: écoulement fluvial, régimes, fleuve Mackenzie, changement climatique, variabilité de l’écoulement fluvial, débit de pointe

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INTRODUCTION

The Mackenzie River drains an area of 1.8 million km^2, about one-fifth of the total land area of Canada. The basin encompasses a diversity of natural environments and possesses abundant potential resources, while the Mackenzie valley corridor facilitates north-south transportation of commodities. The Mackenzie is the largest North American river that brings freshwater to the Arctic Ocean. The freshwater layer maintains a thermohaline gradient that prevents the extrusion of the denser, saline sea water, thus preserving the integrity of the polar ice pack. At the nearshore zone, however, ice breakup is advanced by the massive river discharge in the spring (Searcy et al., 1996). Thus, knowledge of the quantity and the seasonality of freshwater flow of the Mackenzie River not only is

[1] School of Geography and Geology, McMaster University, Hamilton, Ontario L8S 4K1, Canada
[2] Corresponding author: woo@mcmaster.ca
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important to the environment and development within the basin, but has implications for the littoral zone and the broader oceanic and atmospheric circulations (e.g., Rahmstorf, 1994). The Mackenzie Basin is considered in many climatic models as an area likely to be affected by climatic warming (Cohen, 1997). This implies an effect on the water balance and runoff generation in the basin. Furthermore, the issue of inter-basin water transfer is attracting growing interest, and its ramifications upon the Mackenzie cannot be ignored. An examination of the flow conditions of the river and its major tributaries can provide a firm hydrological basis for debates on water resources.

Although streamflow data have been collected for decades along the main trunk of the Mackenzie and for a number of its tributaries, analysis of the discharge data is needed to establish the pattern of how much water is delivered during different times of the year from different environments. This information will be useful in assessing the sensitivity of runoff from various parts of the Mackenzie system to forcing of the climate and to streamflow regulation. The present study analyzes (1) the flow patterns of the Mackenzie and its major sub-basins, (2) the contribution from the major sub-basins to the Mackenzie system, and (3) the possible presence of trends in the recent flow records.

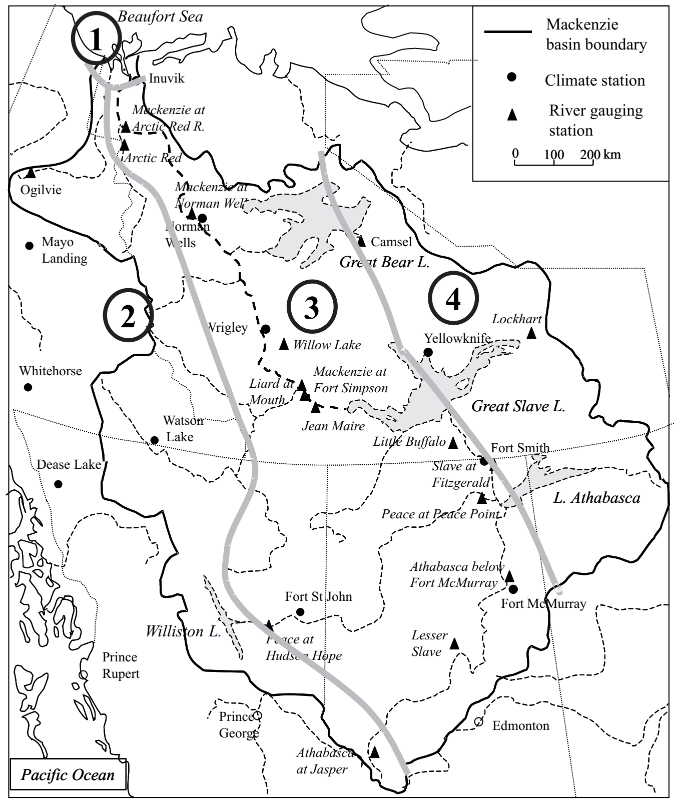

FIG. 1. The Mackenzie Basin and its major drainage areas. Physiographical subdivisions are (1) Delta, (2) Western Cordillera, (3) Interior Plains, and (4) Precambrian Shield. Also shown are the climatic and stream gauging stations that provided data for this study.

THE MACKENZIE BASIN

The Mackenzie Basin extends from central Alberta in the south to the Beaufort Sea coast in the north, and from the continental divide of the Western Cordillera to the Canadian Shield at the eastern border of the Northwest Territories. It encompasses four physiographical regions (Fig. 1). In the west, the Western Cordillera consists of a series of mountain chains and valleys or high plateaus. Many ridges of the Rocky Mountain chain exceed 2000 m elevation, and some have glaciers occupying the mountain tops and high valleys. To the east is the Canadian Shield, a rolling terrain with myriad lakes and valley-wetlands separating upland outcrops of Precambrian bedrock. The central zone is the Interior Plains, with wetlands, lakes, and vegetation that ranges from prairie grassland in the south, through the boreal and subarctic forests, to the tundra in the north. At the mouth of the Mackenzie is its delta, an assemblage of distributaries, levees, wetlands, and lakes.

The basin straddles several climatic regions, including the cold temperate, mountain, subarctic, and arctic zones. Annual precipitation declines notably from west to northeast, ranging from more than 1000 mm in the southwest, and more than 500 mm in the northwest, to low values of about 200 mm along the Arctic coast (Hydrological Atlas of Canada, 1978; Fig. 2a). Snowfall is the major form of precipitation, and it is generally recognized that snowfall is underestimated, particularly in the mountainous and the windswept Arctic terrain (Metcalfe et al., 1994). In many parts of the basin, snow stays on the ground for over half the year, and snowmelt usually triggers major high-flow events. Convectional and frontal rainfall in summer and autumn is also an important source of water for streamflow generation.

DATA AND METHODS

The Water Survey of Canada, through the hydrometric database (HYDAT), provides streamflow and water level data that are used in the present study (Fig. 1). Although many gauging stations in the Mackenzie Basin have only short records with incomplete data, several large catchments (areas > 200 000 km^2) have at least 25 years of monthly and annual flow records. The Mackenzie itself is gauged at several locations along its main trunk. Discharge data for the Mackenzie gauged at the village of Arctic Red River (with a drainage area of 1.68 million km^2), before the river branches into many distributaries, will be used as the total flow for the Mackenzie system.

The Mackenzie drainage is divided into seven major drainage areas with different hydrological characteristics and with discharge values that are measured or that can be estimated (Fig. 1). These include the Athabasca, located in the cold, temperate zone of the southern Mackenzie Basin; the Peace, which is impounded by the Bennett Dam to form

