UNIVERSITÉ DU QUÉBEC

DYNAMIQUE FLUVIO-GLACIELLE
ÉTUDE DE CAS D'UNE FOSSE-À-FRASIL, RIVIÈRE MITIS, BAS-SAINT-LAURENT.

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RÉSUMÉ

Il se forme du frasil dans près de 60 % des grands bassins de l'hémisphère nord, pour des périodes variant de quelques jours à plusieurs mois. Le frasil et les glaces qu'il produit, endommagent les infrastructures submergées, menacent la sécurité civile et modifient considérablement la morphologie des rivières. Les interrelations que partage la dynamique glacielle avec la géomorphologie fluviale et le transport des sédiments sont encore mal comprises, entre autres parce qu'elles s'insèrent dans un champ d'études dépourvu de cadre d'analyse.

Ce mémoire s'intéresse à l'étude des liens entre la dynamique du frasil et la géomorphologie fluviale. Il s'organise autour de deux chapitres-articles. Le premier article propose un modèle conceptuel d'intégration des dynamiques glacielle et fluviale. La « trinité » conceptuelle de Leeder s'impose comme cadre structurel, car elle illustre efficacement les interactions entre les composantes de la dynamique fluviale. Une « trinité » analogue a été développée pour la dynamique glacielle. Le modèle proposé contient six composantes dépendantes et leurs interrelations. Ce modèle est un premier outil d'analyse pour comprendre la complexité des dynamiques physiques sous-jacentes aux rivières froides.

Le second article s'intéresse à la chronologie des ajustements morphologiques d'une fosse à frasil selon l'accumulation de frasil sous couvert de glace et le degré d'obstruction du chenal par cette accumulation. Les fosses à frasil présentent une morphologie « anormalement profonde » résultant d'interrelations non documentées entre la dynamique glacielle et la dynamique fluviale. La fosse à l'étude est située à 14.1 km de l'embouchure de la rivière Mitis, Bas-Saint-Laurent. Les objectifs spécifiques de l'étude sont de trois ordres : documenter les processus qui mènent à l'accumulation de frasil; cartographier l'évolution mensuelle du couvert de glace, de l'accumulation sous-glacielle de frasil et l'évolution du lit; documenter en continu la déformation du lit au cours de l'hiver.

Les premières observations de frasil dans le système fluvial de la rivière Mitis sont associées à une température de l'eau de 0.8 °C, à un abaissement de la température atmosphérique et à d'importantes précipitations neigeuses. L'étude identifie une relation étroite entre la progression du couvert de glace et le gradient de pente du chenal. À l'échelle de la fosse-à-frasil, l'analyse des levés géophysiques et manuels dévoile une géométrie hydraulique sous-glacielle modelée par d'épaisses accumulations de frasil aux formes irrégulières qui occupent jusqu'à 70 % de la section transversale et atteignent une épaisseur de 6.25 mètres. Un système de bornes autonome munies d'accéléromètres, enregistrant la temporalité et la spatialité des déformations du lit, a été développé. L'utilisation de ces bornes sédimentaires dynamiques (DBR), révèle une relation claire entre la déformation du lit et la progression du couvert de glace sur la fosse-à-frasil. Dans les cours d'eau de taille moyenne, l'accumulation suspendue de frasil constitue une composante morphologique imposante qui dépasse le qualificatif de simple accumulation ponctuelle de glace.

CHAPITRE I

INTRODUCTION

1.1 Contexte

Au cours des mois hivernaux, du frasil se forme dans près de 60 % des grands bassins versants de l'hémisphère nord pour des périodes variant de quelques jours à plusieurs mois et contribue à la formation des glaces de rivière. L'action des glaces de rivière et la présence de frasil engendrent des coûts socio-économiques considérables (Beltaos, 2000; Morse et Hicks 2005). Depuis toujours, ces phénomènes climatiques endommagent les infrastructures submergées, menacent la sécurité civile (inondations) et modifient considérablement la morphologie des rivières (Prowse et Beltaos, 2002). Plus récemment, les changements climatiques semblent influencer le régime des glaces fluviales. On observe dans plusieurs systèmes fluviaux de l'Est du Canada, une augmentation des débits hivernaux, une réduction des débits printaniers, des changements dans les dates de gels et de dégels, une réduction des jours avec couvert de glace, une augmentation des événements de dislocation du couvert de glace, ainsi que des changements au niveau de l'épaisseur du couvert de glace (Burrell, 2009). Le frasil constitue l'une des composantes clefs de la dynamique fluvio-glacielle. Considérant sa sensibilité aux conditions hydroclimatiques, il importe de bien comprendre les processus liés à sa formation, son transport et son accumulation dans les rivières des régions froides.

Le frasil apparaît dans un écoulement turbulent en état de surfusion ce qui le distingue de la glace pelliculaire qui se forme à la surface de l'eau dans un écoulement laminaire (Michel, 1971). Le frasil est composé de cristaux aux formes circulaires ou elliptiques de tailles variant de 0.1 à 4 mm de diamètre (Unduche et Doering, 2007; Morse and Richard, 2009). En rivière, on associe intuitivement la production de frasil aux sections de fortes pentes où l'écoulement est turbulent et le couvert de glace est absent (chutes, seuils et rapides) (fig.1.1). La quantité de frasil produite est théoriquement proportionnelle à l'étendue de la surface d'eau libre et au taux de perte thermique (Beltaos, 1995). En transport dans l'écoulement, les particules de frasil croissent, se transforment et évoluent en d'autres formes de glaces. La croissance thermique des cristaux de frasil est gouvernée par les transferts thermiques entre le cristal de glace et l'eau turbulente en surfusion (Hammar et Shen, 1993). Dans une eau en surfusion, le frasil a la capacité d'adhérer à toute surface immergée (substrat, infrastructures, cristaux de glace). Sans être adhésif, le frasil en transit peut également être « harponné » par toute surface irrégulière ou rugueuse. Le frasil initie la formation de glace de fond lorsqu'il entre en collision avec le substrat de la rivière (Daly, 1994). Lié à d'autres particules de frasil, il compose un floculat. Le floculat est une masse d'eau-frasil non uniforme dont la densité est fonction de la concentration de frasil (Tsang, 1993). Alors que la turbulence de l'écoulement œuvre à détruire les agglomérations formées, le culbutage des floculats tend à compacter les floculats entre eux ce qui favorise l'incorporation de grains de frasil et la croissance intrinsèque des cristaux. Cristaux et floculats remontent à la surface lorsque leur taille devient suffisante importante pour augmenter leur flottabilité et contrer les forces de l'écoulement. En surface, ils s'agglomèrent et s'épaississent. pour composer d'autres formes de glace (plaquettes, assiettes, radeaux de glace). La formation dynamique du couvert de glace repose sur la juxtaposition, la progression frontale, la compression ou l'empilement de frasil et de ses formes glacielles parentes. Un couvert de glace dynamique peut occuper jusqu'à 100 % de la largeur du chenal et peut progresser vers l'amont à une vitesse de 40 km/jour (Michel, 1972).

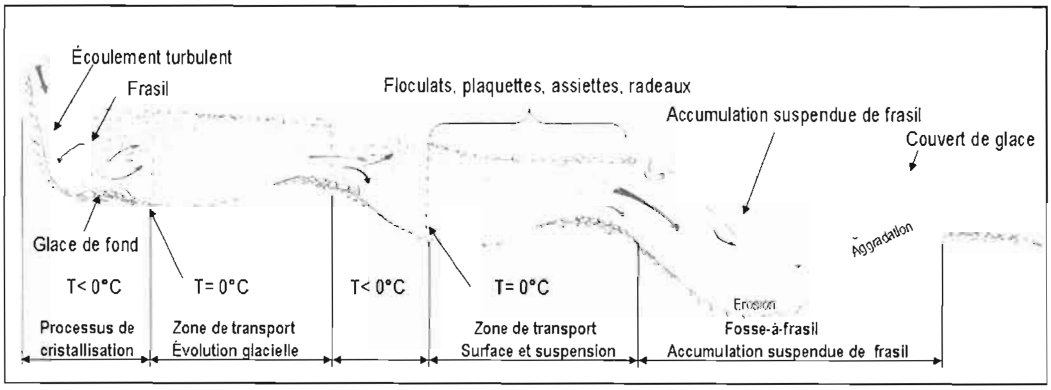


Figure 1.1 Illustration simple des interrelations entre les formes du lit à la dynamique du frasil en rivière, à l'échelle de tronçon.

En présence d'un couvert de glace, lorsque la force de l'écoulement excède la flottabilité des cristaux de frasil et floculats, ceux-ci peuvent être transportés sous le couvert de glace. Ils adhèrent et s'accumulent dans des zones à faible vélocité telles que les mouilles, méandres, exutoires de lacs, etc. Le terme « barrage suspendu » fait référence à une accumulation verticale de glace, de pellicules de glaces de fond, de frasil ou de leurs combinaisons (Shen et Wang, 1995). Un barrage suspendu de frasil peut causer une augmentation de niveau d'eau à l'amont et une augmentation de vitesse près du lit. L'affouillement localisé trouvé sous plus d'un barrage suspendu de frasil, confirme que l'accumulation de frasil sous couvert de glace a des impacts notamment sur le transport de sédiment en charge de fond et indirectement sur la morphologie du chenal (Michel et Drouin 1975; Beltaos, 1995; Sui *et al.*, 2006).

Les interrelations que partage la dynamique glacielle avec la géomorphologie fluviale et le transport des sédiments sont encore mal comprises. Des concepts comme l'altération de la sous-glacielle de la géométrie du lit et les ajustements hivernaux du talweg demeurent faiblement étudiés, principalement parce qu'il s'agit de processus ponctuels et éphémères qu'il peut être difficile de documenter suite à la disparition du couvert de glace. Le transport sédimentaire hivernal est une composante essentielle à la compréhension de la réponse morphologique sous couvert de glace. Pourtant, un nombre limité d'études porte sur le transport sédimentaire sous couvert de glace (Ettema et Daly, 2004; Sui *et al.*, 2000; Milburn et Prowse, 1996, 1998; Muste *et al.*, 2000). De plus, les mesures terrain de transport sédimentaire hivernal sont quasi inexistantes (Beltaos, 1993). Les particules de frasil, comme les particules de glace présentent une résistance au cisaillement de l'écoulement. Il est donc possible de s'intéresser aux interactions frasil-chenal selon la résistance du lit à l'écoulement ou selon la résistance de l'accumulation de frasil à l'écoulement. Des travaux de recherche menés sur le fleuve Jaune ont permit d'établir une relation positive entre l'accroissement vertical d'accumulations de frasil et l'augmentation de la déformation du lit. (Sui *et al.* 2000; 2006). Selon Sui, Wang et Karney (2000), pour une section transversale donnée, les formes du lit et l'accumulation de frasil chercheront à s'ajuster selon un minimum requis d'énergie (*least energy consumption)* afin de faciliter le passage de l'écoulement. Des efforts de recherche sont nécessaires pour aider à comprendre la contribution exacte de chaque composantes glacielles et fluviales impliquées dans la dynamique fluviale hivernale.

Plusieurs études se contentent d'une interprétation intuitive et simpliste de la géomorphologie fluviale lorsqu'il s'agit de comprendre son rôle dans les processus de cristallisation, de transport de glace ou de croissance des formes glacielles. Il en va de même lorsqu'il s'agit de prédire la réponse du chenal à la dynamique glacielle. La création d'un cadre d'analyse qui intègre la dynamique glacielle (cristallisation, transport de glace et morphologie glacielle) à la dynamique fluviale (structure des écoulements, transport de sédiment et formes du lit) pourrait aider à mieux comprendre la complexité des rivières à dynamique glacielle.

Plusieurs fosses anormalement larges et profondes sont présentes dans nos rivières, mais peu d'entre elles sont connues ou documentées (Lapointe, 1984; Gharabaghi *et al.*, 2007; Inkaras *et al*, 2009). La plupart de ces fosses présentent une morphologie difficilement allouable à la dynamique d'écoulement en chenal libre. Il a été proposé que les conditions glacielles pourraient être à l'origine de la morphologie anormalement profonde d'une fosse de 30 mètres de profondeur dans le delta du fleuve Mackenzie (Gharabaghi *et al.*, 2007). Des scénarios de vitesses d'écoulement sous couvert de glace ont montré que la force de cisaillement près du lit diminue lorsque le débit augmente du à l'inversement du gradient de pression. Nous ignorons toujours si l'entretien et l'évolution de telles fosses sont imputables à la dynamique des glaces.

1.2 Structure et objectifs

Ce mémoire s'intéresse à l'étude des liens entre la dynamique du frasil et la géomorphologie alluviale. Il s'organise autour d'un court chapitre de contexte et de deux chapitres-articles. Le choix de cette structure de présentation repose sur l'intérêt de disséminer les résultats de cette recherche à un large éventail de la communauté scientifique car plusieurs disciplines côtoient les interrelations entre la dynamique glacielle, la géomorphologie et la sédimentologie fluviale.

Le modèle conceptuel proposé dans le premier article vise à intégrer les efforts de recherche passés et futurs dans un cadre de travail universel de caractérisation fluvio-glacielle. Les interactions entre les composantes de la dynamique fluviales sont efficacement illustrées par la « trinité » conceptuelle de Leeder (1983) qui s'impose comme cadre structurel. Une « trinité » analogue est développée pour la dynamique glacielle. Le modèle proposé comprend six composantes et leurs interrelations : (1) l'écoulement (2) le transport de sédiments (3) la morphologie du lit (4) le processus de cristallisation (5) le transport de glace et (6) la morphologie glacielle. Les interrelations intrinsèques à la dynamique glacielle ainsi que celles qui unissent cette dynamique aux composantes de la dynamique fluviale sont soutenues par des études expérimentales provenant de la littérature. Ce modèle constitue un cadre d'analyse pour mieux décrire et comprendre la complexité des dynamiques physiques sous-jacentes aux rivières froides. Il se distingue par sa simplicité et sa capacité d'adaptation à une diversité d'environnements et d'échelles fluviales.

Le second article présente la chronologie des ajustements morphologiques d'une fosse-à-frasil suivant l'accumulation sous-glacielle de frasil et le degré d'obstruction du chenal par cette accumulation. Une fosse à frasil est un bas point topographique du profil longitudinal qui présente une morphologie « anomale » résultant d'interrelations entre les dynamiques glacielle et fluviale. La fosse à l'étude, d'une profondeur maximale de 8.2 m (niveau plein bord), est située à 14.1 km de l'embouchure de la rivière Mitis, Bas-Saint-Laurent. Le premier objectif spécifique de cet article consiste à comprendre les processus de production et de transport de glace qui mènent à l'accumulation de frasil dans la fosse d'étude. Au cours de la période de gel, un suivi qualitatif des conditions glacielles a été mené sur un corridor de 24 km à l'amont de la fosse. L'instrumentation de ce corridor et une caractérisation géomorphologique du corridor fluvial permettent de discriminer l'influence des variables climatiques, hydrauliques et géomorphologiques. L'analyse de ces données révèle l'existence d'un lien étroit entre un gradient hydraulique de 5 % et l'absence de couvert de glace sur un tronçon de 5.8 km de la rivière Mitis. Exposé à de basses températures, ce tronçon a la capacité de produire d'importants volumes de frasil au cours de l'hiver. Le deuxième objectif spécifique vise à documenter l'intensité et la spatialité de la réponse morphologique de la fosse à des événements spécifiques de natures hydrologiques ou glacielles. Une matrice de bornes sédimentaires dynamiques (dynamic bed-rods (DBR)), composées d'accéléromètres tridimensionnels a été déployée dans la fosse. Ce système a montré une relation claire entre la déformation du lit et la progression du couvert de glace sur la fosse-à-frasil. Le troisième objectif spécifique consiste à examiner l'évolution morphologique des formes glacielles et des formes du lit d'une fosse-à-frasil, sur une base mensuelle. Des relevés géophysiques et des sondages manuels révèlent l'épaisseur verticale des couches de glace selon 7 profils transversaux. Les résultats obtenus dévoilent une géométrie hydraulique hivernale unique liée à la présence d'épaisses accumulations de frasil aux formes irrégulières qui occupent jusqu'à 70 % des sections transversales de la fosse.

