Spatial and Temporal Patterns and Hydroclimatic Controls of River Ice Break-up in the Mackenzie Delta, NWT

By

Holly Lynn Goulding
B.Eng., Dalhousie University, 2006

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Abstract

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Concern has been expressed regarding the impacts of climate change on the hydroecology of the Mackenzie Delta, thus identifying a need for better understanding of the ice break-up regime. Archived records at hydrometric stations in the delta for the period 1974 to 2006, supplemented with observations and remotely sensed imagery, are used to assemble a break-up chronology and examine spatial and temporal patterns of break-up flooding. Hydroclimatic controls on break-up are assessed by statistical, qualitative, and trend analysis of upstream discharge and downstream ice characteristics. For the most severe break-up flooding, two event types are identified: ice-driven events, with high backwater and high peak levels in the southern, eastern and western delta, and discharge-driven events, with high levels in the mid and outer delta and along Middle Channel. Break-up initiation during ice (discharge) events occurs earlier (later) than the delta average. Severity of break-up water levels is most influenced by upstream discharge, while timing is related to ice conditions and spring hydrograph rise. Rapid upstream melt and lower intensity melt in the delta prior to break-up characterize the most severe events. Trend analysis reveals a tendency toward earlier break-up, a longer prebreak-up melt interval, and a lower magnitude of hydroclimatic controls.

Table of Contents

List of Tables

List of Figures

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Chapter 1: Introduction

1.1 INTRODUCTION

River ice break-up plays an important role in controlling the hydroecology of cold-regions deltas. These ice-dominated deltas (Walker, 1998) are typically characterized by a flat delta front (Are and Reimnitz, 2000) and extensive numbers of lakes influenced by ice-induced recharge. Major examples include the Peace-Athabasca (Peters and Prowse, 2006), the Slave (Brock *et al*., 2008), the Yukon (Dupre and Thompson, 1979), the Colville (Walker and McCloy, 1969), and the Mackenzie (Marsh and Hey, 1989) deltas in North America; the Kvikkjokk (Dahlskog, 1966) and Laiture (Axelsson, 1967) deltas in Sweden; as well as the Yenisey (Burdykina, 1970) and Lena (Antonov, 1969) deltas in Siberia. Often coinciding with the arrival of the spring freshet, break-up and jamming of a floating ice cover can augment water levels in delta channels already receiving large amounts of discharge from snowmelt runoff (Prowse *et al.*, 2006a). Thus, spring break-up ice jams typically produce the largest hydrologic event of the year and the flooding of lakes and riparian landscapes (*e.g.* Prowse and Lalonde, 1996; Marsh and Hey, 1989). The recharge of lakes with sediment-laden river water plays a fundamental role in influencing the aquatic ecosystem (*e.g.* Marsh *et al.*, 1994).

The timing and severity of break-up and ice-jam flooding in these deltas depends on hydroclimatic controls operating at a variety of spatial and temporal scales. Temporally, initial break-up conditions are determined by hydroclimatic conditions during the fall and winter influencing the peak ice thickness and winter snowpack. The final onset and evolution of spring break-up, however, depend on the prevailing hydrometeorology driving melt during the break-up period (Beltaos, 1997), linked primarily to air temperature and short-wave solar radiation (Prowse *et al.*, 1990). As the endpoint for streamflow from the entire basin, arctic deltas are influenced by the combination of sub-basin hydrologic regimes (*e.g.* Woo and Thorne, 2003). Thus, the timing of the northward progression of both melt and the spring floodwave plays a role in determining the severity of break-up (Rouse *et al.*, 1997). The rapid dissipation of energy through the delta reach, such that low channel slopes and extensive channel networks influence the nature and distribution of flow, affects the evolution of freeze-up and winter ice formation, as well as the transmission of the spring hydrograph through the resulting ice cover at break-up (Walker, 1998). Local and basinwide meteorological conditions that influence ice thickness, the depth of snow on the ice, spring ice decay, snow accumulation, snowmelt, and spring runoff (Gray and Prowse, 1993), are important to the development of these intra and extra-delta controls.

Each river basin has unique physiographic, climatic, and hydrologic regimes ultimately influencing the balance of hydroclimatic controls on the ice break-up regime. Greater emphasis has recently been placed on understanding these more complex controls on individual ice regimes, particularly in light of concern related to climate change impacts (*e.g.* Beltaos and Prowse, in press; Prowse and Beltaos, 2002).

1.2 RESEARCH UNKNOWNS

With observed and projected changes to climate most intensely manifested in high latitude regions (Anisimov *et al*., 2007; Walsh *et al*., 2005; Serreze *et al*., 2000) influencing the cryospheric and hydrologic regimes that define deltaic ecosystems in the

Arctic (Bates *et al.*, 2008), increased attention has been focused on the potential impacts of climate change in these environments (*e.g.* Prowse *et al.*, 2006b). Specific effects of broad-scale temperature and precipitation changes to ice break-up regimes in arctic deltas will be diverse and complex, driven by local and regional changes in the magnitude of winter snowpacks and ice thicknesses, and altered timing of the spring freshet and ice freeze-up, ablation and break-up. Thus, a first requirement to assessing possible changes to ice regimes and their influence on the hydroecology in arctic deltas, is an understanding of the specific hydroclimatic conditions controlling the timing and severity of break-up and ice-jam flooding.

Current detailed understanding of such processes in cold-regions deltas is limited primarily to the Peace-Athabasca Delta, the focus of the Northern River Basins Study (NRBS) and the Northern Rivers Ecosystem Initiative (NREI) (*e.g.* Prowse and Conly, 2002; Hydrological Processes Special Issue on the NREI – Hydrology, 2006), while limited research has been devoted to ice break-up in the major arctic deltas of the Mackenzie, the Lena, the Ob, and the Yenisey Rivers. In addition to concerns expressed about the impacts of climate change on the hydroecology of delta lakes in the Mackenzie Delta (Marsh and Lesack, 1996), increased development in the region related to oil and gas exploration and the proposed Mackenzie Valley pipeline (*e.g.* Lawrence, 2004), and recurring flooding of delta communities (*e.g.* Kriwoken, 1983), underline the importance of understanding the driving hydroclimatic processes and the resulting spatial and temporal patterns of break-up and ice-jam flooding in this important delta.

Both the Mackenzie GEWEX Study (MAGS) and the Mackenzie Basin Impact Study (MBIS) have focused on climate change issues in the Mackenzie Basin, with global climate cycles and regional adaptation as respective foci (Rouse *et al.*, 2003; Cohen, 1997). However, neither study explicitly considered the impact of changes to river ice regimes. More recent research has focused on the role of ice break-up in producing peak water level events throughout the Mackenzie Basin (de Rham *et al.*, 2008a) and examined changes to the timing of break-up (de Rham *et al.*, 2008b) and other aspects of streamflow timing (Burn, 2008; Aziz and Burn, 2006; Zhang *et al.*, 2001). While these studies have advanced understanding of river ice hydrology in the Mackenzie Basin, the ice break-up regime in the Mackenzie Delta is still largely unstudied.

Early studies of ice break-up and ice-jam flooding in the Mackenzie Delta have drawn attention to the complexity of break-up patterns and processes in the delta, due to the influence of both the Mackenzie and Peel Rivers and numerous interconnecting channel-lake systems (Brown, 1957; Henoch, 1960; Mackay, 1963a). Ice jams have been observed between Point Separation and Horseshoe Bend in Middle Channel (Brown, 1957, Bigras, 1988) causing diversions of backwater (Henoch, 1960; Kriwoken, 1983) and Mackenzie ice (Brown, 1957; Mackay, 1963b) into the Aklavik and Peel Channels. Apart from these disparate observations and studies, however, very little is known about the extent and recurrence of ice-jam flooding throughout the delta. Equally, details related to the timing, progression, and duration of break-up in the delta associated with different flood magnitudes are lacking.

Investigations into the hydrology of delta lakes have exposed spatial variations in lake flooding regimes throughout the delta (Marsh and Hey, 1989, 1991, 1994; Marsh *et al.*, 1994) related both to levee height distribution and ice-jam flooding. These studies have highlighted the importance of break-up ice jam floods in providing water (Marsh and Hey, 1989) and sediment (Marsh *et al.*, 1999) inputs to delta lakes, ultimately controlling their ecology (Marsh *et al.*, 1994). Nevertheless, investigations of potential climate change impacts on lake hydrologic regimes (Marsh and Lesack, 1996) and studies of recent change (Lesack and Marsh, 2007) have only considered changes to water levels and lake connection times, without assessment of the complex influences on the ice break-up regime responsible for the recharge of delta lakes.

As an important arctic delta facing pressure from climate change and development, and fundamentally influenced by ice processes, better understanding of the ice break-up regime in the Mackenzie Delta is warranted. Because of the complexity of break-up in a deltaic environment, the documentation and assessment of spatial and temporal patterns is needed to provide information on the progression of break-up and flooding through the delta, particularly during the most severe events. Greater understanding of the role of hydroclimatic controls in influencing break-up flood events in the Mackenzie Delta is also needed, such that the impacts of climate change on break-up flooding and the resulting hydrology and ecology of delta lakes can be assessed.

1.3 DATA AND METHODS

A lack of comprehensive continuous data relating to ice break-up exists in the Mackenzie Delta. Observations and measurements associated with ice break-up in the delta have been sporadic and limited, while gaps are present in the network of yearly and seasonal hydrometric data, related to decreases in Arctic ground-based observation networks (Shiklovmanov *et al.*, 2002) and malfunctions or damage to hydrometric stations during break-up events. Thus, there is clear need for the integration of various data and methods in order that a useful assessment of break-up in the delta may be performed. These include hydrometric and meteorological data, ice observations, satellite imagery, air photography, and anecdotal evidence, which are available to varying degrees for the Mackenzie Delta, but have never been compiled and integrated for analysis.

Hydrometric gauge data are a continuous source from which ice break-up characteristics can be inferred (Beltaos *et al.*, 1990) and have formed the basis of several studies of ice break-up (*e.g.* Prowse and Lalonde, 1996; de Rham *et al.*, 2008b). Water levels have been measured at various sites in the Mackenzie Delta by the Water Survey of Canada (WSC) for the last few decades, and represent the main data source utilized. Estimated discharge, ice measurements from winter discharge surveys, and occasional observations of break-up are also available from hydrometric records. Meteorological data, including daily temperature and snow on the ground records, aid in the quantification of hydroclimatic controls.

The use of satellite imagery to track break-up patterns along river courses provides an additional resource for assessing break-up timing. Various studies have been conducted using advanced very high resolution radiometer (AVHRR), moderate resolution imaging spectroradiometer (MODIS), and synthetic aperture radar (SAR) (*e.g.* Dey *et al.*, 1977; Pavelsky and Smith, 2004; Weber *et al.*, 2003; Pelletier *et al.*, 2003). Using such images with appropriate resolutions allows open water, intact ice, and ice jams to be identified with reasonable accuracy (Pelletier *et al.*, 2003), while break-up can be mapped over long distances (Pavelsky and Smith, 2004). Air photography provides more localized information on ice break-up and flood conditions.

Equally, many basinwide and local conditions, required for the exploration of hydroclimatic controls on break-up, are not regularly measured. The derivation of elements of the discharge hydrograph (*e.g.* Cayan *et al.*, 2001), and modelling of ice thickness and ablation (Michel, 1971; Bilello, 1980), represent methods to characterize these hydroclimatic controls despite the dearth of collected data.

1.4 OBJECTIVES

In light of the aforementioned research unknowns surrounding the ice break-up regime in the Mackenzie Delta, and the realities of ice break-up data availability, a variety of data and methods have been integrated to address the two main objectives of this thesis. Each is presented as a stand-alone journal-style manuscript in Chapters 2 and 3:

1) *Characterize spatial and temporal patterns of break-up in the Mackenzie Delta, with particular emphasis on the spatial consistency of intra-delta water level variations and the timing and progression of break-up for the most severe events.* (Chapter 2)

2) *Assess the relative importance of basinwide and intra-delta hydroclimatic controls in influencing the timing and severity of break-up in the Mackenzie Delta, with specific focus on the balance of these forces associated with the most severe flooding events.* (Chapter 3)

A summary of the major findings from this research is presented in Chapter 4 along with a set of recommendations to guide ongoing research into the ice break-up regime of the Mackenzie Delta.

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Chapter 2: Spatial and Temporal Patterns of Break-Up and Ice-Jam Flooding in the Mackenzie Delta, NWT

Abstract

The Mackenzie Delta is populated by numerous freshwater lakes that provide habitat for a myriad of species. The hydrology and ecology of these delta lakes are dominated by cryospheric processes, specifically spring break-up ice jams, which typically produce the largest hydrologic event of the year. Despite the importance of ice-induced flooding in the delta, the patterns and processes characterizing these events are poorly understood. In this paper, archived records at a dozen hydrometric stations in the delta for the period from 1974 to 2006, supplemented with information from remotely sensed imagery, are used to assemble a break-up chronology for the delta, and examine the spatial and temporal patterns of break-up flooding. Analysis of backwater level and discharge at the Mackenzie River at Arctic Red River (MARR) station are used to explain the physical rationale of the resulting patterns. Results highlight years of extensive delta flooding, and within the subset of significant flood years, two event types are identified: (1) ice-driven events, with high backwater levels at MARR associated with high levels in the southern, eastern and western delta, and (2) discharge-driven events, with extensive high water levels in the mid and outer delta and along Middle Channel, despite lower upstream peak water levels. Temporally, the break-up initiation during ice (discharge) driven events occurs earlier (later) than the delta average. These occur later (earlier) in the time-series that is trending toward earlier break-ups. The MARR station is determined to be a suitable index of delta peak water levels for continued investigation into the hydroclimatic controls on extreme hydrological events in the Mackenzie Delta.

Key words: river ice, break-up, floods, spatial and temporal patterns, Mackenzie Delta

2.1 INTRODUCTION

The Mackenzie Delta, located in Western Arctic Canada is the largest cold-regions delta in North America. Over 49,000 lakes populate the deltaic plain (Emmerton *et al.*, 2007), creating a unique aquatic ecosystem that is a haven for fish, waterfowl, and aquatic mammals. The hydrologic and ecological processes in the Mackenzie Delta are intimately tied to cryospheric processes. A nival regime hydrograph is observed for the flow of the Mackenzie River entering the delta at Arctic Red River (Woo and Thorne, 2003), whereby runoff is dominated by high spring flows from snowmelt in the Mackenzie Basin followed by low summer and winter flows (Church, 1974). Given the northern flow direction of the Mackenzie River, the spring freshet progresses downstream with the seasonal advance of warm weather, and the flood wave often encounters an intact and resistant ice cover resulting in the formation of ice jams and flooding (Rouse *et al.*, 1997). Thus, of particular importance for this break-up dominated reach (de Rham *et al.*, 2008a) are spring break-up ice jams occurring concurrently with the spring freshet, which typically produce the largest hydrologic event of the year and the flooding of delta lakes (Marsh and Hey, 1989).

Concern has been expressed regarding the impacts of climate change on the hydroecology of this productive arctic system (Marsh and Lesack, 1996; Prowse *et al.*, 2006b). Increasing development in the region, particularly related to oil and gas exploration and the proposed Mackenzie Valley pipeline, has also highlighted the need for better understanding of the controlling hydroclimatology on ice-induced flooding in the delta. Early studies of ice break-up and flooding in the Mackenzie Delta have drawn attention to the complexity of break-up patterns and processes in the delta, due to the influence of both the Mackenzie and Peel Rivers and numerous interconnecting channel-lake systems (Brown, 1957; Henoch, 1960; Mackay, 1963a). Investigations into the hydrology of delta lakes have exposed spatial variations in lake flooding regimes throughout the delta (Marsh and Hey, 1989, 1991, 1994; Marsh *et al.*, 1994) related both to levee height distribution and ice-jam flooding. Ultimately, for this large and complex system, an examination of spatial and temporal patterns of break-up and flooding in the delta is a necessary first step to understanding intra and extra-delta hydroclimatic controls.

River ice plays a dominant role in controlling extreme hydrologic events in cold regions. The river ice break-up period is particularly noteworthy in arctic regions, as it often coincides with the arrival of the spring freshet (Prowse *et al.*, 2006a). Thus, backwater produced by broken and jammed ice can augment water levels on rivers already receiving large amounts of discharge from snowmelt runoff. In many areas, the resulting high water consistently surpasses open-water maximums (*e.g.* de Rham *et al.*, 2008a), and can result in flooding of riverside communities (*e.g.* Kriwoken, 1983). The physical action of ice during break-up ice jams is also responsible for considerable damage to infrastructure and property, while additional economic costs accrue from disruption to northern transportation networks and hydro-power generation (*e.g.* Prowse *et al.*, 2007).

The ecological impacts of break-up and ice jamming are also considerable (Prowse 2001a,b). High stage and flow velocity combined with the abrasive shoving of ice can cause extensive scouring of high banks, the inside of meander bends, and channel beds, drastically altering channel and floodplain morphology (Scrimgeour *et al.*, 1994).

This channel erosion results in the mobilization of sediment and nutrients; accordingly, suspended sediment concentrations reported during break-up are often orders of magnitude higher than those observed during the rest of the year (*e.g.*, Beltaos and Burrell, 1998; Prowse, 1994; Walker, 1969). Both ice-scour and overbank flooding of sediment-laden water, while disruptive to aquatic and riparian communities, are fundamental to the diversity and productivity of these biological systems (Prowse and Culp, 2003). In particular, ice jam floods play a critical role in the recharge of water to ponds and wetlands in northern river deltas (Prowse *et al.*, 2006a). These include the Peace-Athabasca (Peters and Prowse, 2006), the Yukon (Dupre and Thompson, 1979), the Colville (Walker and McCloy, 1969), and the Mackenzie (Marsh and Hey, 1989) deltas in North America, as well as, the Yenisey (Burdykina, 1970) and Lena (Antonov, 1969) deltas in Siberia. In these systems, spring break-up ice jams typically produce the largest hydrologic event of the year and the flooding of riparian landscapes that usually include extensive numbers of lakes (*e.g.* Prowse and Lalonde, 1996; Marsh and Hey, 1989). High elevation lakes are the most sensitive to changes in the spring water-level regime, particularly during periods of negative summer water balances and infrequent spring flooding (*e.g.* Peters *et al.*, 2006; Marsh and Lesack, 1996).

As the endpoint for streamflow from the entire basin, cold-regions deltas are influenced by the combination of subbasin hydrologic regimes, with basin size moderating minor fluctuations (*e.g.* Woo and Thorne, 2003). Sequential northward progression of warm air temperatures and associated snowmelt runoff drive the spring floodwave in northward flowing nival regimes, while the extra-delta influence of the driving hydroclimatology in trigger tributaries can play an important role in controlling the timing and severity of break-up in downstream reaches (*e.g.* Peters and Prowse, 2006). Equally important in a delta environment are intra-delta effects, particularly the rapid dissipation of energy through the delta reach with several hydrological consequences. Low channel slopes and extensive channel networks affect the nature and distribution of flow, influencing freeze-up and winter ice formation, as well as the transmission of the spring hydrograph through the resulting ice cover at break-up (Walker, 1998).

While air temperature has been shown to have a strong relationship with river ice break-up timing (Williams, 1970; Rannie, 1983; Soldatova, 1993; Magnuson *et al.*, 2000; Zhang *et al.*, 2001; Bonsal and Prowse, 2003), the timing and severity of break-up depend on numerous factors driven by climate but interacting though a variety of geophysical processes (Prowse and Beltaos, 2002). Initial break-up conditions are determined by hydroclimatic conditions during the fall and winter that control freeze-up levels, ice thickness, and the size of upstream snowpack available for runoff. The final onset and evolution of spring break-up, however, depend on the prevailing hydrometeorology during the break-up period (Beltaos, 1997). Air temperature and short-wave solar radiation play a dominant role in influencing spring snowmelt and ice deterioration (Prowse *et al.*, 1990), and ultimately the severity of break-up ice jamming.

Two break-up extremes can be identified on arctic rivers: thermal and dynamic. Thermal break-up is characterized by extensive melting and decay of the ice cover until it deteriorates in place or is easily entrained by river flows. This can occur if flows remain low throughout the break-up period due to a small winter snowpack or protracted melt (Prowse and Marsh, 1989), or given climatic conditions favouring thermal ablation of the downstream cover before the start of upstream snowmelt (Gray and Prowse, 1993). Such conditions include warm air temperatures, high insolation, and reduced albedo (Bigras, 1988). The occurrence of thermal break-up is promoted in areas where water surface slope is relatively flat because the forces acting on the ice cover are fairly small (Hicks *et al*., 1997). Dynamic events occur when a large spring flood wave propagates downstream before any significant decay of the ice cover has taken place, often resulting in the incidence of ice jams (Gray and Prowse, 1993). This results from the rapid melt of a large upstream snowpack, and a competent downstream ice cover due to minimal ablation and radiation inputs. Thus, this type of break-up is controlled by the balance between the upstream driving force, primarily icemelt and snowmelt runoff, and downstream resistance forces, such as ice-cover strength, thickness, and support-boundary conditions (Prowse and Demuth, 1993). The severity of break-up generally increases with river discharge and ice-cover competence (Ferrick *et al.*, 1992).

One of the most thoroughly studied cold-regions deltas is the Peace Athabasca Delta (PAD) in the headwaters of the Mackenzie Basin. Many of the approaches formulated to study the hydroclimatology controlling break-up in the PAD are applicable in the study of other cold-regions deltas. Of particular relevance is the use of an extra-delta index, the Peace Point hydrometric station located upstream of the delta, to identify the historical source of high water levels (Prowse and Lalonde, 1996), investigate the hydroclimatic controls on ice-jam flooding (Prowse and Conly, 1998), and characterize the ice regime (Beltaos *et al.*, 2006), given limited long-term hydrometric records available within the delta reach.

Analyses of break-up patterns and processes in the Mackenzie Delta are equally constrained by a dearth of hydrometric data. Thus, an exploration of break-up patterns and processes in the delta requires the compilation of many resources including hydrometric records, satellite images, aerial photography and break-up observations to create a distinct chronology of break-up. The resulting data can provide insight into the spatial and temporal patterns of break-up flooding, with subsequent links to hydroclimatic controls. To this end, these data can be used to assess the suitability of an upstream hydrometric station, the Mackenzie River at Arctic Red River (MARR) station located 20 km from the delta reach with ~30 years of record, as an extra-delta index. Such an index is required for greater exploration of intra-delta and basinwide break-up controls, given a lack of continuous and relevant station data in the delta. The use of this site is also advantageous given the possibility for linkages to studies of Mackenzie Basin spring flow and any observed/projected changes with climate. Thus, the objectives of the research outlined in this manuscript are: (1) to construct a break-up chronology for the Mackenzie Delta as a means for preliminary investigation of break-up and ice-jam flooding, (2) to explore the spatial consistency and recurrence of intra-delta water level variations occurring during ice-affected extreme floods, (3) to consider the influence of flood severity on the timing and duration of break-up in the delta and examine associated temporal trends, and (4) to assess the suitability of using the Mackenzie River at Arctic Red River hydrometric station as an index for the severity of break-up in the delta.

2.2 STUDY AREA

The Mackenzie Delta is located at the mouth of the Mackenzie River at the Beaufort Sea, covering an area of 12,000 km^2. The delta extends from Point Separation in the south, just north of the confluence of the Mackenzie and Arctic Red Rivers, to Shallow Bay in the northwest and Kittigazuit Bay in the northeast (Figure 2-1), with elevation variations of only 3 to 4 metres over its 200 km length. Levee heights decrease step-wise from over 9 m above late summer water levels in the south to less than 1.5 m in the northern delta, with two rapid drops: approximately 5 m in less than 16 km immediately north of Aklavik and Inuvik, and ~1 m in a few kilometres north of Reindeer station, with the second drop proposed as the limit of break-up induced flooding (Fig. 57 in Mackay, 1963a).

The average annual (1971-2000) air temperature recorded in the delta at Inuvik is –8.8 ˚C, while precipitation averages ~250 mm annually, with 50% falling as snow (Environment Canada, 2008). The Mackenzie River provides a large majority of the discharge passing through the delta, contributing 2.85 x 10^{11} m^3/year (9037 m^3/s) on average annually (Water Survey of Canada, 2008). Covering an area of 1.8 million km^2, the Mackenzie Basin stretches from 52 to 70˚N, extending from central Alberta to the Beaufort Sea and from the Western Cordillera to the Canadian Shield. The basin contains six main subbasins, three major lakes, and three important deltas. Cold temperate, mountain, subarctic and arctic climatic zones occur, with annual precipitation declining from west to northeast (Rouse *et al.*, 1997). Eight of the fifteen ecozones identified in Canada are present in the basin. Also entering the delta, and contributing regional flow from the northwestern portion of the Mackenzie Basin, Peel River sub-basin discharge represents 2.14 x 10^{10} m^3/year (679 m^3/s) on average (Water Survey of Canada, 2008). With headwaters in the Ogilvie Mountains, the Peel River flows northeastward to meet the delta near Fort McPherson, draining an area of 70 600 km^2 between 64 and 67°N. The Peel Basin covers part of the two northernmost ecozones in the Mackenzie Basin.

Published records and observations of break-up and ice-jam flooding in the Mackenzie Delta are unfortunately somewhat sporadic since monitoring of break-up in the delta began in the early 1900s (Mackay, 1963a). They exist largely in the domain of 'grey' literature, primarily driven by development; oil and gas in the 1970s and post-2000 (*e.g.* Lawrence, 2004), and hydroelectric development in the 1980s (*e.g.* Parkinson and Holder, 1982, and many Environment Canada reports). While limited, these have highlighted certain aspects of the spatial and temporal variation of ice jams and flooding.

Break-up patterns and processes in the delta are complex due to the influence of both the Mackenzie and Peel Rivers and numerous interconnecting channel-lake systems (Mackay, 1963a). Early research suggests that the central and eastern sections of the delta are controlled by the Mackenzie River, while the southwestern region is controlled by the Peel and western mountain rivers (Bigras, 1988). Diversions of backwater (Henoch, 1960; Kriwoken, 1983) and Mackenzie ice (Brown, 1957; Mackay, 1963b) into the Aklavik and Peel Channels have also been reported. These are associated with ice jams that have been observed to consistently form between Point Separation and Horseshoe Bend in Middle Channel (Brown, 1957; Bigras, 1988). Due to lower levee heights in the northern delta, extensive flooding has been observed in this region even in years of overmature break-up (Terroux *et al.*, 1981; Bigras, 1988), possibly enhanced by the presence of bottomfast ice (ice frozen to the bed from above) in outer delta channels that promotes overflow. In contrast, it has been postulated that dynamic break-up and ice jamming are required to cause flooding of the southern delta lakes (Bigras, 1988; Bigras, 1990), as this region has the highest levees (Mackay, 1963a). This is confirmed by studies into the flooding hydrology of delta lakes. Specifically, the percentage of high elevation lakes with sill elevations greater than the one-year return period of the spring peak water level (Marsh and Hey, 1989) is highest in the southern delta and decreases northward through the delta (Marsh and Hey, 1991).

Related to break-up timing, Henoch (1960) reported that break-up on the Peel River entering the delta preceded Mackenzie River at Arctic Red River break-up by ten days on average, while Mackay (1963b) calculated an average difference of eight days for 47 years of record between 1895 and 1960. Marsh and Hey (1989) found that the timing of peak water level at Inuvik is remarkably consistent, occurring June 3 on average. Based on a few years of observations Parkinson and Holder (1982) indicated that the progression of the break-up front from Point Separation to the Beaufort Sea can take 3 to 10 days.

2.3 DATA AND METHODOLOGY

To meet the objectives of constructing the break-up chronology, investigating spatial and temporal patterns of break-up, and assessing the suitability of MARR as an extra-delta index of break-up severity, data were extracted from a wide variety of sources and synthesized for the different analyses as described in the following sections.