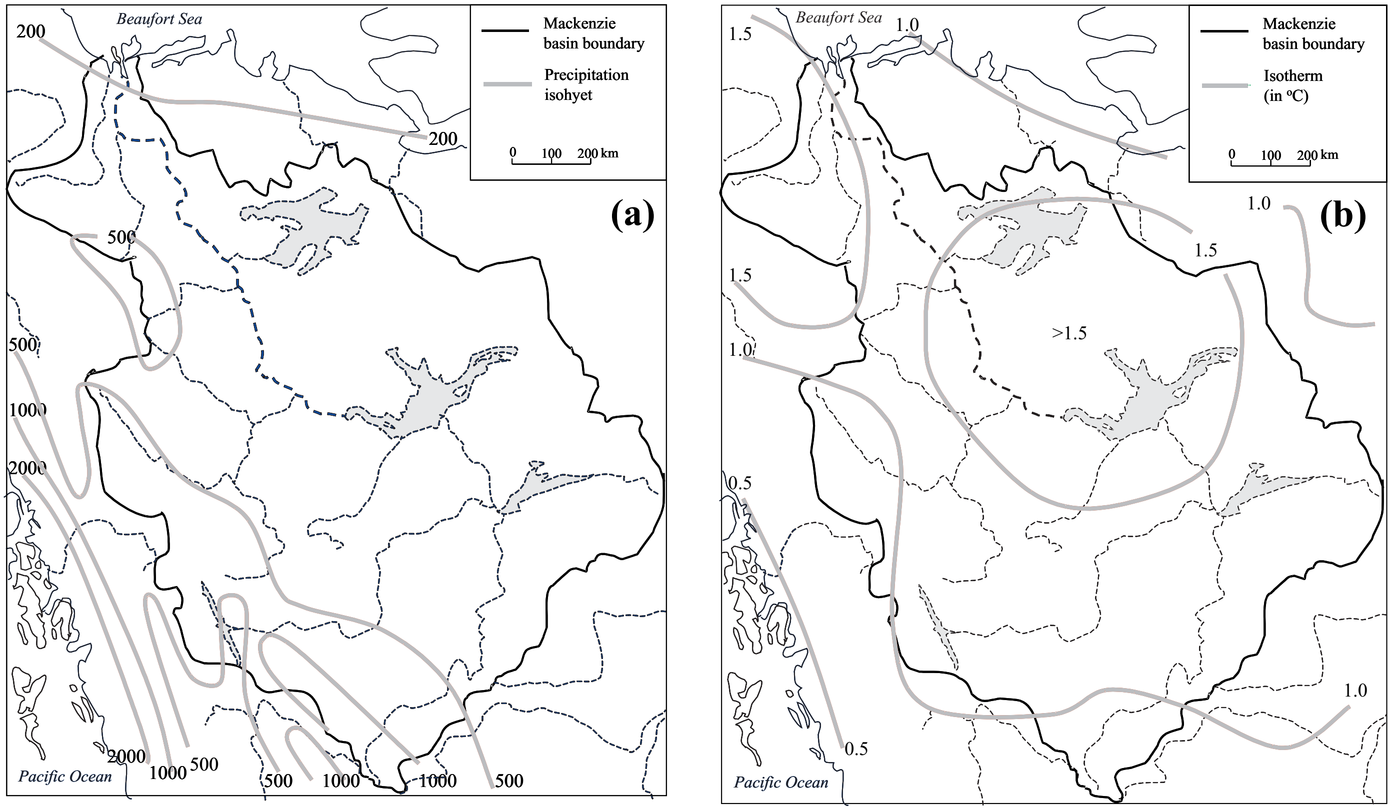

FIG. 2. (a) Annual precipitation in the Mackenzie Basin (source: Hydrological Atlas of Canada), and (b) trends in annual air temperature change between 1950 and 1998 (source: Environment Canada).

the Williston Reservoir; the Great Slave, which includes the drainage from the Canadian Shield, as well as several basins on the high plains (>1500 m); the Great Bear in the Shield region, dominated by the large Great Bear Lake; the low plains, with many basins draining wetlands, small lakes, and northern forests; the Liard, a large, mountainous basin; and the northern mountains, with a collection of smaller catchments in a subarctic, subalpine setting.

Streamflow, reported by HYDAT in m^3/s, is divided by basin area and expressed as specific discharge in $m^3/s/km^2$ or as runoff in mm of water. A water year is considered to span from October to September of the following year, and dates are given as Julian days (Day 1 = 1 January). This study employs annual and monthly data, together with annual maximum daily values from the major sub-basins. Only four major catchments provide flow data at their outlets: the Peace River at Peace Point, the Athabasca River below Fort McMurray, the Liard River near its mouth, and the Great Bear River at the outlet of Great Bear Lake. Monthly and annual discharges for the Great Slave region, the low plains, and the mountains north of the Liard have to be estimated from the records of their sub-basins. Discharge from the Great Slave region is obtained by subtracting from the Mackenzie flow at Fort Simpson the measured flows of the Liard and Slave Rivers and the calculated flows of several rivers on the low plains.

To estimate the flow for the ungauged basins in the mountains and on the low plains, regression relationships were obtained between the basin area and the flow of the gauged rivers with at least seven years of record between 1968 and 1999. The derived empirical equations for glacierized basins are $q = C - Da$ for basins with areas of 15 000 – 30 000 km^2 and $q = Fa^{-G}$ for basins under 15 000 km^2, where q is specific discharge in $m^3/s/km^2$ and a is basin area in km^2. The corresponding equations for low plain basins are $q = Fa^{-G}$ (15 000–30 000 km^2) and $q = C - Da$ (under 15 000 km^2). The numerical values of the regressed coefficients C, D, F, and G are given in Table 1. For individual water years, the parameters were re-estimated (Table 1) using the flows of the gauged rivers for those particular years.

To examine possible trends in the data, we prepared time series for several characteristics of streamflow (annual and monthly discharges, magnitude and timing of peak flow) and climatic variables (monthly temperature and precipitation). These time series were analyzed for trends using the nonparametric Spearman's correlation technique (Appendix 1), which is not affected by the distribution of the hydrological and the climatic data. Although the Mann-Kendall test has recently become popular for many trend studies (e.g., Hirsch et al., 1982), Yue et al. (2002) found that the Mann-Kendall test and Spearman's rank correlation provide almost identical results. The latter is used here because most researchers are familiar with the meaning of the r-value and because Yue et al.'s (2002) study suggests that when a trend exists, the

TABLE 1. Empirical coefficients used in the Data and Methods section to estimate specific discharge for ungauged basins in the mountains and on the low plains.

Average Coefficients[1]	1968–99	1979–80	1987–88
Glacierized basins:			
C	0.2 (r = 0.74, n = 14)	0.02 (r = 0.78, n = 13)	0.02 (r = 0.35, n = 12)
D	-7×10^{-7} (r = 0.74, n = 14)	-7×10^{-7} (r = 0.78, n = 13)	-3×10^{-7} (r = 0.35, n = 12)
F	0.2 (r = 0.87, n = 14)	0.9 (r = 0.91, n = 13)	0.04 (r = 0.51, n = 12)
G	-0.3 (r = 0.87, n = 14)	-0.3 (r = 0.91, n = 13)	-0.1 (r = 0.51, n = 12)
Low plain basins:			
C	0.01 (r = 0.90, n = 15)	0.004 (r = 0.80, n = 14)	0.008 (r = 0.76, n = 14)
D	-2×10^{-7} (r = 0.90, n = 15)	-2×10^{-7} (r = 0.80, n = 14)	-2×10^{-7} (r = 0.76, n = 14)
F	0.2 (r = 0.86, n = 15)	0.5 (r = 0.89, n = 14)	0.3 (r = 0.77, n = 14)
G	-0.5 (r = 0.86, n = 15)	-0.6 (r = 0.89, n = 14)	-0.5 (r = 0.77, n = 14)

[1] r is correlation coefficient; n is number of basins.

power of the Mann-Kendall test is dependent on the distribution type. We estimated the slope of statistically significant trends using the method of Yue et al. (2002), shown in the Appendix.

STREAMFLOW REGIMES

A streamflow regime is the average pattern of seasonal variation in streamflow. Streamflow is influenced by water supply (e.g., snowmelt, rainfall, glacier melt), water losses (e.g., evaporation) and storage modifications (by lakes, wetlands, reservoirs, and groundwater). By interpreting the shape of the hydrographs, guided by physical considerations, we can recognize several principal regime types within the Mackenzie system. We first analyzed the hydrographs of the medium-sized (≤ 20 000 km^2) basins, obtained by calculating the mean, standard deviation, and highest and lowest recorded flows of every calendar day.

The principal seasonal flow pattern exhibited by most rivers is the *subarctic nival regime*, in which snowmelt, often accompanied by river ice breakup, generates high flows (Church, 1974). The nival regime is modified where additional sources of water supply (e.g., glacier meltwater) or significant storage mechanisms (e.g., lakes) alter the streamflow pattern (Woo, 2000). Most of the rivers in the southern basin and at low altitudes peak in early May, but in rivers at higher latitudes and high altitudes, where snowmelt is delayed, spring peaks occur later (e.g., late May for the Ogilvie River in Fig. 3). In glacierized basins, the ablation of glaciers intensifies in the summer and this, together with snowmelt at high elevations, prolongs the high flows into summer (a *proglacial regime*, e.g., the Athabasca River near Jasper). For some basins, autumn rainfall can give rise to secondary peaks that are lower in magnitude than the spring flood (e.g., Arctic Red River, Fig. 3).