Une courte conclusion termine ce mémoire en soulignant les principaux résultats obtenus au cours de ces recherches. Diverses pistes de recherches pour développer nos connaissances sur la dynamique fluvio-glacielle sont également proposées.

CHAPITRE II

A CONCEPTUAL MODEL ON THE INTERACTIONS BETWEEN FLUVIAL AND ICE DYNAMICS

UN MODÈLE CONCEPTUEL D'INTÉGRATION DES DYNAMIQUES FLUVIALES ET GLACIELLES

Abstract

Rivers exposed to fluvial ice processes are complex systems composed of interdependent components that are difficult to study in isolation. Conceptual models are successfully used in fluvial geomorphology to simplify the complex structure of such fluvial systems and to highlight key interactions between the components. Leeder's conceptual « trinity » model describes river dynamics thought three components and theirs interactions: (1) flow (2) sediment transport and (3) bed forms. The strength of this model comes from its simplicity and applicability to nearly all fluvial conditions. Leeder's trinity, like other fluvial dynamics models, fails to integrate the peculiar of ice dynamics (Knighton, 1998; Bridge, 2003). The domain of cold river science has proposed a few conceptual models (Shen, 2003; Michel 1984). These models present the river ice system as a space-time continuum of ice forms and ice processes sharing some interactions with flow processes. Present conceptual models tend to exclude known interactions between ice dynamics, sediment transport and bedforms. No model actually fulfills the need for a conceptual framework to study river ice dynamics as an intrinsic component of fluvial dynamics. In this paper, we propose a conceptual model for the study of the complex interactions between fluvial and river ice dynamics. Strongly influenced by Leeder's conceptual model, the model attempts to integrate both fluvial and ice dynamics within a unified structure of components and interactions. We illustrate the models with several examples from field and experimental studies in order to identify key aspects of river ice processes from past and recent research.

Résumé long

Les environnements fluviaux exposés à la dynamique glacielle sont composés d'une myriade de variables qui interagissent de manière complexe et qui ne peuvent être étudiées isolément. De nombreux modèles conceptuels sont utilisés en géomorphologie fluviale pour simplifier la description et l'étude de la dynamique fluviale. Ces modèles sont composés d'un ensemble structuré de composantes distinctes et interdépendantes. La représentation de l'environnement fluvial sous forme de modèle visuel aide à identifier des relations critiques entre les composantes, à distinguer des manques d'informations et à désigner des sujets d'étude prioritaires.

La « trinité » conceptuelle de Leeder (1983) décrit la dynamique fluviale par le biais de trois composantes et de leurs interactions : (1) l'écoulement (2) le transport de sédiments et (3) la morphologie. Ce modèle se distingue par sa simplicité et sa capacité d'adaptation à la diversité des environnements et échelles fluviales. Malgré l'efficacité de la Trinité à décrire le système fluvial, ce modèle, développé pour des cours d'eau dépourvus de glace, échoue à intégrer la spécificité de la dynamique glacielle des rivières froides. D'ailleurs, aucun autre modèle destiné à la conceptualisation de la dynamique fluviale n'intègre les variables de la dynamique glacielle. Près d'une dizaine de modèles conceptuels de dynamique glacielle ont été proposés au cours des 40 dernières années. Certains de ces modèles, portent sur l'évolution spatio-temporelle des formes et processus de glace alors que d'autres s'intéressent à la relation entre la glace et les processus hydrologiques. Ces modèles sont complexes et les modèles courants excluent les liens connus entre la dynamique glacielle, le transport de sédiment et la morphologie. Aucun n'offre de cadre d'analyse organisationnel permettant de considérer la dynamique glacielle comme étant intrinsèque à la dynamique fluviale.

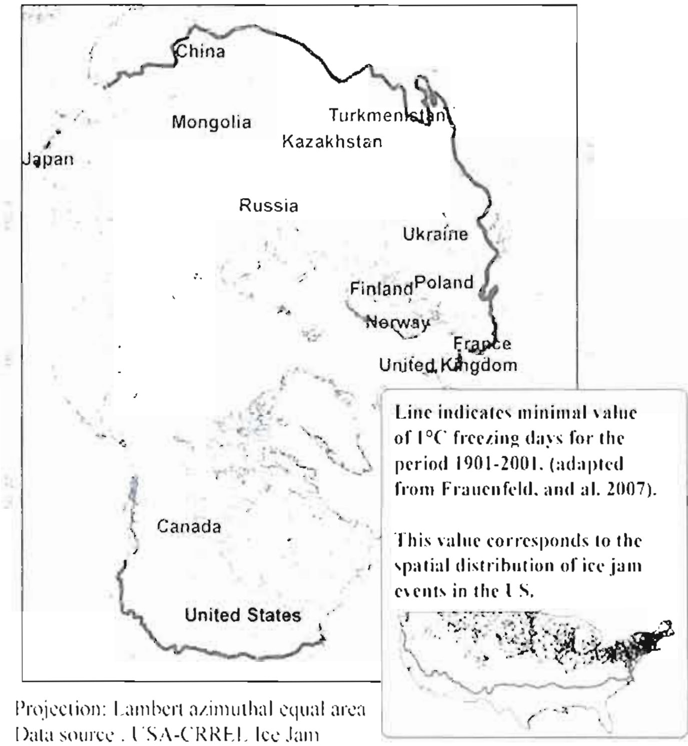
Ce chapitre, composé d'un article de recherche qui sera soumis à la revue scientifique *Progress in Physical Geography*, propose un modèle conceptuel d'intégration des dynamiques fluviales et glacielles. Dans un premier temps, le chapitre-article vise à définir la nature des processus glaciels par la description de la « trinité » de la dynamique glacielle. En analogie à la «trinité» de Leeder, la « trinité » de la dynamique glacielle est composée de trois composantes interdépendantes : (1) le processus de cristallisation (2) le transport de la glace et (3) la morphologie glacielle. Les liens rétroactifs que partagent ces trois composantes sont démontrés à l'aide d'exemples concrets tirés de recherches expérimentales passées et actuelles. Dans un deuxième temps, des liens entre la « trinité » de la dynamique glacielle et chacune des composantes de la dynamique fluviale sont identifiés. Dans un troisième temps, le modèle conceptuel de la dynamique fluvio-glacielle est proposé. Inspiré de la trinité de Leeder, ce modèle tente d'intégrer les deux dynamiques, glacielle et fluviale, en un concept unifié sous la forme de composantes et de liens. Les composantes dépendantes sont contenues dans des boites alors que les liens sont illustrés par des flèches. Les composantes indépendantes telles que les paramètres hydroclimatiques siègent à l'extérieur de l'encadré principal. Le modèle proposé s'adresse à ceux qui cherchent à comprendre les processus et entités abiotiques sous-jacents aux rivières froides. Il offre une approche universelle et simple pour visualiser la complexité des interactions intrinsèques à la dynamique fluviale des régions froides.

2.1 A conceptual model of cold rivers

Based on average monthly minimal temperature data and major basins distribution, it can be estimated that ice processes occur in 60 % of major river basins in the northern hemisphere (fig. 2.1). River ice is known to interfere with river usage (hydropower, water supply, transportation or sanitation), threaten human activities (flooding) and remodel fluvial morphological features. Freezing rivers have received little attention compared to the number of studies devoted to ice free rivers. Literature on river ice treat river ice processes mainly as periods of ice activity (freeze-up, evolution and break up) and as a major component of hydrological processes. Knowledge gaps remains in the domain of how ice and river systems interact as a whole (Shen, 2003; Ettema and Daly, 2004).

Conceptual models are widely used in fluvial geomorphology to help understand the structure and dynamics of idealized alluvial systems (Knighton, 1998; Bridge, 2003; Leeder, 1983). Leeder (1983) proposed a simple, yet efficient, model strictly composed of three dependant variables: (1) flow structure (2) sediment transport and (3) bedforms that interact over a range of time and space scales (fig. 2.2). The « trinity » model illustrates well the continuous feedback system at work between flow and sediment transport can determine the ability of a river to modify its morphology. Easily applicable to nearly any situation, the model is largely acknowledged by multiple citations, including citations coming from Best 1993's adaptation. Authors like Clifford (1993) have brilliantly used and adapted the 'trinity' to present, in simplified terms, the main features of their study.

Albeit the efficiency of Leeder's trinity to describe the dynamics of fluvial system, it fails to adequately integrate the peculiar dynamics of cold rivers. This is also true of many other models of the fluvial system. The factors pertaining to the severity of an ice jams (discharge, channel width and slope, hydraulic resistance, strength and characteristics of a jam, ice volume, water temperature and heat transfer, strength and thickness of ice cover during break-up) enlighten the interactions that exist between ice dynamics, flow, sediments transport and bed forms that are not easily included in actual models of fluvial dynamics (Beltaos, 1995).



Projection: Lambert azimuthal equal area
Data source : USA-CRREL Ice Jam
Database (2008); Global Runoff Data Center (2007); Frauenfeld and al. (2007).

Figure 2.1 Cold rivers subject to freezing in the northern hemisphere. Freezing index is based on a continuous long-term 25 km × 25 km grid monthly temperature dataset for the Northern Hemisphere. Monthly dataset is derived from CRU TS 2.1 daily air temperature observations.

The cold research community offers a comprehensive understanding of river ice dynamics through conceptual models. Most of these models, listed in Table 1, outline river ice processes through three main components (crystallization process, ice transport and iceforms) and relate to the space-time evolution of iceforms (freeze up, ice covered and breakup). Michel (1971) drew a classification scheme based on macroscopic iceforms in lakes and rivers. Michel (1972) discriminated frazil crystallization and frazil transport as an inactive form into a simple schematic representation. Michel's (1984) flow chart distinguishes two conditions of ice cover formation. Gray's (1981) and Shen's (2003) schematics classifications resulted in complex diagrams where an emphasis was given to flow turbulence. Those conceptual models tend relate to a unidirectional bond between flow and ice processes and do not encompass ice retroactive influence on the river system. Lawler's (1993) diagram refers to both the river ice and the fluvial dynamics. Being applicable only to needle-ice processes, it does not fulfill the need for a conceptual frame that considers river ice dynamics as an intrinsic asset to fluvial dynamics. No model presently assesses the scope of interrelation between ice dynamics and morphological processes.

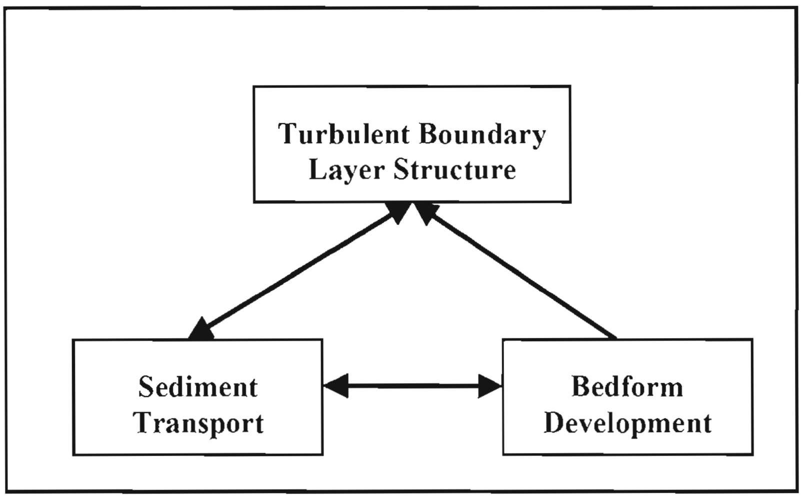


Figure 2.2 The 'trinity' of flow, sediment transport and bedform as a feedback loop (from Leeder, 1983).

Table 2.1
Conceptual models of river ice processes

Authors	Year	Conceptual models	Components: crystallization	transport	iceforms	hydroclimatic conditions	bedforms	flow	sediment transport
Shen H.T.	2003	River ice processes	x	x	x	x		x	
Lawler D.M.	1993	A simple conceptual framework for the study of sediment mobilization associated with ice growth	x		x	x	x	x	x
Nouveau Brunswick River ice manual	1989	River ice formation processes		x	x			x	
Gray D.M.	1981	Schematic classification of processes and form for freshwater ice	x	x	x	x		x	
Michel B.	1984	Flow chart of complete ice cover formation processes during one time interval	x	x	x	x		x	
Michel B.	1972	Development of frazil in a river	x	x	x			x	
Michel B.	1971	Processes of ice formation in rivers and lakes	x	x	x	x		x	

In this paper, we propose a conceptual model applicable to cold river systems. The model seeks to present a new perspective on cold river physical investigation by including morphological components. The model emerges from a comprehensive analysis of the interrelations between two sketched trinities and their six dependant components (fig. 2.3). The first three components belong to Leeder's fluvial dynamics trinity and are: flow, sediment transport and bedforms. The last three belong to the river ice dynamic's trinity and are: the crystallization process, the transport of ice and the iceforms. In the past, both trinities and their interactions have often been investigated in isolation of one another and an integrated model has yet to emerge.

The paper defines the nature of river ice processes first by describing the river ice « trinity » and, secondly by integrating the river ice trinity and the fluvial dynamic trinity within a unified conceptual mode. Several key aspects of these interactions are set against past and recent experimental and field research. The model seeks to provide a better overall understanding of cold river dynamics by identifying clear bonds between fluvial and ice related components, at small and large scales. The advantages and limits of the model are also discussed.

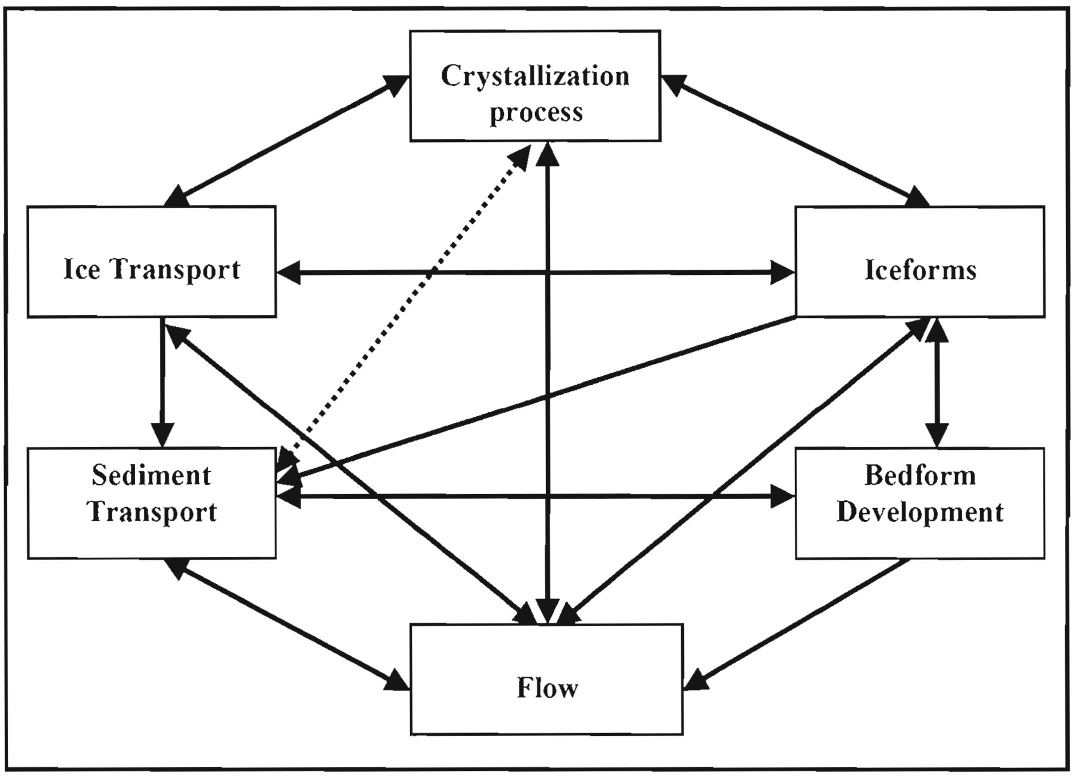


Figure 2.3 The «crystal model» on the interactions between fluvial and river ice dynamics. Dependant components are presented into boxes and interrelationships and feedbacks in arrows. Dotted arrows indicate hypothetical bonds necessitating further research. Independent variables are excluded from the model.