2.3.1 Data Extraction

Data were obtained from various sources for the creation of a break-up chronology and the assessment of spatial and temporal patterns of ice-induced flooding in the Mackenzie Delta. The peak water level during break-up (H_m) and its timing (t_m), in addition to the timing of the initiation of break-up (t_b) and the last ‘B’ date (see below for definition) were gathered for 14 hydrometric stations within and immediately upstream of the delta for the period from 1974 to 2006 (Table 2-1). This time period reflects the availability of Water Survey of Canada (WSC) hydrometric data in the delta, as monitoring at Inuvik and Aklavik started in 1974. The hydrometric variables were extracted from WSC archives, which for each station optimally include: pen recorder charts during the break-up period, yearly station analyses, annual water-level tables, discharge measurement tables, and hydrometric survey notes, in addition to the station description, gauge and benchmark history, and stage-discharge curves. Historical observations, air photographs reconstructed into ~300 mosaics for break-up analysis from National Water Research Institute archives spanning the period from 1973 to 1983, and a series of Landsat, AVHRR and MODIS images dating from 1973 to 2006 were used to confirm and inform the quantification of initial ice movement timing (pertaining to t_b) and ice free status (pertaining to the last ‘B’ date) for the various locations in the delta.

The H_m magnitude and timing correspond to the maximum instantaneous stage during break-up, available from the water-level charts and from published records when it was determined to be the maximum stage for the year. In years when the gauge was damaged or destroyed during break-up, reliable high water marks published by WSC or mean daily water levels, when available, were used as the H_m.

The probable timing of break-up initiation, t_b, was determined following guidelines outlined in Beltaos *et al.* (1990), by considering both the latest time in which a continuous ice cover can be assumed to be present and the earliest time at which broken ice effects are apparent on the stage hydrograph. As shown in Figure 2-2, this is generally fixed as the first significant spike in water level after the stage begins to rise from its winter low, since the steep rise in water level and subsequent peaks are indicative of breaking or broken ice effects. When possible, observations available from site visits, break-up monitoring, air photographs, and satellite images were used to improve interpretation of ice conditions. This was particularly useful for years and stations in which the initiation of break-up was not recorded or not visually striking in the chart recordings, perhaps due to a more thermally driven event or flow diversion of backwater into adjacent channels.

The last 'B' date refers to the end of ice-affected flow conditions at discharge stations. The 'B' designation is used by WSC technicians to identify the period of the year when water levels are influenced by ice, and during which stage-discharge relationships cannot be used (Pelletier, 1990). While this measure is not reproducible using water level charts, as it is based on broader (and potentially more subjective) understanding of any particular station, the last 'B' date can provide a measure of the end of the spring break-up period for that particular river reach. Of stations in the mid- and outer delta, discharge is published only at the East Channel at Inuvik station. To extend this measure to the outer delta, the suite of air photographs and satellite images was used to estimate the first ice-free day in the delta. Given a reasonably good correspondence between the last 'B' date at Inuvik and this visual estimate (Figure 2-3), with an r^2 value of 0.54 (normality assessed using the Kolmogorov-Smirnov test), the former was used to represent the end of the break-up period in the delta.

For the additional delta index analysis, the availability of published discharge values and stage-discharge curves at the MARR station allowed the extraction of maximum break-up discharge (Q_{max}) and the calculation of the backwater level caused by ice for each year. Backwater level is the term used to describe the difference between ice-affected stage and the stage for an equivalent discharge in open-water conditions (Beltaos, 1995).

2.3.2 Peak Break-up Water Levels

Comparisons of peak water levels across hydrometric stations were undertaken using normalized water levels, to account for variations despite site-specific benchmarks at each station. For a given year (i) the normalized peak water level ($\hat{H}_{m_i}$) was calculated as follows:

$$\hat{H}_{m_i} = \frac{\left(H_{m_i} - \overline{H}_m\right)}{\sigma} \qquad [1]$$

Thus, yearly spring peak water levels were compared based on their number of standard deviations (σ) from the long-term mean ($\overline{H}_m$) for the station. Herein, all references of *peak water level* refer to normalized values. These normalized values were averaged for the delta (excluding the source rivers) and a set of west to east and south to north transects (Figure 2-4), to create the break-up chronology and explore regional water level variations, respectively. Due to different station record lengths and break-up events not captured in the record, these averages are based on a different number of stations in any given year, although all stations included were given equal weighting.

Recurrence analysis was performed for the $\hat{H}_m$ values at each station using the Weibull Method (Weibull, 1951) to calculate the probability of exceedance for each peak water level event. The inverse of this value represents the return period of the particular peak water level. Following the approach of de Rham *et al.* (2008a) and standard caution in return period analysis, no extrapolation was undertaken beyond the existing historical record. Peak break-up water levels are influenced by different ice effects in any given year, in addition to the constraints imposed by the height and morphology of channel banks, which prevents accurate estimation of the likelihood or the occurrence of more extreme peak water levels. This limits the return period assessment to the temporal coverage available for each station.

2.3.3 Timing and Duration

The timing of break-up initiation (t_b), peak water level (t_m), and the last 'B' date were compared based on the day of the year (Julian day starting on January 1) of occurrence. The duration of break-up at an individual station (t_3 shown in Figure 2-2) was defined as the number of days between the initiation of break-up and the last 'B' date, while the duration of break-up in the delta (t_D) was determined as the number of days from the earliest initiation of break-up (most commonly on the Peel River, but also sometimes on the Mackenzie River) to the last 'B' date at Inuvik.

Maps of break-up timing in the delta for average and selected years of extreme and large flooding events (outlined below) were created in ArcGIS using Inverse Distance Weighting (IDW) linear interpolation between all stations with records in a given year. T-tests at the 0.05 level were applied to determine if the mean timing and duration of break-up in these extreme and large flood years differed significantly from that for the rest of the series. Trends in the timing of break-up initiation, peak water level, and last 'B' date, as well as the duration of break-up at stations and for the delta were assessed using the Mann-Kendall non-parametric test for trend (Mann, 1945; Kendall, 1975) and Sen's slope estimator (Sen, 1968), contained in the Salmi *et al.* (2002) MAKESENS template in Excel. These methods were used in a similar analysis of break-up timing by de Rham *et al.* (2008b).

2.3.4 Mackenzie Delta Index

The relationship between MARR normalized peak break-up water level and levels at individual stations was explored using standard linear regression. The influence of MARR level on the averaged peak water levels for the delta and the west to east and south to north regional transects was similarly evaluated. The effect of MARR peak discharge was also considered for these same stations and groupings. Maps of r^2 values for delta station peak water level compared to MARR level and discharge were created using IDW linear interpolation.

2.4 RESULTS

2.4.1 Chronology of Break-up Peak Water Levels

To identify years of extreme ice-induced flooding, a chronology of peak water levels during the break-up period was created for the Mackenzie Delta based on the average of available normalized peak water levels recorded at delta stations in each year. The chronology spans the period of available instrumental record in the delta from 1974 to 2006, although more stations are available to inform the delta average peak level later in the chronology (Figure 2-5). The limited availability of stations, particularly prior to 1980, represents a constraint in the analysis. The chronology reveals a multi-year pattern of alternating flood severity, with higher peak levels occurring in the 1980s and since 2004, while lower levels were recorded in the 1970s and for the late 1990s. With peak levels at or approaching a standardized value of 2, 1982, 1992, and 2006 stand out as years with the greatest flood severity, which is confirmed by recorded observations (Kriwoken, 1983; Marsh *et al.*, 1994; George Lennie, pers. comm.). These are subsequently referred to as *extreme* event years. Equally, 1985, 1986, 1989, 1997 and 2005 also have higher than average water levels, above 0.5 and approaching 1, and are herein called *large* event years.

2.4.2 Spatial Variability of Levels for Extreme and Large Events

For the extreme and large peak water level events identified above, a clear signal of above average water levels throughout the delta is apparent. As Figures 2-6 and 2-7 illustrate, west-east and south-north station transects largely agree with the delta average (from Figure 2-5), particularly for the extreme high (low) water events. An exception to this pattern occurs for the peak water levels on the source rivers, the Mackenzie (R1), Arctic Red (R2), and Peel Rivers (R3), which diverge below the delta average in some larger event years (1986, 1992, 2005).

The series of west-east delta transects provide an indication of water level changes in the direction of flow from the source rivers to the outer delta (Figure 2-6). These transects align with Marsh and Hey's (1991) southern, mid, and northern study sites of lake flooding regime along East Channel. For large and extreme events, as previously defined, two distinct and spatially contrasting patterns can be observed:

1. Peak water levels increasing northward through the delta, as exhibited in 1985, 1986 and 1992 (upward arrows in Figure 2-6).
2. Peak water levels highest on the source rivers and in the southern regions of the delta, as occurred in 1989, 1997, and 2006 (downward arrows in Figure 2-6).

Limited station data precludes the identification of any patterns in 1982, while peak water levels in 2005 display a separate pattern, with a northward increase in levels despite lower levels in the mid-delta.

The south-north transects allow an analysis of water level changes for channel groupings (Figure 2-7). The western transect captures stations postulated to be influenced by the Peel and western mountain rivers, the central transect includes stations along Middle Channel, while the stations along East Channel make up the eastern transect. These extend prior work by Marsh *et al.* (1994) on lake flooding along an east-west cross delta transect between Aklavik and Inuvik. The resulting spatial patterns of peak water level for large events appear to be more complex than for the west-east transects. Peak water levels are higher in the central delta than the east for 1985, 1986 and 1992, while the opposite is true for 1997, 2005 and 2006, corresponding to the first and second groups from the previous analysis. However, the average of peak water levels for the western delta transect is lower than both the central and eastern delta transects in all large years except 1997, where it exceeds the central transect, and the extreme years of 1992 and 2006, where the highest levels occurred in the western delta.

Further to the patterns identified above, peak water levels for extreme and large years at individual stations with record lengths exceeding 15 years were analyzed (Table

2-2). Because of their short record lengths, both stations in the south delta transect were excluded. For the extreme years (1982, 1992, and 2006), a majority of the stations have levels near or exceeding 1, while for large years (1985, 1986, 1989, 1997, and 2005), one or several stations in a given region or channel have higher than normal water levels, however the pattern of higher water levels does not apply to all stations.

The spatial patterns of northward-increasing (decreasing) peak water levels are best displayed in the extreme year of 1992 (2006), and are present but less pervasive in the large years. In 1992 levels are not relatively high on the Mackenzie (R1) or Arctic Red Rivers (R2), while levels on the Peel River (R3) and at all stations in the delta are considerably higher than average. The largest peak water level was 2.60 recorded on Reindeer Channel at Ellice Island (N5) in the northern delta, while the Peel Channel above Aklavik (M3) had a level of 2.13 associated with considerable flooding of the town. A similar pattern occurs in 1985 and 1986, with low peak water levels on one or more source river (with Mackenzie River levels consistently less than 1) and higher levels in the mid and northern delta. In contrast, 2006-levels are among the highest recorded for the Mackenzie and Arctic Red Rivers and on the Peel Channel above Aklavik, with considerable ice jamming reported; lower levels were experienced in the northern delta, although flooding could still be widespread in this region for these events given the low levee heights. High levels on the Mackenzie River with above average levels for some mid and northern delta stations characterize the peak water levels in 1989, 1997, and 2005.

The south-north transect patterns are also apparent in the individual station analysis. Higher levels for the western transect in 1992 and 2006 are the result of well above average peak water levels experienced at Aklavik. In addition, the influence of peak water levels at individual stations on the central and eastern transect patterns for the first (1985, 1986, and 1992) and second (1989, 1997, 2005, and 2006) groups is apparent, except in 1985 as the peak water level of 1.88 from the Outflow Middle Channel below Langley Island station (N4) is not included in the table.

The apparent separation of large flooding events in the delta into two groups (1985, 1986, and 1992; and 1989, 1997, 2005, and 2006) based on the observed spatial patterns of peak water levels is an interesting result of this analysis. The outcome suggests that recurring combinations of hydroclimatic drivers may be influencing the spatial variation of peak water levels, likely by determining the location and flow diversions produced by ice jams in the delta.

2.4.3 Recurrence Analysis

A return period assessment for delta and source river stations with records exceeding 10 years, including the delta average, is shown in Figure 2-8. Results indicate that water levels are comparable for low frequency events, while considerable divergence occurs for return periods above 10 years (probability of exceedance of 10%). Events with levels between 1 and 2 have return periods ranging from 10 to 20 years, while the highest levels, between 2 and 3 have return periods of ~30 years. Of particular note, the most recent event, in 2006, resulted in the highest recorded water level, occurring at the MARR station with a return period of 29 years.

2.4.4 Temporal Patterns of Break-up

The average timing of break-up initiation (t_b) and peak water level occurrence (t_m) interpolated across the delta based on individual station records are shown in Figure 2-9, with the average Julian dates also provided for these and the last 'B' dates at discharge stations. Concurrent with historical records, the initiation of break-up begins earliest on the Peel River, followed by the Mackenzie, although the average delay of 3 days is shorter than was reported by Mackay (1963b) for the early 1900s. The southwestern portion of the delta, represented by the Peel Channel above Aklavik station, precedes the central and eastern delta by one day on average. Stations in the outer delta begin breaking-up 5 to 7 days later, although the measure of break-up initiation may be less physically meaningful for these stations given the low hydraulic slope, extensive channel connectivity, and presence of bottomfast ice in the outer delta, such that the effect of ice movements on water level is potentially more complex than for a mechanical break-up in a single river channel. Peak water levels also occur earliest on the Peel River on average, but rise in Middle Channel earlier than both the western and eastern delta, as was observed by Marsh *et al.* (1994).

The average duration of break-up in the delta is 19.4 days. A maximum duration of 42 days resulted in 1994, associated with below average peak water levels in the delta, while a minimum duration of 9 days occurred in 1974 with the delta average nearing 0.5. Average duration of break-up for the Mackenzie, Arctic Red and Peel Rivers are 12.8, 7.7 and 10.7 days respectively, while the duration on East Channel at Inuvik is 8.9 days on average. The association between break-up duration and extreme and large events is explored in a subsequent section.

2.4.5 Timing of Extreme Events

Dynamic break-ups are associated with persistent ice jams capable of producing large-scale flooding (Prowse and Beltaos, 2002). Thus, because this type of break-up typically results from the early movement of a relatively undeteriorated downstream ice sheet (Gray and Prowse, 1993), a natural hypothesis is that earlier break-up will be associated with extreme events in the delta. However, no such pattern emerges when the timing of break-up for extreme events is compared to the average. Break-up initiation occurs 2.5 days later than average at MARR (R1) in 1982 and 1992, and 7.5 days earlier in 2006; similarly on the Peel River (R3) initiation occurs 6 and 8 days later in 1982 and 1992 respectively, and 2 days earlier in 2006. For the t-tests performed comparing mean t_b and t_m for extreme and large events to the rest of the time-series for individual stations with records exceeding 10 years, a statistically significant difference at the 95% confidence level resulted only for Middle Channel at Tununuk Point (N2) for t_b (normality established using the Kolmogorov-Smirnov test; 1 outlier was removed from the N1 t_b data to meet normality assumptions).

A clearer pattern emerges when the time-series of extreme events is separated into the above delineated categories. In the latter years, 1989, 1997, 2005, and 2006, both break-up initiation and peak water level occur earlier or close to the average on both the Mackenzie and Peel Rivers (*e.g.* 0.5, -1.5, -7.5, and -7.5 days from the average respectively for t_b at MARR), while the reverse is true for the former years, 1986 and 1992 (*e.g.* 7.5 and 2.5 days from the average respectively for t_b at MARR). This same pattern results for the south-north and east-west transects of the delta, with later events occurring in 1985, 1986 and 1992, earlier events in 2005 and 2006, and mixed event timing in 1989 and 1997.

Mann-Kendall trend analysis of t_b, t_m and last 'B' dates are provided in Table 2-3 and Figure 2-10a-c, and indicate that many stations in the delta have trends toward earlier occurrences. For the timing of break-up initiation, all stations except Middle Channel below Raymond Channel (M2) have a trend toward earlier break-up (2.09 days/decade), although these trends are only significant at the 90% confidence level for the northeastern side of the delta (Figure 2-10a). The greater slope estimate for the Mackenzie River (R1) over that of the Peel River (R3) (Figure 2-11, not significant at the 90% level), could explain the lower time delay calculated between these two stations compared with historical observations. A trend of earlier occurrence of peak break-up water level results for 8 of the 13 stations (-1.46 days/decade), however only the East Channel at Kittigazuit Bay station (N1) has a trend that is significant at the 90% level (Figure 2-10b). For the last 'B' date, no dominant trend was observed with half the stations showing earlier/later occurrence (Figure 2-10c).

The dominance of trends toward earlier occurrences of break-up initiation and peak water level is notable, particularly given the nature of the extreme and large event groups identified. Namely, the later-occurring events (1985, 1986, and 1992) are centered during the mid-1980s to mid-1990s, while the earlier events (1989, 1997, 2005, and 2006) span a later period from 1989 to 2006. This may reveal trending of large and extreme events in the delta toward the occurrence of the second, and earlier, event type. Confirmation of such a trend should be revisited as more data become available.

2.4.6 Duration of Extreme Events

Focusing on extreme events, two alternate hypotheses regarding the duration of break-up in the delta *(t_D)* can be formulated. During dynamic events, the driving force of discharge associated with the steep rising limb of the spring hydrograph fractures and moves the ice cover forming ice jams (Gray and Prowse, 1993). Waves generated by the release of these ice jams have the energy to dislodge and set in motion considerable lengths of ice cover (Beltaos, 2007), thus continuing the cycle downstream. Given that the evolution of break-up is sustained in this way, a shorter duration might be expected for these dynamic events, compared to thermal events in which the ice cover essentially disintegrates in place. Alternatively, if dynamic events produce long- lasting ice jams due to a strong competent ice cover then a longer duration would be expected.

A standard linear regression exploring the relationship between break-up duration and average delta peak stage, as represented by the delta average normalized peak water level, did not reveal a significant association (r^2 =0.08). Consideration of the two above-identified event types reveals that the former grouping (1985, 1986, 1992) had shorter than average durations, while the latter group had both shorter (2005 and 1997) and longer (2006 and 1989) than average durations (Figure 2-12). The absence of a clear pattern in these durations underscores the complexity of break-up in the delta, with both of the above described progressions likely occurring in large and extreme events, the combination of which ultimately determines the duration of break-up in the delta.

Trend analysis of break-up duration at individual stations (t_3) and for the delta (t_D) revealed a tendency toward longer break-up durations, except for East Channel at Inuvik (M1), for which had no trend (Figure 2-10d). These trends were significant for both the Mackenzie River (R1) and Peel River (R3) stations, and likely associated with the tendency toward earlier break-up initiation on these rivers (Figure 2-11).

2.4.7 Mackenzie Delta Index

The regression analysis to determine the suitability of the MARR station as an index for delta flood severity resulted in an r^2 value of 0.57 when peak water levels (H_m) at MARR were compared with the average peak water level in the delta. A pattern of decreasing r^2 values with distance from the MARR station is apparent for the relationship of peak water levels both for individual stations and east to west transects (Figure 2-13a,c). Thus, the MARR station provides the best index for stations in the southern delta, although a good amount of the variability in peak water levels at mid and northern delta stations are also explained by MARR peak water levels. As the first (delta average peak water level) and second (MARR peak water level) bars in Figure 2-14 show, water levels on the Mackenzie and in the delta consistently deviate in the same direction from the series mean in extreme and large years, although the magnitudes of the water levels differ.

For outer delta stations, regressions with the MARR maximum discharge (Q_{max}) yielded better relationships than for level, while the opposite is generally the case in the mid-delta (Figure 2-13c). A better relationship is also noted for Q_{max} than H_m in the southern delta, although this is mainly based on the S2 station record. The strong relationship of Mackenzie peak discharge with peak water levels in Middle Channel is displayed in Figure 2-13b; higher r^2 values result for the central transect water level relationship with MARR Q_{max} than for the east or west transects, as well as for the relationship with MARR H_m in the central transect.

2.4.8 Driving Processes

A comparison of the delta average peak water level to the peak water level, backwater effect, and discharge at MARR (Figure 2-14) provides some insight into the processes likely driving the observed high water patterns and regression relationships. MARR is the entry point station to the delta, delivering the dominant hydrologic signal from the Mackenzie Basin; here it is also shown to fairly accurately represent delta average peak water levels, particularly for years with considerably higher than average levels. With the availability of backwater and maximum break-up discharge data at this site allowing the influence of ice and discharge to be isolated, a set of patterns emerges complementary to those noted above. In years when the peak water level at MARR exceeds the delta average (1989, 1997, 2005, and 2006), the backwater effect shows a comparable deviation above the average, indicating that ice resistive effects (such as ice jams) are a dominant driver of increased water levels. Conversely in years when the delta average is comparable or larger than the value at MARR and discharge is high (1985, 1986, 1992), the backwater effect is generally less pronounced while discharge levels are well above average. With the identification of these processes, the event types described above can thus be referred to as *ice-driven* and *discharge-driven,* respectively.

The above patterns, and the processes they describe, provide a link between spatio-temporal patterns of break-up in the delta and hydroclimatic drivers (Table 2-4). High water levels in the southern and mid-delta, and in eastern and western channels, associated with early break-up timing for ice-driven events substantiate the importance of ice jamming between Point Separation and Horseshoe Bend to stage increases from backwater diversions in these years. In addition, high water levels in Middle Channel and the outer delta, both regions more linked with MARR discharge, and later break-up initiation during discharge-driven events highlight the dominant influence of upstream discharge to these flood years. The results are particularly striking as they suggest that the spatial expression of peak water levels in the delta and the temporal patterns of break-up are ultimately controlled by a recurring progression of hydroclimatic controls during the spring break-up period.

2.5 DISCUSSION AND CONCLUSIONS

This analysis is the first comprehensive work chronicling spatial and temporal patterns of spring break-up in the Mackenzie Delta. The results highlight years with greater flood severity, with 1982, 1992 and 2006 standing out as having extreme peak water levels. For these years, a consistent pattern of higher than normal water levels at a majority of stations is observed, suggesting wide-scale flooding. These events have water levels with return periods greater than 10 years (probability of exceedance less than 10%), although the range of return periods at different stations is quite large for events with peak water levels above 1.

The examination of spatial peak water level variations and temporal patterns of break-up revealed two dominant event types for large and extreme years that can ultimately be linked to driving physical processes. Discharge-driven events, when above average peak water levels are associated with larger levels of discharge on the Mackenzie River as in 1985, 1986, and 1992, display increasing peak water levels in a northward direction in the delta. Water levels are highest in the central delta along Middle Channel, with lower levels on the eastern and western delta channels, except in 1992 when significant flooding occurred on the Peel Channel above Aklavik (M3). Coupled with these patterns, a higher regressional relationship with MARR discharge than water level occurs for both the outer delta stations and Middle Channel stations in the central region. In the outer delta, discharge likely plays a greater role in controlling peak water levels because low channel slopes and low levee heights in this region inhibit the formation of ice jams and associated backwater. As a result flow increases are the primary mechanism for rising water levels and flooding. These results reinforce early observations of break-up and flooding in the outer delta (Mackay, 1963a, Terroux *et al.*, 1981, Bigras, 1988). Middle Channel in the central region, carries the majority of the flow from the Mackenzie River (80% during peak flow; Fassnacht and Conly, 2000), while all the stations in the transect are downstream from Horseshoe Bend and the known sites of recurring ice jam formation, thus explaining the stronger relationship of peak water levels to MARR discharge than MARR water levels.

Break-up initiation and peak water level occur later than average during discharge-driven years. For these later events a greater amount of snowmelt runoff from the basin is likely able to reach the delta during the break-up period and augment water levels, despite a potentially more deteriorated ice cover in the delta. A shorter than average duration of break-up in the delta is observed in discharge-driven years.

Ice-driven events, so labelled because high backwater levels are observed at MARR, are characterized by high water levels in the southern and mid-delta, while levels are not especially high in the outer delta (although flooding of low-lying land may still take place). Equally, levels are higher in the eastern delta and higher or similar in the western delta than the central delta. Given that ice conditions conducive to ice jamming at MARR also trigger ice jam formation between Point Separation and Horseshoe Bend, the strong relationship of mid and southern station peak water level to MARR level can be explained due to backwater diversions of Middle Channel ice and flow upstream and into secondary channels (Kriwoken, 1983). This also explains the higher water levels in the eastern and western delta, particularly on the Peel Channel above Aklavik (M3).

Earlier break-up initiation and peak water level in the delta occur in ice-driven years. This timing would suggest that less ice decay from solar radiation and warming took place prior to break-up, resulting in a more competent ice cover and the possibility of ice jam formation given adequate discharge. Such a result would agree with Bigras' (1988) deduction that ice-cover strength in the delta is a central factor controlling the formation of ice jams, in addition to influencing the length of time a jam will remain static and the amount of backwater builds-up. No pattern of break-up duration was observed for ice-driven events, ultimately highlighting the complex controls on the progression of break-up and ice jamming in the delta.

Trend analysis of break-up timing revealed a widespread tendency toward earlier break-up initiation and occurrences of peak water level in the delta. A similar trend was also noted for the Mackenzie Basin at large (de Rham *et al.*, 2008b). These results underscore another trend in the occurrence of large and extreme events; namely a tendency toward the occurrence of ice-driven events, which are associated with earlier break-up timing. Following this trend, it could be postulated that a continued retreat of break-up timing could eliminate the occurrence of discharge-driven events; however, better understanding of the hydroclimatology controlling break-up in the delta is necessary to adequately test this hypothesis.

Trends in the timing of break-up initiation at MARR and on the Peel River prompt additional questions about hydroclimatic controls. In contrast to early break-up observations, the delay between break-up initiation at these two sites has decreased, associated with a greater slope for the retreat in break-up timing at MARR than on the Peel River. These results are surprising given the different hydroclimatic characteristics of the two basins contributing to these sites: the Mackenzie Basin influenced by a variety of hydrologic, climatic, and ecological conditions over a large latitudinal gradient, and the smaller Peel Basin representing a more localized hydroclimatic response in an area of pronounced warming (Rouse *et al.*, 1997, Prowse *et al.*, 2006b), such that a greater response would be expected for the Peel River.

Ultimately, the above results highlight the need for a better understanding of the extra and intra-delta hydroclimatic controls of break-up in the Mackenzie Delta. For these subsequent analyses the MARR station has been shown to be a suitable index of delta peak water levels, given fair explanatory power of water levels in the delta based on MARR levels, with consistent representation in large and extreme years.

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Table 2-1. Number of peak break-up water level, H_m, timing of peak level, t_m, timing of break-up initiation, t_b, and last 'B' records extracted in the period from 1974 to 2006 for the 14 Water Survey of Canada hydrometric stations in and upstream of the Mackenzie Delta.

ID	Station	Start	End	n, H_m	n, t_m	n, t_b	n, last 'B'
R1	Mackenzie River at Arctic Red River [a]	1974	2006	28	27	32	33
R2	Arctic Red River near the Mouth [a]	1974	2006	27	25	28	32
R3	Peel River above Fort McPherson [a]	1974	2006	33	33	29	32
S1	Mackenzie River at Confluence East Channel	1992	2006	11	11	8	
S2	Peel River at Frog Creek	1998	2006	6	5	9	
M1	East Channel at Inuvik [a]	1974	2006	33	33	34	33
M2	Middle Channel below Raymond Channel	1983	2006	17	16	17	
M3	Peel Channel above Aklavik	1983	2006	22	15	22	
N1	East Channel above Kittigazuit Bay	1984	2006	21	17	15	
N2	Middle Channel at Tununuk Point	1984	2006	21	19	13	
N3	Kumak Channel below Middle Channel	1998	2006	7	7		
N4	Outflow Middle Channel below Langley Island	1983	2006	10	9	4	
N5	Reindeer Channel at Ellice Island	1984	2006	18	17	15	
N6	Napoiak Channel above Shallow Bay	1998	2006	9	9		

[a] stations with published discharge values

Table 2-2. Normalized peak water levels for extreme and large years at stations with record lengths exceeding 15 years. Values greater than 1 are highlighted in medium grey; values greater than 2 are highlighted in dark grey. Refer to Figure 1 for locations.