Wetlands have little effect in modifying the spring high flows because of their low storage capacity when frozen (Woo, 1988), and therefore rivers with a *wetland regime* show prominent snowmelt peaks (e.g., the Little Buffalo River in Fig. 3). After the ground thaws, however, the unfrozen soil in the wetlands have an increased capacity to retain water and to retard the summer flows, leaving a large amount of moisture available at or near the ground surface to support evaporation. Large lakes are highly effective in providing large storage capacities to reduce the high flows and to extend the low flows. Thus, basins with a *prolacustrine regime* tend to have fairly even runoff during the year (e.g., the Lesser Slave River, Fig. 3). When the flow is modified by reservoir operation to generate hydroelectric power, as is the case of the Peace River at Hudson Hope, the natural flow regime is strongly altered (Peters and Prowse, 2001), though the total annual flow volume may not be seriously modified.

Rivers from the major sub-basins combine the flow regimes of their tributaries. The Liard exhibits a nival regime (Fig. 4), but it attains its peak more gradually than the medium-size basins (cf. Ogilvie). This is attributed to its large elevation range, which causes an extended melt contribution period as snowmelt progresses from the valleys to the mountain tops. The Athabasca below Fort McMurray (Fig. 4) has an early hydrograph rise that is due to snowmelt in the lowlands and a summer peak, possibly sustained by glacier and high-elevation snowmelt in its headwater areas (cf. Athabasca at Jasper, Fig. 3). The Peace River at Peace Point combines the regulated flow effect of the Williston Reservoir at Bennett Dam (cf. Peace at Hudson Hope in Fig. 5) with the nival regime runoff of its tributaries downstream of the dam. A comparison of the pre- and post-damming periods shows that reservoir operation maintains a large winter flow but eliminates the spring peak and reduces the summer discharge (cf. Peace at Hudson Hope). This effectively reduces the snowmelt and summer flows at Peace Point, which receives its spring freshet and rainfall-induced high flows only from the tributaries downstream of the dam (Fig. 5). The reservoir operations may substantially influence the regime of water level fluctuations in the Great Slave Lake. Kerr (1997) noted that since the dam construction, the long-term mean amplitude of the lake has declined and its high water level period occurs sooner. In subsequent analysis, we consider

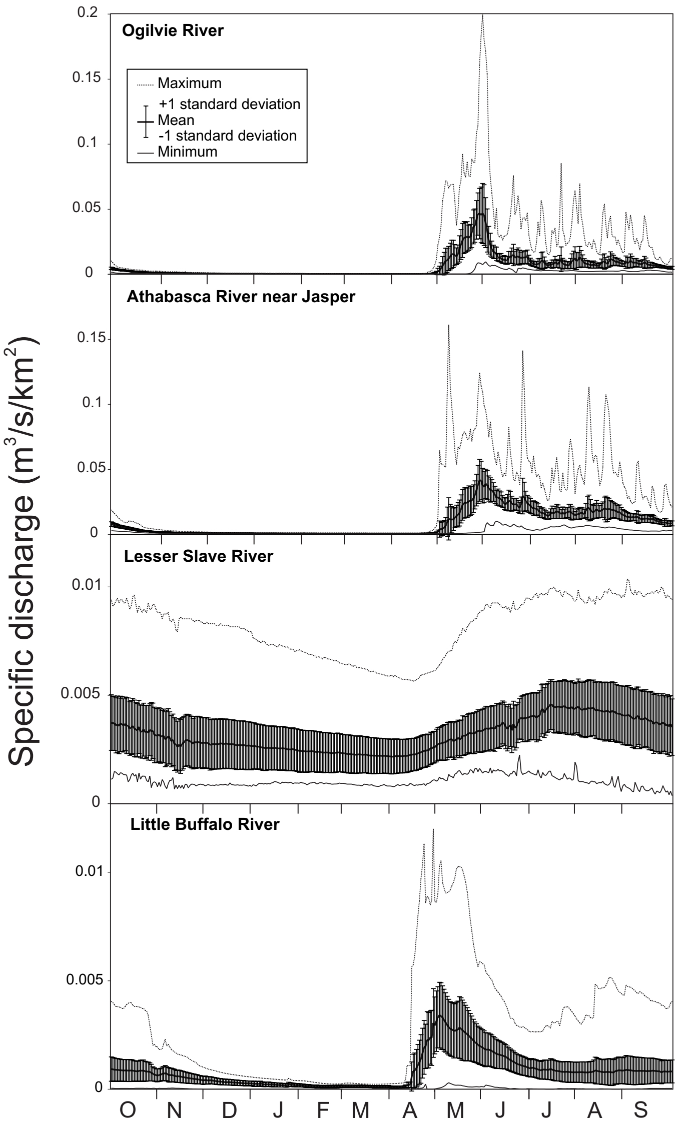

FIG. 3. Streamflow regimes, indicated by the average daily discharges, of rivers draining medium-size basins: nival regime (Ogilvie River), proglacial regime (Athabasca River near Jasper), prolacustrine regime (Lesser Slave River), and wetland regime (Little Buffalo River).

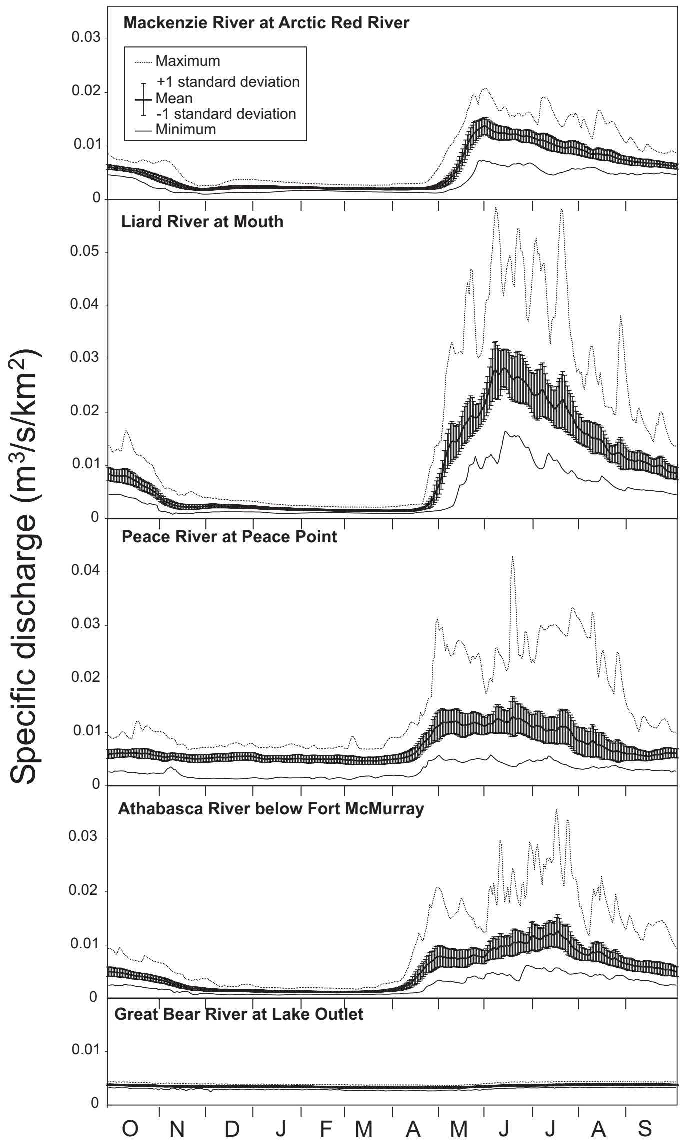

FIG. 4. Regime of the Mackenzie River at the Arctic Red River station and regimes of its major sub-basins: the Liard River at its mouth, the Peace River at Peace Point, the Athabasca River below Fort McMurray, and the Great Bear River at the outlet of Great Bear Lake.

only the period during which the Peace River has been regulated by reservoir operations.

The flow of the Mackenzie River at Arctic Red River combines the regimes of its sub-basins; in addition, the large basin size has a moderating effect that smoothes out the minor fluctuations, leaving a nival regime hydrograph that is dominated by peak flow in the snowmelt period followed by declining flows in the summer and low flow in the winter (Fig. 4).