2.2 The trinity of crystallization process, ice transport and iceforms

The trinity of ice dynamics appears when independent variables, such as hydroclimatic parameters, allow the appearance of ice. The crystallization processes initially require the cooling of the water surface close to the freezing point and an unlimited presence of cores of nucleation. Crystallization is the process by which ice appears when initial conditions are met. Any change in external parameters might induce a change in the river ice dynamics « trinity ». Ice transport is when ice crystals or any iceforms are being moved with the flow in a river system. The agglomeration of transported ice crystal or the *in situ* growth of ice might induce an iceform. The detailed description of each « river ice trinity » components is followed by a definition of the nature and strength of their interrelations (fig. 2.4).

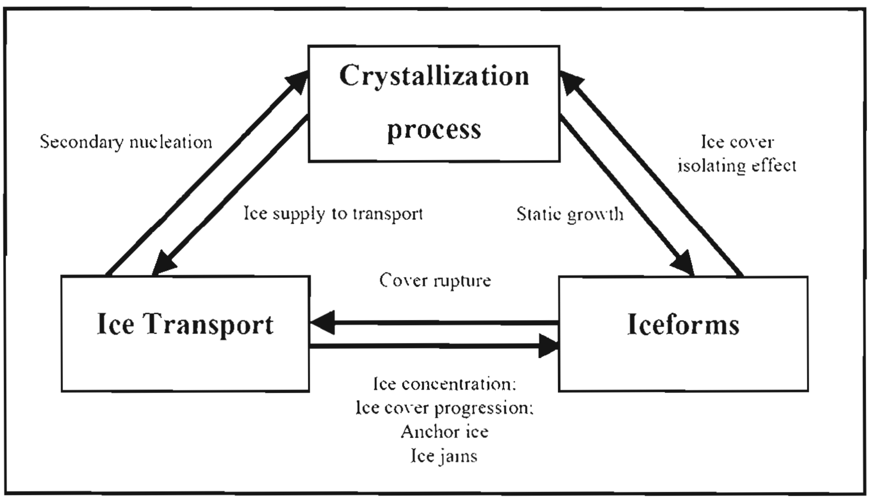


Figure 2.4 The «ice trinity» model between crystallization process, ice transport and iceforms components.

2.2.1 Crystallization process

Each type of ice crystal has its own genesis and its own hydraulic effects. Three types of ice grains are described in the literature: skim ice, frazil ice and snow crystal. Skim ice appears in laminar flows at Froude number lower than 0.24 and is formed by nucleation at water surface (Unduche and Doering, 2007). It grows in number rapidly and is transported downstream, on the surface or into the flow. Skim ice can cover calm waters even if the temperature of the principal flow is above the freezing point. It differs from frazil ice because it cannot be initiated from frazil crystals while the contrary is known to occur.

The frazil ice starts with the process of supercooling (when water temperature is below the freezing point) and the formation of frazil crystals on floating nucleus (air bubbles, fine sediments or frazil crystal). Supercooling conditions and frazil crystallization are closely related to air temperature and flow turbulence. Once generated in a turbulent flow the frazil crystal will grow, flocculate or collide with other crystals. Secondary nucleation is known as a frazil multiplication process caused by crystal collision. The size of the particles produced by this process is among the smallest sizes of frazil found (Hammar and Shen, 1993). While in supercooling conditions, frazil ice particles are considered «active» or «sticky» and attach to almost any object including other ice crystals. Frazil ice particles transported out of supercooling conditions become «non-active» or «passive» frazil crystals. Snow crystals blown onto the water surface of a river form the third source of river ice grains that lead to the process of « slushing ». Snow crystals, skim ice and frazil ice crystals constitute the fundamental grains of all ice forms.

2.2.2 Ice transport

Subject to the critical velocity concept, ice transport is initiated when ice particles are being displaced with the flow in a river system. Ice transport addresses all scales of transports from grain mobilization to the transport of a dislocated ice cover. Ice transport usually implies the transport of ice in the downstream direction but upstream ice transport can result from backflow conditions. Ice can be transported in two different modes: in suspension or near the surface. Suspended ice transport takes place when ice particles or iceforms are maintained in the flow by turbulent mixing processes. The transport of ice particles is analogous to the transport of sediment particles. Bedload transport formulas for low density sediments have been adapted to calculate the ice-transport capacity of the flow although the density difference between frazil ice and water is 0.92:1 while it is 2.65:1 for sand and water (Shen and Wang, 1995; Ye and Doering, 2001). Surface ice transport begins when ice buoyancy overcomes the flow turbulence. Surface ice transport is controlled by how much surface area is available for the transport of ice. It is influenced by several internal and external factors such as wind drag, secondary flow current, emerging flow from tributaries. An interchange exists between the two modes of transport depending on 1) the carrying capacity of the flow and 2) particle shape and size.

A distinction can be made between three transport periods: the freeze-up ice runs, the undercover transport, and the surface ice cover movement. The undercover transport is, when the ice-cover creates an additional boundary to ice-loaded stream flow and its underside surfaces (sinks) become available to frazil ice accumulation. Ice cover movement occurs when surface ice develops as a moving cover or when a primary cover is transported downstream following break-up. Transport transitions are observable in time and space and each transport state is associated to certain ice forms.

2.2.3 Iceforms

Iceforms are the product of the accumulation of transported ice and of the *in situ* growth of ice. They can be explained by their stage (initial, transitional and final forms), source, magnitude, shape, migration pattern, position into the flow and stability limits. They can also be static accumulations, moving accumulations, iceforms bonded to the ice cover or not. Static ice accumulations produce iceforms 1) anchored to the substrate and/or 2) non-available for transport. Moving accumulations are available for transport. Table 2 defines several examples of iceforms. A schematic representation (fig. 2.5) of a broad classification of river ice illustrates the complexity of iceforms evolutions. Iceforms urgently necessitate a more detailed classification to ease our understanding of its complexity and diversity.

2.2.4 A complete river ice trinity

Figure 2.4 sketches the river ice trinity and highlights several bidirectional bonds between the three components. To fully describe the river ice trinity, observations concerning these bonds are drawn together from existing scientific knowledge. The process of secondary nucleation first serves to comprehend the works of a unidirectional bond between the Ice transport and Crystallization components. Secondly, a plural relation between the three ice components is explained with the formation and growth of anchor ice. Thirdly, the delicate retroactive bond between ice load and ice transport capacity provides an example of a well defined bidirectional bond.

Table 2.2
Iceform types and definitions

Static ice forms

- **Anchor ice** is composed of submerged frazil crystals growing or adhering to, underwater objects, vegetation of to the bed substrate into flocculent masses. It is a transitional form composed of frazil. It migrates downstream in flocculent masses.

- **River ice mounds or hummocks** are localized mounds or rounded knoll of ice rising above the general level of an ice cover. They can result from icing or from the grounding of surface ice to the bed;

- **Border ice**, a form of skim ice cover that grows from the lateral accumulation of surface ice pieces.

Moving accumulations

- **Surface ice sheets** results from skim ice growth without anchorage to the banks. It grows downward by freezing at the ice-water interface and grows laterally by particle growth and interlocking (Unduche and Doering, 2007);

- **Slush ice** is floating clusters of ice. It can be formed by anchor ice released from anchorage or by the agglomeration of frazil flocs on surface. **Frazil flocs** are the product of frazil crystals adherence to other crystals. They are compacted by the tumbling action of the flow and are carried downstream into large packs of slush ice on surface (Kerr *et al.* 2002; Tsang, 1993).

- **Frazil floes** result from the freezing of the surface slush layer into shaped floes with edges or ridges like ice pans. Floes rely on bonding between frazil particles with some degree of internal strength. This effective cohesion can help the floes resist to failures. Floes may sinter together and form interfloe bonding which differs from interparticle bonding (White, 1993).

Bonded iceforms

- **Frazil dunes or bars**, undercover frazil accumulations shaped like small sand dunes that do not obstruct the flow in the manner of a frazil ice jam. It is still uncertain to how a frazil dune migrates but it does not seem to move like sediment in sand dunes (Tsang and Scuz, 1972; Lawson *et al.*,1986)

- **Frazil ice jams** are static undercover accumulations of frazil particles obstructing the flow. Frazil ice jams can be interpreted as a hanging dam but this term applies to ice and/or frazil undercover accumulations (Shen and Wang, 1995).

- **Icings** form on existing ice covers and are sheet-like mass of layered ice formed from successive overlying flows of water during freezing temperatures.

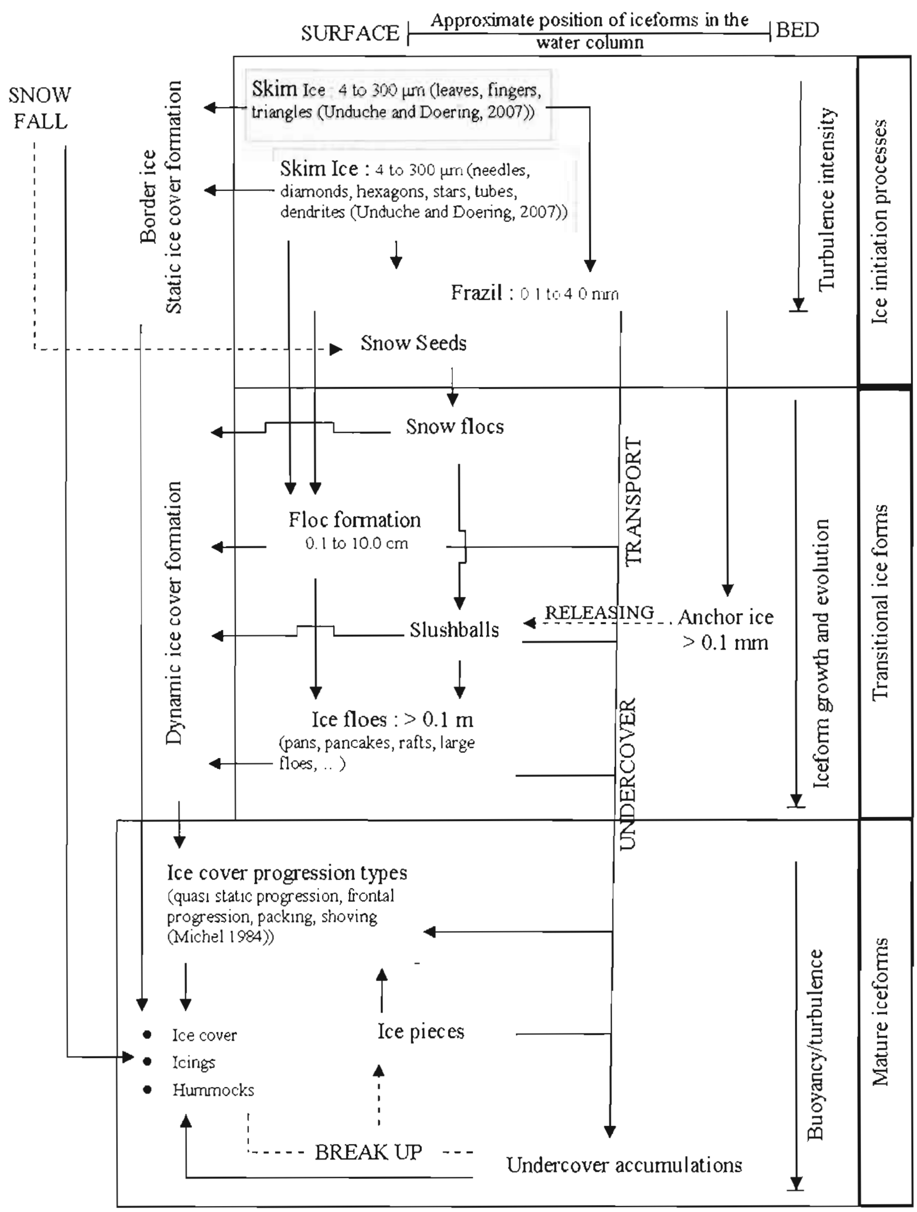


Figure 2.5 Evolution of iceforms in cold rivers, from ice initiation to ice cover break-up. Length scales are from Morse and Richard, 2009.

Let us first examine how ice transport relates to ice crystallization processes. Ice crystallization, or primary nucleation, is the formation of an initial crystal of sufficient size where no crystal was found. Secondary nucleation involves the crystallization of a new layer of ice on the existing flat crystal surface. This process takes place during ice transport which consequently influences the ice crystal growth capacity. Secondary crystallization is attributable to collisions between already existing crystals with either a solid surface or with other crystals (Clark and Doering, 2007). Secondary nucleation is known to be a very effective and common mode for nucleation. In rivers, ice concentration and turbulence enhances secondary nucleation by the compaction of frazil crystal or by the fragmentation of iceforms (Shen, 2003). Ice transport therefore enhances the formation of ice crystals through 1) the concentration of ice particles and 2) the process of secondary nucleation.

A second, intricate example is the formation of anchor ice, an iceform that grows and evolve on the riverbed substrate or on any underwater objects. Anchor ice is a derivative form of active frazil crystals that accrete on submerged surfaces under supercooling conditions. Anchor ice grows to considerable thicknesses and can emerge to become a major component of the solid surface ice cover (Dubé *et al.*, 2009). Laboratory experiment has shown the initiation of anchor ice by frazil attachment. The majority of anchor ice grows by massive frazil attachment but laboratory evidence has shown that anchor ice also grows by thermal growth based on heat balance (Kerr *et al.*, 2002; Qu and Doering, 2007). Kempema and Ettema (2009) obtained field evidence that anchor ice growth is related to the fusion and growth of agglomerated frazil crystals. Near bed frazil concentration, which occurs in transport phase, is intimately tied to the accumulation of frazil ice in the form of anchor ice and anchor ice could not exist without the presence of active frazil crystals (Morse and Richard, 2009). This example shows how an anchored iceform clearly relate to all trinity components.

Ice transport controls the iceform load through a delicate equilibrium between supply-limited and capacity-limited transport. The ice transport capacity is the volume of ice that can be carried by the flow under given conditions. This implies a close bond between the amount of ice moved by the flow and ice availability. Ice availability comes from spontaneously crystallized ice particles and from evolving moving iceforms. When ice supply overcomes the ice transport capacity at a location, the ice particles collect into surface iceforms or accumulate under an ice cover. The ice transport capacity therefore controls iceform evolution through ice deposition and iceform erosion.

These short statements illustrate the relevance of the bidirectional bonds between the three ice trinity components. This river ice trinity is a never ending cycle that cannot be neglected when studying cold river environments. Some interactions are less obvious than others but it is a reality that they exist and have to be taken into account.

2.3 Cold river dynamics

The challenge of this paper is to develop an analytical framework to enlighten the interrelationships between geomorphologic processes and river ice dynamics using a simple schematic representation. Attention is strictly given to physical processes excluding thermal, chemical, and biological processes. A conceptual model integrating both fluvial and ice dynamics is proposed (fig. 2.3). This section develops on the relationships highlighted by the model to display its usefulness.