Year	R1	R2	R3	M1	M2	M3	N1	N2	N5
1982[a]		1.87	1.36	2.11					
1985		1.00	0.85	1.27		0.50	0.62	1.12	
1986	0.79		-1.05	0.53	1.27	0.89	1.35	1.19	
1989	2.01	0.96	0.58	0.84				0.82	
1992[a]	0.71	0.25	1.39	1.72	1.30	2.13	1.40	1.82	2.60
1997	1.57	-0.15	0.58	1.18	0.11	0.47	0.40		0.32
2005	1.34	-0.15	-1.10	0.68	0.28	0.19	1.08	0.64	0.85
2006[a]	2.70	2.18	0.97	1.74	1.21	2.15	0.89	0.31	

[a] years with the greatest delta average of normalized peak water levels

Table 2-3. Mann-Kendall trends for timing of peak level, t_m, timing of break-up initiation, t_b, last 'B' date, and break-up duration, t_3, at Water Survey of Canada hydrometric stations in the Mackenzie Delta and delta break-up duration, t_D, for the period from 1974 to 2006. Earlier/short and later/longer columns give the number of sites with the trend, while bracketed values are the number of sites for which the trend is significant at the 90% level.

Parameter	Number of stations	Earlier/ shorter	Average days/decade	Later/ longer	Average days/decade	No trend
t_b	11	10(3)	-2.09	0		1
t_m	13	8(1)	-1.46	3(0)	1.76	2
last 'B' date	4	2(0)	-0.85	2(1)	1.63	0
t_3	4	0		3(2)	2.06	1
t_D	1	0		1(0)	0.47	0

Table 2-4. Summary of spatial and temporal characteristics of peak water event types in the Mackenzie Delta.

Characteristic	Discharge-driven	Ice-driven
Spatial expression of H_m	- Increasing northward - Higher in central delta than east - Higher in central delta than west, except in 1992	- Highest on source rivers and southern/mid stations - Lower in the central delta than east - Lower or similar in the central delta than west
Timing of t_b, t_m	Later	Earlier
Duration of break-up, t_D	Shorter	Mixed
Years	1985, 1986, 1992	1989, 1997, 2005, 2006
Location in time-series	Mid (1980s)	Later (~1990 - present)

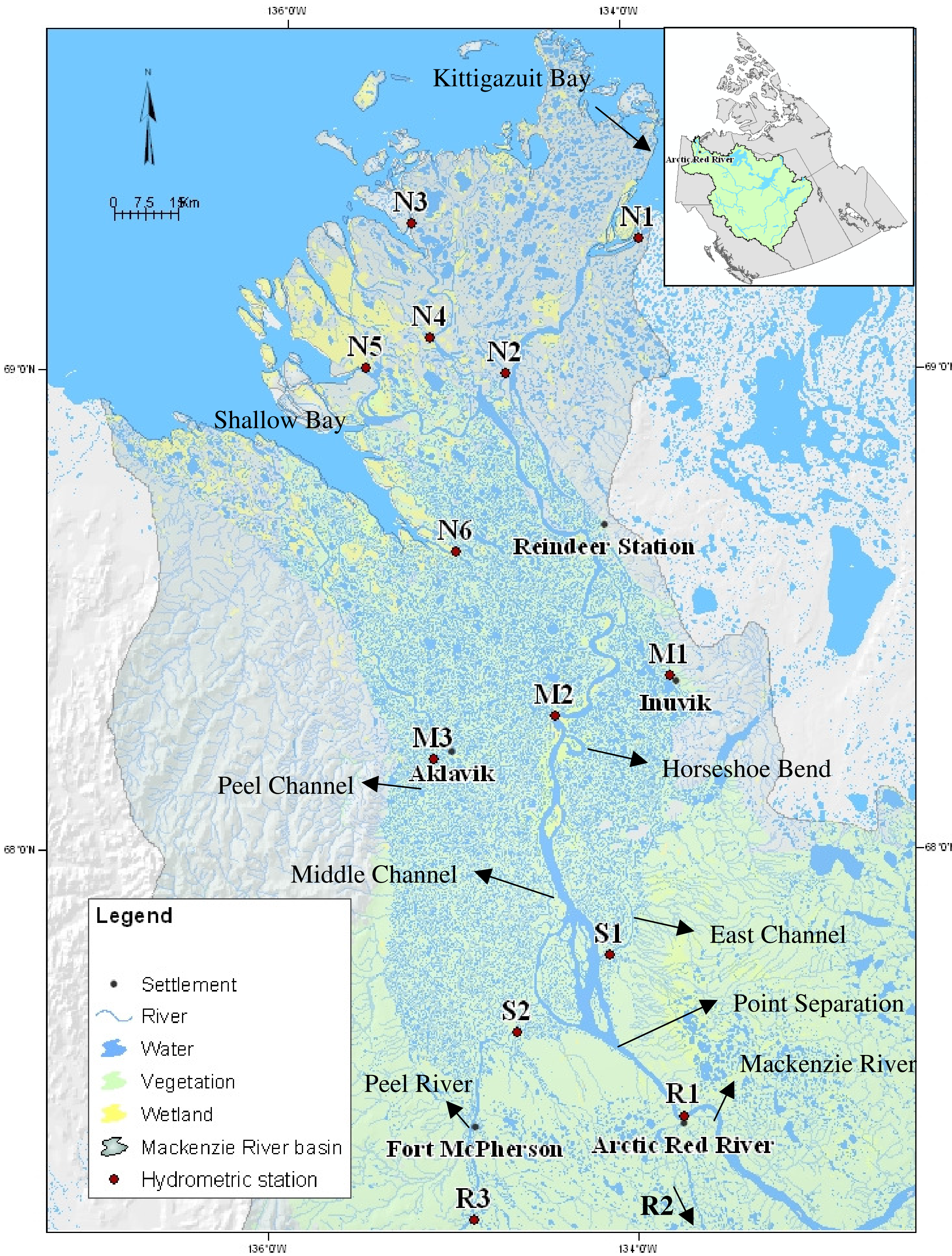


Figure 2-1. The Mackenzie Delta, with locations of Water Survey of Canada hydrometric stations and locations relevant to historical ice jam observations (Mackenzie Basin inset). The R2 station is located on the Arctic Red River approximately 80km upstream of the confluence with the Mackenzie River.

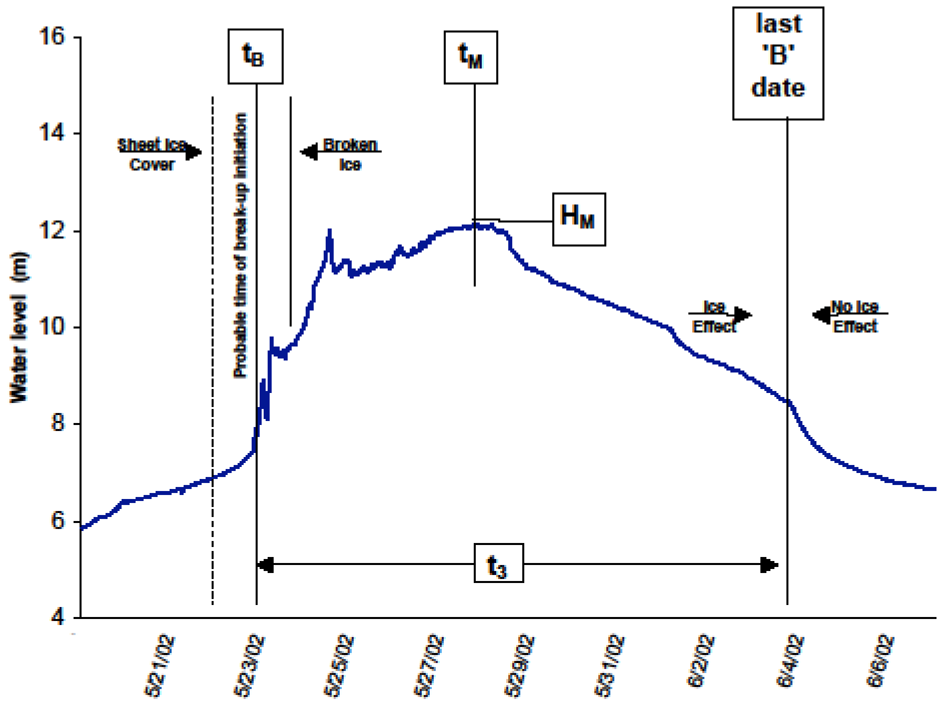


Figure 2-2. Schematic of instantaneous water level record at the MARR station (R1) during the break-up period in 2002. The timing of break-up initiation, t_b, peak water level, t_m, and the last 'B' date, in addition to the duration of break-up, t_3, are shown given interpretation of break-up initiation (Beltaos *et al.*, 1990), and WSC published peak water level (H_m) and 'B' dates.

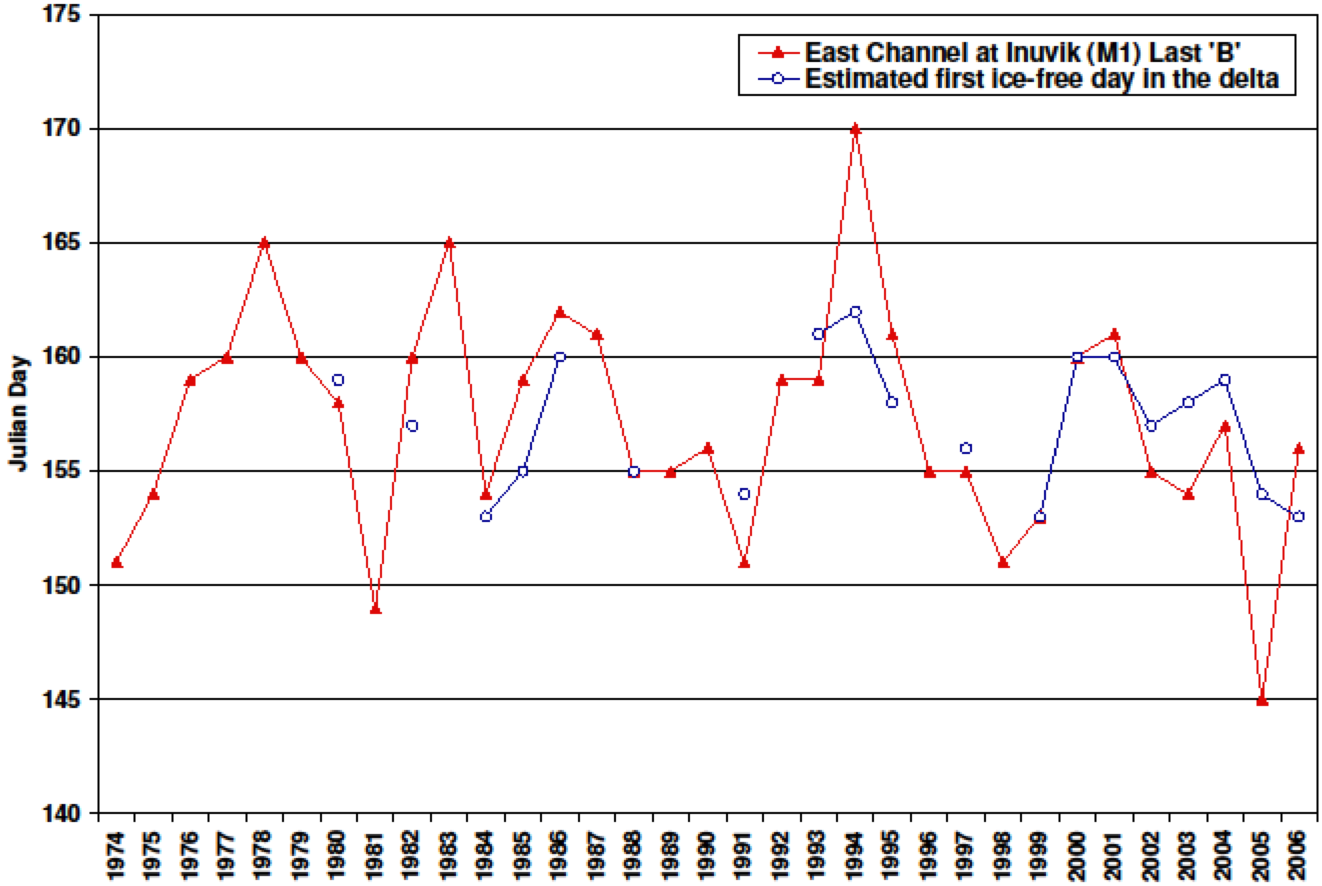
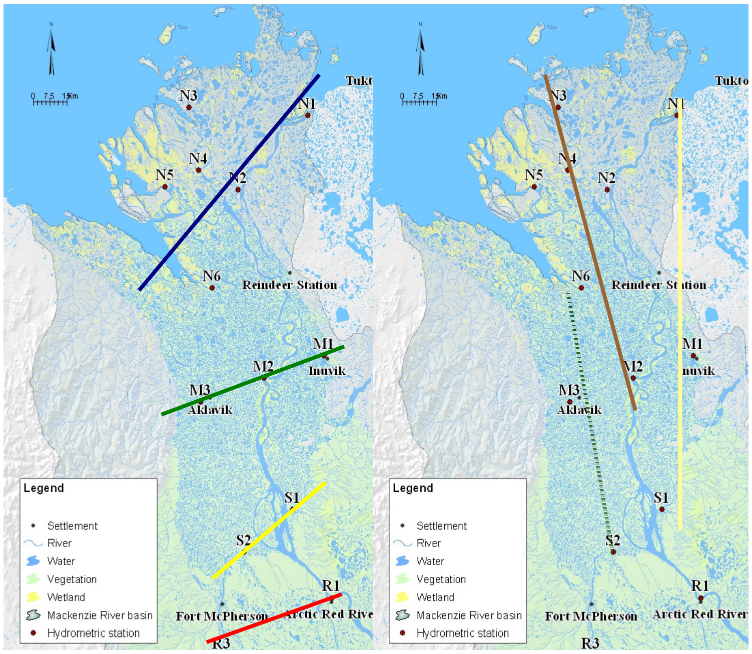


Figure 2-3. Comparison of last 'B' dates at East Channel at Inuvik hydrometric station and estimates of the first ice-free day in the delta from air photographs and satellite images (r^2=0.54).



c)

		Transect	Stations
West to East		Source Rivers	R1, R2, R3
		South	S1, S2
		Mid	M1, M2, M3
		North	N1, N2, N3, N4, N5, N6
South to North		West	S2, M3, N6
		Central	M2, N3, N4, N5, N2
		East	S1, M1, N1

Figure 2-4. a) West to east delta transects, b) south to north delta transects, and c) station composition of delta transects used for the analysis of spatial peak water level patterns.

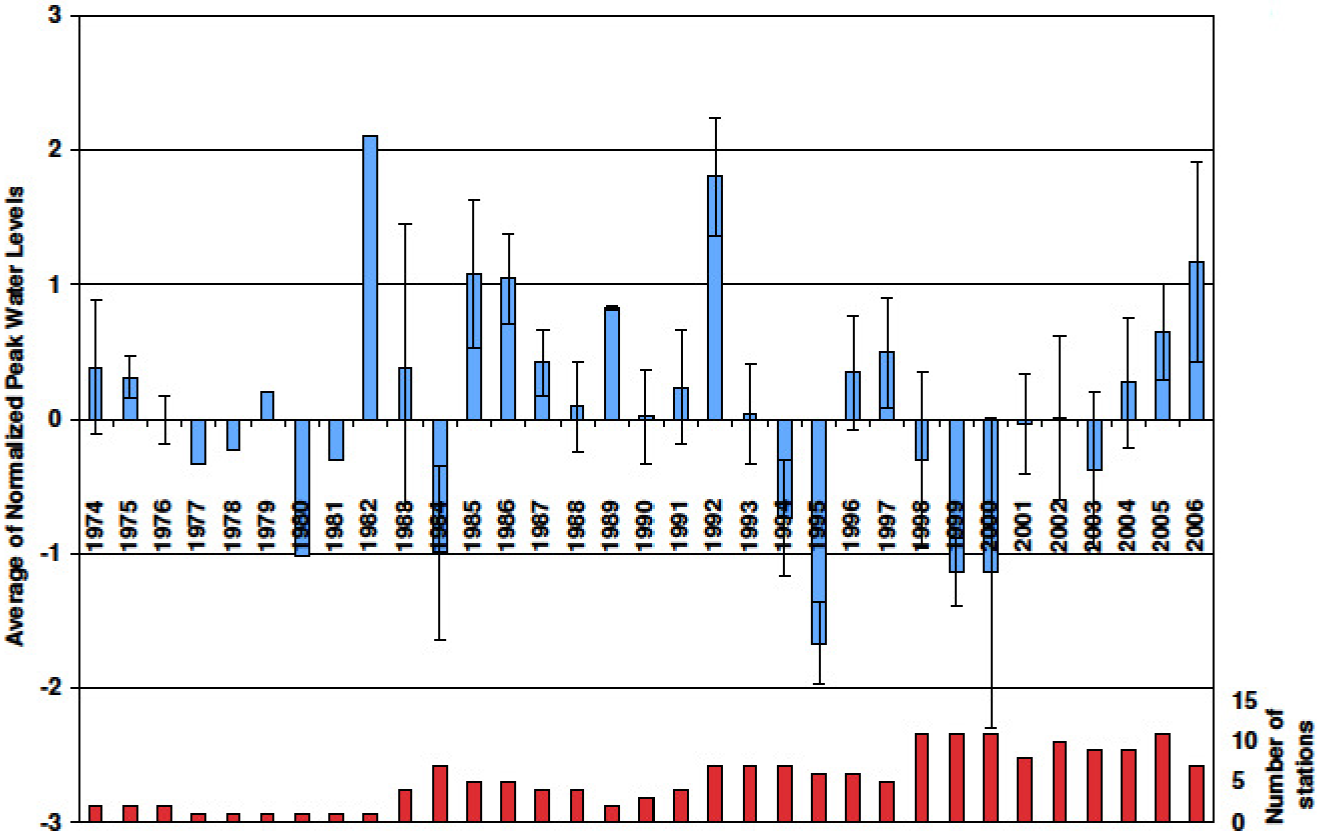


Figure 2-5. Average of normalized peak water levels in the Mackenzie Delta from 1974 to 2006 (with bars depicting one standard deviation for years with more than one station record), and histogram of stations contributing to the delta average.

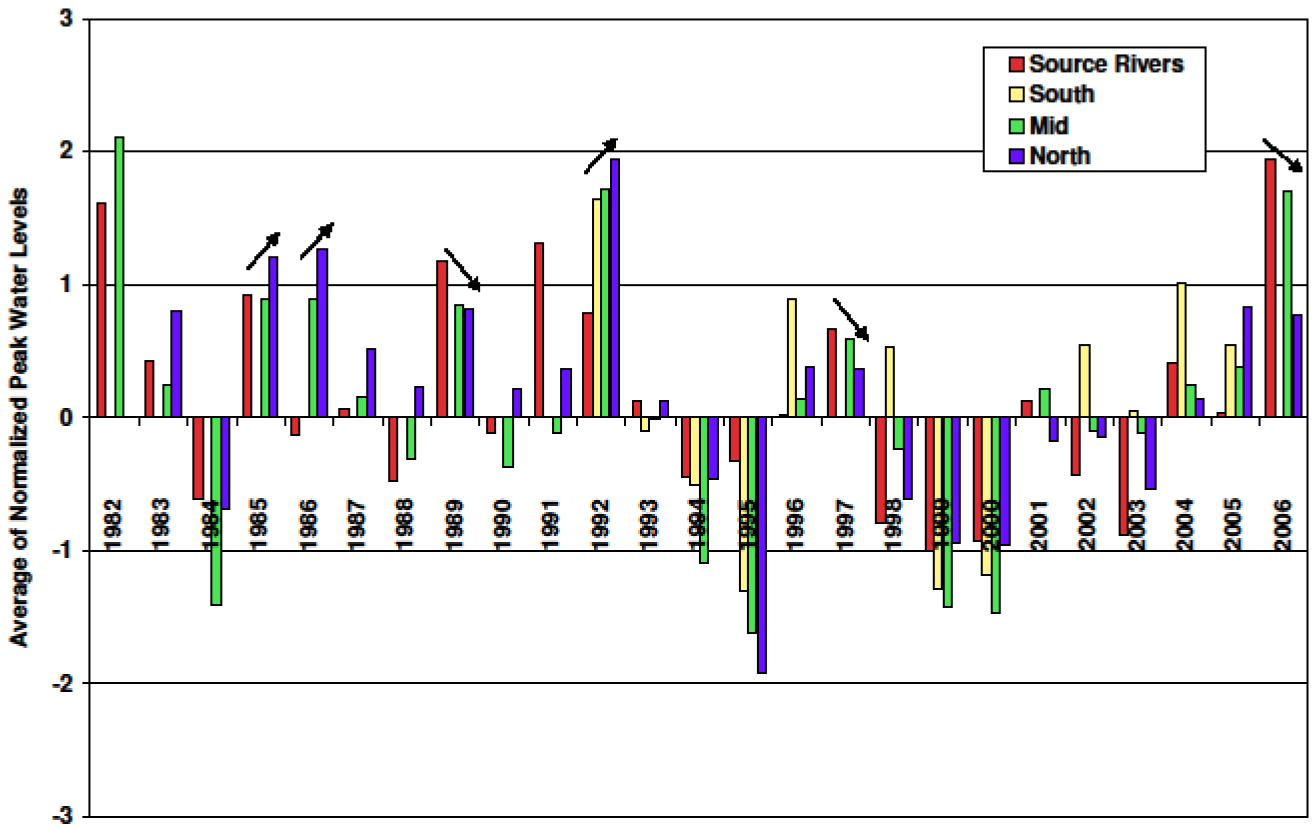


Figure 2-6. Average of normalized peak water levels for west-east transects of source rivers, and south, mid and north delta stations from 1982 to 2006, inclusive of large and extreme years. Upward (downward) arrows indicate events with northward increasing (decreasing) water levels.

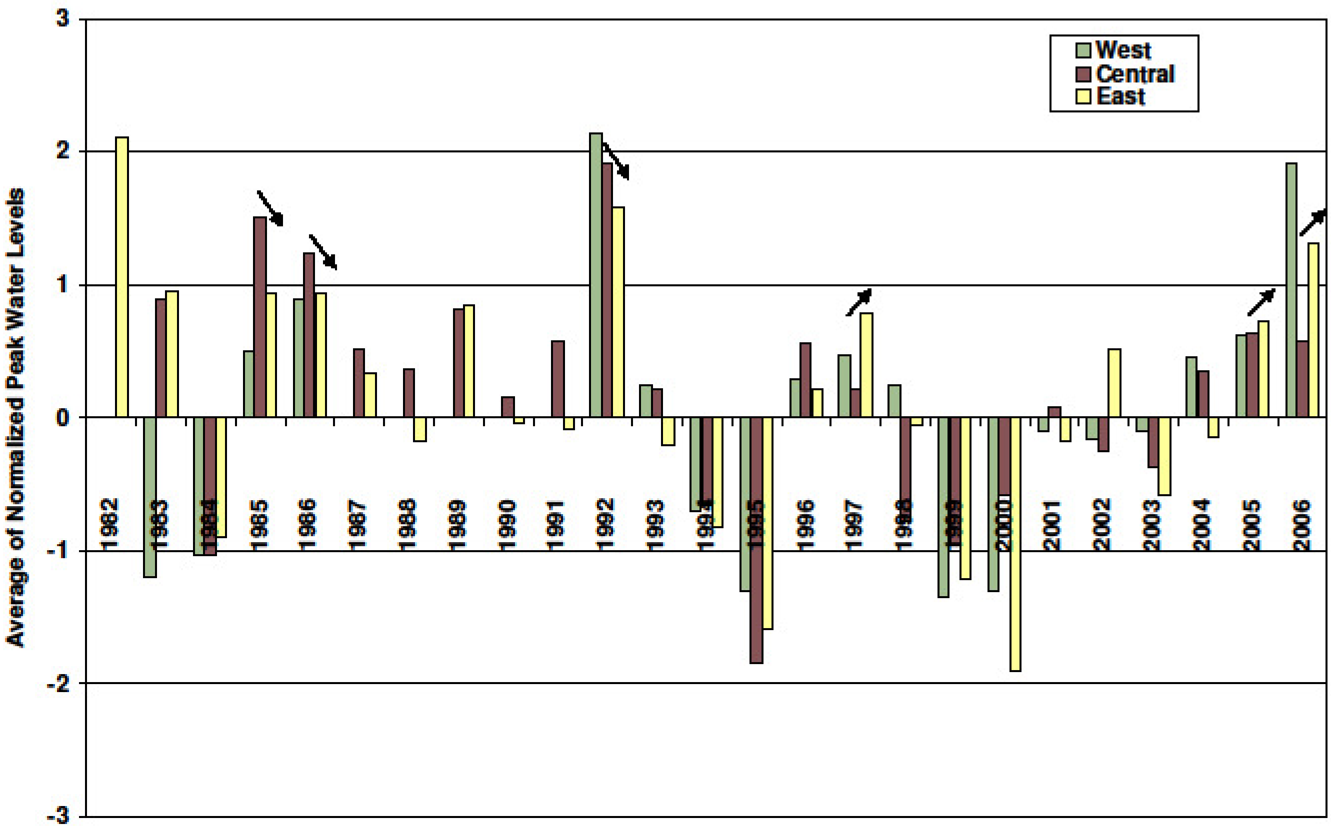


Figure 2-7. Average of normalized peak water levels for south-north transects of west, central, and east delta stations from 1982 to 2006, inclusive of large and extreme years. Downward arrows indicate events with peak levels higher in the central delta than the east, upward arrows indicate the events with the opposite trend.