SUB-BASIN FLOW CONTRIBUTIONS

Different sub-basins play varying roles in terms of flow contribution to the Mackenzie. On a per unit area basis, the headwater catchments of the glacierized basins yield the highest flow, followed by the non-glacierized basins in the mountainous areas. Both the Shield and the plains have low runoff. This accounts for the large flow from the mountainous sub-basins of the North, the Liard and the Peace, and the low flows from the eastern sub-basins of the Shield, the plains, the Great Bear, and the Great Slave areas (Table 2). The Athabasca has intermediate flow values, as it has mountainous headwaters combined with high plains and Shield provinces in its lower course.

In terms of the percentage contribution of a sub-basin to the Mackenzie system, both the runoff intensity and the sub-basin area have to be considered: ($100 \times qa/QA$), where q and Q are runoff (in mm, which is equivalent to specific discharge except for the difference in units) from the sub-basin and from the Mackenzie at Arctic Red River, while a and A are their respective drainage areas. The long-term mean contributions are given in Table 2, and the

TABLE 2. Sub-basin annual runoff (in mm) and flow contribution to the Mackenzie system (%, in parentheses) for an average water year, a high-flow water year (1987–88) and a low-flow water year (1979–80) of the Mackenzie at Arctic Red River.

Basins	Basin area (km^2)	Mean runoff and % contribution	1987–88 runoff and % contribution	1979–80 runoff and % contribution
N. Mountains	112 037	307 (10)	334 (11)	225 (10)
Liard	275 000	279 (27)	401 (34)	220 (25)
Peace	293 000	223 (23)	253 (23)	164 (20)
Athabasca	307 000	159 (17)	114 (11)	131 (17)
Great Bear	145 000	114 (6)	107 (5)	101 (6)
Low Plains	138 452	104 (6)	152 (7)	51 (3)
Great Slave	404 470	103 (14)	43 (5)	63 (10)
Mackenzie at Arctic Red River	1 680 000	169 (103)	195 (96)	114 (91)

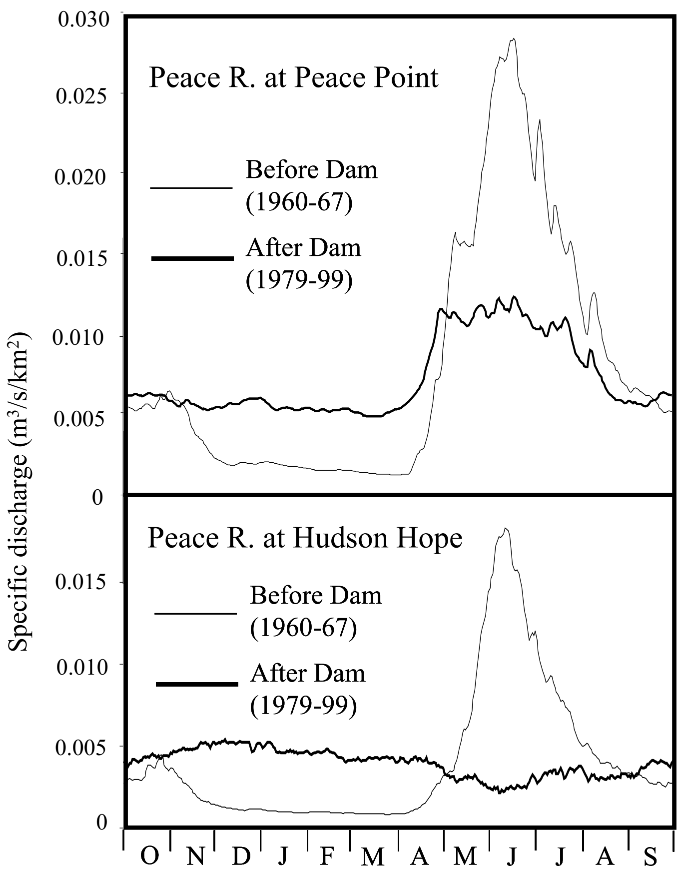

FIG. 5. Mean daily flows of the Peace River at Hudson Hope and Peace Point, before and after the construction of the Bennett Dam. The after-dam lines show the elimination of spring freshet (but augmentation of winter discharge) at Hudson Hope due to reservoir operation and a reduction in spring peak flow at Peace Point, which now receives snowmelt high flows only from the tributaries downstream of the dam.

monthly portions of contribution are depicted in Figure 6a. In spring, much of the flow is contributed by the southern basins with early melt. In summer, the Liard basin is the main flow contributor. During autumn, the central parts of the basin yield the majority of the Mackenzie discharge. In winter, the Athabasca and the regulated Peace River sustain much of the low flow.

The total annual flow from the main sub-basins (i.e., the sum of qa for all sub-basins) is remarkably close (within 5%) to the annual flow of the Mackenzie River at Arctic Red River (or QA). On a monthly basis, however, the Mackenzie River has lower flow than the combined sub-basin discharge in May, but higher flow in summer (Fig. 6a). Possible reasons include (1) errors in discharge measurement and calculations (the former are particularly likely during the spring breakup period, when ice jams interrupt regular gauging procedures); (2) temporary storage effects due to the river ice, in which the ice-induced hydraulic storage can represent 15–19% of the spring freshet volume (Prowse and Carter, 2002); (3) storage along the channel and in the riparian zones during the spring season, such as ponding and ice jam flooding; and (4) groundwater that flows directly into the Mackenzie, but is not included in the gauged surface flows.

To examine the deviations from the mean conditions, we analyzed sub-basin flow contributions during a high-flow year and a low-flow year of the Mackenzie River. A review of the discharge record reveals that data are frequently missing for many gauging sites. Only between 1979 and 1989 were complete records available for the five major basins and for at least 15 of the smaller sub-basins. Indeed, the hydrometric network of northern Canada was limited before the 1970s, and it has suffered serious attrition in recent decades (cf. Shiklomanov et al., 2002). Consequently, we had to select an above-average flow year (1987–88) and a below-average flow year (1979–80) from this period of record. Using calculations based on the discharge of the Mackenzie River at the Arctic Red River station, the flow exceedance probabilities are 0.07 for 1979–80 and 0.9 for 1987–88. For these two water-years, the flow contributions from the major sub-basins to the Mackenzie system are given in Table 1. The occurrences of negative flows from the Great Slave catchment are possibly an artifact of flow calculation, since its flows are obtained as the difference between the measured flow of the Mackenzie at Fort Simpson and discharges from other sub-basins (see Methods section). This possible error is confined to the winter period, when under-ice discharge values are known to have limited accuracy. In any case, the values are small: the differences between Mackenzie discharges and total sub-basin discharges for the entire year remain below 10%.

In l987–88, the year with above-average flow for the Mackenzie, the mountainous catchments all yielded above-average runoff, but all other catchments except the low