2.3.1 Flow structure and ice dynamics

Interactions between river ice dynamic and flow hydraulics are numerous and evolve constantly over a winter cycle. Some interactions are clearly visible - the collapse of an ice cover resulting from stage variations - while others are less obvious - the resistance effect of frazil granules on flow. These interactions are well documented in the literature and they are addressed here in two parts. First, we look at three examples of ice-induced flow changes: the effect of ice dynamics on flow stage, flow dampening by ice and flow diversion by ice. Secondly, some effects of flow on ice dynamics are described.

River ice effect on flow

A backwater effect is an increase of the water surface elevation upstream from and as a result of a significant reduction in channel cross-sectional area or an increase in the hydraulic roughness of the channel. A constriction to flow can be caused by many types of iceforms: grounded ice (anchor ice, icing, ice cover grounding), shore ice, hanging dams, and massive ice crystal presence in the water column. The backwater effects associated with them are generally localized. The upstream extent of backwater depends upon the scale of the iceform and the slope of the channel. Blachut (1988) classified ice-induced backwater effects into two categories: a) the 30-40 cm stage rise related to anchor ice, frazil ice and margin ice; b) >1m stage rise related to the ice cover. The effects of backwatering include increased flood levels, (increases in upstream flow depth and wetted perimeter), flow redistribution, water table elevations, and reduced reach transport of sediment. Other effects associated with reduced sediment transport include channel aggradation, channel widening, bank erosion and increased channel meandering. Backwater effects extend much further on low-gradient streams than on high gradient streams. In this case, iceforms share a direct bond with stage variation, with indirect consequences on sediment transport riverbank and bed deformation.

The progression of a waterwaves out ahead of the icewaves is well documented in the literature (Henderson and Gerard 1981; Daly 1993, 1994; Beltaos 2004, 2005, 2007; She and Hicks, 2006). Waterwaves are the propagation of crests and troughs upstream or downstream as a consequence of unsteady flows. A surge is the most violent and spectacular type of riverwave events that follow within minutes the release of a jam. The surge wave may rise by a few meters presenting an abrupt front. It propagates downstream at high velocity posing a risk to downstream structures, people, and aquatic life (Beltaos, 2005). Waterwaves also result from the transition of a moving cover to a stationary cover during the formation of an ice jam (Henderson and Gerard, 1981). The duration and magnitude of a river are function of a balance between friction, inertia and ice properties (Ferrick, 1985). Interestingly, it has been proposed that low amplitude water-waves could be responsible for the fracture and break-up of ice covers (Daly, 1994). Work on waterwave amplitude by Beltaos (2004) have shown that the only wave type capable of generating wave-fractures in an ice cover is that of singular high amplitude waves (or surge) that result from an jam release, thereby explaining the ice clearing capacity of surges. This case shows how an ice breakup event has significant effects on water levels with retroactive impacts on downstream iceforms and throughout the river system.

The effects of the ice trinity components (crystallization processes, ice transport and iceforms) on flow are diversified and are here only sparsely presented. Other examples could easily be used to the display the relevance of the illustrated interrelations. A simple list of some of the most evident cases to our minds follows: 1) channel roughness is increased with the presence of an ice cover or of surface ice. The ice subsurface forms a boundary layer at the top of the flow velocity profile, reducing cross-sectional flow velocities; 2) Channel conveyance is reduced by iceform accumulation and growth and portions of the channel may become unavailable to flow. Such reduction in channel conveyance can require adjustment of the channel to its effective roughness by an increase in flow depth. 3) Dampening of turbulence and enhancement of flow viscosity could occur as a result of frazil ice concentration. The resistance effect of frazil particles on natural flow regime is not yet understood (Shen and Wang, 1995); 4) Ice jams and ice covers are known to divert flow lines, icings to reroute channel flows elsewhere, and anchor ice to causes short duration flow diversions. Ice-induced flow diversions result in channel flow velocities reduction to values below those required to mobilize the channel's bed and in localized flow velocity concentration and scour.

Effect of flow on ice dynamics

The ways in which flow affects the trilogy of ice dynamics is depicted from three examples. The first example concerns the interrelation between turbulence and crystallization. High turbulence intensity is related to frazil crystallization and anchor ice formation whereas low velocity flows govern skim ice formation and static ice forms. Frazil crystallization is known to be governed by water temperature and turbulent structures. Turbulence significantly affects the supercooling processes which closely relate to frazil crystal geometry, frazil particles diameter and the vertical distribution of frazil ice concentration (Ye and Doering, 2007). The exact contribution of turbulent structures to this relation remains unexplained but is nonetheless existent.

The second example refers to the movement of ice particles and iceforms along distinctive flow lines, known as ice trains (fig. 2.6). Ice train trajectories are influenced by flow hydraulics, wind drag on surface ice and emerging flows from tributaries. Flow withdrawal from stream can also affect the ice trajectory (Calkins, 1993). Laboratory analysis on particle trajectory in river bends clearly showed the effect of centrifugal force on drifting ice (Urroz and Ettema, 1992). This is depicted by a radially outward drift of iceforms constrained to a narrow band toward the outer bank. The entrainment of ice grains and iceforms into larger flow dynamics enhances iceform concentration, flocculation and floes collision.

The last example of flow's effect on ice is the variability of ice jam thickness in relation to flow intensity. Healy and Hicks (2007) present a clear relationship between Froude numbers and jam thickness profile where the jam thickness diminishes with a stronger Froude number. Results show ice jams thinning in highly dynamic flow conditions compared to steady base flow conditions for similar discharges.

Interactions between river ice dynamics and flow hydraulics are diverse and well covered in the literature. A bidirectional bond resides between the flow component and each of the three ice trinity components. Several examples served to validate the interrelations outlined in the proposed model.

Figure 2.6 Ice trains at river bend PK21 from estuary, Mitis River, Québec. Photography by Jerôme Dubé.

2.3.2 Sediment transport and ice dynamics

Plenty of work has been done in the domain of sediment transport in open channel flows but sediment transport in ice loaded flow is still at its infancy. Tuthill (2005) identified « a lack of adequate analytical or numerical models to predict sediment transport under ice covers ». Some studies have addressed the interactions between suspended ice crystals and small sediment particles (Eidsvick, 1998), the ice-cover influences on sediment transport (Ettema and Daly, 2004; Sui *et al.*, 2000; Milburn and Prowse, 1996, 1998; Muste *et al.*, 2000) and the transport of sediments into drifting ice (Kempema *et al.*, 2002). The complexity of sediment transport under ice conditions is described and illustrated in the proposed model.

Ice rafted sediments

Ice-rafted sediment transport is the initiation of sediment movement in response to the uplifting (grounded ice) or pushing of sediments by ice. It includes the displacement of sediments embedded into the ice cover and the transport of sediment-laden ice slush. Ice rafting is known to occur but there are too few and unreliable measurements of ice-rafted sediment transport rates at present. Ettema and Daly (2004) produced a unique report dedicated to the impacts of river ice on sediment transport. The report concludes that drifting ice is an important transport mechanism for sediments. Sediment-laden anchor ice releasing is an underestimated transporting agent. Anchor ice forms over different material and is observed mostly on boulders, stones, gravel, coarse sands and aquatic weeds. Anchor ice is rarely observed over sand, silt and clay substrate because small size substrate is easily lifted off under the rapidly growing buoyancy of anchor ice. Also, compact substrates are more prone to geothermal heat, limiting the anchorage of ice crystals. Anchor ice releasing is capable of moving substantial quantities of sediments in some rivers (Ettema and Daly, 2004). Observations from Kempema (2002) show evidence of anchor ice rafting of sediment particles of sizes up to 1700g (cobble). Anchor ice releasing often occurs at low flow, a counterintuitive period for the transport of large sediment particles. Chain reaction anchor ice releasing was documented in laboratory as a result of the detachment and impingement of singular anchor ice pieces (Kerr *et al.*, 2002).This dislodgment of large particles indirectly affects sediment transport by destroying the upper surface layer of coarse sediment (pavement or armored layer), reducing the degree of « consolidation » of the bed and exposing the underneath layer to the flow. Under supply-limited condition, this can lead to a net scouring of the river bed (Vericat *et al.*, 2006). Much of the ice-entrained sediments become included in the ice cover. It is stored over the winter and it is displaced at break-up. Conversely, anchor ice also prevents sediment transport, before releasing, by covering the bed with a blanket of ice. Ice rafting is a sediment transporting agent in cold rivers. This relation is illustrated by a single-headed arrow in the proposed model.

Sediment scavenging by ice crystals

It is believed that high quantities of suspended sediments increase frazil formation. Calkins (1993) noted a decrease in water turbidity following frazil crystallization and speculated that frazil ice had scavenged the suspended particles causing the turbidity of the water. Sediment entrainment models for oceanic conditions suggest that sediment entrainment is an intermittent process limited by supercooling conditions (Eidsvik, 1998). Studies performed by Mueller and Calkins (1978) showed no evidence that suspended organic material had enhanced nucleation. The hypothesis that suspended sediment transport is necessary for nucleation remains to be proven in the field. In the model, the dotted arrow illustrates this hypothetical bond in need of further research.

Complex relationships between sediment transport and iceforms

The proposed model has the capability of illustrating complex relationships encompassing multiple components and bonds. The following case makes use of multiple components (iceform, flow, sediment transport, bedform) and linking arrows to comprehend the indirect effects of undercover iceforms on sediment transport and on bedform change.

Undercover frazil accumulation and ice jams can redistribute flow velocities causing sediment transport and local bed degradation where the bulk velocity of flow is increased (Ettema and Daly, 2004). Undercover frazil accumulations can significantly enhance the overall thickness of an ice cover. In the Mitis River, up to 70% of the cross-sectional area was occupied by a thick and irregular undercover frazil layer (Allard *et al.*, 2009b). There are many observations of localized bed scouring under massive undercover ice accumulations known as hanging dams or suspended dams (Michel, 1971; 1975; Lapointe 1984; Beltaos, 1995; Sui *et al.* 2006). Extensive studies in the Yellow river established a positive reinforcement between frazil jam thickness and riverbed deformation through ice-induced hydrodynamic conditions (Sui *et al.*, 2000). Scour forms are found where the flow is concentrated by the frazil ice masses whereas the calmer parts of the channel are protected from deformation by the thick frazil accumulations. Research efforts are needed to comprehend the exact involvement of each component in this complex relationship.

It is often necessary to imply the flow component to explain the relation of ice dynamics to sediment transport. The following cases could also serve to support the illustrated bonds. Although it is rarely documented, ice jams have lead to the partial or full avulsion of rivers (Schumm, 2005). In the Niobrara River, ice blocks and ice sheets were seen at the time of avulsion (Ethridge *et al.*, 1999). The movement of large ice pieces during break-up surges erodes great quantities of material from the bed and banks. High celerity surge flows are also known to set the bed in motion, moving particles up to 20 cm in diameter and causing considerable scour in unexpected locations (Beltaos, 1993).

2.3.3 Bedforms development and ice dynamics

What is often called ice impact relates to the effect of ice dynamics upon bedforms. It focuses on the ways whereby flow and ice interact with bedform development and vice versa. The degree to which river ice dynamics lead to channel response depends on an interplay between: flow discharge, sediment transport, ice characteristic and mobility, and channel morphology and strength. A channel's response to ice-dynamics occurs over a large range of scales in space and time. Ice influence might be punctual, seasonal, annual or durable. It extends from the displacement of a sediment particle to the modifications of the profile gradient.

The reverse relationship is also true and in some case, is indivisible from the first. Fluvial morphology plays an important role in frazil ice production, ice jams formation and general ice dynamics. Certain ice processes are clearly associated to specific fluvial morphologies or transitions. Some important morphological features regarding ice dynamics are: bed geometry and slope, channel width, confluence location, bed material and sediment transport, bank and overbank characteristics.

Fluvial geomorphology (cross-sectional form, slope, bed configuration and planform) reflects flow pattern and sediment transportation (Leeder, 1983). The ways in which ice influences the fluvial dynamic components can be regarded in two ways: 1) the influence of ice on the relationships between flow and fluvial morphology or 2) the direct effects of ice on channel morphology. The delineation of scales of interactions provides an effective basis for the examination of geomorphologic features and their relation to ice dynamics.

Ice on large scale morphology

The concept of ice as a geomorphic agent is not new but, the ever-present question of cause and effect between morphology and ice, keeps the understanding of large-scale and long-term effects of ice dynamics on channel morphology at its infancy. Several studies have been and are still conducted at this scale. Ettema (2002) proposed that, for a typical range of ice cover thickness, the ice cover influence on morphology may vary with channel size. Therefore there must be a middle size range of channels for which the ice cover would be more morphologically active. This remains unproven. In 2005, Best, McNamara and Liberty attempted to explain the co-evolution of channel morphology and long-term average ice condition. A clear relationship was found between hydraulic geometry and iceform transitions. Taylor *et al.* (2009) proposed that fluvial transitions might be more susceptible to ice jamming than homogeneous channel segments. Based on this hypothesis, ice scars should be more abundant in heterogeneous stretches, where strong slope breaks are found.

Surface ice on bedforms

The indirect, and complex, influences of river ice on bedforms are numerous. As previously discussed, the presence of ice can generate a local imbalance between sediment supply and sediment transport capacity resulting in localized scouring or deposition. It is well understood how ice jams relate to lateral variations in size and shape of cross-sectional form by a redistribution of the flow (Smith, 1979; Sui *et al.*, 2006). What remains unclear is to what extent river ice induces irregularities in a channel planform. Based on the concept that a thalweg is more sensitive to flow variability than its bankfull channel, Ettema (2002) attempted to predict changes in thalweg sinuosity and alignment for several fluvial styles. In presence of an ice cover, a meandering channel thalweg is expected to straighten and meander-loops to shorten; for sinuous-braided channel, a decrease in effective energy gradient may cause the flow to concentrate, increasing the thalweg sinuosity; in braided channels, ice-cover presence may concentrate flow into the larger subchannels. Zabilansky *et al.* (2002) documented alternating thalweg shifting and deepening following ice cover formation. The authors hypothesized that thalweg switching is a recurrent process taking several winters to complete. Although not observed, thalweg entrenchment is also possible. Hicks (1993) investigated the contribution of ice jams to the development of an anastomosing river reach. River ice observations led to the conclusion that the development and subsequent enlargement of new channels in an anastomosing reach of the Mackenzie River were attributable to the rare and extremely high water levels associated to recurrent ice accumulations.

Moving ice on riverbed and banks

In a meander bend, river ice runs are believed to accelerate the downstream migration of meander loops, by concentrating the ice charged flows toward the apex of the loop and eroding the outer bank. The direct contact of moving ice abrades, pushes, and scours riverbed and banks, removing large quantities of sediment and vegetation and creating unique bedforms like the Bechevnick, introduced by Hamelin in 1979. The Bechevnick is an ice maintained lower-bank inclined shelf resulting partly from ice-abrasion and partly from the deposition of sediments and debris left by the melting of ice rubbles after ice runs. Moving ice generates several other types of erosional and depositional features. There are ice-push features (ridges, boulder lines, boulder pavement, debris piles), commonly found along the banks, at upstream end of islands, and outside of channel bends and ice-scour features (gouges, ploughing tracks, scratching and polishing of bedrock) (Ettema, 2002). It is believed that the most significant abrasive features are formed during dynamic events, such as mechanical break up, when moving ice sheets are forced to the banks at maximal ice strength, flow velocity and stage (Scrimgeour *et al.*, 1994). Certain channel segments more sensitive to disturbances are more likely to be disturbed by river ice. Like meandering and braided rivers, irregular rivers tend to be more eroded by ice due to shore configuration (Ettema, 2002).