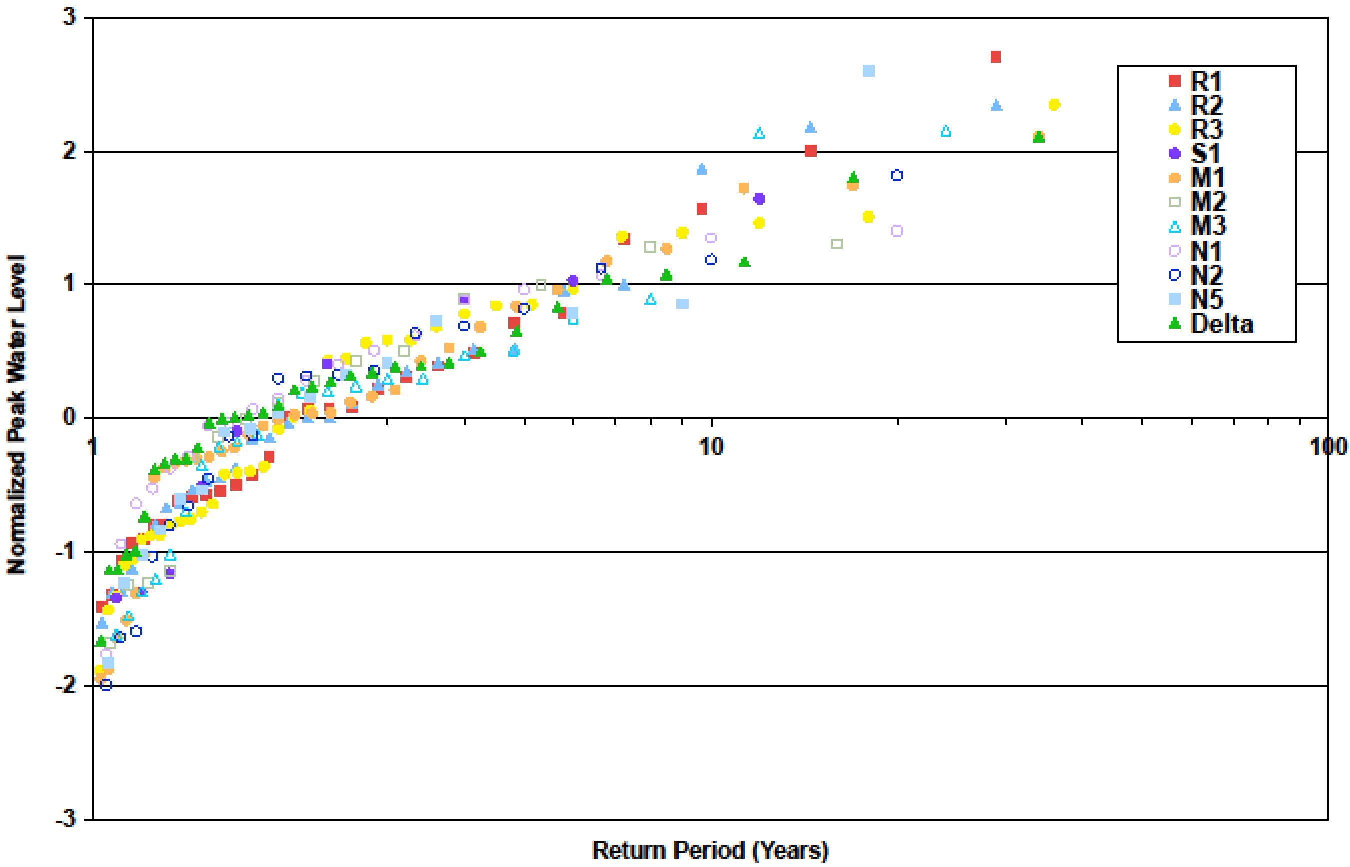
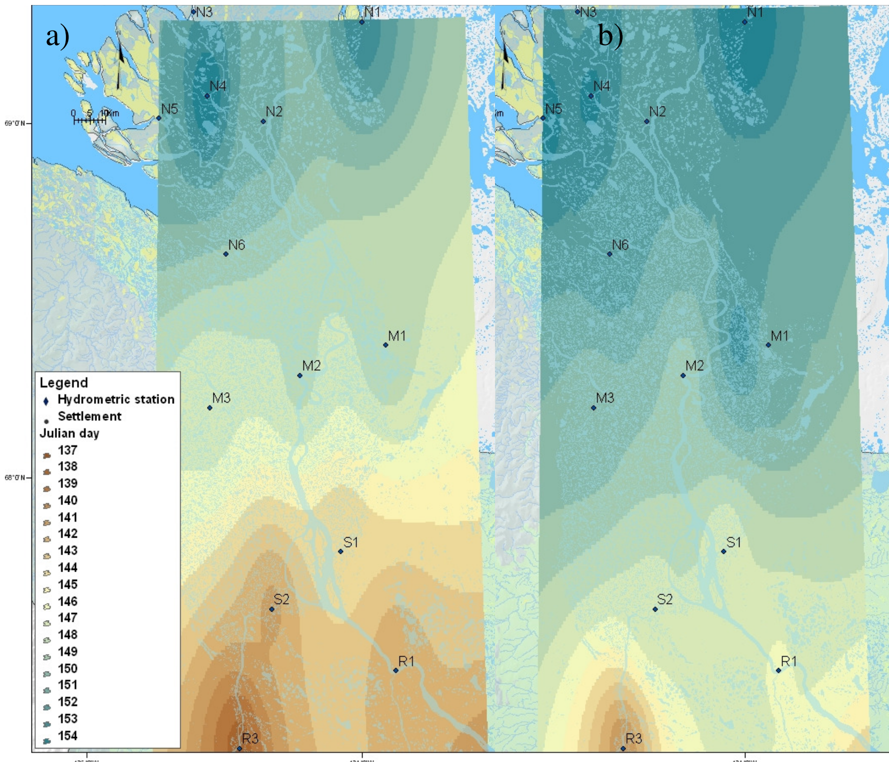


Figure 2-8. Return period plot for delta and source river stations with record lengths exceeding 10 years, including the delta average.



c)	t_b	t_m	last ‘B’
R1	141	146	154
R2	137	139	143
R3	138	141	149
S1	144	147	
S2	141	147	
M1	148	151	157
M2	148	149	
M3	147	150	
N1	153	154	
N2	151	152	
N3		152	
N4	154	154	
N5	152	154	
N6		152	

Figure 2-9. a) Map of average timing of break-up initiation, t_b, for hydrometric stations in the delta, b) map of average timing of peak break-up water level, t_m, for hydrometric stations in the delta, and c) table with average t_b, t_m and last ‘B’ dates.

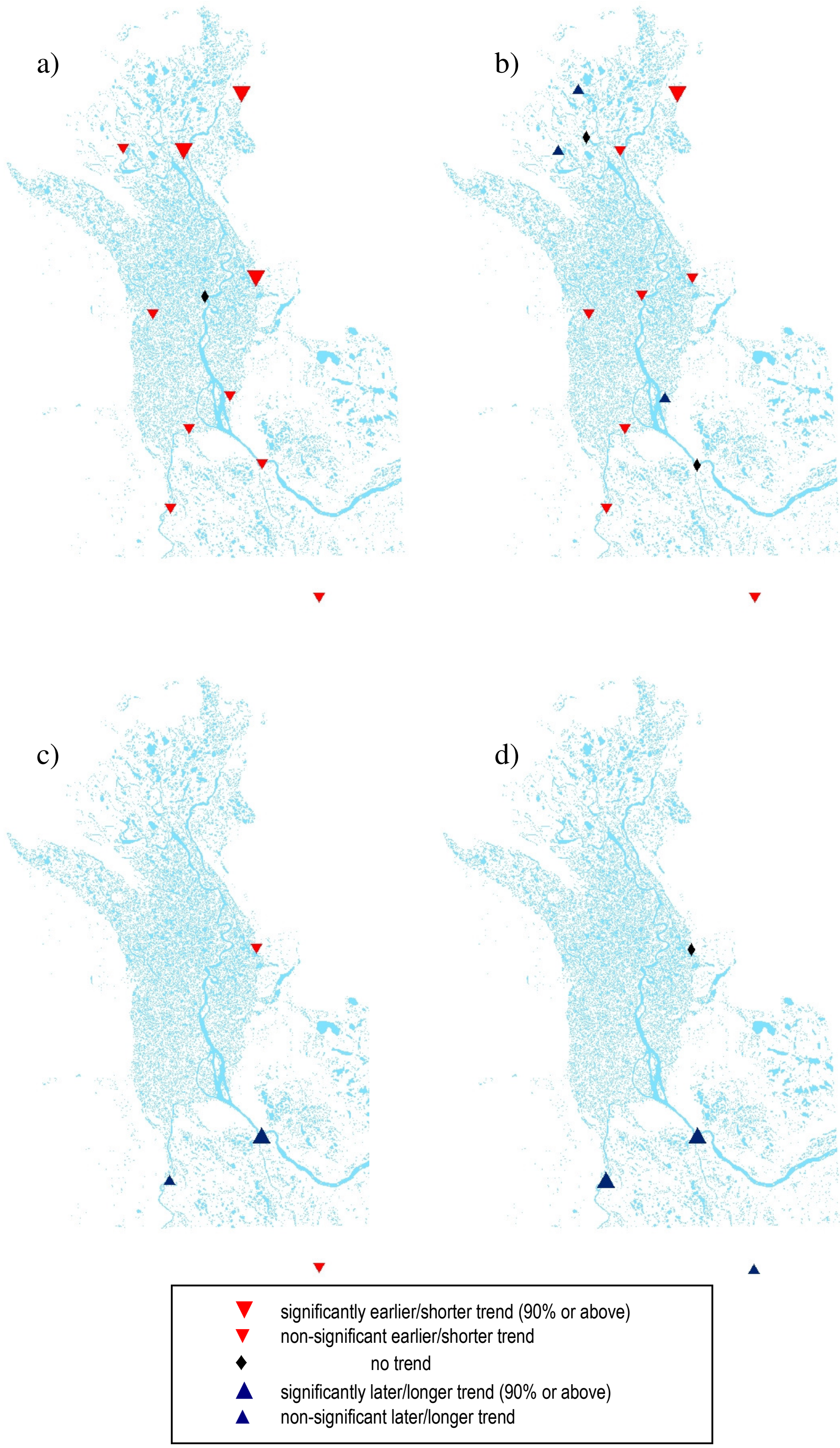


Figure 2-10. Mann-Kendall trends at Water Survey of Canada hydrometric stations in the Mackenzie Delta for a) timing of break-up initiation, t_b, b) peak break-up water level occurrence, t_m, c) last 'B' date, and d) break-up duration, t_3. The number of stations included in c and d is limited by the number of operating discharge stations in the delta (4).

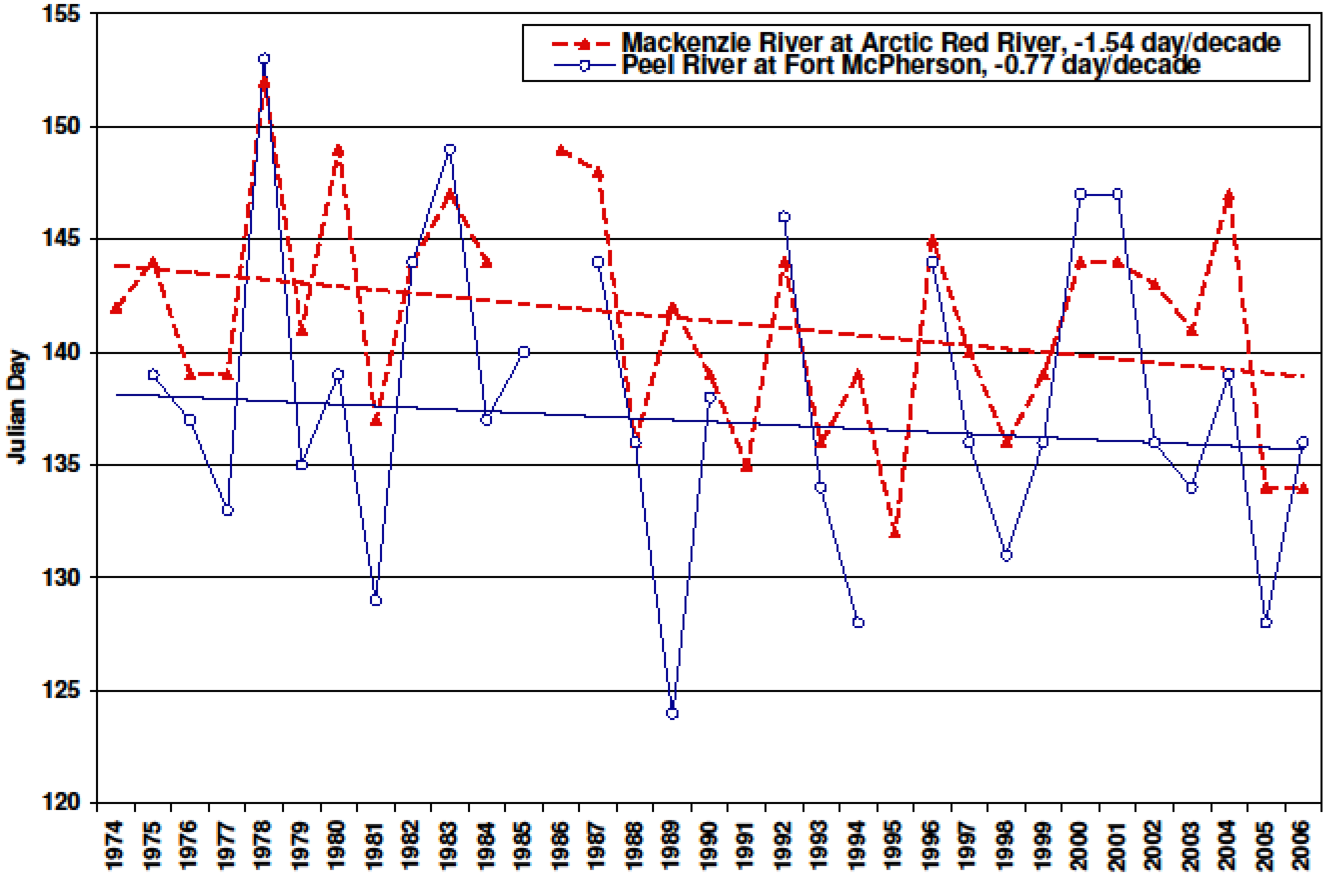


Figure 2-11. Trend of break-up initiation timing, t_b, for the Mackenzie and Peel Rivers, with Sen's slope estimate included in the plot (trends not significant at the 90% level).

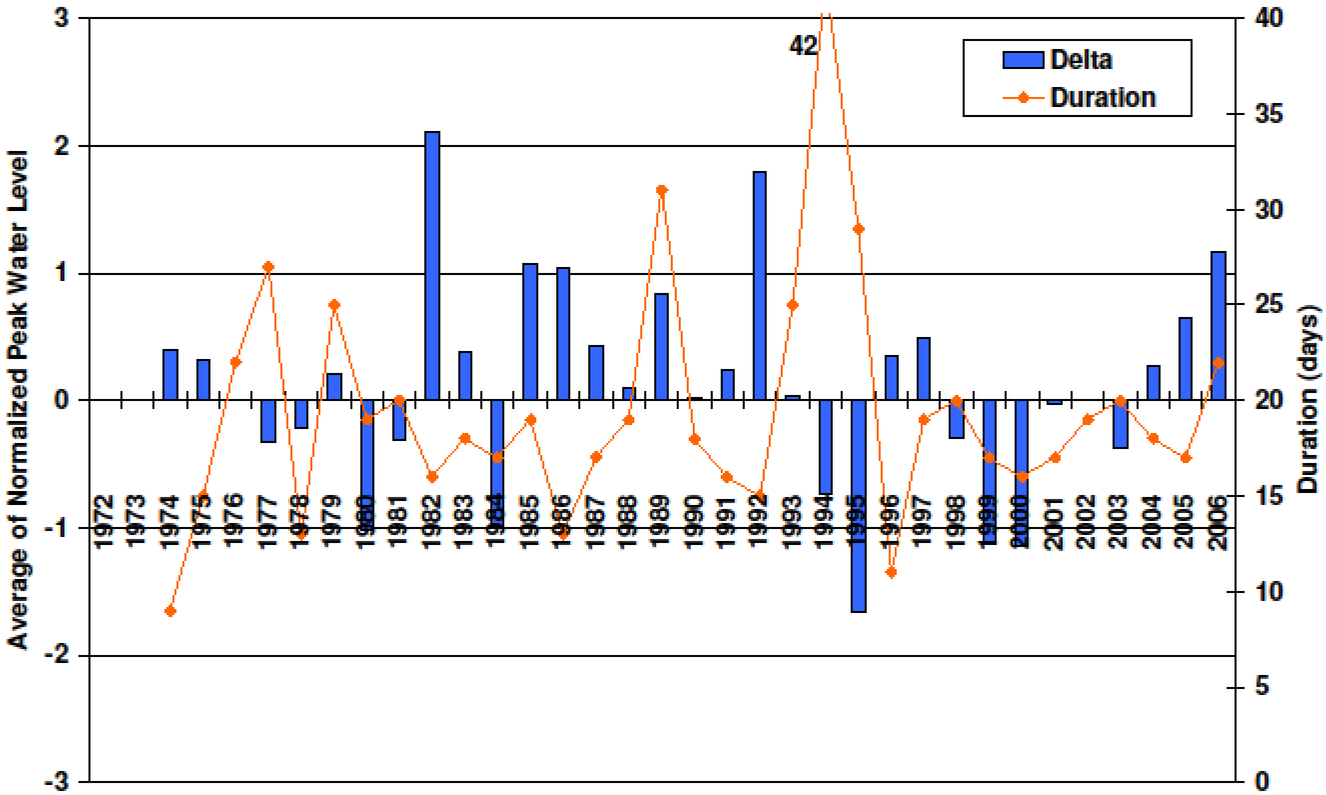
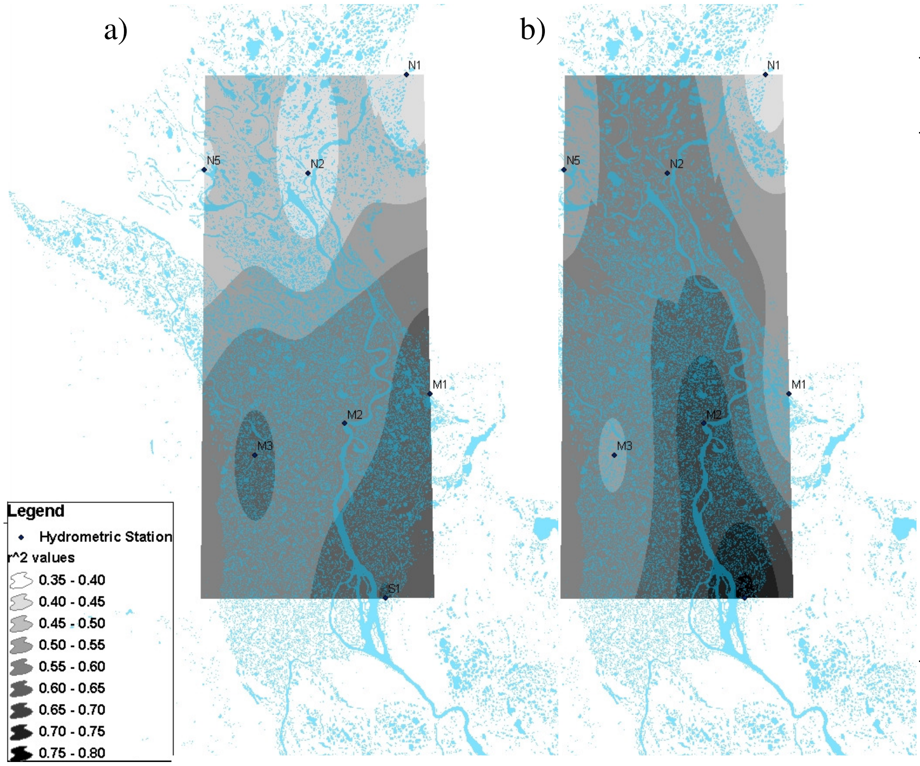


Figure 2-12. Comparison of break-up duration, t_D, and delta average normalized peak water level for 1972 to 2006. Duration refers to the period from the earliest break-up initiation in the delta to East Channel at Inuvik last 'B' date.



c)	MARR H_m	MARR Q_{max}
Delta	0.57	0.48
South	0.71	0.74
Mid	0.60	0.51
North	0.52	0.61
West	0.64	0.56
Central	0.43	0.56
East	0.63	0.51
S1	0.65	0.76
M1	0.63	0.51
M2	0.59	0.68
M3	0.61	0.55
N1	0.42	0.43
N2	0.42	0.59
N5	0.45	0.54

Figure 2-13. Map of r^2 values for the relationship between a) MARR peak break-up water level, H_m, and peak water level at delta stations, b) MARR peak break-up discharge, Q_{max}, and peak water level at delta stations, and c) table with r^2 values for the delta, transects, and individual stations.

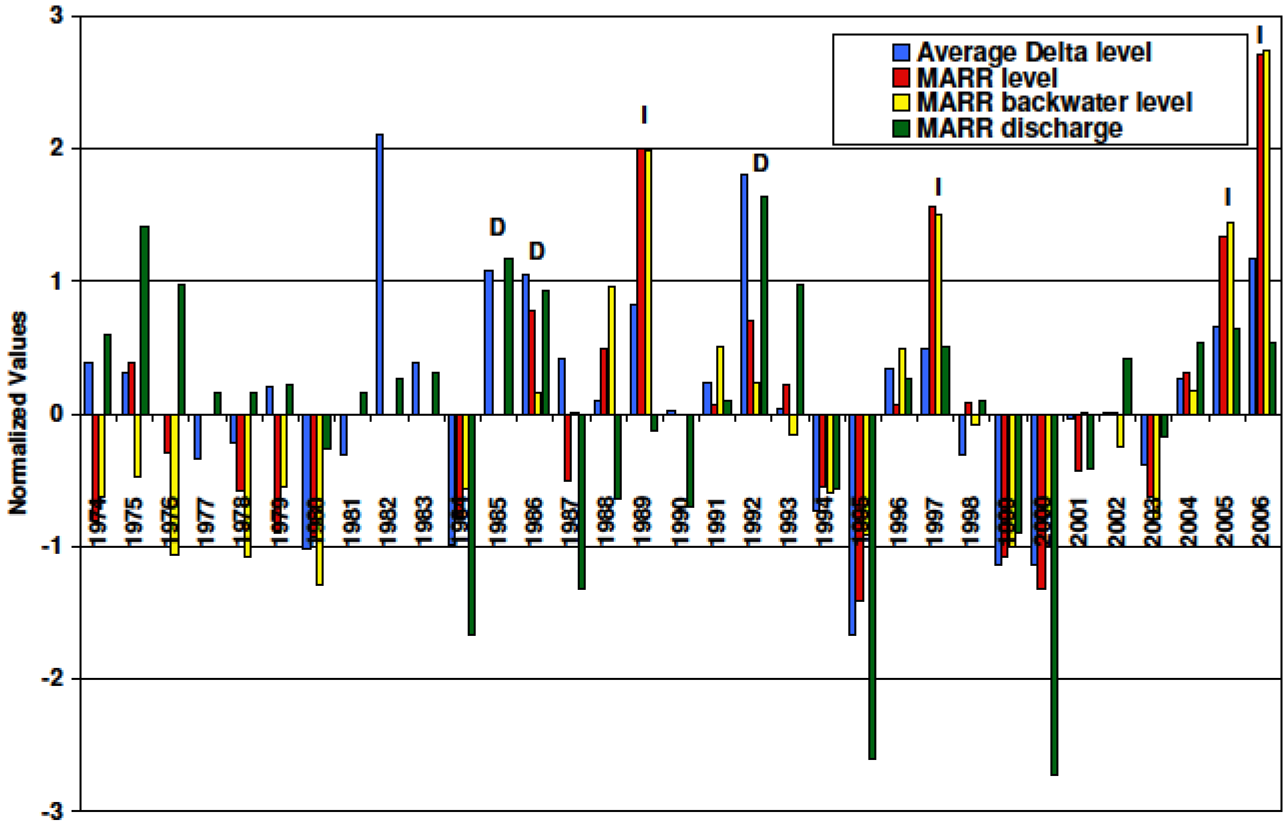


Figure 2-14. Comparison of average normalized peak water level for the delta with normalized peak water level, backwater level, and discharge at the MARR hydrometric station. The symbol D (I) identifies discharge (ice) dominated years. Normalized discharge values of 2, 0, and -3 represent spring peak discharges of 36700, 27500, and 14000 m^3/s respectively.

Chapter 3: Hydroclimatic Controls on the Occurrence of Break-Up and Ice-Jam Flooding in the Mackenzie Delta, NWT

ABSTRACT

Concern has been expressed regarding the impacts of climate change on freshwater aquatic ecosystems in arctic regions. Populated with lakes controlled by flooding from spring break-up ice jams, arctic deltas such as the Mackenzie Delta in northern Canada are particularly sensitive to changing ice break-up conditions and the hydroclimatic controls on break-up and ice-jam flooding. An understanding of these controls is necessary for assessing future effects. This paper presents an assessment of hydroclimatic conditions controlling break-up over the period 1974 to 2006, with a focus on extreme flood events. Both the upstream driving force, capturing elements of the spring discharge hydrograph, and the downstream resistance force, describing the competence of the downstream ice cover, were quantified with reference to the Mackenzie River at Arctic Red River hydrometric station such that the contribution of each to the severity and timing of break-up could be explored. Results show that the severity of peak break-up stage is most influenced by upstream discharge and the balance between upstream and downstream melt, while timing is related to delta ice conditions and the rise of the spring hydrograph. The highest peak stage events require a rapid rise in discharge and high peak discharge. Minimal downstream melting degree-days and greater ice thickness are also important, although no relationship of these appears to control the level of backwater produced from broken ice and ice jamming effects. The pattern of rapid (protracted) upstream melt and lower (higher) intensity melt in the delta characterizes the highest (lowest) break-up events. For the most severe events, upstream forces are important in controlling discharge-driven events, while an altered hydrologic response occurring for ice-driven events was noted, meriting future examination. Finally, trends toward a longer prebreak-up melt interval, lower peak discharge, rate of rise in discharge, and ice thickness, and higher freeze-up stage were observed, with greater variability of these controls and break-up severity in the most recent decade.

Key words: river ice, break-up, ice jamming, hydroclimatic controls, Mackenzie Delta

3.1 INTRODUCTION

The impacts of climate change on freshwater aquatic ecosystems in arctic regions are increasingly being reported (*e.g.* Prowse *et al.*, 2006a,b), as observed and projected warming most intensely manifested in high latitude regions (Anisimov *et al.*, 2007; Walsh *et al.*, 2005; Serreze *et al.*, 2000) influences the cryospheric and hydrologic regimes that define these ecosystems (Bates *et al.*, 2008). Arctic deltas are particularly sensitive to such changes, as these dynamic and productive systems often comprise extensive numbers of lakes controlled by channel water levels. Spring break-up ice jams, and the resulting flooding of riparian landscapes, are a critical hydrologic feature and the main source of recharge for the highest elevation lakes (*e.g.* Marsh and Hey, 1989), as these produce water levels much higher than could be generated by an equivalent discharge under open-water conditions (Beltaos, 1995).

The timing and severity of break-up and ice-jam flooding depend on numerous factors driven by climate, but interacting though a variety of geophysical processes (Prowse and Beltaos, 2002) and operating at a variety of spatial and temporal scales (Prowse *et al.*, 2007; Beltaos and Prowse, in press). These hydroclimatic controls on ice break-up are more complex than the simple empirical relationships established between air temperature and break-up timing (*e.g.* Soldatova, 1993; Magnuson *et al.*, 2000; Bonsal and Prowse, 2003), particularly for deltas, influenced both by basinwide and intra-delta processes. Appropriately, increased emphasis has been placed on these controls, given the questionable future reliability of temperature index approaches (Prowse *et al.* 2007) and a need for the development of physically based river-ice models for future prediction of changes in river-ice regimes (Prowse *et al.*, 2008).

Concern has been expressed regarding the impacts of climate change on the hydroecology of the Mackenzie Delta (Marsh and Lesack, 1996). Preliminary work suggests that the annual connection times of the highest elevation lakes may have declined over the last 50 years (Lesack and Marsh, 2007); this is suggested to be related to reduced river ice jamming. Investigation of spatial patterns and trends associated with the largest break-up floods in the delta over a similar time period, however, reveals a tendency toward break-up events characterized by high peak water levels in the southern delta associated with ice jamming, termed *ice-driven* events, as opposed to more *discharge-driven* events, which result in delta-wide high water levels (Goulding *et al.*, 2009). Thus, an understanding of the hydroclimatic controls influencing these break-up events is a first requirement to determining the impacts of climate change on break-up flooding and the resulting hydrology and ecology of delta lakes.

Little attention has been given to the controlling hydroclimatology of break-up in the Mackenzie Delta, despite its recognized importance. Research is limited to postulations about the influence of ice thickness on break-up and ice jamming in the delta in early reports (*e.g*. Henoch, 1960; Mackay, 1963a; Bigras, 1988), based on limited observations and only narrow consideration of upstream conditions. Thus, the objectives of this study are: (1) to evaluate the hydroclimatic characteristics associated with break-up in the Mackenzie Delta over the 33-year instrumental record, (2) to determine the controlling hydroclimatology of break-up and ice-jam flooding, specifically assessing the relative importance of basinwide and intra-delta forces in controlling the severity and timing of break-up for the most severe events, and (3) to determine the nature of trends and variability within the major hydroclimatic conditions controlling break-up. These objectives also allowed exploratory links to be identified with the larger scale atmospheric and surface temperature conditions prior to the onset of break-up, although a full analysis was beyond the scope of this study.

The paper begins with a description of the Mackenzie Delta and an overview of existing research related to ice break-up in the delta. A detailed background on river ice break-up processes is then given to provide context for the methodology. This is followed by a description of the data and methodology. Results and discussion are then presented followed by conclusions and future recommendations.

3.2 STUDY AREA

The Mackenzie Delta, at the mouth of the Mackenzie River in Western Arctic Canada, is a productive lake-rich environment (Emmerton *et al.*, 2007) covering an area of 12,000 km^2. The Mackenzie River enters the delta at Point Separation north of the confluence of the Mackenzie and Arctic Red Rivers, and travels 200 km through a maze of meandering channels to reach the Beaufort Sea at Shallow Bay in the northwest and Kittigazuit Bay in the northeast (Figure 3-1). The Peel River flows into the delta below Fort McPherson, while several smaller mountain rivers enter the delta on the western edge.