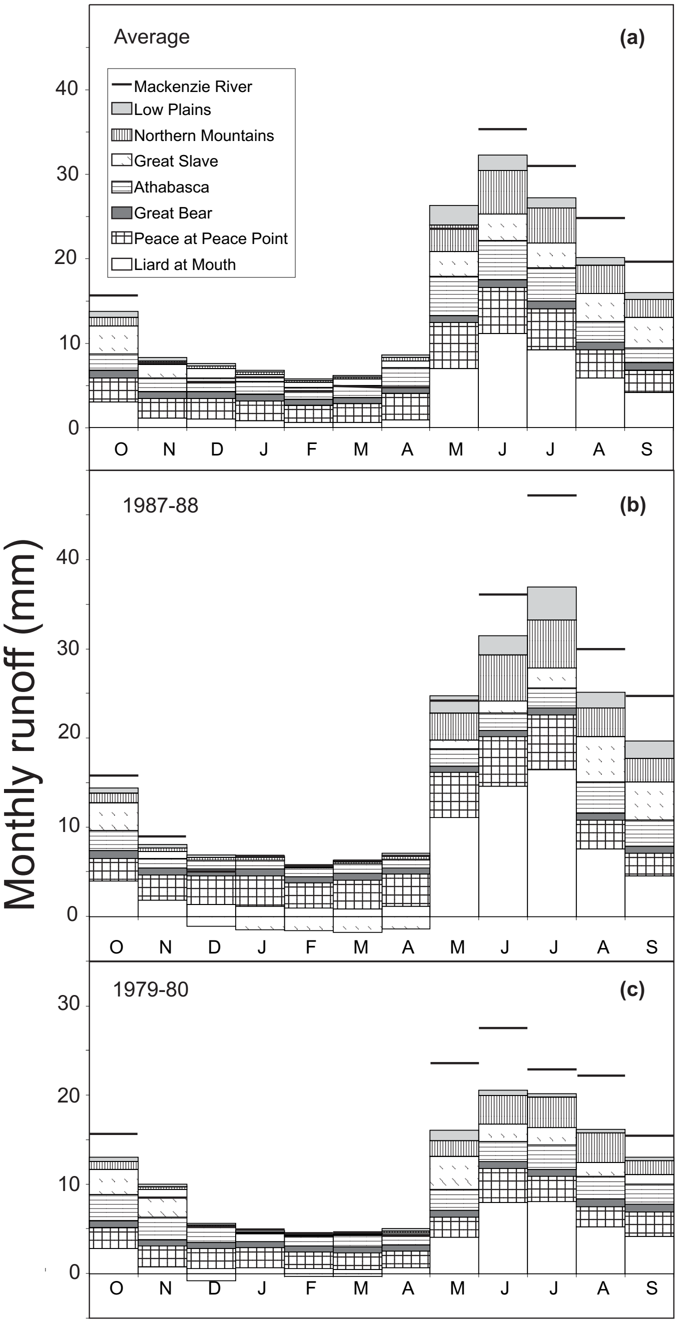

FIG. 6. Monthly runoff contribution from the main sub-basins of the Mackenzie and the runoff of the Mackenzie at Arctic Red River (thick horizontal bar), for (a) the average of water years 1967–99, (b) the high flow water year 1987–88, and (c) the low flow water year 1979–80.

plains produced below-average runoff. The monthly values (Fig. 6b) further indicate that the Peace provided more winter releases than normal, but the other mountainous sub-basins generated higher-than-normal summer flows, likely in response to high precipitation (above-average precipitation for that year was reported for Watson Lake, but not for Norman Wells or Fort Smith, located east of the mountains). The Liard was the major flow contributor, accounting for about one-third of the total Mackenzie discharge. The annual peak of the Mackenzie, like that of the Liard, was delayed until July in 1988.

In 1979–80, the below-average flow year for the Mackenzie, runoff from all sub-basins was reduced, though the proportion of sub-basin contributions to the Mackenzie remained similar to that of the mean situation (Fig. 6c). However, the sum of all sub-basin flows added up to only 91% of the Mackenzie discharge (Table 2), and this discrepancy may result either from a combination of discharge gauging errors, or from groundwater inputs to the Mackenzie not measured in the surface flows.

The three examples above demonstrate that the mountain sub-basins are the principal contributors to the Mackenzie under all flow conditions. Where lakes prevail (e.g., Great Bear basin), fluctuations in runoff contribution are dampened by lake storage, but the lake catchments in the east have low precipitation and provide limited portions of the total Mackenzie flow. The mountainous sub-basins can experience large fluctuations in precipitation and therefore wide variations in runoff. It follows that any change in the spring high flows in the mountainous region will directly affect the timing and the rate of seasonal freshwater delivery through the Mackenzie River to the Beaufort Sea.

FLOWS AND LAKE LEVELS OF THE GREAT SLAVE SYSTEM: A CASE STUDY

To examine in detail how natural and artificial forcing and storage influence the discharge and lake level fluctuation patterns, we analyzed the 1995–99 variations of the Great Slave system. In mid-1996, the highest artificial release of flow on record for the Peace River was discharged at Bennett Dam, sustaining considerable summer flows throughout the Peace-Slave drainage (Fig. 7). At the same time, the summer high-flow regime of the Athabasca River raised the level of Lake Athabasca. Large inflow from the Peace River, together with storage release from Lake Athabasca, produced high flows in the Slave River that subsequently caused the level of Great Slave Lake to rise. The peak flow for the Mackenzie was delayed by Great Slave Lake storage, though the recession flow was interrupted by the freeze-up. In the following spring, snowmelt-related high flows were generated in the Athabasca basin and in the Wabasca and Smoky River catchments downstream of Hudson Hope (indicated by the records of high flow at Peace Point, but not at Hudson Hope), causing a rise in the level of Lake Athabasca. The high flow cascaded down the Slave River to Great Slave Lake and contributed to the large discharge of the Mackenzie. It is of interest to note that this large Mackenzie discharge was detected in 1997, as a freshwater plume advected to the Beaufort Sea, by the oceanographers of the SHEBA (Surface Heat Budget of the Arctic) project (Macdonald et al., 1999). Streamflow modification caused by human activities has also been reported for the major Russian rivers that enter the Arctic Ocean (Ye et al., 2003). Regime modifications, like those of the Russian

TABLE 3. Spearman's correlation coefficients (r-values) of temperature and precipitation characteristics versus time (year), for selected weather stations in and near the Mackenzie Basin.

Stations	Years of record	Day of first temperature rise above 0˚C	April temperature	May temperature	June temperature	Annual precipitation
Fort McMurray	51	-0.39**	0.33*	0.20	0.35*	-0.09
Fort Smith	51	-0.16	0.39**	0.20	0.31*	0.14
Wrigley	42	-0.29	0.37**	0.22	0.32*	0.11
Norman Wells	51	-0.18	0.34*	0.31*	0.46**	-0.40**
Fort St. John	51	-0.35*	0.41**	0.02	0.19	-0.24
Watson Lake	48	0.01	0.27	0.06	0.04	-0.23
Dease Lake	50	-0.11	0.40**	0.16	0.17	0.22
Whitehorse[1]	48	-0.37*	0.33*	0.07	0.02	0.10
Mayo[1]	49	-0.42**	0.43**	0.28*	0.32*	0.07

* Correlation significant at 0.90 probability; ** correlation significant at 0.95 probability.
[1] Station outside the Mackenzie Basin.

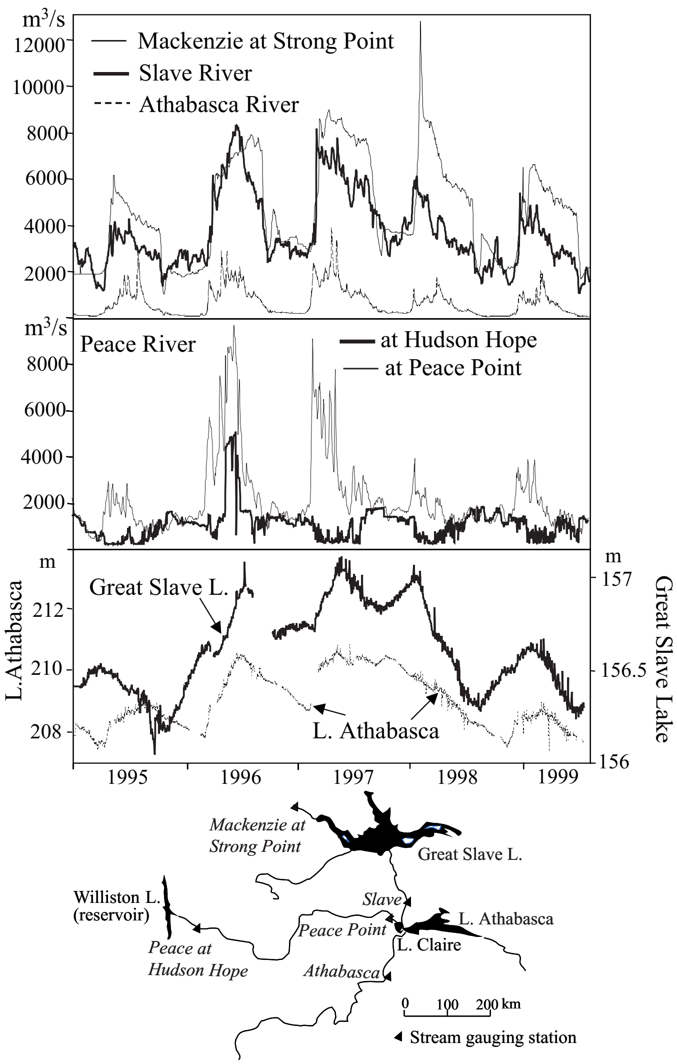

FIG. 7. 1995–99 daily streamflows and lake levels in the Great Slave system. Inset map shows the location of the gauging stations.

rivers and the Mackenzie, can have impacts on oceanic circulation.