Channel enlargement by moving ice has been discussed by many. In 1979, Smith proposed that ice runs could enlarge channel cross-sections at bank-full stage by as much as 2.6–3 times. This statement was strongly questioned by Kellerhals and Church (1980). In small streams, Boucher *et al.* (2009) found that, where ice jams were frequent, channels appeared enlarged and presented typical two-level ice-scoured morphologies. Contrastingly, in meander bends developed in more cohesive sediment, ice jams could be responsible for the narrowing of channels by overbank flows. Overbank flow induces sediment deposition increasing bank heights and reinforcing meander loop. Further studies are however needed to better understand the influence of ice in such geomorphic adjustments in regard to overall environmental processes.

Anchor ice on bedforms

The relation of anchor ice to bedforms is regarded at three scales: the reach scale, the particle structural arrangement scale and the grain scale. Anchor ice dams (or ice-riffles) are localized solid accumulations of submerged ice anchored to the bottom that raises the water level. Anchor-ice dams alter the channel roughness by raising the effective channel elevation leading to backwater effect upstream, and plunging flows downstream (fig. 2.7). The backflow of water promote the deposition of fine sediments in the channel while the plunging flows create localized scouring on the bed. At a smaller scale (particle structural arrangement), anchor ice strongly affects channel morphology in opposite ways. During the initial stage of anchor ice formation, the roughness of the bed is increased by ice-induced grain size exaggeration. After initial growth, the bed-ice surface roughness decreases as ice grows out of gravel-bed, smoothening the surface. The anchor ice protrudes into the flow during the transitional growth stage. At this stage the drag on anchor ice is increased and leads to flattening. When the bed is covered by a blanket of anchor ice, the bed sediment mobility is reduced. Finally, the release of anchor ice clusters causes an increase in drag (Kerr *et al.*, 2002). At grain scale, in porous substrate rivers, anchor ice fills the voids at the substrate surface cutting off longitudinal and vertical exchanges between channel and hyporheic zone. This blocking of advected flow has tragic impacts on spawning activities, limiting oxygen supply and exposing eggs and embryos to freezing temperature (Calkins, 1993). A feedback mechanism might exist between anchor ice and hyporeic flows where hyporheic flow paths influence the distribution of anchor ice on the streambed (Kempema and Konrad, 2004).

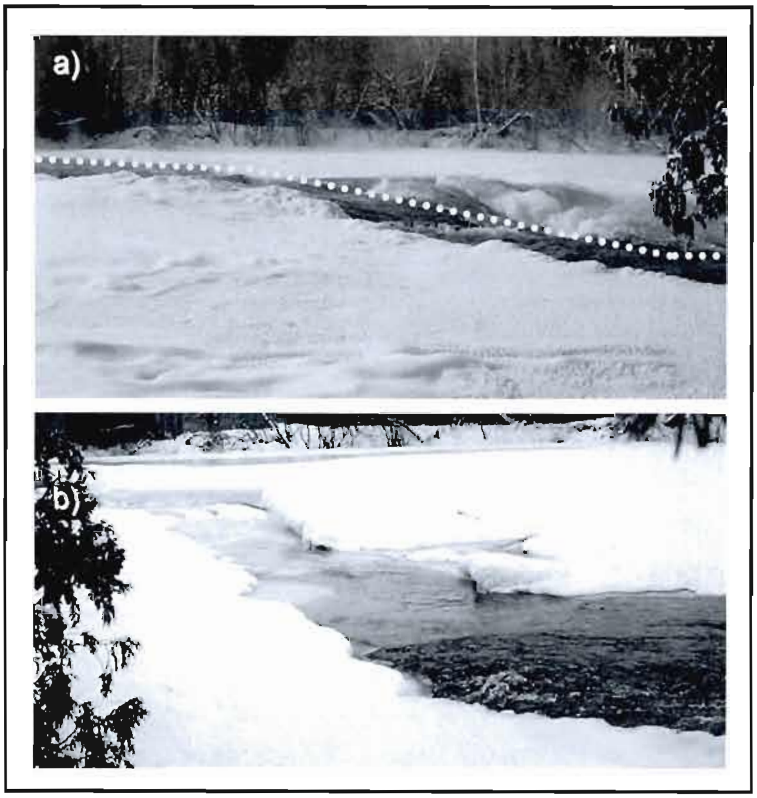


Figure 2.7 Anchor ice riffle formation at straight reach PK33.6 from estuary, Mitis River, Québec. a) Anchor ice riffle formation 01/25/2008 b) Anchor ice riffle at lower discharge 02/02/2008.

The understanding of the relation of ice dynamics to bedforms is evident and unidirectional in the case of the physical abrasion of moving ice on the riverbed and banks. Yet, it becomes more complex at a planform scale. Although it varies greatly with time and scale when it comes to the relation of anchor ice to bedforms, the relationship remains in all cases well illustrated in the model. The ways in which channel configuration and riverbed gradient exert a control on ice dynamics are also well illustrated by the model.

Channel configuration and ice jams

Ice jams are likely to form in a large diversity of channel configurations: in tight meander loops, in subchannels, around bars or islands, at channel constrictions, in pools, at lake outlets or at river confluences. Ice jams occur at locations were the drifting of ice is limited. Their formation can be triggered by the localized growth of iceforms (ice bridging), by a sudden change in bedforms (change in slope gradient) or by human infrastructures (bridge piers, dikes). The pattern of flow convergence found in meanders makes this type of channel form very sensitive to ice jamming. In laboratory experiments the jam thickness distribution behaved similarly to the bed topography. The jam being thicker towards the inner bank, it becomes more susceptible to potential grounding and associated flooding. From several years of observations, Gay *et al.* (1998) documented frequent meander loop cut-off as a result of ice-jam formation. The formation of the jam resulted in upstream water levels rises and the flow was diverted across the neck of the meander loop. Ice jams were found to be more susceptible to form in meanders with the tightest curvature. The authors observed that successive ice jams, separated by years or decades, and could be necessary to complete the breach. In this regard, the net effect of ice jamming is to reduce channel sinuosity, increase the channel gradient and increase the local transporting ability with associated consequences upstream and downstream of the cutoff reach.

Riverbed gradient on frazil ice

It is intuitive to attribute the appearance of frazil ice to steep fluvial morphology. It is also intuitive to associate the formation of static ice and the undercover accumulation of ice to low gradient reaches.

Pools are notable morphological features found in most channels, varying in shape, size, and causative factors. They are typically found in meander bends, in riffle-pool sequences and at confluences. At a smaller scale, pools are found downstream from large roughness elements and in step-pool steep mountain streams. In winter conditions, pools are usually covered by the lateral extension of static ice, the dynamic accumulation of incoming ice, or a combination of both. Any covered reach is subject to the undercover transport and accumulation of ice granules and floes. Deep pools are ideal sites for undercover ice accumulation because of their gentle water slope gradient; low level of near-bed turbulence activity due to greater depth and; reduced flow velocities and bed shear stresses caused by decreasing winter discharge and flow dampening by ice. The cold research community has documented impressive ice accumulations, such as hanging dams, with little analysis of geomorphologic evolution. Ice dynamics was identified as a potential key morphological process in the formation of « anomalous deep-pool » (Gharabaghi *et al.* 2007). To this day, research has found it difficult to impute the maintenance and evolution of such deep pools to ice dynamics (Allard *et al.*, 2009b).

Frazil crystals are known to form in turbulent open-water reaches with fast flows, steep water surface slope and coarse bed material. The limitations of bedforms and anthropomorphologic features in their relation to crystallization have yet to be examined. The intrinsic relation between bedforms and iceforms is often forgotten. The proposed model fulfills its purpose by illustrating the bonds between the components of both trilogies in this complex relationship.

2.4 Discussion

The proposed model attempts to integrate both fluvial and ice dynamics within a unified conceptual structure of components and bonds. The model is strongly influenced by Leeder's conceptual model of fluvial dynamics. Dependant components are presented into boxes and interrelationships and feedbacks in arrows. Dotted arrows indicate hypothetical bonds necessitating further research. Independent variables such as hydroclimatic parameters are left outside the main frame.

Hydroclimatic parameters (precipitation, air and water temperature, wind) are external variables with unilateral relations to the bonds and components of the model. It is understood that frazil crystallization is closely related to heat loss, that thermal breakup is induced by sunlight absorption and that mechanical break-up often results from precipitations, snowmelt event or from dam operation (Shen, 2003). External variables are translated in this schematic representation by the river's physical response (increase in discharge). This model also leaves aside biological, thermal and chemical interactions with physical components, like changes in sediment heat exchange rate due to the presence of anchor ice. It must not be forgotten that such indirect variable action (heat exchange) can enhance or reduce the strength of certain bonds.

The models strengths and weaknesses are cited hereafter. Inspired by Leeder's trinity, the model illustrates simply the complex interactions between river ice and fluvial dynamics. Conceptual modeling generates a simplified view of reality. It is an essential and inseparable part of all scientific activity, constructing insightful and useful formal system and uncovering aspects that may have escaped the notice of others. This model presents cold river physical dynamics as an 'integrated whole' embodying a set of relationships. It enhances our ability to understand the underlying dynamics of a complex system of entities and processes.

The model provides a framework for discussing and establishing research needs and prospects in the study of rivers in cold regions. By drawing attention to several key processes, models are a necessary visualization tools for pedagogic use. This cold river dynamic model is intended to find out about the causal relations that hold between certain components and processes. It is these relations that demonstrate theoretical claims about cold river internal mechanism.

Like Leeder's conceptual model of fluvial dynamics, the proposed model is relatively easy to remember. However, the non-consistent use of double-arrows and the absence of certain bonds between components might induce errors of interpretation for reproduction. Certain bonds, like the relation of frazil to turbulence are intrinsic to other components, in occurrence to bedform. Such complex relations take a little more understanding of the model capability.

This model can be used to comprehend multiple scales of interactions, spatially as well as temporally. Most ice events are studied at a specific scale range, from the grain scale to the large planform scale. The temporality of ice and fluvial dynamics might be difficult to visualize with this type of conceptual model. The illustrated dynamics are mostly of causal-effect nature occurring over a certain timescale, from the short duration of the secondary nucleation to the interannual change scale.

The influence of permafrost was not treated in this paper. Permafrost can exert an important control on fluvial forms: scour process are retarded by frozen bed material, cap ice in deeper channels alter stream hydraulics, bottom ice suppresses bedload transport, thermoerosional processes lead to enhanced bank undercutting, permafrost prevent geomorphic responses to summer floods (McNamara, 1999; 2009). Other components, not included in the proposed model, might better explain permafrost river dynamics.

As the climate changes, we can expect changes to the river-ice regime leading to alteration of ice thickness and duration, or to severity and timing of ice jams and floods. Such changes could potentially lead to dynamic adjustment of fluvial geomorphology and consequent changes in hydrologic response and sediment delivery. To date, future predictions of changes in river-ice hydrology are rare and limited to the extrapolations of timing of events from hydrologic and atmospheric data (Beltaos and Prowse, 2009). Unfortunately morphological responses to ice dynamics are scarcely taken into account and remain to be successfully included into river-ice modeling. Changes caused by a new climate will undeniably have repercussion on each of the six presented components in retroactive ways. Methods need to be developed to evaluate the extent of a river's physical response to climate change under cold climates.

2.5 Conclusion

Cold rivers are subject to parameters interacting and mutually influencing in feedback loops. A simple change in parameters, such as ice conditions, affect the state of the fluvial dynamics and result in changes in fluvial physical environment. Conceptual modeling appeals to a universals approach of fluvial environments. The few models applicable to cold rivers relate to space-time evolution of iceforms and processes. No model presently assesses the scope of interrelation between ice dynamics and morphological processes. The specificity of cold river systems must be regarded as a whole. From the point of view of management of environmental problems, it is paramount to obtain analysis tools making it possible to simplify river complexity. The object of this paper is to presents a conceptual model specific to ice affected rivers on local and regional scales. This cold river dynamics model is highly inspired by the trinity of Leeder. The components and interactions suggested are illustrated by past and ongoing studies. This model will help avoid potential future impact assessments. It is intended for those interested in understanding the underlying dynamics of this complex system of entities and processes. Components are well understood but causal relations that hold between certain components require detailed experimental work. Concentrate efforts to realize more complete and realistic models of interactions will lead to rapid and fundamental advances from detailed and comprehensive experiment.

CHAPITRE III

FLUVIAL AND ICE DYNAMICS AT A FRAZIL-POOL

DYNAMIQUE FLUVIO-GLACIELLE D'UNE FOSSE-À-FRASIL

Abstract

This paper addresses the bidirectional linkages between frazil ice dynamics and alluvial bedforms. Frazil forms in open water area of high turbulent intensity and accumulates downstream along flatter river segments. Frazil accumulation is largely dependent on the overall frazil production which is governed by hydroclimatic and morphological parameters upstream from the frazil sinks. Past field investigations have described strong relationships between riverbed scour and undercover frazil accumulation. However, only few studies have documented in continuous the bed deformation in relation to undercover frazil accumulation over an entire winter season.

Here, we present the chronology of an undercover frazil accumulation event at a pool section of the Mitis River (Québec, Canada) over the 2007-2008 ice period. The characteristics of upstream frazil ice production, ice cover growth, undercover frazil ice accumulation behavior, water level variation, hydroclimatic conditions and riverbed deformations are described. Innovative *dynamic bed-rods* integrating three-axis accelerometer pendant loggers were deployed at the bed to obtain dynamic measurements of erosional or depositional activity over the pool section. Bedform scour occurred during the formation and evolution of the ice cover over the frazil-pool. Ground penetrating radar and manual sounding investigations of the ice cover were conducted on a monthly basis. Analysis reveals the irregular, but stable, morphology of the frazil ice layer. Thicknesses ranged from 0 to 6.25 meters and frazil ice occupied up to 70% of a cross-sectional area. Collected data is used to evaluate riverbed deformation in relation to iceforms.

Résumé long

Ce chapitre s'intéresse aux liens qui unissent le frasil et la géomorphologie fluviale. En rivière, le frasil cristallise dans un écoulement turbulent en surfusion et il s'accumule dans des environnements préférentiels à faible vélocité. L'approvisionnement de ces lieux d'accumulations repose sur la capacité de production de frasil, qui est gouvernée par la géomorphologie fluviale et par les conditions hydroclimatiques de l'amont. L'appellation, «fosse-à-frasil» utilisée dans ce chapitre fait référence à une fosse alluviale anormalement profonde dans laquelle il s'accumule du frasil sous-glaciel la majorité des ans. Plusieurs études ont montré l'existence de relations étroites entre l'accumulation suspendue de frasil et un affouillement localisé du lit. Cependant, peu d'études ont documenté en continu l'évolution spatio-temporelle des formes du lit et des formes glacielles durant l'hiver.

Ce chapitre est composé d'un article de recherche qui sera soumis à la revue scientifique *River Research and Application*. Il présente la chronologie des ajustements morphologiques d'une fosse-à-frasil selon l'accumulation de frasil sous le couvert de glace au cours de l'hiver 2007-2008. La fosse concernée par cette étude est anormalement large et profonde en par rapport aux autres fosses du système alluvial. D'une largeur maximale de 75 mètres et d'une profondeur maximale de 8.2 m (niveau plein bord), elle est située à 14.1 km de l'embouchure de la rivière Mitis, Bas-Saint-Laurent.

Ce chapitre-article vise dans un premier temps, à comprendre les processus de production et de transport de glace qui mènent à l'accumulation de frasil dans la fosse d'étude. Un suivi des conditions glacielles en eaux libre, des variations de niveaux d'eau, des conditions hydroclimatiques et une caractérisation géomorphologique d'un corridor fluvial de 24 km permettent de discriminer l'influence des variables climatiques, hydrauliques et géomorphologiques. Dans un deuxième temps, une matrice de bornes sédimentaires dynamiques, composées d'accéléromètres tridimensionnels a été déployée dans la fosse afin de documenter l'intensité et la spatialité de la réponse morphologique de la fosse à des événements spécifiques de natures hydrologiques ou glacielles. Dans un troisième temps, des relevés géophysiques et des sondages manuels servent à examiner l'évolution morphologique des formes glacielles et des formes du lit sur une base mensuelle.