The Mackenzie Basin extends from 52 to 70˚N and 103 to 140˚W, stretching 1.8 million km^2 from central Alberta to the Beaufort Sea and from the Western Cordillera to the Canadian Shield. The Peel River contributes regional flow from the northwestern portion of the Mackenzie Basin, draining an area of 70 600 km^2 between 64 and 67˚N. Within the six main Mackenzie sub-basins, cold temperate, mountain, subarctic, and arctic climatic zones occur. The average annual (1971-2000) air temperature in the delta at Inuvik (~68˚N) is -8.8˚ C, while near the headwaters of the basin at Athabasca (~54˚N) it is 2.1 ˚C. Annual precipitation declines in the basin from west to northeast (Rouse *et al.*, 1997); at Inuvik average annual precipitation is ~250 mm, with 50% falling as snow (Environment Canada, 2008). An increasing trend in winter and spring air temperatures has been observed throughout the Mackenzie Basin over the latter half of the 20th century (Aziz and Burn, 2006), with an average increase of over 1.5°C reported for the period from 1950-98 (Zhang *et al.*, 2000). However, the magnitude of warming has not been uniform throughout the basin (Woo and Thorne, 2003). Trends in precipitation are not consistent, although decreases in total fall and winter precipitation and increases in total spring precipitation have been noted, along with decreases in December, January, and annual snow in the basin and a shift in spring precipitation with decreased snow and increased rain (Aziz and Burn, 2006).

The Mackenzie River provides the majority of delta discharge, contributing on average 2.85 x 10^{11} m^3/year (9037 m^3/s), while Peel River discharge represents 2.14 x 10^{10} m^3/year (679 m^3/s) (Water Survey of Canada, 2008). The main seasonal flow pattern observed for the Mackenzie River entering the delta is characterized as subarctic nival (Woo and Thorne, 2003). Thus, runoff is dominated by high spring flows (Church, 1974), as more than half the annual precipitation melts in a short period and substantially increases river discharge. The highest annual water levels are observed in spring for the Peel and Mackenzie Rivers entering the delta (de Rham *et al.*, 2008a), occurring during the eight-week break-up period in the Mackenzie Basin (de Rham *et al.*, 2008b). No significant trends have been detected for annual flow in the Mackenzie Basin or sub-basins (Woo and Thorne, 2003; Zhang *et al.*, 2001), although increasing trends in December to April monthly streamflow are statistically significant in the basin for 25 to 40 year periods ending in 2000 (Aziz and Burn, 2006), and decreases in annual maximum flows are significant for many sites when considered over certain periods (Aziz and Burn, 2006; Burn *et al.*, 2004). Earlier onset of the spring freshet has also been observed, strongly linked with warming spring temperatures (Aziz and Burn, 2006; Burn, 2008), and likely reflecting earlier onset of snowmelt in the basin. Equally, earlier occurrence of sharp hydrograph rises were noted along the mainstem of the Mackenzie associated with an advance of 3 days/decade for flow entering the delta (Woo and Thorne, 2003). Decreasing trends of ~1.5 days/decade on average were also observed for break-up event variables (onset of break-up, maximum water level) throughout the basin (de Rham *et al.*, 2008b).

Observations of break-up and ice jamming in the Mackenzie Delta dating back to the early 1900s confirm the importance of spring ice break-up (Brown, 1957; Henoch, 1960; Mackay, 1963a; Bigras, 1988) and its role in extreme flood events (Kriwoken, 1983). Equally, investigations into the hydrology of delta lakes underscore the importance of ice-jam flooding, in addition to levee height distribution, in controlling lake flooding regimes (Marsh and Hey, 1989, 1991, 1994; Marsh *et al.*, 1994). Floods from spring break-up also have an influence on the bioclimate in the Mackenzie Delta, as the overflow of sediment laden waters onto the snow- and ice-covered landscapes, channels, and lakes, in addition to the rapid flushing of ice from the delta, drastically alters the surface albedo, thus increasing solar radiation receipt (Gill, 1974). While the resulting increases in air temperature during break-up are small, they occur at a critical time for terrestrial and aquatic plant growth (Hirst, 1987).

Limited focus has been placed on the hydroclimatic controls on break-up in the delta, however several hypotheses have been made about the influence of ice thickness on break-up and ice jamming. Henoch (1960) suggested that ice thickness in the Mackenzie Delta's contributing channels influences the time of break-up and contributes to flooding. Thus, thicker ice on the Arctic Red and Mackenzie Rivers, given greater exposure to snow sweeping northwesterly winds, results in later break-up than on the more protected Peel River. Mackay (1963b) suggested that ice thickness influences the rapidity of break-up, while Bigras (1988) proposed that ice-cover strength in the delta is a central factor controlling the formation of ice jams, in addition to influencing the length of time a jam will remain static and the amount of backwater that builds up.

Break-up patterns and processes are complex in this deltaic environment, due to the influence of both the Mackenzie and Peel Rivers and numerous interconnecting channel-lake systems (Mackay, 1963a), and the varying spatial and temporal expressions of different break-up event types (Goulding *et al.*, 2009). An extra-delta hydrometric station, Mackenzie River at Arctic Red River (MARR), located immediately upstream of the confluence of the Mackenzie and Arctic Red Rivers (R1 in Figure 3-1, photo of gauge site in Figure 3-2), however, provides an opportunity for the analysis of the hydroclimatic controls of break-up, as this station essentially links the basin and the delta. Discharge at MARR represents 95% of the total outflow of the basin and captures the dominant hydrologic signal entering the delta (Woo and Thorne, 2003). Equally, the severity of break-up in the delta is reflected in conditions at MARR, as peak break-up water levels recorded at MARR explain a substantial amount of the variability in peak water levels recorded throughout the delta (r^2=0.57, Goulding *et al.*, 2009).

3.3 RIVER ICE BREAK-UP PROCESSES

Hydroclimatic controls at a variety of spatial and temporal scales influence the nature and progression of break-up events in cold regions. Initial break-up conditions are determined by hydroclimatic conditions during the fall and winter that control freeze-up levels, the thickness and composition of ice, and the size of upstream snowpack available for runoff (Beltaos, 1997). All things being equal, lower freeze-up levels, thicker ice and a larger snowpack would tend to produce conditions conducive to a more severe break-up event. The onset and evolution of spring break-up, however, depend more on the prevailing hydrometeorology during the break-up period. Air temperature and short-wave solar radiation play a dominant role in influencing spring snowmelt and the melting and decay of the ice cover (Prowse *et al.*, 1990).

During the prebreak-up period, increased solar radiation induces snowmelt both on the ice and the land, while higher air temperatures increase heat fluxes to the snow-air and ice-water interfaces (Gray and Prowse, 1993). Once reflective snow and surface ice are ablated, deterioration of the ice from solar radiation rapidly reduces its mechanical strength (Ashton, 1985). Thinning occurs at both the top and bottom of the ice cover, associated with atmospheric heat inputs (Prowse, 1990) and heat gains from increases in water temperature and flow velocity (Marsh and Prowse, 1987). The reduction in the strength and thickness of the ice cover, and its contact to the bed and banks, make it more susceptible to fracture and movement. At the same time, snowmelt runoff from contributing sub-basins increases river discharge. The onset of break-up then occurs when higher water levels fracture the ice cover, while higher flow velocities break the ice sheets into smaller blocks and set the broken ice into motion (Beltaos, 1997).

Within this development of break-up, two extremes can be identified on arctic rivers: thermal and dynamic break-up. Thermal break-up is characterized by extensive melting and decay of the ice cover until it deteriorates in place or is easily entrained by river flows. This can occur given low break-up flows due to a small winter snowpack, protracted melt (Prowse and Marsh, 1989), or climatic conditions favouring thermal ablation of the downstream cover before the start of significant upstream snowmelt (Gray and Prowse, 1993). Dynamic events occur when a large spring floodwave, produced by the rapid melt of the upstream snowpack, propagates downstream before any significant ablation or decay of the ice cover has occurred, often resulting in the incidence of severe ice jams (Gray and Prowse, 1993). Between the two extremes, break-up may involve variations in the timing and size of the spring floodwave and in the deterioration of the downstream ice cover.

Thus, break-up is controlled by the balance between the upstream *driving* force of discharge, and the downstream *resistance* of the ice cover (Prowse and Demuth, 1993). It should be noted that the term *force* employed in this analysis is consistent with everyday-language usage, used to describe opposing influences, rather than with the classical scientific definition. The severity of break-up generally increases with river discharge and ice-cover competence (Ferrick *et al.*, 1992). As such, the most dynamic events are promoted on rivers where upstream melt drives the break-up front into colder downstream reaches with the northern advance of spring warming. Even in such regimes, however, a wide range in break-up severities occurs, due to variability in upstream and downstream hydroclimatic controls (*e.g.* Prowse and Conly, 1998).

The upstream force controlling the initiation of break-up and the occurrence of ice jamming is the spring floodwave. Attributes of the spring discharge hydrograph characterizing the upstream force, and that can be extracted from daily streamflow records, include the magnitude and timing of the peak flow, and the timing of hydrograph rise (Woo and Thorne, 2003). Equally important is the quantification of the steepness of the rising limb of the hydrograph, capturing the rate of melt runoff occurring in the basin.

The downstream resistance of the ice cover is mainly determined by ice thickness, mechanical strength and support-boundary conditions at the time of break-up (Prowse and Demuth, 1993). Ice thickness is initially influenced by the winter growth environment; however, the conditions during the spring ablation period ultimately determine how much resistance it provides at break-up. Thus, growth of the ice cover continues from freeze-up until spring melt commences, after which ablation reduces the thickness of the ice from its yearly peak. Ice thickness measurements are recorded at Water Survey of Canada (WSC) gauge stations during winter flow measurements (Pelletier, 1990), however these are limited during most of the prebreak-up melt interval because of logistic and safety problems associated with collecting discharge and ice measurements over a weakened ice cover (Prowse and Conly, 1998). To extend measurements to the time of break-up, degree-day models are used to estimate ice growth and ablation. For ice growth over the winter period, models based on a simplified Stefan equation have been shown to provide good results when varied to account for different growth environments, while ablation models provide only a crude index of the heat available for melt (Gray and Prowse, 1993).

The strength of the ice cover at the onset of break-up is also an important element of the downstream resistance force. The main driver of ice strength decay is radiative melt from shortwave solar radiation, which increases ice porosity, resulting in the deterioration of the ice cover (Ashton, 1985; Prowse *et al.*, 1990). Unfortunately, no regular measurements of ice strength are conducted and research in this area is limited. Prowse *et al.* (1996) attempted to create an ice strength index based on ice porosity development resulting from inputs of modelled shortwave radiation, however lack of validation data and the absence of a clear relationship with break-up severity limited the utility of results.

The support boundary-conditions of the ice represent the final element of the downstream resistance, however, after hinge cracks have developed, the attachment of the ice to the bed and banks is only a minor factor contributing to resistance (Prowse and Conly, 1998). An important factor determining the development of hinge cracks is the freeze-up level of the ice. Defined as the average level during the week following the formation of a complete ice cover (Beltaos, 1997), the freeze-up level is thought to give an indication of the spring water level above which the ice is detached from the banks. Higher freeze-up levels are postulated to limit the severity of spring break-up, as all things being equal, higher break-up levels would be required to detached and fracture the ice cover (Prowse and Conly, 1998; Beltaos *et al.*, 2006)

An assessment of the relative contribution of these forces to the severity of historical break-up events can provide insight into the importance of each in influencing the occurrence of break-up and ice-jam flooding. For example, in the Peace Athabasca Delta (PAD) in the headwaters of the Mackenzie Basin, investigation into the importance of these hydroclimatic drivers in controlling break-up and ice-jam flooding has significantly increased understanding of the delta flooding regime (Prowse and Lalonde, 1996; Prowse and Conly, 1998; Beltaos *et al.*, 2006; Peters and Prowse, 2006; Peters *et al.*, 2006; Romolo *et al.*, 2006a,b; Wolfe *et al.*, 2006). Thus, the characterization of the upstream driving force, capturing elements of the spring discharge hydrograph, and the downstream resistance force, mainly describing the thickness and degree of deterioration of the downstream ice cover, provide a basic framework within which to assess the contribution of extra and intra-delta hydroclimatic forces to the severity of break-up in the Mackenzie Delta.

3.4 DATA AND METHODOLOGY

To meet the objectives of evaluating the hydroclimatic characteristics of break-up, assessing the importance of basinwide and intra-delta forces in controlling the severity of break-up, and exploring trends in hydroclimatic controls, data from a wide variety of sources were used to characterize the upstream and downstream hydroclimatic controls. The quantification of these controls, and of break-up severity and timing, along with subsequent statistical, qualitative, and trend analysis of their role in influencing break-up in the Mackenzie Delta are described in the following sections.

3.4.1 Hydroclimatic Data

Three main elements of the spring discharge hydrograph and five descriptors of downstream ice conditions were used to statistically and qualitatively analyze the importance of upstream and downstream controls to break-up severity and timing, and investigate time-series trends. The maximum break-up discharge (Q_{max}), the timing of the spring streamflow pulse (t_p), and the rate of rise in discharge of the rising limb of the spring hydrograph (k_Q), were used to quantify the upstream driving force of spring discharge, while the yearly peak ice thickness (h_a), the ice thickness at the time of break-up (h_b), the timing of the initiation of spring melt (t_a), the degree-days of thaw above –5ºC leading up to the initiation of break-up (S_5), and the freeze-up level (H_f) were used to characterize downstream ice competence. These were extracted, calculated, or modelled using hydrometric and meteorological data over the period of hydrometric record in the delta from 1974 to 2006.

The upstream elements, Q_{max}, t_p, and k_Q, were directly extracted and calculated from daily discharge records at MARR. Q_{max} represents the daily maximum discharge within the break-up period in m^3/s. Characterizing the hydrograph rise, the timing of the spring streamflow pulse, t_p, is defined as the day when the cumulative departure from the year's mean flow is most negative (Cayan *et al.*, 2001) as illustrated in Figure 3-3. Finally, the rate of rise in discharge (k_Q) in m^{3}s^{-1}day^{-1} is calculated as:

$$k_Q = \frac{Q_{\max} - Q_p}{t_{\max} - t_p} \qquad [1]$$

where Q_p is the discharge at the onset of the spring streamflow pulse, and t_{max} is the timing of maximum break-up discharge. Missing t_p and k_Q values in 1997 and 1998 are the result of missing discharge data. An important caveat to be made with the use of discharge-based factors is the uncertainty related to the measurement and computation of discharge under ice conditions (Pelletier, 1990), with the relative error of daily discharge reaching as high as 45% in extreme cases (Shiklomanov *et al.*, 2006). Thus, the quality of WSC discharge estimates prior to and during the break-up period represents an important constraint on the accuracy of these factors.

The downstream elements of yearly peak ice thickness (h_a) and the ice thickness at the time of break-up (h_b) were modelled using mean daily air temperature from the Inuvik airport (Meteorological Service of Canada (MSC) ID 2202570). The Stefan equation (Michel, 1971) was used for ice growth to the peak ice thickness, h_a, just prior to the initiation of spring melt:

$$h_i = \alpha\sqrt{D_f} \qquad [2]$$

where h_i is ice thickness (m), α is a coefficient varied to account for conditions of exposure and surface insulation (m °C$^{-1/2}$ day$^{-1/2}$), and D_f is the accumulated degree-days below freezing (°C day) from the onset of freeze-up. The onset of freeze-up is determined from water level charts and daily water level tables as the peak water level after backwater conditions are apparent on the charts, subsequently followed by decreasing winter water levels (Beltaos *et al.*, 1990). After the initiation of spring melt, ice thickness reductions from h_a to h_b were modelled using Bilello's (1980) approach:

$$\Delta h_i = \tau D_t \qquad [3]$$

where Δh_i is the change in ice thickness (m), τ is an empirical coefficient found to range between 0.004 to 0.01 for northern (i.e. > 60°N) Canadian and Alaskan rivers (m ºC^{-1} day^{-1}), and D_t is the accumulated thawing degree-days (ºC day) above a base of –5ºC. Ice thicknesses recorded at MARR and East Channel at Inuvik (WSC ID 10LC002, M1 in

Figure 3-1) hydrometric stations were used to determine suitable values of the constants α and τ. Given limited data available for the validation of the ablation model, spring degree-days of thaw above –5ºC at the onset of break-up (S_5) were also considered independently within the downstream factors, representing the degree of prebreak-up melt to the ice cover.

To identify the date of the initiation of spring melt, required to separate the periods of ice growth and ice ablation, a procedure outlined by Prowse *et al*. (1996) was used. Daily snow on the ground records collected at the Inuvik airport were examined to identify the point associated with pronounced snow melt. Maximum, mean, and minimum daily air temperatures from the Inuvik airport were analyzed to ascertain when air temperatures begin to rise and remain above 0ºC. Daily discharge for small index streams (characteristics given in Table 3-1) obtained from WSC were also assessed to determine the beginning of significant stream discharge increases. These data were then plotted together such that a common day, using the above criteria, could be identified as the date of melt initiation, as seen in the example of 1992 (Figure 3-4).

Gaps in the temperature and snow on the ground records used for ice modelling represented less than 5% of the data. Infilling of missing values at Inuvik was undertaken using an anomaly-based approach. Using records from the Aklavik (MSC ID 2200100) airport (~50km away), daily climate normals were created for each parameter (*e.g.* maximum air temperature) over the period of available record from 1981 to 2006. Missing values at Inuvik were then estimated by applying the existing anomaly for that day at the Aklavik airport to the Inuvik daily normal.

The freeze-up level (H_f) of the ice at MARR is included in some parts of the analysis, and was determined following methods outlined in Beltaos (1997). Ice strength was not considered given a dearth of atmospheric and ice data available for modelling.

3.4.2 Severity and Timing of Break-Up Flooding

The severity of break-up was represented by the peak water level observed during break-up (H_m) at the MARR station, while the backwater level (ΔH) and the maximum break-up discharge (Q_{max}) were also used to delineate ice and discharge-driven events as defined in Goulding *et al.* (2009). The timing of break-up initiation (t_b) and its stage (H_b) were also considered.

The peak break-up water level (H_m) corresponds to the maximum instantaneous stage during break-up, available from WSC water-level charts and from published records when it was determined to be the maximum stage for the year. Alternatively, reliable high water marks published by WSC or mean daily water levels, when available, are used as the H_m in years when the gauge is moved or damaged during break-up. H_m values were missing for the years 1981, 1982, 1983 and 1985.

Based on the break-up chronology assembled for the Mackenzie Delta from 1974 to 2006 (Goulding *et al.*, 2009), the years 1982, 1985, 1986, 1989, 1992, 1997, 2005, and 2006 had the highest average peak water levels in the delta. These correspond with MARR peak break-up water levels exceeding a threshold of 14 masl, as shown in Figure 3-5, compared with the station open-water stage-discharge curve.

Two distinct types of events, ice and discharge-driven, were identified in Goulding *et al.*'s (2009) analysis of spatial and temporal patterns and key processes for these high water years. The highest peak stages occurred at MARR for ice-driven events (1989, 1997, 2005, 2006; 1982 anecdotally) associated with high levels of backwater (Figure 3-6, years denoted by the letter 'I'). Backwater level (ΔH) refers to the difference between ice-affected stage and that for an equivalent discharge under open-water conditions (Beltaos, 1995). The high backwater levels associated with these events thus indicate that ice resistive effects (such as ice jamming) are a dominant driver of increased water levels. Conversely, discharge-driven events (1985, 1986, 1992) were characterized by high maximum break-up discharge (Q_{max}) and lower peak water and backwater levels (Figure 3-6, years denoted by the letter 'D').

The timing of break-up initiation (t_b) was determined following guidelines outlined in Beltaos *et al.* (1990), informed by observations available from site visits, break-up monitoring, air photographs, and satellite images (Goulding *et al.*, 2009). The water level observed at the onset of break-up (H_b), while not used as an indication of break-up severity, is also considered in the hydroclimatic analysis.

An important assumption is the homogeneity of the MARR station over the 33 years of record. Prior to 1985, records were collected ~16km upstream of the current MARR hydrometric station (WSC ID 10LC014) at the Mackenzie River *above* Arctic Red River station (WSC ID 10LA003). Within the WSC archives, however, a water level adjustment factor was applied to the pre-1985 data to create a homogenous data series.

3.4.3 Hydroclimatic Controls on the Timing and Severity of Break-up

To assess the importance of hydroclimatic conditions in controlling the timing and severity of break-up, correlation analysis was conducted between H_m and each of the upstream (Q_{max}, t_p, k_Q) and downstream (h_a, t_a, h_b, S_5) forces. Multivariate regressional relationships were also explored using stepwise and backward elimination methods. Additional analysis was conducted for sub-populations based on the severity of break-up water levels. These were defined based on the *normalized* peak water level ($\hat{H}_{m_i}$) for each year (i), calculated as:

$$\hat{H}_{m_i} = \frac{\left(H_{m_i} - \overline{H}_m\right)}{\sigma} \qquad [4]$$

Thus, yearly spring peak water levels were compared based on their number of standard deviations (σ) from the long-term mean ($\overline{H}_m$). Three groups were defined as:

High H_m events	($\hat{H}_m > 0.5$)
Medium H_m events	($-0.5 < \hat{H}_m < 0.5$)
Low H_m events	($-0.5 > \hat{H}_m$)

This classification was also used for comparison between upstream and downstream value (Q_{max}, h_a, h_b) and rate (k_Q and S_5) based variables, while classifications of timing (t_b, t_p, t_a) were defined based on normalized values of the date of occurrence, for example, t_b:

Early events	($-0.5 > \hat{t}_b$)
Average events	($-0.5 < \hat{t}_b < 0.5$)
Late events	($\hat{t}_b > 0.5$)

This separation of events, particularly the peak break-up water level, H_m, allowed specific consideration of the hydroclimatic drivers in the years of highest peak water level. A Mann-Whitney U test was applied to determine if the average variables representing hydroclimatic conditions during high H_m events differed significantly from the mean for low H_m events.

The balance between the upstream and downstream forces for high H_m events were considered qualitatively; specifically, the nature of the ablation period in the upstream portions of the basin and in the delta downstream was inferred from comparisons between the rate of rise in discharge and the degree-days of thaw**.** The assessment was made using the rank of each value, based on the above separation of events, such that: high k_Q and S_5 values were ranked 1, medium events were ranked 2, and low events were ranked 3. This evaluation also allowed exploratory links to be made with larger-scale atmospheric and surface temperature conditions prior to break-up. In addition, differences in the hydroclimatic conditions coinciding with ice and discharge-driven years were investigated qualitatively using normalized values.

3.4.4 Trend Analysis

Trends in the time-series of hydroclimatic controls were assessed using the Mann-Kendall non-parametric test (Mann, 1945; Kendall, 1975) and Sen's slope estimator (Sen, 1968), contained the Salmi *et al.* (2002) MAKESENS template in Excel.

3.5 RESULTS AND DISCUSSION

3.5.1 General Characteristics of Hydroclimatic Controls

The data extracted, calculated and modelled for the analysis of upstream and downstream forces are given in Table 3-2. The least severe break-up events, with the lowest peak break-up water levels observed at MARR, occurred in 1995, 2000, and 1999 (listed by rank), while the highest peak stages were observed during break-up in 2006, 1989, 1997, and 2005.

Peak break-up discharges recorded at MARR have ranged from 15200 m^3/s in 2000 to 35000 m^3/s in 1992 (also one of the more severe events), with an average of ~27600 m^3/s. The lowest rates of rise in discharge occurred in 2000 and 1987, while the large events in 1985 and 2005 saw the highest rates, an order of magnitude larger. The shapes of the spring hydrograph that these rates of rise represent (in addition to two medium years 1977 and 1996) are shown in Figure 3-7. Considerable variations in the timing of the spring streamflow pulse are noted even for events with comparable rates of rise (*e.g.* t_p = 129 and 137 for 2005 and 1985). The highest rates of rise in discharge occur in years of above average peak discharge (Figure 3-8a). However, a range of k_Q values is noted even for events with the highest Q_{max} (Figure 3-8b), while low peak discharges are always characterized by low rates of rise.

Results of the ice thickness modelling carried out for MARR, and corroborated with modelled ice growth and ablation at East Channel at Inuvik, are shown in Figure 3-9. For winter ice growth, ice thickness measurements recorded at MARR are plotted against the associated value of $\sqrt{D_f}$ (Figure 3-9a). The best fit was obtained with α=0.02 (r^2=0.55); for Inuvik the best results were obtained when α=0.018 (r^2=0.87, not shown). Both of these values are reasonably close to the range of 0.014 to 0.017 suggested to represent conditions on an “average river with snow” (Michel, 1971), and to the value of 0.018 obtained at Peace Point in the southern reaches of the Mackenzie Basin (Prowse and Conly, 1998; Beltaos *et al.*, 2006). For validation of the ablation model, a limited number of ice thickness measurements available during the ablation period required that the fit for τ be determined by comparing the modelled ice thickness (with growth to a peak thickness at t_a, and ablation thereafter) to these measurements (Figure 3-9b). A reference 1:1 line is included in the plot, showing a considerable amount of scatter. Such a result could be expected, since degree-day models do not reflect all the heat fluxes driving the ablation of an ice cover, including radiative and convective fluxes. The most suitable value of τ for the MARR data was 0.0027 with an r^2 of 0.34, while at East Channel at Inuvik it was 0.0029 with an r^2 of 0.39 (not shown). Both of these values are below Bilello's (1980) suggested range of 0.004 to 0.01, while being comparable to recent results in more southerly locations (Table III in Beltaos *et al.*, 2006).

The resulting peak ice growth and ice thickness at break-up initiation are shown in Figure 3-10 with the associated degree-days of thaw. The average peak ice thickness at MARR is 1.26 m, however of particular note in the graph, ice thickness does not exceed ~1.30 m from the early 1990s onward, while regularly exceeding this value in the 1970s and 1980s. Ice thickness at the time of break-up varies considerably from the average of 0.97 m, ranging from 0.51 m in 1998 associated with the highest amount of melt and a thickness reduction of ~55%, to 1.22 m in 1983 and 2000 with thickness reductions of 12 and 4% respectively. The average reduction from the peak thickness is ~20%. Spring degree-days of thaw ranged from ~50 to 250 over the period of record. As with peak ice thickness, greater degree-days of thaw and melt have been observed in the ablation period since 1990. Accordingly, years with high spring degree-days of thaw and related ice ablation generally have the earliest occurrences of melt initiation.

A comparison between the freeze-up stage and the break-up onset stage (Figure 3-11), shows all values exceeding the 1:1 reference line, indicating that higher break-up levels are required to dislodge the ice cover formed the previous fall. On average, H_b is 3.8 m higher than H_f, although the difference ranges from 1.2 m (in 2002) to 7.5 m (in

2006). The large amount of scatter, and the resulting r^2 value of 0.11, are indicative of the many additional effects influencing the outcome of the spring break-up event.

3.5.2 Statistical Analysis of Controlling Hydroclimatology

Correlation analysis reveals that the best relationships with break-up severity were obtained for upstream discharge related factors, while for the timing of break-up, downstream ice resistance forces and the rate and timing of spring hydrograph rise tended to produce the highest correlations (Table 3-3). Correlations between H_m and the discharge factors Q_{max} and k_Q had Spearman's R values of 0.61 and 0.55 respectively, significant at the 95% confidence level. The correlations with h_b and S_5 were 0.15 and –0.20. These results suggest that discharge is a dominant driver of high break-up water levels, necessary as an initial condition for ice-induced flooding. The direction of the relationships substantiate physical theory suggesting that the most severe break-ups causing the largest rises in water level occur when a large spring floodwave ($\uparrow Q_{max}$) caused by rapid upstream melt ($\uparrow k_Q$) meets a thick ($\uparrow h_b$) minimally decayed ($\downarrow S_5$) ice cover.