The spring of 1998 yielded low snowmelt runoff from the Peace and Athabasca Rivers, but these flows were superimposed on the storage release from Lake Athabasca, so that the Slave River maintained a moderate inflow to Great Slave Lake. This elevated Great Slave Lake to another high level and its outflow at Strong Point attained the highest recorded peak (for the 1996–98 period). The 1997–98 levels of Lake Athabasca and Great Slave Lake also reached their record highs. However, after one long open-water season with exceptionally high evaporation—Rouse (2000) calculated the loss to be 506 mm between 3 July 1998 and 8 January 1999, a value that is about 100 mm higher than average—the lake level fell and the non-exceedance probabilities for the two lakes were only 0.1 and 0.2, respectively. Reflecting the reduction in lake outflow, the discharges of the Slave and the Mackenzie at Strong Point declined accordingly.

This example demonstrates several features of flow generation within the Great Slave system, which constitutes half of the entire Mackenzie catchment. (1) This drainage system receives important runoff contributions from both natural and regulated water sources. (2) Given the vastness of the region, the timing and the magnitude of water supply can be highly variable in space, as evidenced by the 1997 spring flows of the Peace River. (3) Lake levels rise in response to inflows, but can be lowered substantially by evaporation losses, as was the case in 1998. (4) The large lakes of the Slave system offer textbook examples of lake storage effect, delaying the high flow and extending the recession flows downstream. (5) A combination of moderate inflow and storage release from the lake can generate high outflows. (6) Lake levels and streamflows of the Great Slave system can vary greatly from year to year (cf. 1995 and the subsequent three years), and interpretation of the 1995–99 data enables unraveling of the major mechanisms that cause such large variability.

TRENDS AND VARIABILITY IN STREAMFLOW

Records of air temperature in the Mackenzie Basin show an average increase of over 1.5˚C for the period 1950–98

TABLE 4. Spearman's correlation coefficients (r-values) of annual streamflow and variability of streamflow versus time (year) for selected streamflow stations in the Mackenzie Basin.

Stations	Area (km^2)	Annual flow		Flow variability[1]	
		r	n[2]	r	n[2]
Camsel	31 100	0.26	32	0.26	23
Jean-Marie	1310	-0.04	27	0.67*	18
Willowlake	22 000	0.16	20	-0.41	11
Lockhart	26 600	0.12	31	0.84*	23
Liard	275 000	-0.16	27	-0.41	18
Peace	293 000	0.18	32	0.02	21
Athabasca	133 000	-0.31	30	0.69*	20
Mackenzie	1 680 000	-0.15	26	-0.15	16

[1] Variability is calculated as the running standard deviations for consecutive ten-year periods.
[2] n is number of paired data used in the correlation.
* Correlation significant at 0.95 probability.

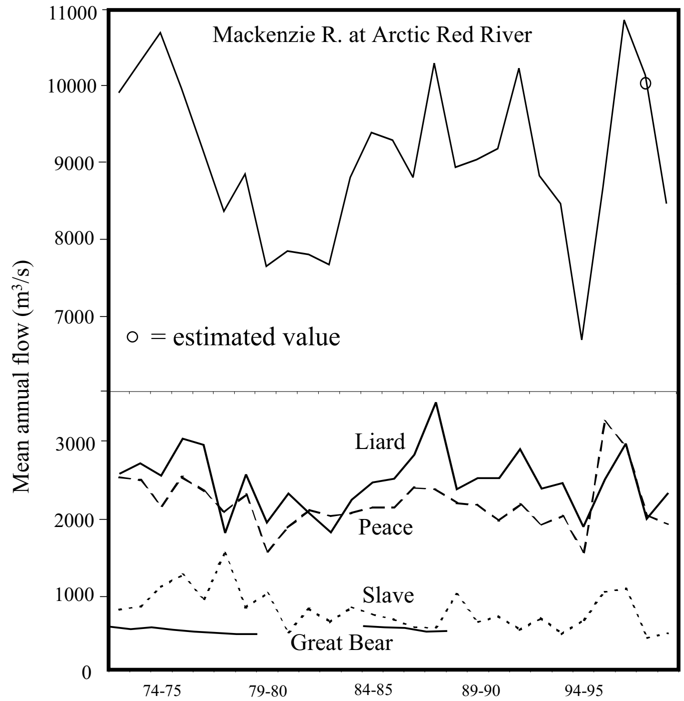

FIG. 8. Measured annual flow of the Mackenzie River and its major sub-basins (Great Bear, Liard, Peace, and Slave) during water years from 1972–73 to 1998–99.

(Zhang et al., 2000). It should be cautioned, however, that such warming "is no larger than the observed interdecadal range in high-latitude temperatures during this century" (Serreze et al., 2000:197). Furthermore, given the vastness of the study area, warming is not felt uniformly across the basin (Fig. 2b). For instance, Fort McMurray and Fort Smith in the southeast experience significant temperature increase in January (1.2 – 1.4˚C per decade), but no statistically significant increase occurs at that time for Watson Lake in the western mountains or for Norman Wells on the northern interior plain. In July, Norman Wells shows a significant temperature increase, but the other three stations do not. It is in April that most stations exhibit significant warming (Table 3), and this warming can advance the arrival of snowmelt in northern basins. Annual precipitation displays a significant decrease for Norman Wells, but this trend is not found in the other stations (Table 3). Zhang et al. (2000) suggested that there is an increase in the ratio of total snowfall to annual precipitation in the north. It is to be cautioned, however, that snowfall data are prone to measurement error (Metcalfe at al., 1994), and since most stations are located in valleys or lowlands, their data are unlikely to reflect the average condition of the entire basins (Woo et al., 1983). Nevertheless, both temperature and precipitation changes can affect the water balance through rainfall input, snow accumulation and melt, and evaporation, hence influencing the magnitude and timing of runoff. This section will examine recent changes (or lack of change) both in air temperature and in the annual, monthly, and maximum flows of the Mackenzie system.

Annual and Monthly Flows

The scarcity of long-term data precludes conclusive statements about the streamflow trends. Several investigators have made use of the Reference Hydrometric Basin Network data to study streamflow trends in Canada (Whitfield and Cannon, 2000; Zhang et al., 2001; Burn and Hag Elnur, 2002), but with limited reference to the Mackenzie Basin because of the paucity of data from the Network. The more extensive HYDAT dataset offers 27 to 32 years of record for the western basins of the Mackenzie catchment, though its data for the catchment's eastern sector are insufficient for this study (Fig. 8). Nevertheless, as the previous analysis indicates, the western regions yield the bulk of runoff for the Mackenzie, and any signals of their flow trend or flow variability will be transmitted downstream to the main trunk.

In terms of annual flows, no statistically significant trend can be detected within the record for the Mackenzie or for its main sub-basins (Table 4). A similar result was reported by Zhang et al. (2001) for selected basins in the Mackenzie system. Such findings are consistent with Serreze et al.'s (2000) conclusion that there are no apparent trends in precipitation-minus-evaporation north of 70˚N, though the Mackenzie lies considerably farther south than their study region. While the annual flows do not reveal any obvious trend (Table 4 and Fig. 8), there appears to be a recent change in the year-to-year variations in streamflow in some sub-basins. The variability of annual flows is obtained as 10-year running standard deviations in streamflow (Fig. 9). These standard deviations are regressed against time (in years). The Athabasca in the southeast and some small catchments farther north (e.g., the Jean-Marie on the interior plains and the Lockhart in the Shield) show significant increases in the variability of their annual flows (Table 4).