Les principaux résultats à l'échelle du corridor fluvial indiquent que l'initiation du transport glaciel 2007-2008 est associée à une température de l'eau de 0.8 °C, à un abaissement de la température atmosphérique et à d'importantes précipitations neigeuses. La première observation de transport de frasil, qui date du 23 novembre, est associée aux eaux froides de sont tributaire, la rivière Mistigougèche. Ce n'est que le 7 décembre, à 124 °C DJGC, que la rivière Mitis se refroidit suffisamment pour produire du frasil à l'amont de son tributaire principal. Ce même jour, la production combinée de frasil par la rivière et son tributaire favorise la formation d'un couvert de glace à l'aval du corridor fluvial. Le couvert de glace progresse alors sur une distance de 5 km en un seul jour et couvre totalement la fosse d'étude.

La progression du couvert de glace est contrôlée entre autres par le gradient de pente du chenal. Le couvert de glace cesse de progresser vers l'amont alors que la pente du chenal

atteint le fort gradient hydraulique de 5.09%. Un tronçon de 5.8 km, historiquement sujet à embâcle de frasil, demeure exempt de glace tout l'hiver. Ce tronçon a la capacité de produire d'importants volumes de frasil lorsqu'exposé à de basses températures.

À l'échelle de la fosse-à-frasil, les résultats montrent la formation d'un embâcle de frasil quatre (4) jours suivant la mise en place du couvert de glace. Le transit de frasil vers l'aval y est interrompu durant les trois (3) jours suivant la constitution de l'embâcle malgré un apport continu de frasil de l'amont. L'analyse des levés géophysiques et manuels menés au cours de l'hiver dévoile une géométrie hydraulique sous-glacielle modelée par d'épaisses accumulations de frasil aux formes irrégulières. En opposition au couvert de glace qui présente une morphologie uniforme d'une épaisseur maximale de un mètre, l'accumulation de frasil occupe jusqu'à 70 % de la section transversale et atteint une épaisseur de 6.25 m. Malgré un approvisionnement constant de frasil (de décembre à mars), l'accumulation est demeurée morphologiquement stable au cours de l'hiver.

L'utilisation des bornes sédimentaires dynamiques (DBR), révèle une relation claire entre la déformation du lit et la progression du couvert de glace sur la fosse-à-frasil. Cependant, les déformations enregistrées sont de courtes durées et aucune n'a pu être associée à l'accumulation suspendue de frasil. Les DBR, sont des instruments autonomes qui ont montré leur efficacité à détecter les déformations du lit. Connaissant l'étendue de l'accumulation de frasil, il est désormais possible d'utiliser ces instruments pour documenter en continu l'accumulation de frasil dans la fosse. Le relevé géophysique de l'interface frasil-glace a été réalisé avec succès. Bien que les résultats présentent une résolution verticale comparable aux sondages manuels, les nombreux biais méthodologiques nuisent à l'interprétation quantitative des épaisseurs sondées.

Il est essentiel à une gestion éclairée des corridors fluviaux, de tenir compte de la dynamique fluviale hivernale des rivières exposées à des climats froids. Le frasil est une composante essentielle de la dynamique glacielle et sa sensibilité aux conditions climatiques en fait un sujet prioritaire. Ce projet de recherche documente l'effet morphologique d'un remplissage imposant et durable de l'aire du chenal par le frasil. Dans les cours d'eau de taille moyenne, l'accumulation suspendue de frasil doit être considérée comme une composante morphologique intrinsèque au système plutôt qu'une simple accumulation ponctuelle de glace.

3.1 Introduction

Ice processes occur in 60 % of major river basins in the northern hemisphere and frazil ice is one of the key processes contributing to fluvial dynamics. Localized undercover frazil accumulations, or hanging dams, result from the accumulation of advecting frazil ice forms in river reach with low flow velocities or flow recirculating patterns, such as meander bends, deep pool and lake outlets. Hanging dams have been investigated for their distribution and thicknesses because they produce water level fluctuations and localized scouring. Michel and Drouin (1975) documented a 16 km long hanging dam on LaGrande River (Quebec) comprising some 60 million cubic meters of frazil slush and ice pans. Gold and Williams (1963) reported a hanging dam thickness of 90 m on the Ottawa River. In 1975, a recurrent frazil hanging dam in the Smokey River (Alberta) grew to an approximate 20 m depth (Beltaos and Dean, 1981). In Finland, a 4.6 m deep frazil jam filled 50% of the Taivalkoski River cross-sectional area (Kuuskoski, 1972). In the Tanana River (Alaska), Lawson *et al.* (1986) measured a 60% obstruction of sub-ice channels by longitudinal frazil bars.

Although most documented hanging dams were clearly associated to deep scouring morphologies, a stable core of research interest focuses on crystallization processes, iceform hydraulics and ice transport mechanism. Often invisible from the surface and mostly difficult to measure, the complicated interactions between frazil ice accumulations and bedforms are rarely addressed. Few studies have documented riverbed alteration in relation to ongoing undercover frazil accumulation. Among these, extensive studies in the Yellow river showed how frazil jam thickness and riverbed deformation reinforce each other through ice-induced hydrodynamic conditions (Sui *et al.*, 2000). In 2006, a R^2 correlation of 0.70 was established between bed scour area and ice accumulation area at a cross-section (Sui *et al.*, 2006).

However, a clear understanding of the temporality and spatiality of bed scouring beneath recurrent frazil accumulations has yet to emerge. In the last few years, key research needs have been identified that could develop our understanding: to investigate the interactions between bedload transport and ice cover load transport (Shen and Wang, 1995); to delineate the physical processes of freeze-up ice runs and transitions between different freeze-up regimes (Shen, 2003); to investigate an entire reach encompassing a suspended accumulation of frazil to develop more sophisticated relationships with bed deformation (Sui *et al.*, 2006). Furthermore, most documented frazil accumulations are located in large rivers of fine substratum (bed composition is often unnoted) leaving a large spectrum of fluvial environments unknown. Figure 3.1 illustrates the interactions between fluvial and river ice dynamics according to the conceptual model from Allard *et al.* (in review). Dependant components are presented in boxes and interrelationships and feedbacks with arrows. The black arrow enhances the interrelations between frazil ice and bedforms in a complex system of entities and processes.

From a bedform perspective, the existence of « anomalous deep pools » in an alluvial system is intriguing. Several anomalous deep pools or scour holes are known to exist in meandering channel sections, but not many have been documented (Lapointe, 1984; Gharabaghi *et al.*, 2007). Such deep meander pools can present a scour-related morphology hardly attributable to open flow conditions. Facing the uncertainty of the origin of a 30 m deep scour-hole in the Mackenzie delta, it has been noted that ice-covered conditions could be a potential key morphological process (Gharabaghi *et al.*, 2007). In theoretically-based ice-covered flow scenarios, the bed shear stress decreased as the flow rate increased due to adverse pressure gradient in the pool (Inkratas *et al.*, 2009). So far, it remains unknown if such deep pool maintenance and evolution is imputable to ice dynamics.

This study aims to document in continuous the morphological adjustments at the bed of a large « frazil-pool » in relation to frazil accumulation over a winter period. Here, we use the appellation « frazil-pool » to define a river pool in which an accumulation of frazil ice is known to occur most years. To provide a complete understanding of the pool's geomorphic relation to frazil accumulation, this study deployed innovative field work methodologies allowing 1) to describe the processes leading to the infilling of the frazil-pool; 2) to map the evolution of the ice cover, the undercover frazil accumulation and the bed morphology on a monthly basis; and 3) to document in continuous the morphological changes occurring at the bed over the winter period. The results not only provide information on the channel response to frazil disturbance but it also provides insights on innovative undercover instrumentation and investigative methods.

3.2 Study location

The Mitis River flows from the Appalachian plateau to the Saint-Lawrence maritime estuary, draining a 1805 km^2 basin with an average elevation of 342 m (fig. 3.2). The river has a length of 51 km from headwater reservoir Lac Mitis to Mitis Bay with two significant tributaries, the Mistigougèche River (82.3 km) and the Neigette River (117.1 km). The flow regime of the Mitis River is nivo-pluvial with the highest discharges occurring in mid-May. Mean annual discharge of 33.3 m^3/s (1921-1984) is submitted to upstream control from the retaining structures Mitis and Mistigougèche. Run-of-river generating stations, Mitis-1 and Mitis-2 are implanted respectively at 1.8 and 2.6 km from the estuary. The backflow of water from Mitis-2 dam is estimated to be 2 km long.

The studied river corridor can be broken into two different sections: a riffle-pool dominated reach and a well developed meander reach. The riffle-pool section is 13 km long with an average width of 38 m. It is relatively straight (sinuosity of 1.1) and possesses a cobble-gravel bed substrate with poorly developed riffles-pools. Pools are more developed (2-3 m deep at low flow) when lithologic constrictions are present. The meander section is 11 km long with an average width of 40 m and flows in a wide alluvial valley cut into postglacial deposits. It presents an irregular meandering planform (sinuosity 2.4) and encompasses 5 deep pools (> 4 m deep at low flow). The channel gradient progressively decreases from 4.27 m/km at the upstream end of the river corridor to 2.26 m/km at the valley entrance to 0.67 m/km downstream from Ste-Angèle-de-Mérici. Evidences of freeze-up and break-up ice jams have been recorded from 1987 to 2004. Recent frazil jams occurred mostly in the high gradient riffle-pool stretch.

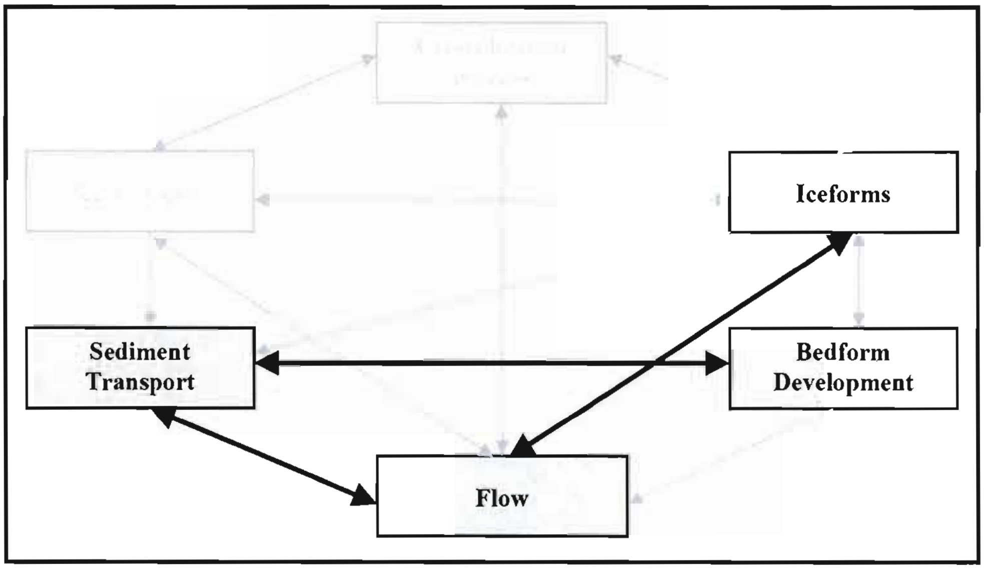


Figure 3.1 A conceptual model for the complex interrelations between frazil ice and bedforms from Allard (2009a). The black arrows show the interrelations and feedbacks between the frasil Iceform component and Bedform component through the Flow and Sediment transport dependant components.

The selected frazil-pool is located 14.1 km upstream from the estuary (fig. 3.2). The pool is located downstream from a short cobble riffle and widens in a sharp 160° meander. Five (5) morphological units define de pool. The first unit, the bed entrance slope, is mainly composed of cobbles. The second unit, the pool-center, lies towards the inner bank and presents a maximum bankfull flow depth of 8.2 meters. The main channel bed consists of cobble and gravel throughout, but sand is present within the secondary flow separation zone. The third unit is a 75 m long clay-cliff constriction. It runs parallel to the pool-center and reaches a height of ±3.5 meters from channel bed. The cliff is cut into Goldthwait Sea blue-gray marine clay underling Mitis terrace's silty intertidal deposits. The fourth unit is the outer bank recirculation plateau. It is composed of sand-gravel alluvial deposits overlying Mitis terrace's silt deposits. The fifth unit is a wide pool exit-slope with strong diverging flows.

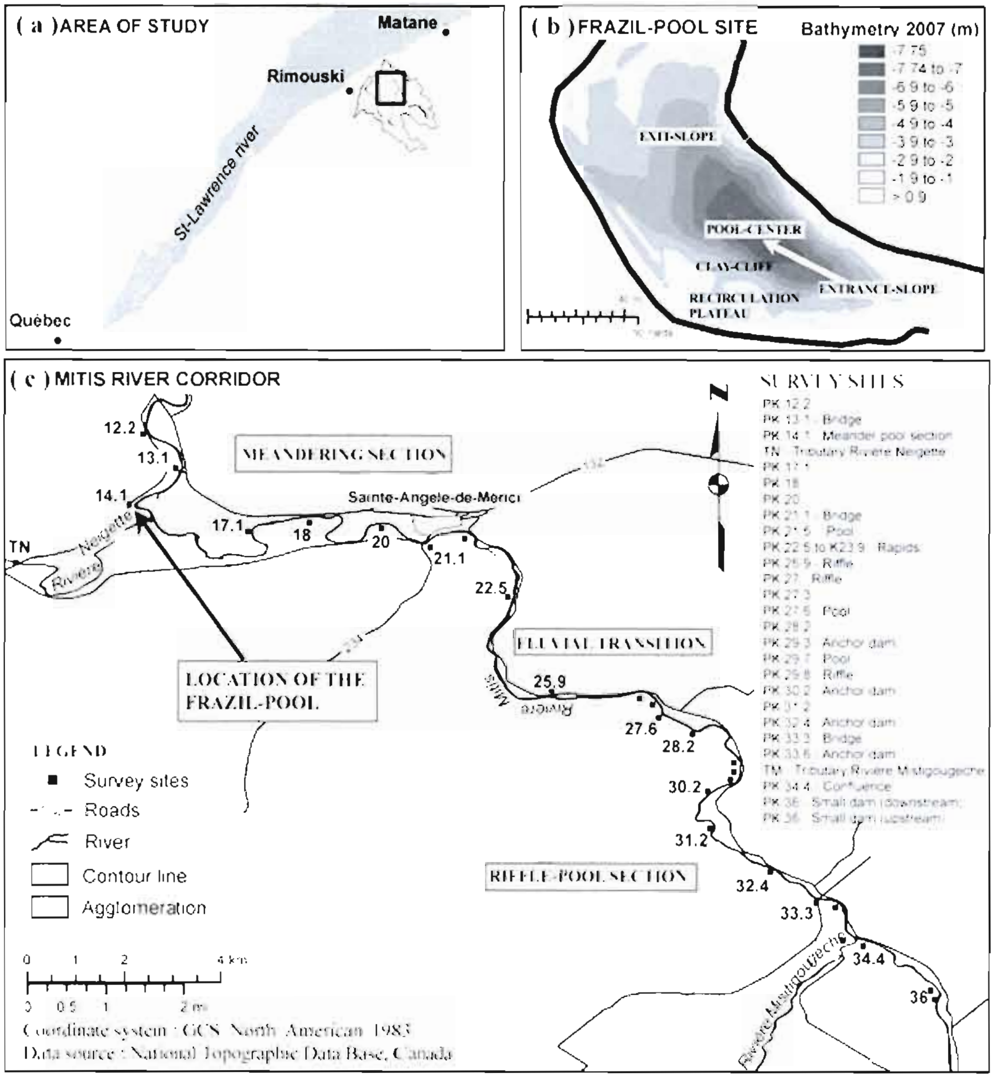


Figure 3.2 (a) Map showing Mitis River basin, adjacent basins and location of Mitis River survey corridor. (b) Illustration of the frazil-pool study site. Arrows indicates flow direction. (c) Physiographic map showing freeze-up survey sites along the survey corridor. The arrow indicates the location of the pool study site.