For break-up timing, correlations with the downstream ice factors h_b, S_5, and h_a were 0.44, -0.37, and 0.37 respectively, all significant at the 95% level. The timing of the spring streamflow pulse, t_p, however, had a stronger relationship with t_b than the initiation of spring ablation, t_a (R=0.41 at the 95% confidence level and R=0.30 at the 90% level, respectively), while the correlation with k_Q was -0.27. These relationships imply that later break-ups generally occur with thicker ice covers. Similarly, fewer degree-days of melt in the spring, which would effectively limit ablation of the ice cover, and a late and slow rise in discharge from late or protracted snowmelt in the basin, are also associated with later break-up onset.

These relationships were further explored using multivariate regression, although several omissions and transformations were required to meet model assumptions. Given significant multicollinearity between h_b and S_5 (R=-0.91, correlation matrix not shown), the derived variable h_b was omitted. Assumptions of normality were valid for most k_Q values, although three large outliers, evident in Figure 3-8b, appeared to exceed a threshold above which a different set of relationships dominated. These were removed from the model, but their hydrologic significance merits further investigation. One extreme high outlier was also removed from the S_5 values. Finally, Q_{max} was included in the model using a logarithmic transformation, given values and a range that were orders of magnitude larger than those of the other hydroclimatic factors, while three low outliers were also removed. With these adjustments, the resulting multivariate analysis tended to focus primarily on hydroclimatic drivers in non-extreme circumstances.

The multivariate regression performed for H_m identified S_5 and k_Q for inclusion in the stepwise model (Table 3-4, r^2=0.32). These were also the only significant variables in the backward elimination model. These results indicate that beyond the dominant influence of discharge, particularly its influence for the lowest and highest severity events, the relative amounts of upstream and downstream melt play an important role in determining the severity of break-up. For break-up timing, both the stepwise and backward elimination methods incorporated t_p and k_Q (r^2=0.35), thus highlighting the importance of the timing and rate of upstream melt to break-up timing.

Mean values for the elements of upstream and downstream hydroclimatic controls for the entire series and the three sub-groupings of break-up water level are shown in Table 3-5. Of note, the average values of hydroclimatic controls for high H_m events are consistently larger (smaller for S_5) than the series mean and that of medium and low H_m events. The Mann-Whitney U test for difference of means between high and low H_m events was significant at the 95% confidence level for Q_{max}, k_Q, and S_5, and at the 90% level for h_b. The difference between groupings is further illustrated in Figure 3-12, showing the relationship between both the maximum break-up discharge and the ice thickness at break-up initiation with the peak stage during break-up. Apparent in the scatterplot, all high H_m events are associated with discharge and ice thickness exceeding or close to the mean for the entire series. A similar outcome recurs for all upstream and downstream controls; thus, a set of thresholds can be defined that must simultaneously be exceeded (unsurpassed for S_5) to produce a high H_m event:

Q_{max}	= 27000 m^3/s	h_b	= 1.00 m
t_p	= Julian day 129 (May 9th)	t_a	= Julian day 125 (May 5th)
k_Q	= 1000 $m^3s^{-1}day^{-1}$	S_5	= 100 ◦C day

Not included in the delineation of these thresholds are the anomalously high S_5 and low h_b values calculated in 1997 (163 °C day and 0.87 m respectively), which do not agree with the above described observations of low downstream melt for high H_m events. Regardless, this composite threshold effectively isolates high H_m events; only one non-high H_m event is associated with controls exceeding all defined thresholds (1993).

Equally noteworthy in Figure 3-12, is that the relationships of increasing break-up water level with increasing hydroclimatic controls do not persist for the highest H_m events. Correlations between H_m and the hydroclimatic controls are all negative for this group, as the highest peak water level events in the high H_m grouping tend to be associated with lower magnitude hydroclimatic factors.

Thus, discharge above 27000 m^3/s and a rate of rise above 1000 $m^3s^{-1}day^{-1}$ are necessary conditions for the occurrence of high H_m events, however downstream ice conditions ultimately determine its effectiveness in producing the highest peak water levels. Particularly, the combination of upstream and downstream melt is an important determinant on peak break-up water level, although no upper threshold of melt or lower limit of ice thickness, appears to control the level of backwater produced from broken ice and ice jamming effects. In fact, a noticeable change in physical processes is evident within the highest severity events, as evidenced by different dominating statistical relationships between the peak break-up stage and hydroclimatic factors, and within hydroclimatic factors. These could be attributed to an altered hydrologic response given near bankfull conditions. The strength of the ice cover, influenced by shortwave solar radiation, could equally provide a missing piece of the puzzle, while the freeze-up stage also appears to play a role in influencing break-up water levels. In terms of break-up timing, thicker ice covers and later protracted melt are responsible for later break-up events.

3.5.3 Comparisons of Upstream and Downstream Melt

The relationship between the rate of rise in discharge and the degree-days of thaw for each of the break-up severity groupings is shown in Figure 3-13, representing the balance between the rate of ablation of the upstream snowpack and the intensity of downstream melt during the prebreak-up period. High H_m events cluster to the upper left side of the graph with a combination of high k_Q values and low S_5 values. When the respective ranks are compared, (k_Q, S_5), all seven high H_m events have a (1,2) or (2,3) combination (Table 3-6), indicating that high peak water events in the delta require the presence of a strong thermal gradient between the upstream and downstream reaches. As such, a south to north progression of warm temperatures and melt appears to drive ice break-up in these years. The pattern is also present for break-up in 1993, (1,3), although this event did not produce a large break-up stage at MARR or in the delta, while high levels did occur on the Peel River at Fort McPherson (R3 in Figure 3-1).

Interestingly, most yearly break-up events in the delta do not appear to be driven by this strong gradient in the sequential warming pattern. Medium H_m events span the range of rising discharge types, but always with an equivalent or lower rank of thawing degree-days, signifying greater levels of downstream melt. Low H_m events are characterized by low or medium rates of rise in discharge combined with high and medium intensity melt in the delta. In addition, the pattern of low discharge rise and minimal melt for several low and medium events is suggestive of colder conditions across the entire basin, limiting prebreak-up melt. The combination in 1995 (3,1) represents an extreme exception to the latitudinal thermal gradient, as early protracted melt of a small snowpack and high intensity melt of delta ice produced the earliest and least severe break-up recorded in the Mackenzie Delta (Goulding *et al.*, 2009). Corroborating this result, de Rham *et al.* (2008) found that in 1995 only 4 days separated break-up initiation at the outlets of the Peace and Mackenzie mainstems, ~1400 km apart.

3.5.4 Large-Scale Controls

A preliminary examination of large-scale surface temperature controls shows that the sequential progression of melt from south to north (upstream to downstream) occurring during years of high peak stage coincides with a large temperature gradient between the basin and the delta regions. Conversely, in years with the lowest peak stage events, the south to north thermal gradient is still apparent but greatly reduced. This is expected given the importance of temperature to the opposing controls of snowmelt runoff in the basin and ice melt in the delta. Furthermore, these temperature gradients appear to be linked to mid-tropospheric circulation prior to the onset of break-up. In particular, the largest north-south gradients are associated with westerly and/or southerly components in the direction of upper-level flow indicating origins from a milder Pacific or a warm continental interior. However, the warm air advection is limited to the southern (upstream) portion of the basin with cold air streams dominating northern (downstream) areas. Prior to the less severe events, the mid-tropospheric height patterns suggest that the warm air is advected into the entire basin, thus resulting in a smaller south to north temperature gradient. Beyond these common characteristics, however, a large amount variability in average and daily upper atmospheric circulation in the prebreak-up period exists even within similar types of peak water events. Thus, a more comprehensive analysis of large-scale controls, including mid-tropospheric synoptic patterns and their linkages to the delta ice regime, is warranted.

3.5.5 Hydroclimactic Controls for Ice and Discharge-Driven Events

Within the high H_m events, a clear delineation of ice and discharge-driven events is less apparent when upstream and downstream hydroclimatic controls are considered. The ice-driven years (2006, 2005, 1997, 1989, 1982) with the highest peak stages and backwater levels at MARR, occur the earliest of the high H_m events, while t_p and t_a also occur earlier than average for most of these years. The opposite condition occurs for the discharge-driven years (1992, 1986, 1985), with later t_b, t_p and t_a.

For the discharge driven years, both 1992 and 1986 have a similar balance of forces, with high Q_{max} and h_b and medium to low k_Q and S_5; however, the pattern in 1985 with an exceptionally large k_Q is more similar to that in 2005 (Figure 3-14). The balance in the ice-driven years of 2006 and 1982 is also quite similar to 1992 and 1986, although these have a smaller ice thickness and peak discharge respectively. Low Q_{max} and k_Q in 1989 set this event apart as being more linked to ice factors. Conversely, 1997 is characterized by low h_b and high S_5 and thus possibly more controlled by discharge factors.

Overall, upstream forces tend to be more important in controlling discharge-driven events, which are inherently characterized by high peak discharge. Equally, given the inclusion of 2005 as a partial discharge-driven year, with several overlapping spatial and temporal characteristics of delta peak water levels, particularly above average flooding in the outer delta (Goulding *et al.*, 2009), high rates of rise in discharge would also exclusively describe discharge-driven years. Despite low and medium melt and thicker ice covers in the delta for these events, the later occurrence of break-up suggests that radiative strength decay of the ice could explain the absence of severe ice jamming and limited backwater levels. Fewer common characteristics could be identified for the ice-driven events, as varying combinations of upstream and downstream controls occurred in each of these years. For each event at least one of the upstream or downstream controls was smaller in classification compared with the combinations observed in discharge-driven events, although this occurred for the range of factors over the series of ice-driven events. As noted above, the balance of forces during these events, which produce the highest break-up water levels at MARR despite lower magnitude hydroclimatic factors than for discharge-driven events, suggests the presence of a threshold above which an altered hydrologic response occurs. Thus, the identification of the specific hydroclimatic conditions responsible for considerable ice jamming and high backwater levels in ice-driven years will require detailed investigation, however, the small sample size (5 events) precludes current analysis. Thus, greater exploration of these patterns should be pursued when additional data are available.

3.5.6 Trends in Hydroclimatic Controls

Trend analysis of hydroclimatic controls over the 1974 to 2006 period revealed a dominance of trends toward earlier occurrences and smaller values for the upstream and downstream factors (Table 3-7). The spring streamflow pulse and the timing of melt initiation have both advanced by 1.11 and 2.00 days/decade over the period of record (both not significant at the 90% level), similar to the advance of 1.54 days/decade for the timing of break-up initiation reported by Goulding *et al.* (2009) (Figure 3-15). Of note in these trends, the ablation period, from melt initiation to the onset of break-up, has increased over the 33 years of record. This agrees with Smith's (2000) results for Russian arctic rivers, wherein a trend toward a longer prebreak-up was noted and suggested to be a driver of more frequent thermal break-ups in the future. The decrease in peak ice thickness of 4 cm/decade is significant at the 95% confidence level, while maximum discharge, rate of rise in discharge, ice thickness at the onset of break-up, and the number of thawing degree days have all decreased, although the trends are not significant and increased variability is present since the 1990s. Only the freeze-up stage has increased by

0.27 m/decade, significant at the 90% confidence level. While these trends would be expected to promote the occurrence of more frequent low peak stage events, increasing trends have occurred over the same period for peak break-up water level and backwater level, both significant at the 90% level. These trends are inherently influenced by the lack of high peak break-up water levels recorded in the 1970s at the start of the time series, and a comparable abundance in recent years. Nevertheless, the three lowest severity years also occurred in the most recent decade, wherein increased variability is observed both in break-up severity and hydroclimatic controls.

3.6 CONCLUSIONS AND RECOMMENDATIONS

River ice break-up and the occurrence of ice-jam flooding are considered a vitally important hydrological phenomenon in the Mackenzie Delta. This analysis shows that the severity of peak break-up stage is most influenced by upstream discharge, driven by snowpack size and ablation in the Mackenzie Basin, and comparative upstream and downstream melt, while timing is controlled predominantly by delta ice conditions and the nature and timing of hydrograph rise. Higher discharges and rates of rise in discharge are a necessary condition for the occurrence of high peak stage events, however, downstream ice conditions and the melt gradient in the basin ultimately determine the effectiveness of this high spring discharge in producing flooding in the delta. The expression of melt during high peak stage events concurs with physical theory of break-up on northward flowing rivers, suggestive that the sequential progression of warming from upstream to downstream effectively drives river ice break-up in these years. Contrary to this pattern, break-up events with the lowest peak stages are often characterized by protracted melt in the basin combined with higher intensity melt in the delta. Within the high peak stage years, upstream forces tended to be more important in controlling discharge-driven events, while an altered hydrologic response occurring for ice-driven events was noted. Finally, a dominance of trends toward earlier occurrences and smaller values for the upstream and downstream factors was observed, with greater variability of these controls and break-up severity in the most recent decade.

The results of this first study into the hydroclimatology influencing break-up in the Mackenzie Delta prompt many additional questions regarding the controls on break-up and ice-jam flooding. Further investigation into the delineation and quantification of upstream controls is recommended, particularly the accumulation and ablation of snowpack in the Mackenzie Basin, and the importance of trigger tributaries. Exploration of the influence of ice strength on break-up severity in the delta could provide greater understanding of the controls in high peak water level years, particularly ice-driven events. In addition, as more years of data are available, further examination of the hydroclimatic controls of ice-driven events could clarify the relative significance of the hydroclimatic drivers of these important flood events.

In light of trends toward earlier occurrences and lower magnitudes of hydroclimatic controls combined with increased variability in the time-series of hydroclimatic controls and resulting severity of break-up in recent decades, continued investigation into both the drivers and effects of these changes is warranted. As such, the particular cause of the significant increasing trend in freeze-up levels merits further examination. Most significantly, further analysis of the linkages of upstream and downstream controls to synoptic patterns could better elucidate the large-scale controls on break-up in the Mackenzie Delta and improve understanding of the potential effects of climatic changes on the break-up regime. These must also be informed by investigations into observed and projected changes to sea ice conditions in the Beaufort Sea, since changes in the extent of sea ice and grounded ice and their role in producing water level effects at this downstream boundary of the Mackenzie Delta have the potential to alter ice-affected water levels at outer delta stations.

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Table 3-1. Characteristics of small index streams used for the identification of the initiation of spring ablation.

WSC ID	Station	Start	End	Years of record	Drainage area (km^2)
10ND001	Hans Creek near Inuvik	1977	1987	10	337
10ND002	Trail Valley Creek near Inuvik	1977	2006	28	68.3
10LC010	Boot Creek near Inuvik	1981	1990	10	28.2
10LC017	Havikpak Creek near Inuvik	1995	2006	11	15.21
10LC009	Cabin Creek above Dempster Highway	1984	1996	13	150
10LC007	Caribou Creek above Dempster Highway	1975	2006	32	590
10LC003	Rengleng River below Dempster Highway	1976	2006	31	1300

Table 3-2. MARR ice regime data, pertaining to severity and timing (H_m, ΔH, t_b, H_b), upstream (Q_{max}, t_p, k_Q) and downstream forces (h_a, t_a, h_b, S_5, H_f) during the spring and previous fall of the given break-up event. Missing data are denoted by '-'.

Year	H_m (m)	ΔH (m)	t_b (Julian day)	H_b (m)	Q_{max} (m^3/s)	t_p (Julian day)	k_Q (m^{3}s^{-1} day^{-1})	h_a (m)	t_a (Julian day)	h_b (m)	S_5 (°C day)	H_f (m)
1974	9.90	3.70	142	8.88	30300	137	1515	1.34	128	1.05	117	5.92
1975	13.20	4.08	144	12.70	34000	131	1406	1.36	119	1.07	111	-
1976	11.33	2.62	139	10.40	32000	128	1299	1.35	117	1.01	149	4.95
1977	-	-	139	-	28300	133	1980	1.27	110	0.94	127	4.75
1978	10.51	2.59	152	10.28	28300	137	1075	1.21	105	0.93	109	3.33
1979	9.64	3.88	141	7.40	28600	138	1546	1.31	112	0.86	173	-
1980	9.55	2.08	149	9.49	26400	132	1056	1.22	131	0.91	120	4.71
1981	-	-	137	-	28300	134	2130	1.19	117	0.85	130	5.07
1982	-	-	144	-	28800	142	1628	1.31	137	1.20	56	3.33
1983	-	-	147	7.05	29000	129	888	1.38	112	1.22	69	4.08
1984	9.93	3.86	144	8.94	20000	134	881	1.31	120	0.93	146	4.55
1985	-	-	141	-	32900	137	4033	1.34	122	1.07	106	4.40
1986	14.29	5.65	149	8.23	31800	139	1693	1.32	120	1.15	65	4.58
1987	10.74	5.25	148	9.18	21600	128	411	1.31	139	1.11	91	5.27
1988	13.48	7.61	136	7.37	24700	132	833	1.19	107	0.89	126	4.87
1989	17.65	10.14	142	11.42	27000	130	1067	1.25	114	1.07	90	5.81
1990	10.53	-	139	8.25	24400	132	513	1.31	116	1.11	99	5.03
1991	12.33	6.48	135	9.37	28000	133	3250	1.31	104	0.69	230	4.89
1992	14.08	5.81	144	8.55	35000	134	1422	1.30	133	1.12	83	4.67
1993	12.72	4.85	136	8.34	32000	131	1953	1.20	131	1.08	78	5.44
1994	10.62	3.77	139	8.01	25000	126	1094	1.25	113	0.78	180	4.67
1995	8.23	2.99	132	6.49	15800	123	885	1.21	112	0.61	226	3.25
1996	12.31	6.45	145	7.38	28800	139	1647	1.28	130	0.98	117	4.23
1997	16.45	8.95	140	8.89	29900	-	-	1.27	114	0.87	164	6.36
1998	12.35	5.05	136	10.55	28000	-	-	1.12	104	0.51	258	7.42
1999	9.17	2.78	139	8.52	23500	129	740	1.23	127	1.06	84	4.98
2000	8.48	3.36	144	7.77	15200	130	304	1.27	128	1.22	29	5.76
2001	10.93	5.28	144	9.57	25700	135	558	1.23	130	1.15	50	4.66
2002	12.16	4.64	143	7.02	29500	139	2251	1.19	108	0.71	184	5.80
2003	10.42	3.30	141	8.25	26800	125	643	1.15	105	0.82	125	4.93
2004	12.97	5.67	147	9.70	30000	137	1581	1.24	99	0.95	114	5.24
2005	15.82	8.82	134	10.20	30500	129	3875	1.23	113	1.00	97	5.09
2006	19.58	11.99	134	12.97	30000	130	1755	1.20	124	1.05	62	5.48

Table 3-3. Correlation coefficients (Spearman's R) between the severity and timing of ice break-up at MARR and upstream (Q_{max}, t_p, k_Q) and downstream forces (h_a, t_a, h_b, S_5). Bolded values are significant at the 90% confidence level, bolded italicized values are significant at the 95% confidence level.

	H_m	t_b
Q_{max}	***0.61***	0.06
t_p	0.15	***0.41***
k_Q	***0.55***	-0.27
h_a	-0.03	***0.37***
t_a	-0.09	**0.30**
h_b	0.15	***0.44***
S_5	-0.20	***-0.37***

Table 3-4 Results of stepwise multivariate regression for the severity of peak break-up water levels, H_m, and the timing of break-up, t_b.

Dependant Variable	Included Variables	Adjusted r^2
H_m	S_5, k_Q	0.32
t_b	t_p, k_Q	0.35

Table 3-5. Average values of upstream (Q_{max}, t_p, k_Q) and downstream (h_a, t_a, h_b, S_5) hydroclimatic controls for all break-up events and high, medium and low H_m groupings. Significant differences between high and low H_m events using the Mann-Whitney U test are shown with bolded values denoting significance at the 90% confidence level and bolded italicized values significance at the 95% level.

Average	H_m	Q_{max}	t_p	k_Q	h_a	t_a	h_b	S_5
Entire series	12.12	27578	132.7	1481	1.26	118.2	0.97	120.1
High H_m events	16.31	***30737***	134.4	***2210***	1.28	122.1	**1.07**	***90.3***
Med H_m events	12.23	28572	133.3	1519	1.25	117.1	0.92	137.0
Low H_m events	9.72	***24027***	131.2	***932***	1.26	117.9	**0.94**	***128.0***

Table 3-6. Rate of rise in discharge, k_Q, and degree-days of thaw, S_5, rank combinations for high, medium and low H_m events. The numbers in each column represent the number of break-up events exhibiting the associated (k_Q, S_5) combination.

(k_Q, S_5)	High H_m	Med H_m	Low H_m
(1,1)		2	
(1,2)	2		
(1,3)		1	
(2,1)		1	2
(2,2)		3	3
(2,3)	5		
(3,1)			1
(3,2)		1	2
(3,3)		1	3

Table 3-7. Results from Mann-Kendall analysis for trend and Sen's slope estimate for severity and timing (H_m, ΔH, t_b), upstream (Q_{max}, t_p, k_Q), and downstream forces (h_a, t_a, h_b, S_5, H_f). Bolded values are significant at the 90% confidence level, bolded italicized values are significant at the 95% confidence level.

Parameter	Trend	Slope estimate	Units	Number of years
H_m	+ve	0.59	m/decade	28
ΔH	**+ve**	**0.74**	m/decade	27
t_b	-ve	-1.54	days/decade	33
Q_{max}	-ve	-667	m^3/s/decade	33
t_p	-ve	-1.11	days/decade	31
k_Q	-ve	-92.5	m^3s/day/decade	31
h_a	***-ve***	***-0.04***	m/decade	33
t_a	-ve	-2.00	days/decade	33
h_b	-ve	-0.03	m/decade	33
S_5	-ve	-5.64	°C day/decade	33
H_f	**+ve**	**0.27**	m/decade	31

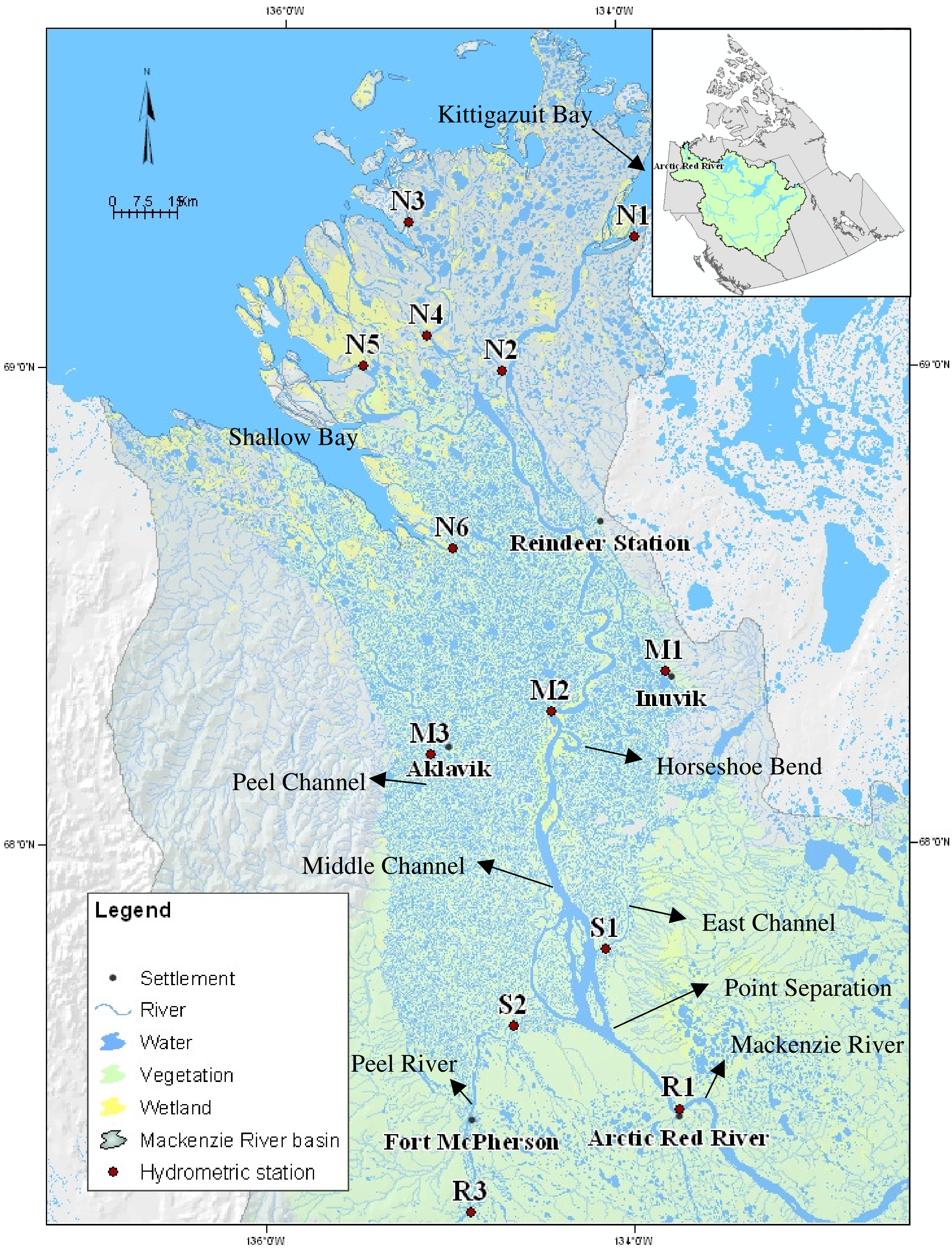


Figure 3-1. The Mackenzie Delta, with locations of Water Survey of Canada hydrometric stations and locations relevant to historical ice jam observations (Mackenzie Basin inset).

Figure 3-2. Photo of Mackenzie River looking upstream from the ferry crossing one day after ice clearance, May 29, 2007. MARR hydrometric ‘station’ is the small white building in the upper left.

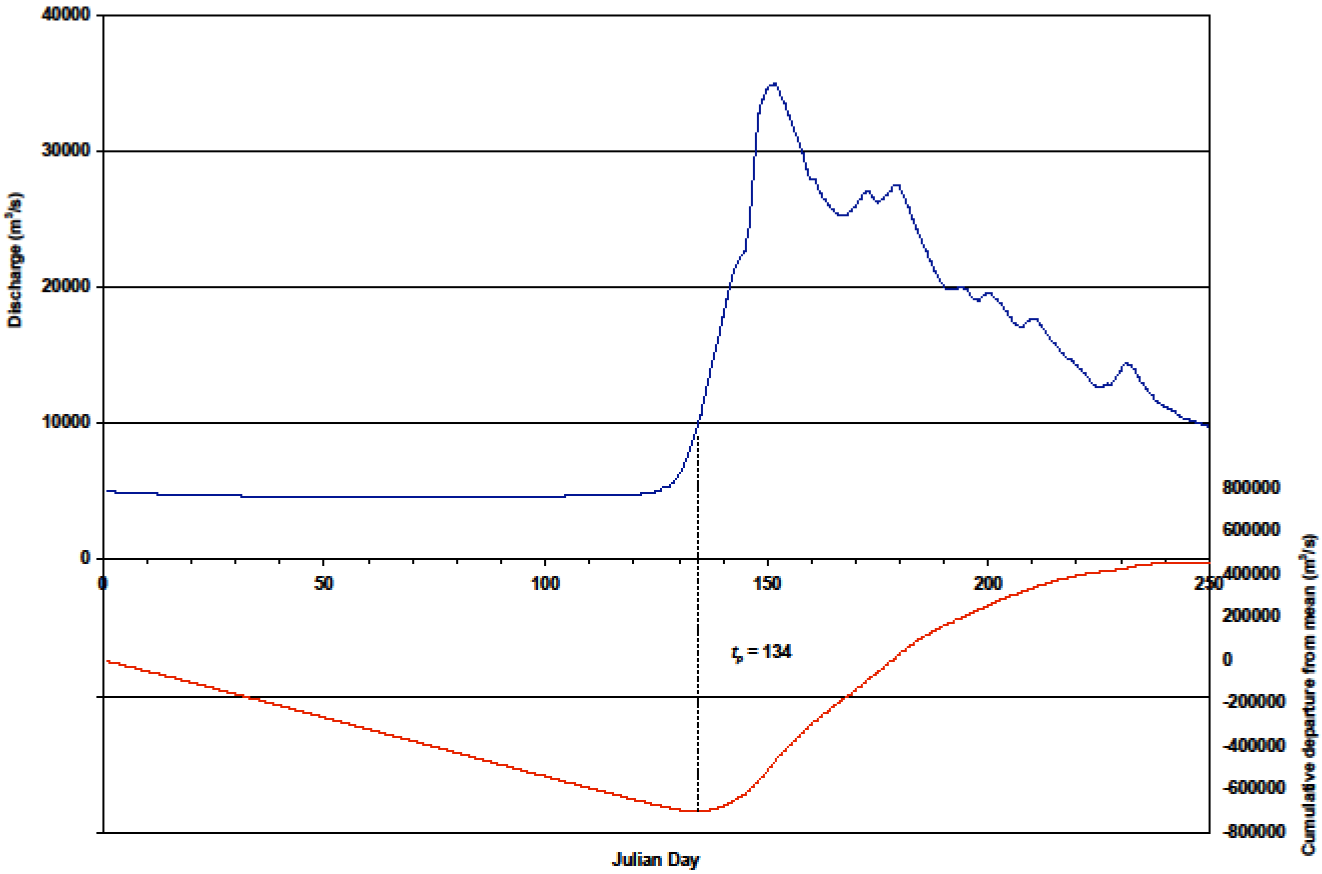


Figure 3-3. Daily discharge (upper curve) and cumulative departure from the year's mean flow (lower curve) at MARR from Julian day 1–250 in 1992. The onset of the spring pulse of streamflow is defined as the day when the cumulative departure is most negative (vertical line), here on Julian day 134.