Annual values mask the changes in monthly flows. On a monthly basis, the Liard River and the regulated Peace

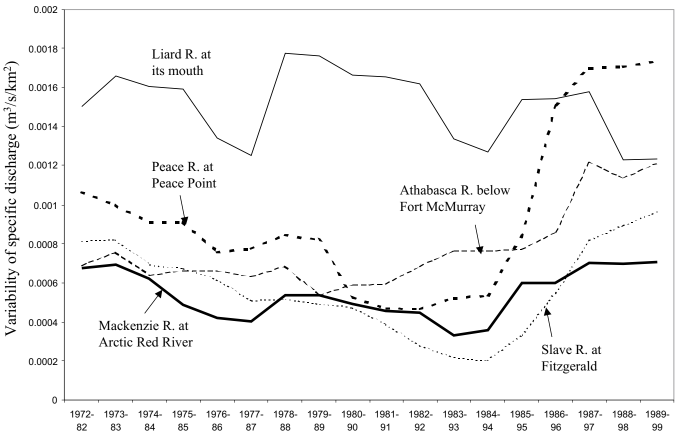

FIG. 9. Variability of annual flows, represented by the running standard deviations of 10-year periods, for the Mackenzie and its major sub-basins.

River show increases for December and April only (Table 5a). In terms of variability, the Peace displays increased variations in its April, June, and autumn (September to November) discharges (r-values in Table 5b). This phenomenon may be related to streamflow regulation for hydroelectric power generation. The Athabasca River experiences a significant increase in flow variability in March and August, and a significant decrease in September. The Liard reveals a significant increase in flow variability in April and September, but a reduction in variability during the intervening summer months of July and August. A general reduction in flow variability at all the major sub-basins in May is followed by an increase in June. This tendency is also reflected in the Mackenzie discharge (Table 5b). On the other hand, the Mackenzie shows a general reduction in flow variability in its October to December flows, and this pattern is largely contrary to the variability trends of its major sub-basins. These discharge trends cannot be explained easily because of differential river ice growth rates (Prowse and Carter, 2002) and because flow data obtained during the ice-covered period are notoriously unreliable.

Annual Peak Flows (Annual Floods)

An annual flood is defined as the highest flow that occurs in a water year. For the Mackenzie system, these extreme high flow events are important because they often cause flooding and river ice damage to the riverside communities, interrupt traffic, and alter the morphology and ecology of the channel and river plain, as well as injecting large quantities of freshwater into the sea within a period of days. Three attributes of the flood can be derived from the daily streamflow data for this study: the first day of major hydrograph rise (defined as the day when discharge is double that of the previous day), the day when the annual peak occurs, and the magnitude of this peak.

Climatic warming can advance the dates of snowmelt initiation and snow cover depletion (Kane et al., 1992), and any such effects on streamflow will be best detected during the spring high flow period. Given the vastness and the topographic contrasts of the Mackenzie Basin, it is difficult to identify the beginning of the melt period for particular sub-basins. For generalized purposes, we make use of atmospheric warming as a surrogate indicator of the initiation of snowmelt. We infer that the arrival of the melt season will follow three consecutive days of air temperatures above 0˚C. Although this does not take account of the energy balance or the ripening of the snow pack, air temperature remains as the only climatic indicator that is measured at different parts of the basin. Using such information, it is suggested that spring warming occurs early in the southern zones (e.g., in early April in northern Alberta, around 10 April on the high plains and in the Great Slave Lake vicinity, and after 20 April on the lowlands north of Fort Simpson). The southern sector (Fort McMurray and Fort St. John) and the western sector (Mayo Landing and Whitehorse, immediately west of the Liard drainage) show significant trends of earlier arrival of air temperatures above 0˚C (Table 3), though most other stations have a weak tendency toward early warming that is not statistically significant. For Fort St. John and Fort McMurray, the arrival of above-freezing conditions has advanced by 3.3 and 3.8 days/decade, and it advanced by 3.2–4 days/decade at Mayo and Whitehorse, but for the other sites, the change was around 1 day/decade. In addition to an earlier rise above the freezing point, there has been a general increase in the air temperature for April (positive r-values in Table 3). Climatic stations in the Mackenzie Basin, from Inuvik in the north, through Yellowknife, to Fort McMurray in the south, show statistically significant warming trends of 0.9˚, 0.7˚, and 0.5˚C per decade, respectively (Fig. 10). Such trends can cause an early warming of the snow cover in the north and advance the melt date in the south. Following melt initiation, the intensity of melt is another consideration in the production of meltwater runoff. The northern sector and Mayo have become warmer in May, and this tendency persists in June (Table 3). These results are in accord with Whitfield and Cannon (2000) and Zhang et al.'s (2000) finding of a general increase in spring temperatures for the region.

A large influx of meltwater runoff often induces sharp hydrograph rises. The Mackenzie does show statistically significant earlier occurrence of sharp hydrograph rises along its main trunk from the Slave at Fitzgerald to Arctic Red River. Between 1973 and 1999, the date of spring hydrograph rise for the Mackenzie at Arctic Red River advanced by three days per decade (Fig. 11) and that for the Slave by five days per decade. A weak and statistically insignificant trend of earlier hydrograph rise (by about three days per decade) is exhibited by the Liard and the Peace (Table 6). If sharp hydrograph rise indicates spring breakup, the evidence from most parts of the Mackenzie system agrees with the breakup trends reported for a number of lakes and rivers in the Northern Hemisphere by Magnuson et al. (2000), even though different definitions for breakup are used. Our definition suggests that the

TABLE 5. Spearman's correlation coefficients (r-values) for (a) monthly streamflow versus time (years) and (b) flow variability versus time (years) for the Mackenzie River and its major tributaries, calculated from the 1972–99 records.

Station	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.
(a) Monthly Flow												
Liard	0.36	0.31	0.24	0.41*	0.08	-0.17	-0.23	-0.06	-0.11	0.16	0.08	0.46**
Peace	0.14	0.14	0.14	0.36	-0.22	-0.21	-0.26	-0.27	-0.31	-0.26	-0.15	0.27
Athabasca	-0.38*	-0.41*	-0.18	0	-0.25	-0.11	-0.14	-0.16	-0.50**	-0.34	-0.33	-0.31
Mackenzie	0.25	0.23	0.20	0.22	0.15	-0.29	-0.28	-0.19	-0.10	-0.26	-0.38*	0.07
b) Variability of Monthly Flow												
Liard	-0.33	-0.32	-0.37	0.80**	-0.11	0.39	-0.54*	-0.59**	0.72**	-0.20	0.25	0.17
Peace	-0.30	-0.27	0.34	0.79**	-0.23	0.84**	0.17	0.36	0.62**	0.65**	0.82**	0.20
Athabasca	0.01	0.02	0.86**	0.24	-0.45*	0.35	0.32	0.94**	-0.51*	0.37	0.27	-0.31
Mackenzie	0.03	0.12	0.06	-0.83**	-0.80**	0.79**	0.29	0.09	0.27	-0.18	-0.04	-0.74**

* Correlation significant at 0.90 probability; ** correlation significant at 0.95 probability.

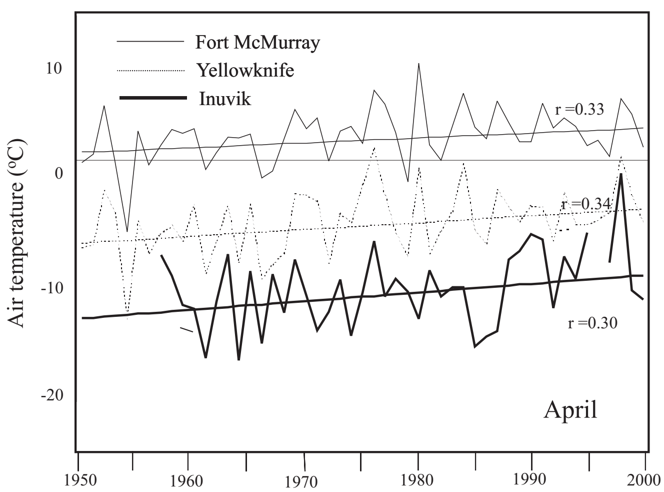

FIG. 10. Mean April air temperatures for Inuvik, Yellowknife, and Fort McMurray. Fitted linear trend lines show warming in recent decades. Spearman's correlation coefficients are also given.