3.3 Methods

Methods specifications are detailed in table 3.1 and most significant elements are described hereafter.

3.3.1 Freeze-up survey

From November to December, freeze-up data was collected along a 24 km river corridor between point kilometer (PK) 12 and PK 36 from estuary. Field sites were chosen on the basis of accessibility and geomorphic/anthropomorphic representativeness. Photos were taken on each qualitative field reconnaissance. A time-lapse digital camera monitored ice conditions at the frazil-pool section to identify changes in ice conditions that may have contributed to local changes in the river channel, as well as for monitoring ice transport transitions. From collected observations, frazil ice presence and evolutionary state was classified into four iceform categories:

(1) Frazil fragments: frazil crystals, ice particles, frazil granules and poorly agglomerated frazil;
(2) Ice pieces: hydraulic resistant ice aggregates of <10cm surface diameter (frazil flocs and floes, floating slush, ice clusters and ice cover fragments);
(3) Anchored ice forms;
(4) Partial and full ice cover (static and dynamic).

Air and water temperatures were recorded hourly at the frazil-pool site, in the Neigette tributary, at PK22.5 and PK33.3. At the pool site, water temperature and water pressure loggers served to monitor backflows caused by ice jams (fig. 3.3). Hydro-Québec water discharge data from Mitis-2 dam was used to estimate the discharge conditions at the pool.

The mean discharge profiles of total turbinated and released waters present a coherent profile shape with adjacent basins.

Regional air temperature, atmospheric pressure and precipitations data were extracted from *Environment Canada* climate data archive. The weather of the 2007-2008 winter was slightly cold with mean accumulated freezing degree-days of 1152°C, in contrast to the mean historic record of 1076°C. Warm periods occurred in December (12/22 to 12/30) and January (01/04 to 01/19), and cold periods occurred in January (01/19 to 01/27) and March (03/21 to 03/30). Rainfalls and snowfalls remained below average, except on March 9th and 21st when snowfalls slightly exceeded average.

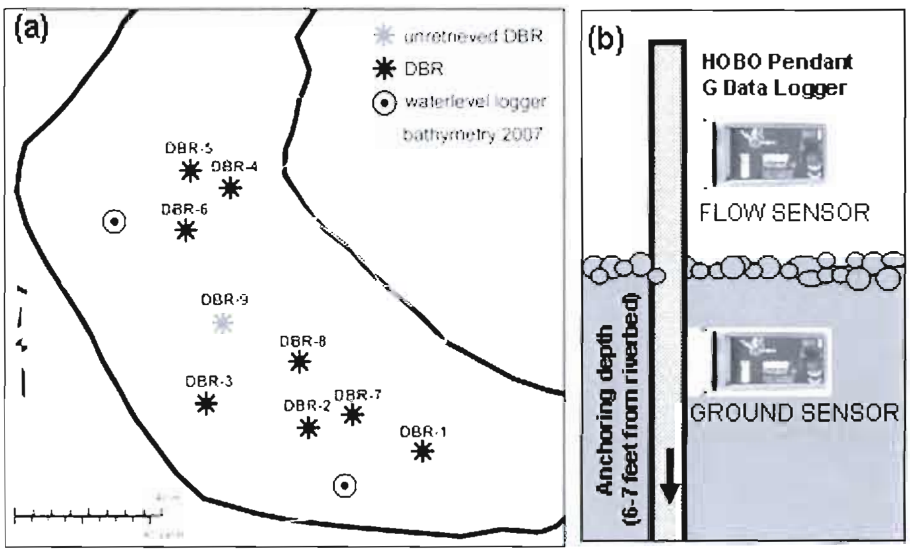


Figure 3.3 (a) Map showing DBR and water level logger deployment at the frazil-pool site. (b) Dynamic bed-rod arrangement.

A broad morphological characterization served to delineate homogeneous river reaches of the 24 km river corridor. Channel pattern, entrenchment, sinuosity and sedimentary features provided sufficient information to describe the range of river geomorphic conditions. The stream reaches were divided into water slope categories that reflect profile morphology. This river-length approach was designed to correlate the frazil-ice regime of the river with the broad plan-view morphology.

3.3.2 Bed monitoring

In order to gain an insight on the geomorphic role of frazil accumulation, we developed an autonomous (unwired) sensor system for continuous monitoring of bed deformation. The *dynamic bed-rods* (DBR) prototype is based on data logged movement detection. The DBR is made of a « scour pin » vertically equipped with three-axis accelerometer G-Pendant logging sensors from Onset Computer Corporation. The DBR is inserted into the riverbed with « ground sensors » buried to expected scour depth and « flow sensors » emerging to anticipated sediment deposition height (fig. 3.3). Only one flow-sensor and one ground-sensor per DBR were used in this study, due to cost limitation. Nine (9) DBR monitored the frazil-pool site from September 2007 to July 2008. DBR #9 was not retrieved because it was buried under a pile of woody debris. The DBR height above bed was noted on installation and removal as another measure of erosion or deposition.

Acceleration data acquisition was recorded at a 70-min sampling frequency. Acceleration data was converted into a binary time series with on-off values as a response or not to flow. To do so, a 5-lag moving window was used on the real acceleration values to smooth out short-term fluctuations and highlight longer-term trends. The RMS time series was converted into a motion-no motion time series by applying a 0.3 threshold. This was done for both X and Y acceleration axis and resulted in a continuous motion (on), alternate motion (on-off) and no-motion (off) time series. Upon installation, the flow sensors could either respond or not to the surrounding flow velocity and turbulence. A motion response from the buried sensors could only be the result of scouring. The scour/deposition thickness accuracy is estimated within 5 cm of sediment depth.

Bathymetric surveys from a theodolite were conducted at low summer flow using random pole-soundings while navigating up and down the pool in a canoe. Bathymetric maps were created using natural neighbor interpolation. Mean Euclidian distance between all pair of points equal to 13 meters.

3.3.3 Ice monitoring

Ice and frazil layer profiles and cross-sectional bed profile of the channel were documented using two methods: (1) direct drilling method and (2) indirect ground penetrating radar (GPR) investigations. The purpose of using both methods served to establish the feasibility of ground probing radar analysis as a tool to detect the frazil ice layer in an undercover environment.

Field survey

We collected GPR data at the pool on January 29th, after 10 weeks of frazil production. The ice had attained sufficient thickness for ice field work and discharge was low (18 m^3/s). Working with georadar GSSI *SIR-3000*, we used two different high frequency antenna arrangements: the 400 MHz antenna, to detect frazil-water and water-bed interfaces, and the 900 MHz antenna to detect the snow-ice layer thickness. Several GPR profiles were acquired following seven (7) cross-sections (fig. 3.4). Each transect was covered on foot with one person carrying the control unit and another towing the antenna across the ice-cover on a sled. The XY coordinates of each transect were positioned using an electronic theodolite.

Figure 3.4 (a) Outer bank view of frazil pool site on March 19th 2007. (b) Frazil-pool site from aerial photograph Q01805-162 (Ministère des ressources naturelles et de la faune). The arrows show the direction of the flow and the lines the approximate location of the cross-sections.

Drill-hole soundings were made along each cross-section on February 29th and March 11th. Mean discharge was 15 m³/s on both occasions. Thickness measurements were obtained in drilled holes using a rigid graduated J-rod. The J-rod was lowered through the hole to the bottom of the ice layer or until a frazil-water density change was felt by the operator. Drill-hole soundings revealed that a highly variable frazil density stratification between drill-holes. Bed profile measurements have an accuracy error of ±0.5 m vertically and horizontally due to flow strength.

Data analysis

Temporal and spatial filtering are basic processing steps of GPR signal processing. For the visualization of the ice/frazil frazil/water and water/sediment interfaces, a spatial high-pass filter applied to the wavenumber domain. It removed the low frequency waves often generated by system noise. A temporal low-pass filter applied down the individual traces in time removed the horizontal banding noise, displaying mostly horizontal reflections of short length. A continuous wavelet (Symlet) transformation served to remove the effect of a composite reflection trace. Figure 3.5a shows an example of a 400 Mhz original dataset after basic filter processing. The band corresponding to the air/ice interface has been removed by adjusting the signal position to the place in time where the radar pulse leaves the antenna and enters the subsurface. The top band represents the radar reflection from the ice/frazil; the following band, the frazil/water interface and the lower band, the water/bed interface. Figure 3.5b shows an example of 2D spatial visualization of the cross-section based on filtering procedures. Image processing includes using a depth-conversion algorithm that locates the top and bottom interface reflections and calculates the layer thickness using known velocities.

Drill-hole measurements of thicknesses provided ground truth verification of the radar data and velocities were adjusted consequently. We used a signal velocity of 0.17 m*ms^{-1} for the snow-ice layer, and of 0.033 m*ms^{-1} for fresh water. Frazil velocity measurements ranged from 0.053 to 0.065 m*ms^{-1}, we used a velocity of 0.055 m*ms^{-1}. The 900 MHz antennae had a maximum penetration depth of ±2 m in ice-frazil-water medium, whereas the 400 MHz antennae had a penetration depth of ±4 meters. The signal penetration limitations provided only a good coverage of the recirculation plateau. A bias in vertical measurement could be present in the calculated ice thickness data due to snow coverage, ice surface roughness and ice cover lamination. The ice cover roughness was fairly smooth, slightly elevated toward the banks where drifting snow accumulated to a maximal 0.85 meter.

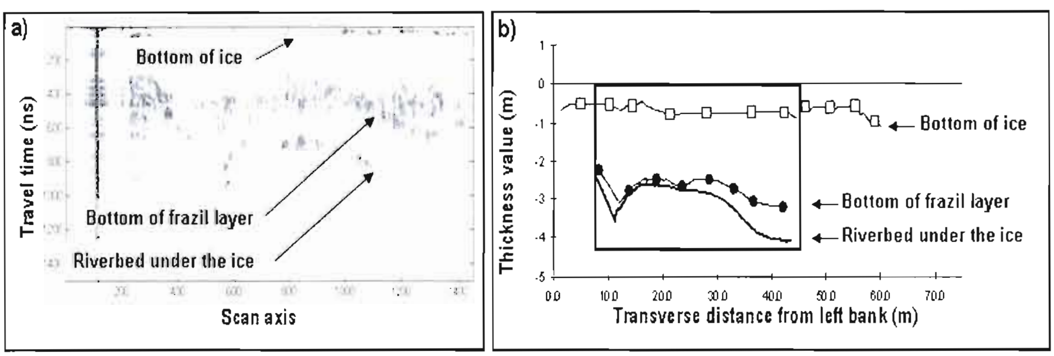


Figure 3.5 (a) a 400 MHz GPR reflexion profile after filtering procedures. (b) Processed GPR profile of cross-section 2.

Table 3.1
Methods specifications

Survey	Specification
Freeze-up survey	
Visual on-site observations	21 field reconnaissance with photos 27 survey sites
Digital camera at pool site	RECONYX Silent ImageTM – Professional, model PM35 (standard) passive infrared (PIR) motion detector; 100ft, 40° angle field view; 256MB picture resolution; 10 minutes time lapse setting.
Geomorphologic characterization	Lalonde, Girouard, Letendre et associé (1991) field reconnaissance report. Missing data was completed with 1:20,000 scale topographic contour lines from Softmap.
Hydroclimatic data	
Air temperature	Digital temperature loggers iButtons® DS1922L/T from DALLAS Semiconductor. Air temperature measurements are made 1.5 meters above ground within 10 meters of river bank. Accuracy is of ±0.5°C (8-bit) from -10 to 65°C (0.9°F over 14F° to 149°F). Chosen logging interval: 20 min.
Water temperature	HOBO® Pro v2 Water Temperature Data Logger from Onset Computer Corporation. Logger is positioned at a depth of ±1m, ±10cm above river bed. In water accuracy is of 0.2°C over 0° to 50°C (0.36°F over 32° to 122°F). Chosen logging interval: 20 min.
Water level	HOBO® 30-Foot Depth Data Logger U20-001-01. Logger is positioned ±10 cm above riverbed. Water level range from 0 to 9 m (0 to 30 ft). Accuracy typical error is ± 0.5 cm (0.015 ft). Temperature measurements with same accuracy and range as Pro v2. Loggers were positioned 132 cm deep upstream and 44 cm deep downstream. Chosen logging interval: 70 min.
Water discharge	Hydro-Québec, Mitis-2 discharge data. Cumulative Turbinated and Released discharges.
Regional climate data	Canada's national climate archives for station Mont-JoliA. Historical meteorological data (1943-2008) of daily average temperature, precipitation, freezing degree days (FDD), and atmospheric pressure. Hourly regional air temperature data.
Ice and bed morphology	
Bathymetric surveying	Random pole-sounding measurements. 1124 sounding in 2007 and 682 soundings in 2008. Mean Euclidian distance is 13 m. Accuracy error is ± 30 cm vertically and horizontally due to flow strength.
Dynamic Bed Rods accelerometers	G-Pendant loggers from Onset Corporation. Measurement range: ± 3 g; 29.4 m/s^2 (96.5 ft/s^2). Accuracy: ± 0.105 g; 1.03 m/s^2 (3.38 ft/s^2) from -20°C to 70°C (-4°F to 158°F). Resolution: 0.025 g; 0.245 m/s^2 (0.8 ft/s^2). Weight: 18 g (0.6 oz) Dimensions: 58 x 33 x 23 mm (2.3 x 1.3 x 0.9 inches). Chosen logging interval: 70 min.
Ground profiling radar	GSSI *SIR-3000 system* from Geophysical Survey Sytems Inc, owned by INRS-ÉTÉ equipped with a 400MHz antenna and a 900 MHz antenna. T_Rate : 100 khz , Mode : Time, GPS : None, Scan samples : 2048, Format : 16 bits, Range : variable, Diel : 3.1, Rate : 32, Scn/unit : 100-200, Gain : auto
GPR analysis	
Basic handling and filtering using Matlab	Adjust signal position, FIR wavenumber filter high pass, FIR frequency filter low pass, Symlet wavelet sym 4
Signal wave concordance with drill-hole measurements	Velocities used : Ice 0.17 $m{*}ms^{-1}$, Fresh water 0.033 $m{*}ms^{-1}$, Frazil 0.055 $m{*}ms^{-1}$

3.4 Results

3.4.1 Freeze-up dynamics

Frazil ice runs were observed at the pool section from November 23rd to January 25th. The beginning of frazil ice production is associated with low water temperature, decreasing air temperature, and above average solid precipitation (Table 3.2). Although water and air temperatures were very much alike on November 11th and 20th, these conditions were only met on November 23rd. Subsequently, ice runs were apparent when water temperatures were of 0.8°C or below.

Figure 3.6 summarizes the freeze-up observations along the river corridor. The first frazil runs were triggered by the Mistigougèche cold water inputs to the warmer Mitis River. Until December 7th, mature ice aggregates (flocs) were observed strictly downstream from the Mistigougèche-Mitis confluence. Throughout the study, the Neigette tributary, smaller tributaries and the 1m concrete dam waterfall at PK 36 showed negligible production capacity. On December 7th, at 124°C of AFDD, the Mitis River's water had sufficiently cooled down to produce frazil ice upstream from the Mistigougèche confluence. This is illustrated by the thermal gradient of water temperature in figure 3.7. Frazil ice generation was only observed during cold periods of high AFDD while other observed iceforms proved to be less sensitive to the weather variability.