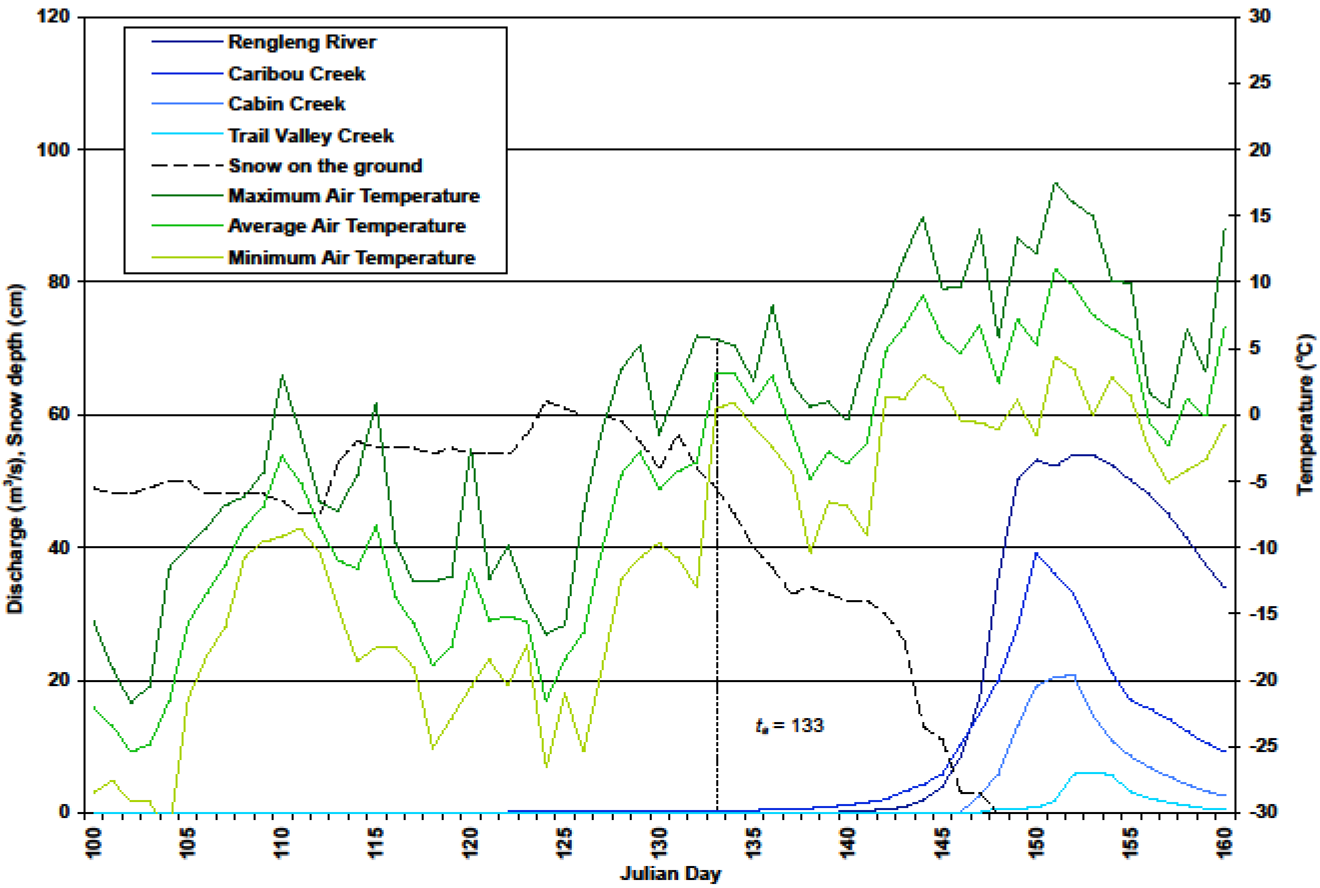


Figure 3-4. Snow on the ground and air temperature records from Inuvik Airport and small stream discharge in the Mackenzie Delta from Julian day 100-160 in 1992. From these records, the initiation of spring melt was estimated to occur on Julian day 133.

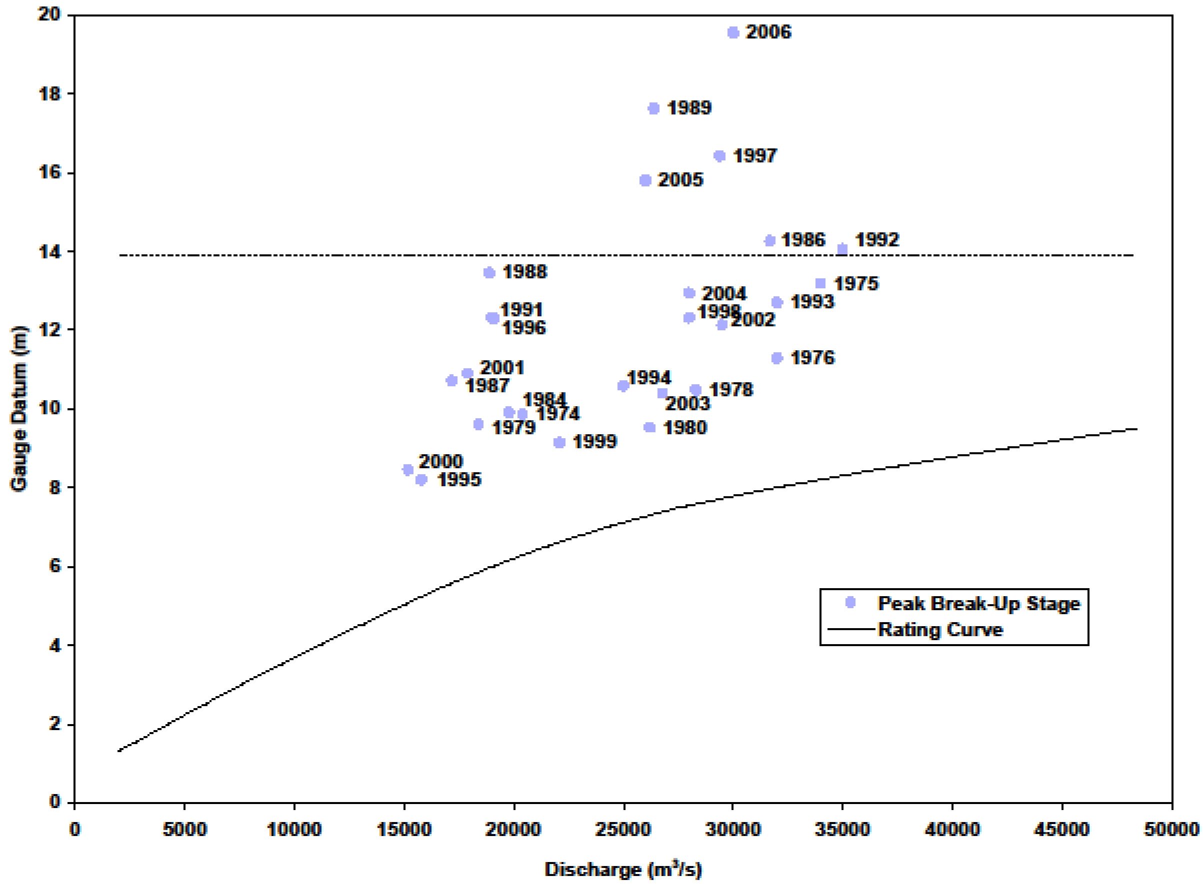


Figure 3-5. Peak break-up water level at MARR compared with the stage-discharge curve for open water conditions. The highest averaged peak water levels in the delta correspond with MARR peak levels exceeding a threshold of 14 masl.

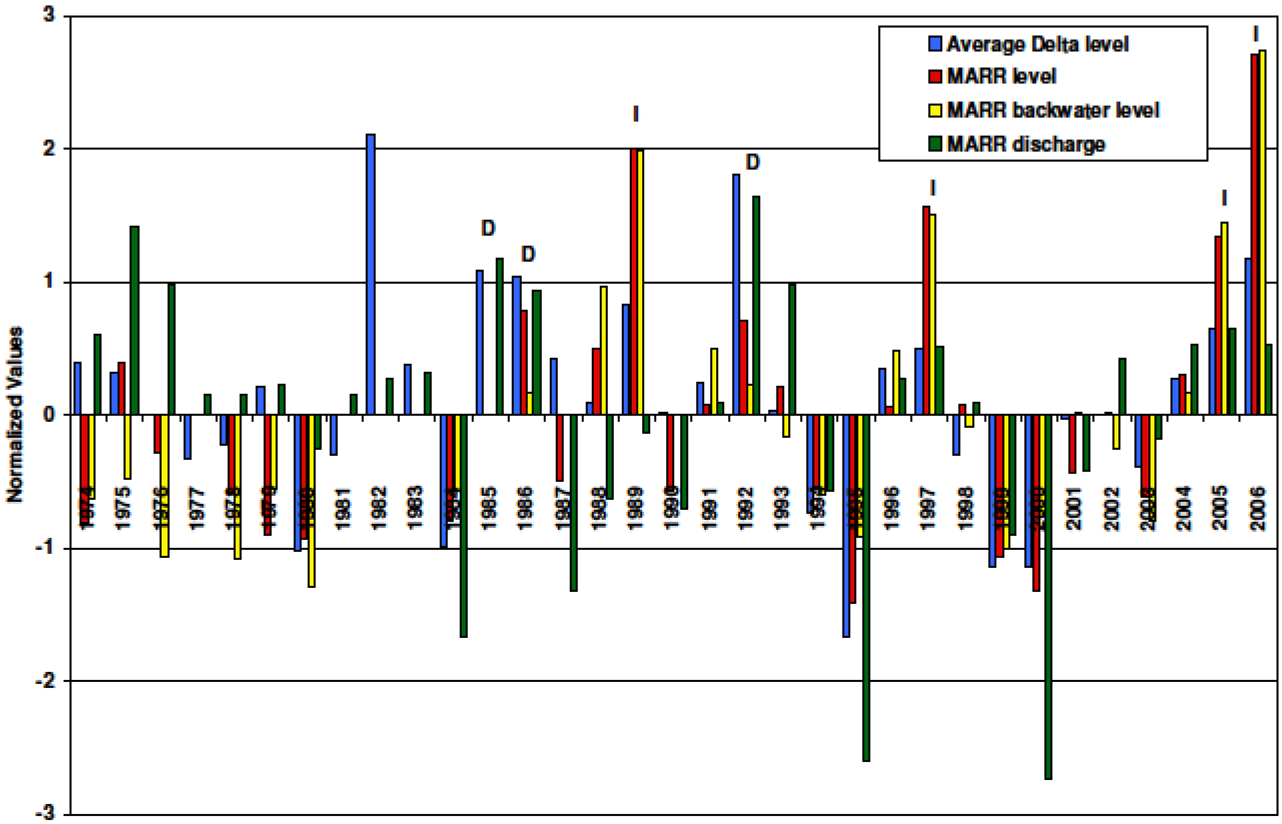


Figure 3-6. Comparison of delta and MARR peak water levels with backwater and maximum break-up discharge at MARR (all normalized). The symbol D (I) identifies discharge (ice) dominated years.

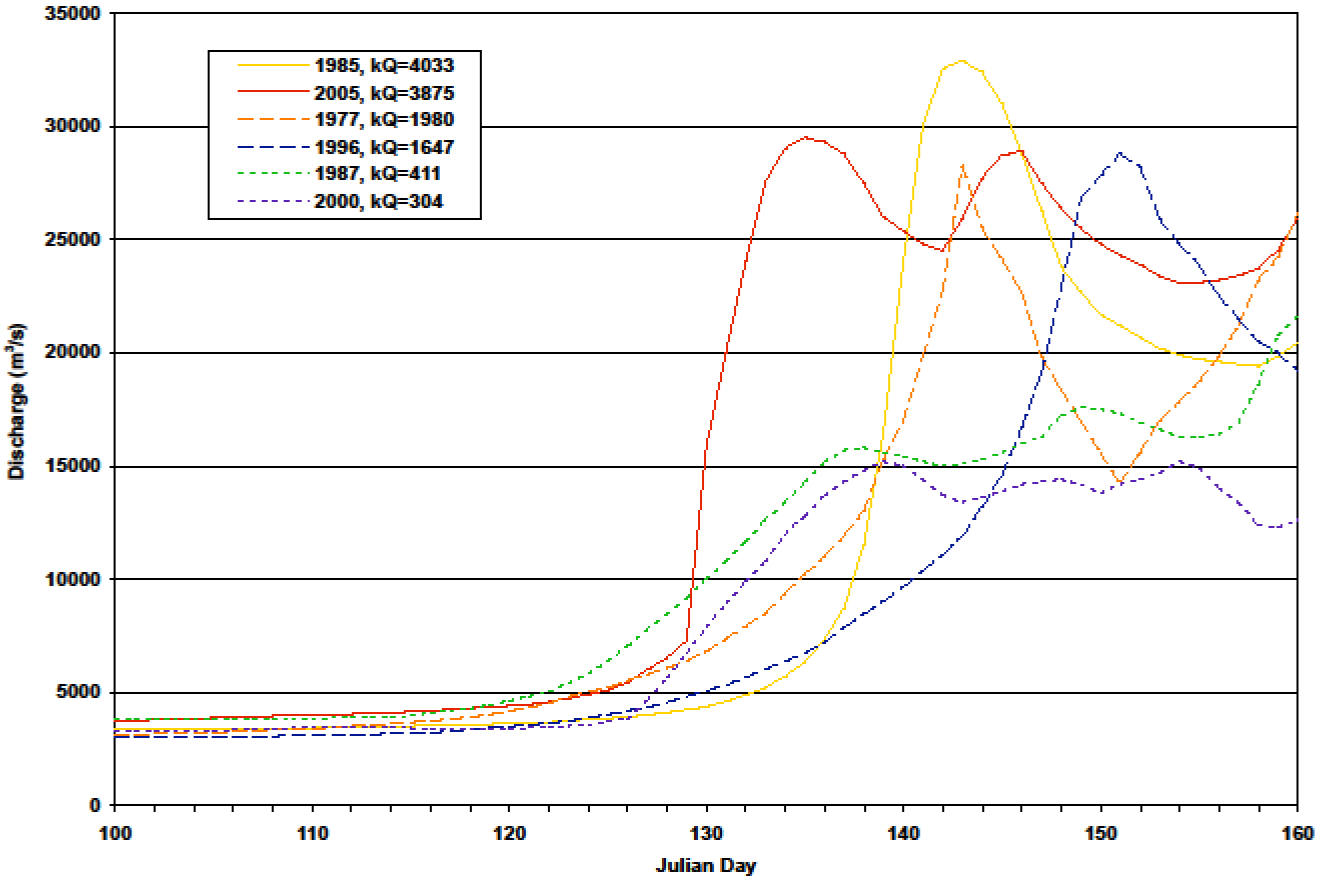


Figure 3-7. Spring hydrographs for selected years of high (1985, 2005), medium (1977, 1996), and low (1987, 2000) rate of rise in discharge, k_Q, values.

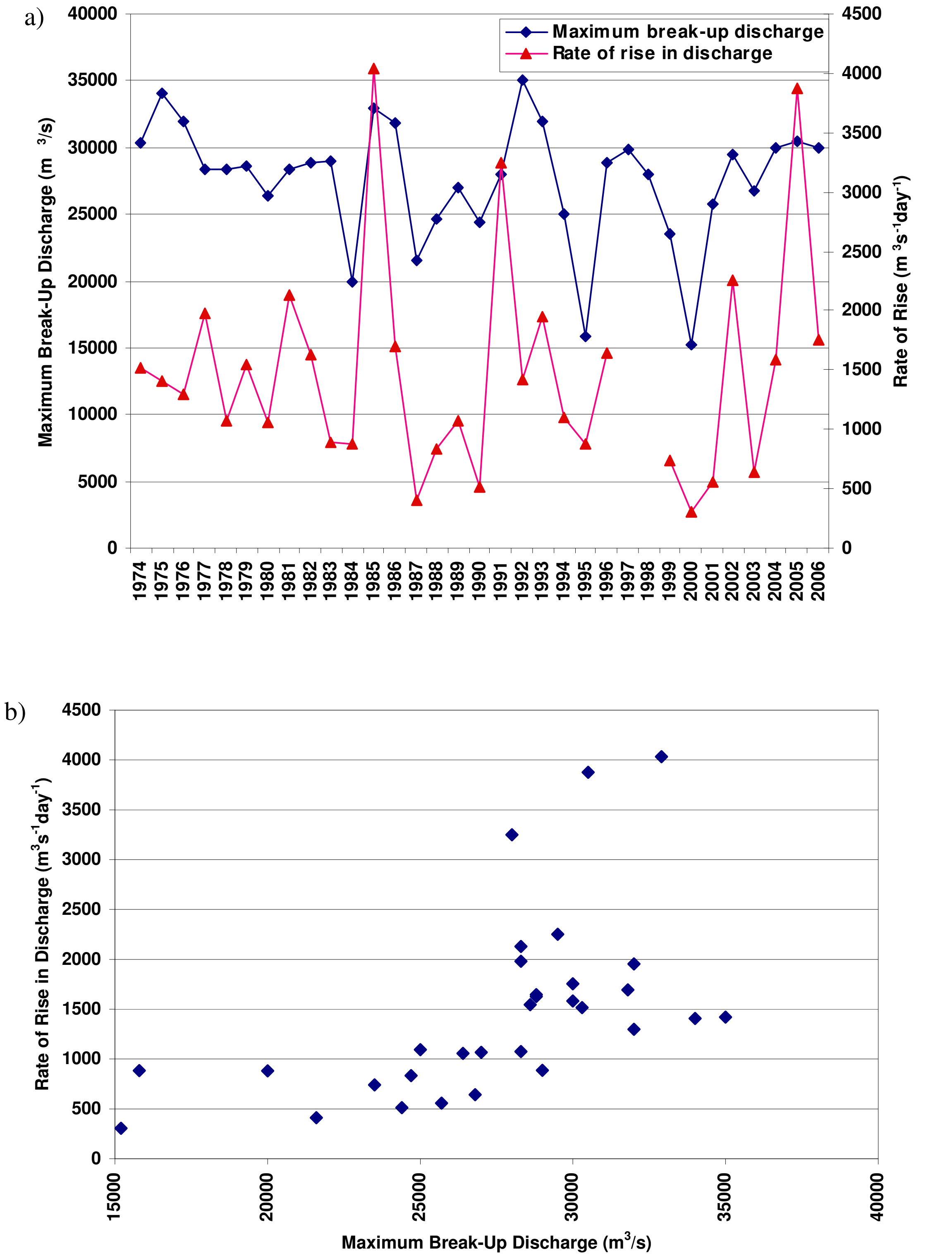
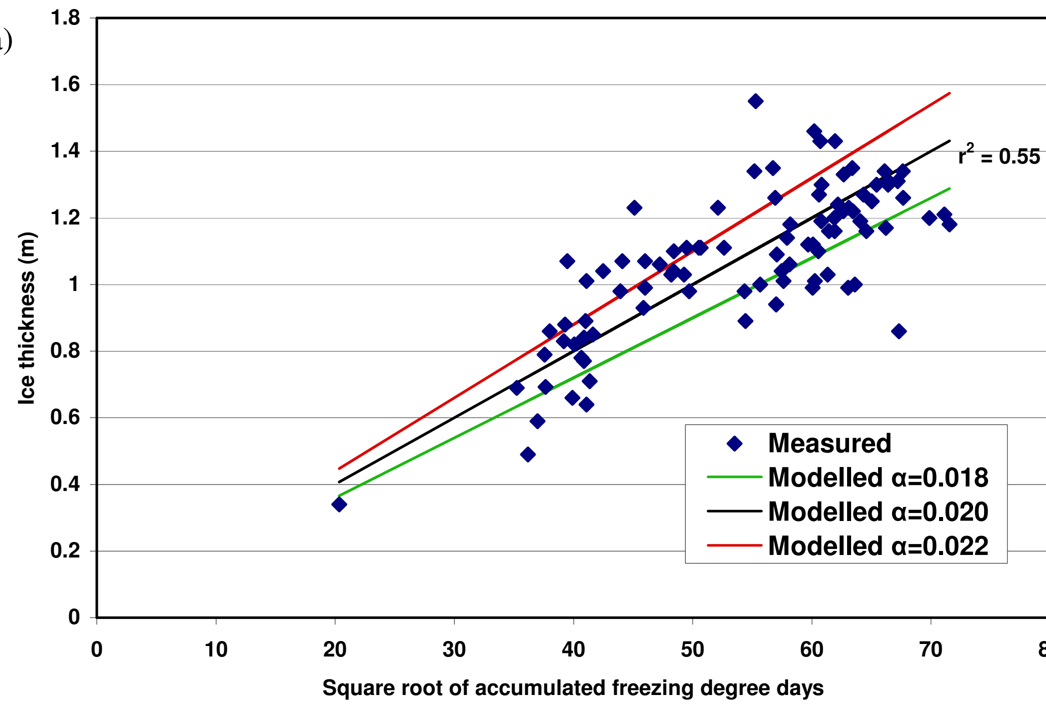


Figure 3-8. a) Maximum break-up discharge and rate of rise in discharge for the period from 1974 to 2006 at MARR, and b) maximum break-up discharge compared with the rate of rise in discharge.

a)



b)

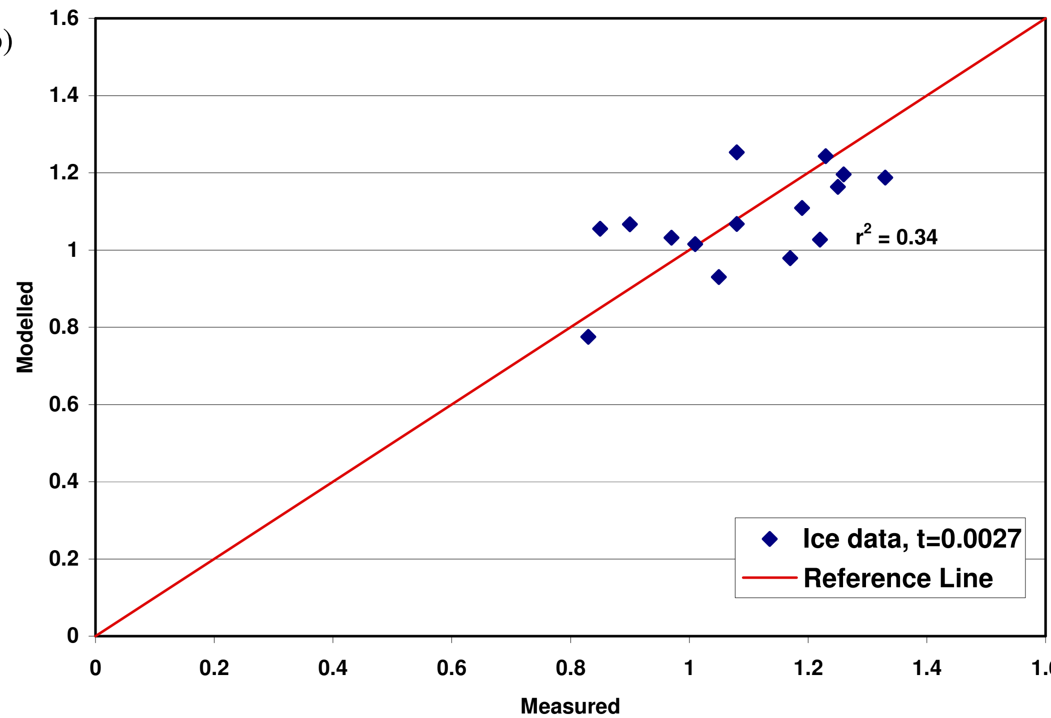


Figure 3-9. Comparison of measured and modelled a) winter ice growth and b) spring ice ablation at MARR. The best model results were obtained with α=0.02 and τ =0.0027 with r^2 values of 0.55 and 0.34 respectively

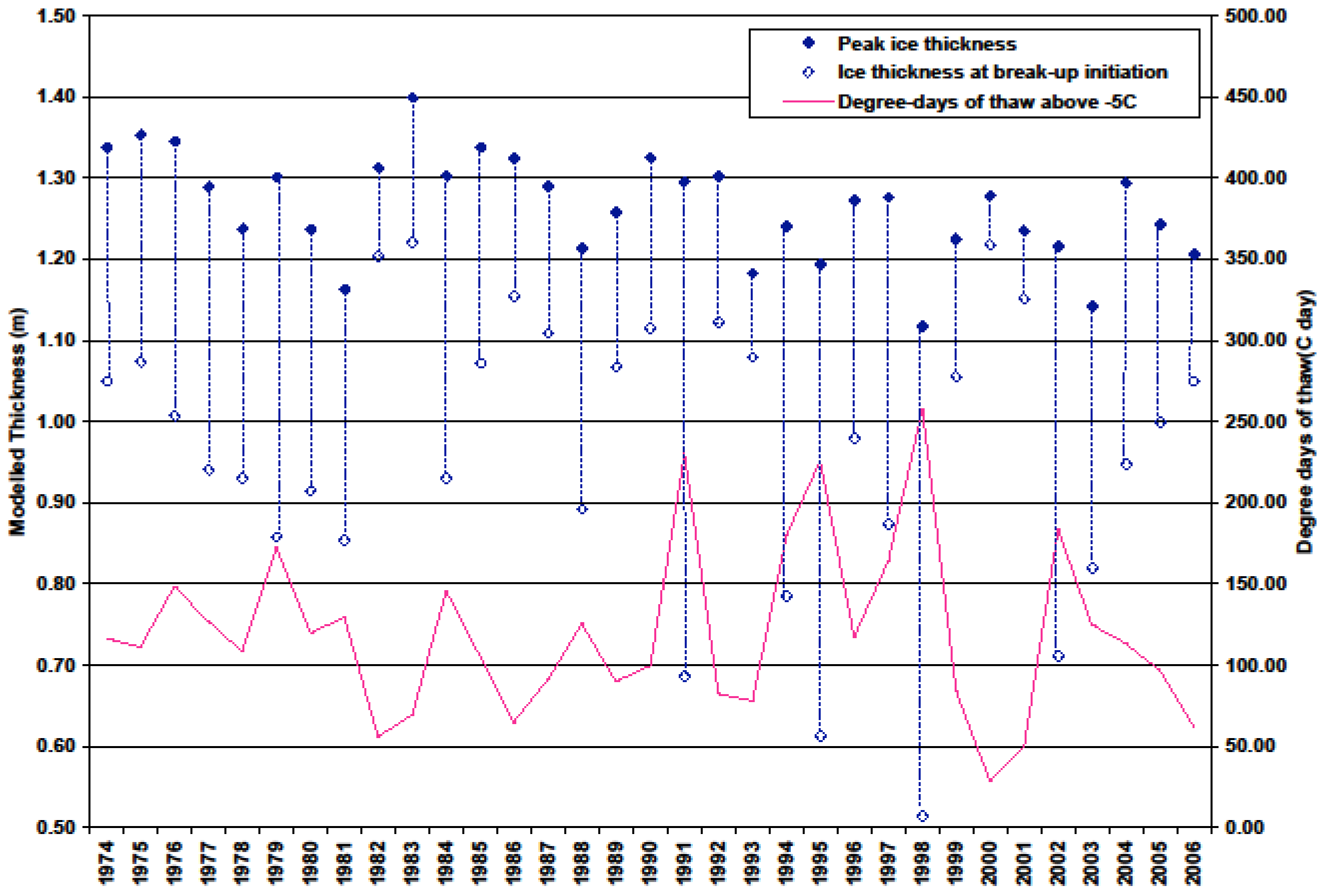


Figure 3-10. Peak ice thickness, h_a, ice thickness at break-up initiation, h_b, and degree-days of thaw above –5ºC, S_5, for the period from 1974 to 2006 at MARR.