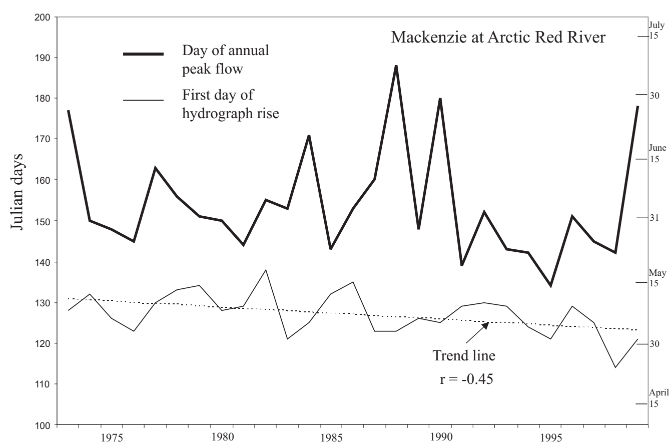

FIG. 11. First day of prominent hydrograph rise (at least double the discharge of the previous day) and day of annual peak flow for the Mackenzie River at Arctic Red River station, 1973–99.

beginning of breakup corresponds with a sharp rise in the hydrograph, while Magnuson et al. (2000:1743) focus on the end of the process, defining "breakup" as "the date of the last breakup observed before the summer open water phase." They caution, however, that the breakup date is related to a multiplicity of factors, being "strongly influenced by the timing, magnitude, and rate of spring runoff as well as by the nature of the freezing process and ice stratigraphy."

The mean date of annual peak occurrence for the Mackenzie is 3 June at Arctic Red River (Fig. 11), but it is 7 June at Norman Wells and 19 June at Fort Simpson. The fact that the lower course of the river peaks earlier than its upstream sections suggests that the local and not the basin-wide spring melt inflows give rise to the peak flows along the river. This is understandable, since the occurrence of annual peaks is complicated by several considerations besides spring snowmelt. The Liard basin traverses a large altitudinal range of 2700 m, so that the snow at the higher elevations may still be melting while the lowland snow cover has long disappeared. High flows are thus prolonged, and the arrival of the annual peak is delayed until mid-June (mean date: 18 June). The Athabasca, fed by glacier melt and snowmelt at high altitudes, does not peak until July (mean date: 5 July). No statistically significant trends can be discerned in the date or the magnitude of the annual peaks that occurred in the Mackenzie and its major sub-basins (Table 6). However, the lower Mackenzie and the Peace show trends toward increasing variability in both annual peak date and peak discharge. For the Peace, the dual influence of natural forcing and flow regulation of Williston Reservoir (Peters and Prowse, 2001) can account for these trends.

DISCUSSION AND CONCLUSIONS

The flow of the Mackenzie River reflects the contributions from its major sub-basins at different times of the year. The seasonal flow exhibits essentially a subarctic nival regime: high flows that occur during the snowmelt and river ice breakup period are followed by a steady decline, sometimes raised by summer and autumn rain events, until the winter, when low flow prevails. This

TABLE 6. Timing and magnitude of annual peak flow, correlated with year, for selected streamflow stations during the period 1973–99. Unless otherwise indicated, all values are Spearman's correlation coefficients (r-values).

Stations	First day of hydrograph rise[1] vs. year	Trend (no. of days per decade)	Day of annual peak flow vs. year	Variation of annual peak day vs. year	Magnitude of annual peak[2] vs. year	Variation of annual peak magnitude[2] vs. year
Liard at mouth, Fort Simpson	-0.34	earlier (1.8)	-0.15	0.20	-0.35	-0.26
Peace at Peace Point	-0.31[3]	earlier (3.3)	-0.32	0.63**	-0.21	0.95**
Athabasca at Fort McMurray	0.03	no change	-0.02	-0.41	-0.19	0.09
Slave at Fitzgerald	-0.41*	earlier (4.8)	-0.28	0.03	-0.28	0.33
Mackenzie at Fort Simpson	-0.44*	earlier (2.6)	-0.14	0.62**	-0.26	-0.36
Mackenzie at Norman Wells[4]	-0.52*	earlier (3.6)	0	0.76**	-0.11	0.85**
Mackenzie at Arctic Red River	-0.38*	earlier (2.7)	-0.24	0.77**	-0.23	0.94**

[1] First day of hydrograph rise is the first day in a year when discharge is double that of the previous day.
[2] Variability is calculated as the running standard deviations for consecutive ten-year periods.
[3] The first of day of hydrograph rise for Peace River cannot be defined with certainty, as the flow increase is often gradual.
[4] Only 22 years of record are available.
* Correlation significant at 0.90 probability; ** correlation significant at 0.95 probability.

regime is manifested in most of the sub-basins, though there are modifications, including the proglacial regime with summer high flows sustained by glacier melting and the wetland regime whereby summer runoff is reduced through storage and enhanced evaporation losses. The prolacustrine regime provides a strong contrast to the nival pattern, in that large lake storage strongly attenuates high flows and maintains steady runoff during the winter. Human interference also modifies the natural flows, so that the highly regulated Peace River has reduced seasonal flow variations due to hydroelectric power generation. Sub-basins with large natural lakes (e.g., Great Slave, Great Bear) or artificial lakes (e.g., Williston Reservoir on the Peace River) support most of the winter discharge for the Mackenzie River.

Within the Mackenzie Basin, a northward decrease in runoff reflects the spatial trend in precipitation, which shows a decline towards the northern plains and the Shield areas. The mountainous basins generate the highest annual discharges in the Mackenzie Basin and their flows display the largest annual variations. On the other hand, basins in the Shield and on the low plains have the lowest discharge, but they still have large seasonal variations. Large lakes even out the seasonal flow variations, but the lake catchments contribute a low percentage of the total Mackenzie flow. Sub-basins in the mountainous region, such as the Peace, the Liard and the rivers from the northern mountains, constitute only 40% of the drainage area, but they usually yield over half of the Mackenzie flow. They have overwhelming influence on the Mackenzie discharge.

Although the available streamflow series are too short to warrant definite statements regarding the sensitivity of the Mackenzie system to the climatic warming suggested by the temperature records of the past decades, it is worthwhile for predictive or planning purposes to recognize the variations in the magnitude and the timing of flow contributions from the major sub-basins. Despite the warming signal revealed by the air temperature records, streamflow of the Mackenzie system does not indicate any obvious trend at either an annual or a monthly time scale. On the other hand, significant changes in the flow variability have occurred for several rivers in different months. There is also evidence of an earlier breakup in the past few decades, and this may be related to the increasing air temperature trends for the snowmelt months of April to June. However, the date and the magnitude of peak flow show no trend. Of interest is that the date and the magnitude of peak flow have become more variable for the lower Mackenzie and the Peace, the latter river having been regulated by the operations of the Williston Reservoir. Streamflow in the past three decades indicates no general trend but suggests a tendency towards greater variability in several flow characteristics. Confirmation of the Mackenzie drainage system's responses to the climatic warming signal awaits further investigation when longer records from a denser hydrometric network become available.

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APPENDIX

1) Calculation of Spearman's Correlation Coefficient (r):

Rank the two variables X and Y that are to be correlated, so that their ranks are represented by RX_1, RX_2, ... RX_n and RY_1, RY_2, ... RY_n, with n being the number of pairs of observations. Then, $r = 1 - [6D / (n^3 - n)]$ with $D = \Sigma(RX_j - RY_j)^2$

2) Calculation of the Slope of a Linear Trend (b):

$b = \text{median } [(x_j - x_k) / (j - k)]$ for all $k < j$ where x_j is the j^{th} observation in the time series.