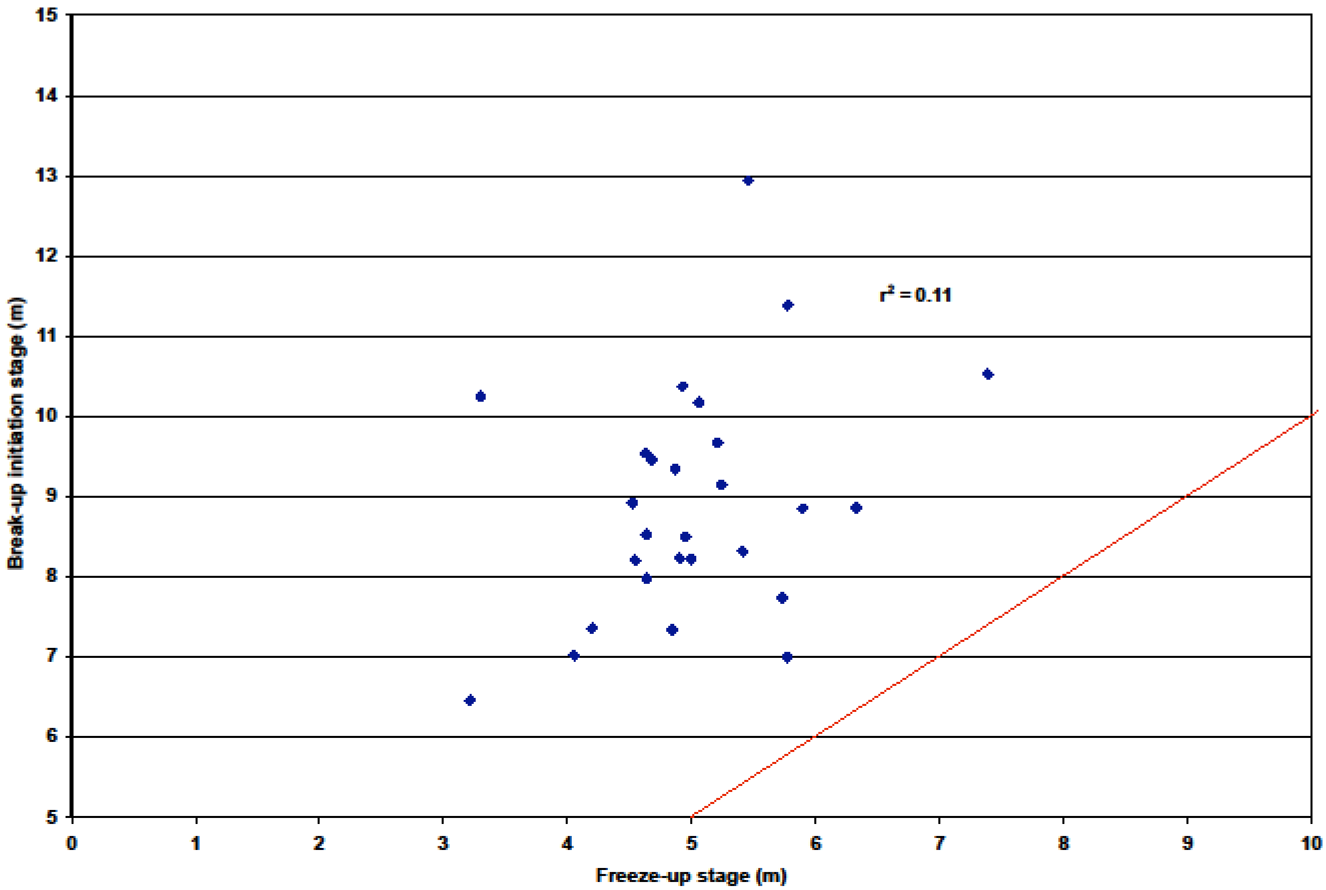


Figure 3-11. Stage at the initiation of break-up, H_b, compared with freeze-up stage, H_f, at MARR (r^2 =0.11). A 1:1 reference line is also shown in the plot.

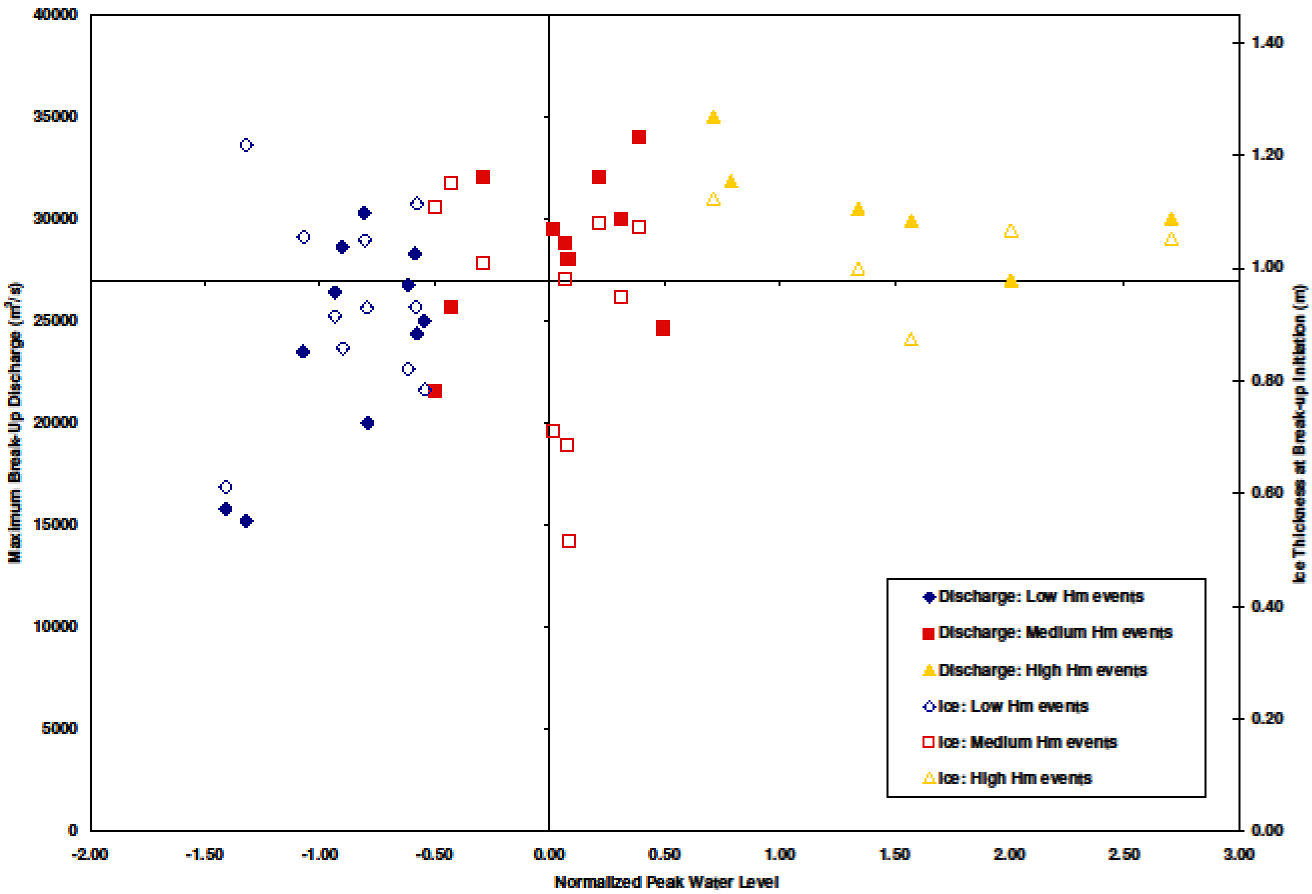


Figure 3-12. Comparison of maximum break-up discharge, Q_{max}, and ice thickness at break-up initiation, h_b, with severity of break-up for low, medium and high H_m events. The horizontal line represents average discharge (~27600 m^3/s) and ice thickness (0.97 m).

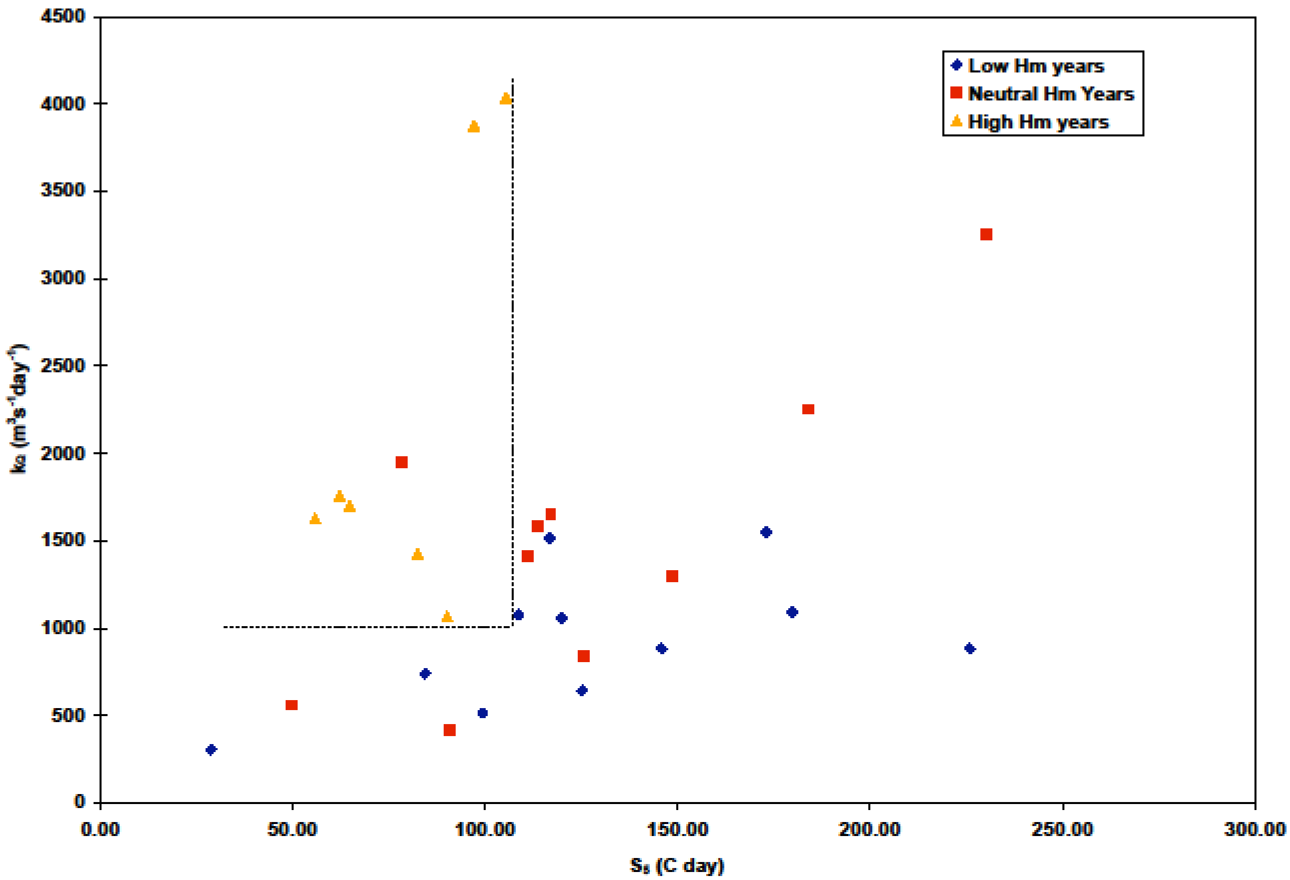


Figure 3-13. Rate of rise in discharge, k_Q, compared with degree-days of thaw above –5ºC, S_5, for low, medium, and high H_m events. The range for the occurrence of high H_m events is bounded by the dotted lines.

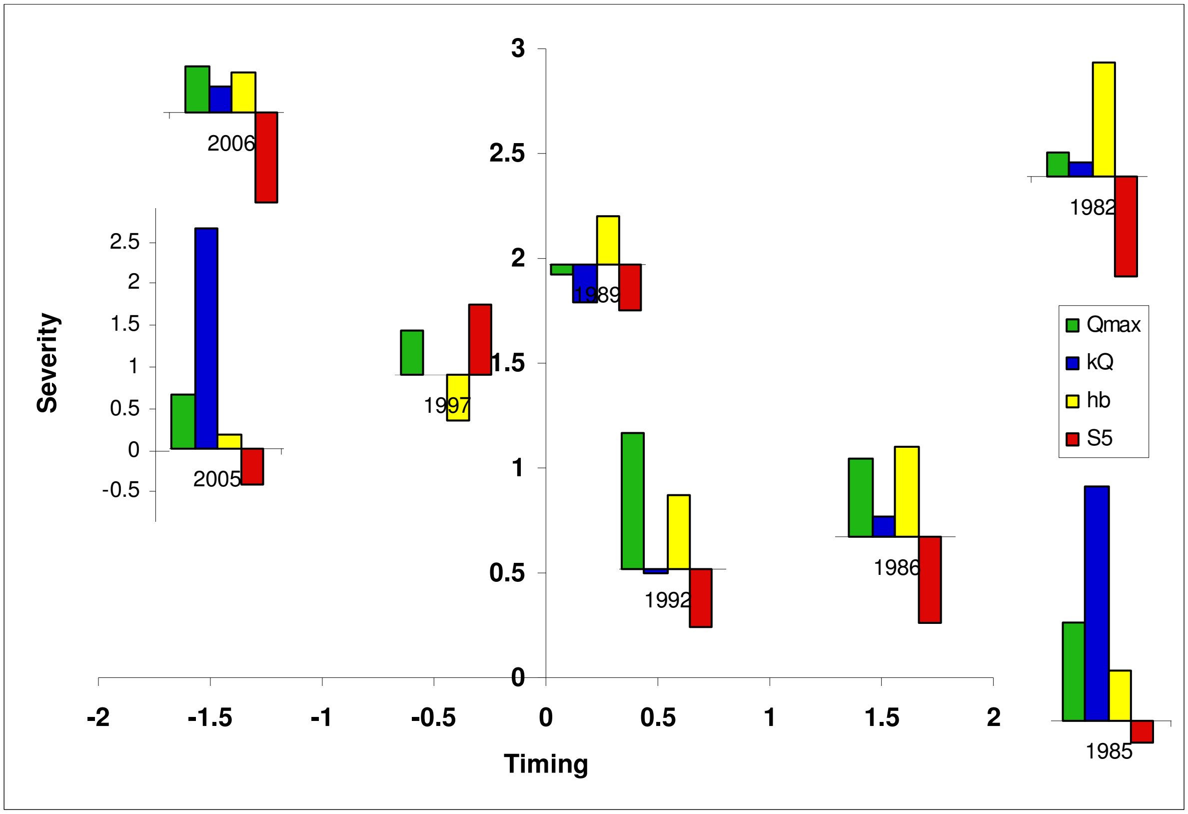


Figure 3-14. Selected normalized upstream (Q_{max}, k_Q) and downstream (h_b, S_5) hydroclimatic controls of ice (1982, 1989, 1997, 2005, 2006) and discharge-driven (1985, 1986, 1992) high H_m events associated with the normalized timing and severity of break-up. The scale shown for 2005 applies to all years. 1982 and 1985 are included beside the plot to show hydroclimatic controls despite missing peak break-up water level in these years.

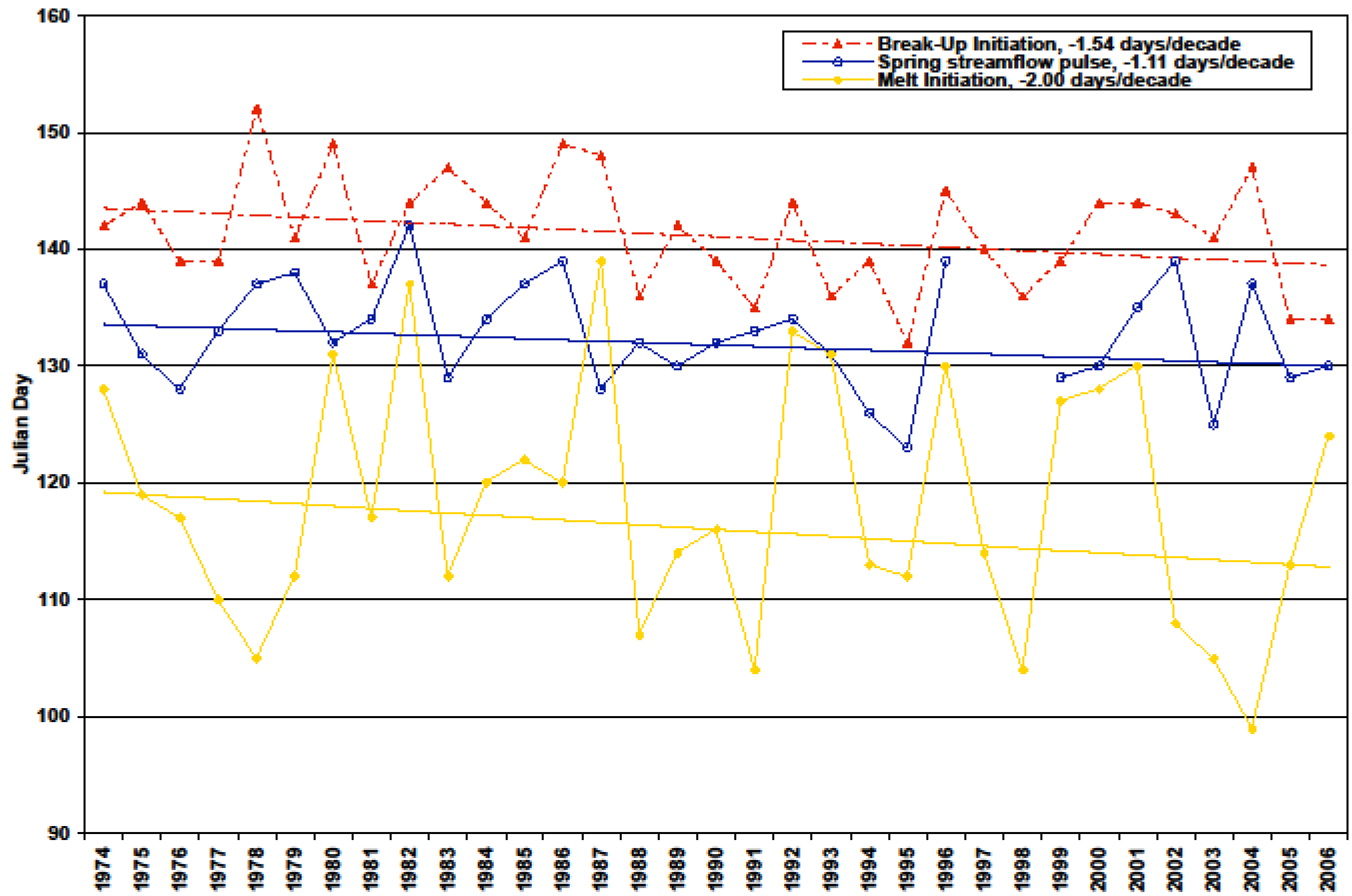


Figure 3-15. Trend of break-up initiation timing, t_b, timing of spring streamflow pulse, t_p, and timing of the onset of spring melt, t_a, including the Sen's slope estimate (trends not significant at the 90% level).

Chapter 4: Conclusion

River ice break-up and the occurrence of ice-jam flooding in the Mackenzie Delta are believed to be fundamental to the hydroecology of delta lakes. Better understanding of the historical break-up regime in the delta and the controlling hydroclimatology of extreme break-up floods can provide a strong basis for future predictions and the modelling of changes to this complex hydrological phenomenon. A break-up chronology was created for the Mackenzie Delta for the 1974 to 2006 period of instrumental record, based primarily on Water Survey of Canada (WSC) hydrometric records, but informed and constrained by ice and break-up observations, satellite imagery, and air photography. This chronology allowed the examination of spatial and temporal patterns of break-up in the delta, outlined in Chapter 2 of this thesis. Equally, the Mackenzie River at Arctic Red River (MARR) hydrometric station, immediately upstream of the delta and capturing the dominant hydrologic signal from the basin, was assessed for its suitability as an extra-delta index of flood severity. The quantification of basinwide and intra-delta controls on break-up, based on upstream discharge characteristics and delta ice conditions extracted and modeled from WSC and Meteorological Service of Canada (MSC) data for the representative MARR station, was used to assess the influence of these forces on the timing and severity of break-up in the delta, particularly related to the most severe flooding events. These findings are presented in Chapter 3 of this thesis.

Over the 1974 to 2006 period, three years stood out in the break-up chronology with the highest flood severity in the delta; in 1982, 1992, and 2006 a consistent pattern of higher than normal water levels at a majority of stations in the delta occurred, suggesting wide-scale flooding. Among the subset of the highest peak water years, two types of events were identified, each with a respective set of spatial and temporal patterns.

Discharge-driven events, when above average peak water levels are associated with high discharge on the Mackenzie River, occurred in 1985, 1986 and 1992. Spatially, these events are characterized by a northward (downstream) increase in peak water levels through the delta, and higher levels through the central delta along Middle Channel than the eastern and western delta channels. In 1992, however, significant flooding on the Peel Channel above Aklavik resulted in the highest levels being recorded in the western delta. Coupled with these patterns, a stronger statistical relationship with MARR discharge than with water level occurs for both the outer delta stations and Middle Channel stations in the central region, underlining the importance of discharge to the occurrence of high water levels in these regions. During discharge-driven years break-up initiation and peak water level occur later than average, while a shorter than average duration of break-up in the delta is observed, suggesting that the large discharge effectively flushes the ice out of the delta.

Ice-driven events, as occurred in 1982 (anecdotally), 1989, 1997, 2005, and 2006 are so labelled because high backwater levels are observed at MARR. These are characterized by high peak water levels in the southern and mid-delta, with relatively lower levels in the outer delta. Equally, peak levels higher in the eastern delta and higher or similar in the western delta than the central delta are observed. A strong relationship of mid and southern station peak water level to MARR peak stage is noted, highlighting the association between ice jamming at MARR and ice jam formation between Point Separation and Horseshoe Bend, observed to be responsible for localized flooding in the southern delta. Earlier break-up initiation and peak water level in the delta occur in ice-driven years, while no pattern of break-up duration was evident.

Trend analysis of break-up timing revealed a widespread tendency toward earlier break-up initiation (~2 days/decade) and occurrences of peak water level (~1.5 days/decade) in the delta over the study period. This underscores another trend in the occurrence of large and extreme events; that is, a tendency toward the occurrence of ice-driven events, which are associated with earlier break-up timing. Equally, the delay between break-up initiation at MARR and on the Peel River below Fort McPherson has decreased, contrary to expectations of a greater response to pronounced northern warming in the smaller Peel Basin.

The results of the hydroclimatic analysis showed that the severity of peak break-up stage is most influenced by upstream discharge, and thus snowpack size and ablation in the Mackenzie Basin, and the balance of upstream and downstream melt, while timing is controlled predominantly by delta ice conditions and the nature and timing of hydrograph rise. Thus, upstream discharge is a dominant driver of high peak stage events, necessary as an initial condition for ice-induced high water levels. However, downstream ice conditions and the melt gradient in the basin ultimately determine the effectiveness of high spring discharge in producing flooding in the delta, although no upper threshold of intra-delta melt or lower limit of ice thickness appears to control the level of backwater produced from broken ice and ice jamming effects.

The combination of rapid upstream snowmelt and low melt intensity of the downstream ice during high peak stage years is consistent with physical theory of break-up on northward-flowing rivers, suggesting that the sequential progression of warming from upstream to downstream effectively drives river ice break-up. In contrast, most break-up events occurring in the delta do not appear to be driven by a strong gradient in the sequential warming pattern, and break-up events with the lowest peak stages are often characterized by protracted melt in the basin combined with higher intensity melt in the delta. These patterns are corroborated by large-scale controls, namely in the surface temperature gradient between the basin and the delta regions, and in mid-tropospheric synoptic patterns prior to the onset of break-up.

Within the high peak stage years, upstream forces were most important in controlling discharge-driven events. No dominant control could be identified for ice-driven events, as varying combinations of upstream and downstream controls were observed in each of the different ice-driven years. These events, which produce the highest break-up water levels at MARR with lower magnitude hydroclimatic factors than for discharge-driven events, appear to exhibit an altered hydrologic response.

A dominance of trends toward earlier occurrences and smaller values was observed for the upstream and downstream factors. Advances in the spring streamflow pulse and the timing of melt initiation occurred over the study period; a greater advance in the latter, compared with the above-noted advance in the timing of break-up initiation, has resulted in the lengthening of the prebreak-up melt interval. A tendency toward decreasing peak discharge, rate of rise in discharge, and ice thickness, and increasing length of the prebreak-up melt period and freeze-up stage was observed, with greater variability of these controls and break-up severity in the most recent decade.

The results of this first study into the hydroclimatology influencing break-up in the Mackenzie Delta prompt many additional questions regarding the controls on break-up and ice-jam flooding. Further investigation into the delineation and quantification of upstream controls is recommended, particularly the accumulation and ablation of snowpack in the Mackenzie Basin, and the importance of trigger tributaries. Exploration of the influence of ice strength on break-up severity in the delta could provide greater understanding of the controls in high peak water level years, particularly ice-driven events. In addition, as more years of data become available, further examination of the hydroclimatic controls of ice-driven events could clarify the relative significance of the hydroclimatic drivers of these important flood events.

Additional context of the pre-instrumental flood record could be obtained through the incorporation of traditional knowledge with regards to the timing of break-up and the magnitude of flooding for various regions of the delta. Paleolimnological studies of delta lake stratigraphy could further enhance understanding of long-term flood frequency in the delta and the spatial extent of historic flooding, helping to place current trends and variability in hydroclimatic controls and break-up flood levels within the context of the longer-term record.

In light of trends toward earlier occurrences and lower magnitudes of hydroclimatic controls combined with increased variability in the time-series of hydroclimatic controls and resulting severity of break-up in recent decades, continued investigation into both the drivers and effects of these changes is warranted. As such, the particular cause of the significant increasing trend in freeze-up levels merits further examination. Most significantly, further analysis of the linkages of upstream and downstream controls to synoptic patterns could better elucidate the large-scale controls on break-up in the Mackenzie Delta and improve understanding of the potential effects of climatic changes on the break-up regime. These must also be informed by investigations into observed and projected changes to sea ice conditions in the Beaufort Sea, since changes in the extent of sea ice and grounded ice and their role in producing water level effects at this downstream boundary of the Mackenzie Delta have the potential to alter ice-affected water levels at outer delta stations.