Changes in frazil ice presence and evolutionary state relate to river channel morphology. Frazil crystals were observed in rapids and riffle reaches (PK 22.5, 25.9, 29.8) and in less turbulent reaches immediately downstream from riffles (PK 14.1, 21.5, 29.7, 33.3, 34.4, 36). Anchor ice was observed in riffles, on substrate surrounding anthropomorphic entities (bridges and a low concrete dam), and at the Mistigougèche-Mitis confluence. Anchor ice dams first appeared on December 2nd at 88°C AFDD. They formed in uncovered riffles of low water slope and in shallow, fast flowing rectilinear reaches (PK 27, 29.3, 30.2, 32.4, 33.6). Ice aggregates were observed in riffles, pools, meanders and rectilinear stretches and could hardly be related to specific morphologies.

Table 3.2

Summary of hydroclimatic conditions prior to the first frazil ice runs at the frazil-pool site and hydroclimatic conditions when full ice cover formed on the pool.

Daily hydroclimatic conditions (2007)	Nov. 11th	Nov. 20th	Nov. 23rd	Dec. 7th
Water temp. min (°C) (at the pool)	0.8	0.8	0.8	0.8
Air temp. min (°C)	-6.1	-5.9	-8.4	-8.8
Mean air temp. difference with following day (C°)	+4.2	+0.3	-5.4	+1.6
FDD mean (°C)	3.8	3.3	5.3	8.8
ADD mean (°C)	13.2	24.2	33.4	124.2
Solid precipitation (cm)	0	4.4	10.8 (25.6 on 11/22)	0

On December 7th the combined ice inputs from the Mitis and the Mistigougèche rivers were sufficient for the initiation of a permanent ice cover. The ice cover progression upstream relate well with the slope gradient. Downstream from PK17, where the river slope gradient is near zero, a quasi-static ice cover formed over 5 km within a single day. Upstream from PK 17, the migrating speed of the ice cover decreased significantly from 203 m/day (PK 17 to 21.5) to a slow 31 m/day (PK 27.5 to 30). A mix of juxtaposition and frontal progression was prevalent in the 1-2% meandering reach. At the riffle-pool reach (upstream from PK 23), the progression speed slowed as cold periods and uncovered riffle were needed to produce the necessary amount of frazil ice to overcome 2-3% slopes. Past PK 30, the slope gradient attained a high 5.09% and frazil ice was forced to pass underneath the ice cover. This 5.8 km of open water in subzero weather was capable of producing massive quantities of frazil ice. This is corroborated by the presence of three long-lasting ice dams in the uncovered reach and by historical records of freeze-up jams in this high gradient segment. Upstream from the concrete dam (PK 36), a permanent ice cover formed by bridging on December 22nd.

A permanent ice cover had formed from bank to bank at the frazil pool on December 7th. Frazil ice runs remained apparent through a large opening or polyneas, maintained by upwelling flows upon the pool-exit slope. The incoming frazil ice jammed under the ice cover 4 days following full coverage of the pool. Image analysis showed a 3-days interruption in frazil and ice transport at the exit-slope opening from December 11th to 13th (fig. 3.8). The jam was recorded by both water level loggers, as a decrease in water level at the pool. A regional decrease in discharges was also recorded in the Mitis and adjacent basin, possibly as a result of river ice storage. During the jam, Mitis water temperature loggers recorded sustained near 0°C temperatures. Ice transport was visible again on December 14th. Frazil transport at the pool was last seen on December 22nd as the exit-slope opening sutured to full closure. Intense undercover frazil ice transport has however been felt during drill-hole soundings in January, February and March and was also recorded using an underwater video camera on March 11th.

3.4.2 GPR and soundings

Figure 3.9 shows January's GPR analysis results and February-March's drill-hole measurements results at the pool cross-sections (CS). This figure illustrates the widespread distribution of frazil ice in this medium size river. A fairly uniform, 1 meter thick, snow-ice layer covered the pool from January to March. During the same period, a highly asymmetrical frazil layer obstructed the pool to considerable depth. The frazil layer reached a maximum thickness of 6.25 m in March and filled up to 70% of the channel cross-sections. It molded the shallow bed and acted as lateral walls of a main sub-ice water channel. Probing with a rod revealed the general characteristics of the frazil ice layer. The main frazil ice layer enclosed several discontinuous sub-layered structures of various densities occasionally separated by water lenses. Bore hole samples revealed irregular singular ice grains, with rounded edges, of <1 cm diameter. The frazil layer spatial distribution remained fairly stable over the three months period. Two temporal changes are of interest: 1) a migration of the sub-ice channel from mid-channel to left bank over the exit-slope between the January GPR measurements and March; 2) an apparent frazil-fill of the pool center at cross-section 5 between February and March.

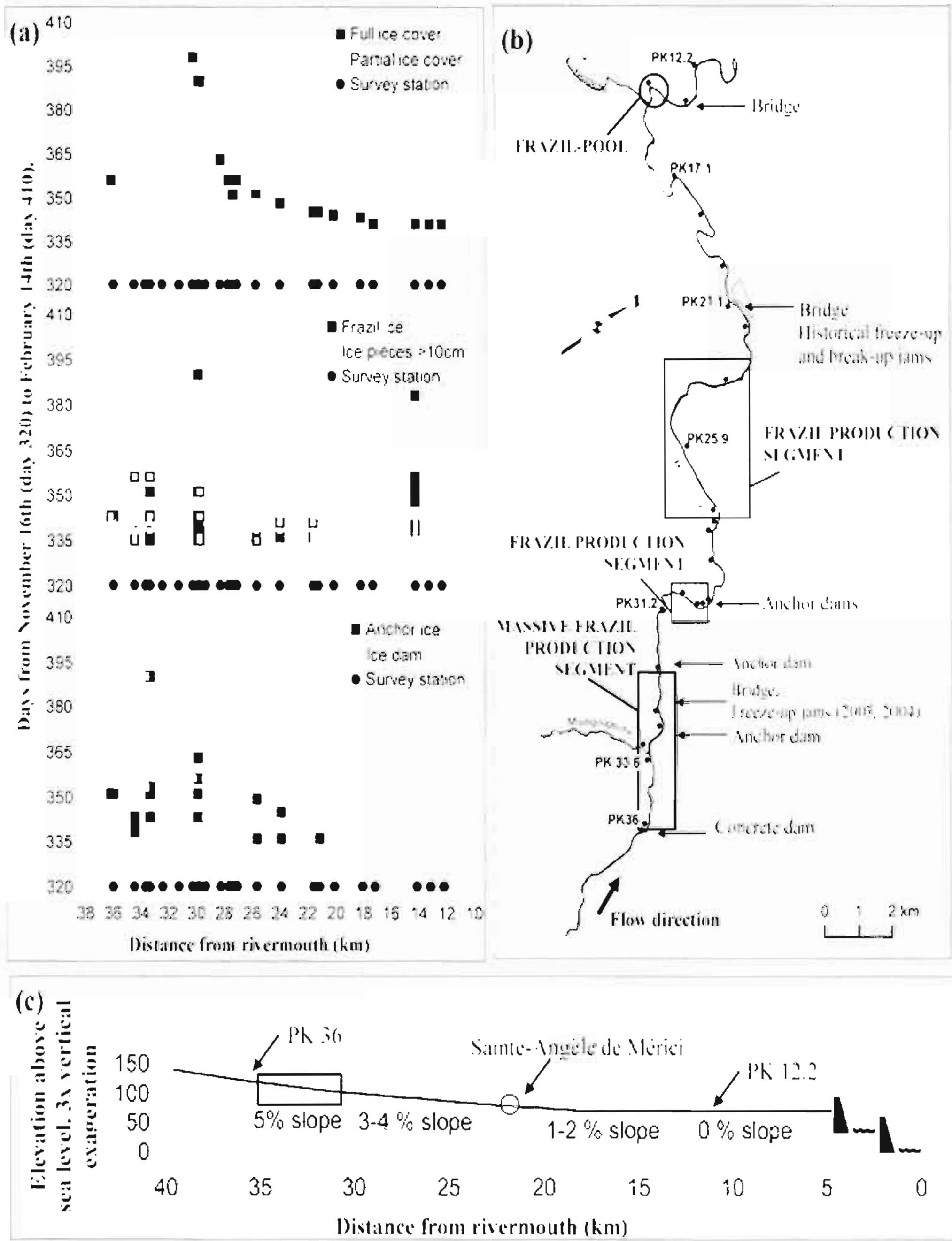


Figure 3.6 (a) Freeze-up survey results for the period Nov. 16th to Feb. 14th. (b) Broad geomorphologic characterization of river corridor and localization of frazil producing river segments. (c) Longitudinal profile and water slope gradient of the river corridor.

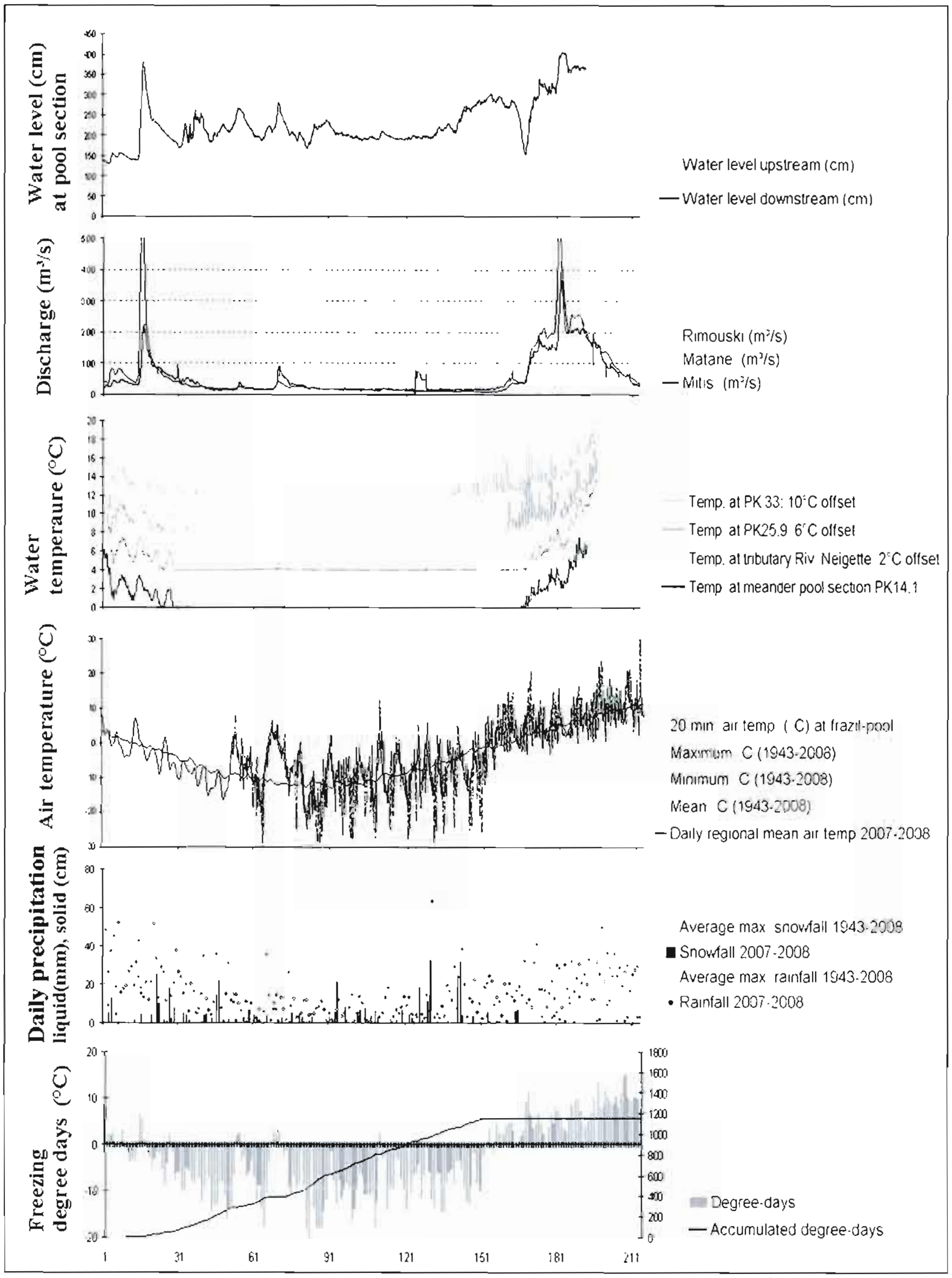


Figure 3.7 Daily hydroclimatic conditions for the period November 1st, 2007 (day 1) to June 1st, 2008 (day 214).

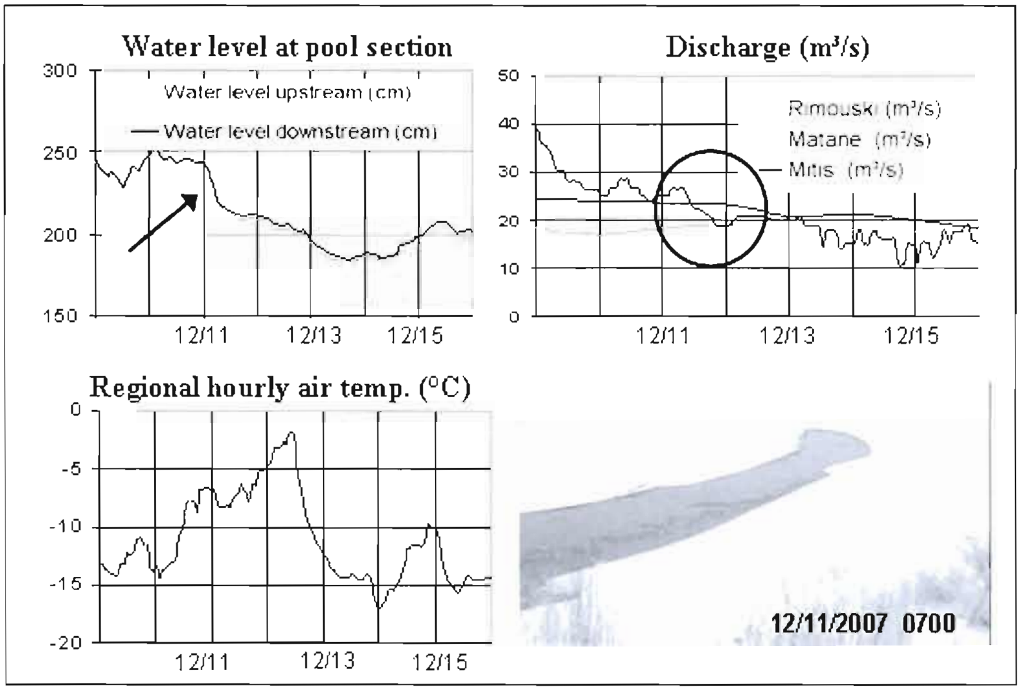


Figure 3.8 Hydroclimatic conditions at the frazil-pool site during frazil jam. Period shown is from December 9^{th} to December 16th. The arrow is pointing at a strong decrease in water level at the pool followed by a decrease in discharge at Mitis-2 dam. Right-bottom photo shows continuous ice runs through downstream opening shortly before ice jamming.

Figure 3.10a illustrates the spatial distribution of March 11^{th} sub-ice flows, beneath the frazil layer. The main sub-ice channel shape and orientation is typical of open-channel flow patterns at meander bends. It moves from the inner bank at the bed entrance to the outer bank at bend exit, crossing the channel in the zone of greatest curvature. The main channel extends to a maximum height of 2.5 meters along the clay-cliff constriction. Common opposite circulation patterns at the outer bank and weaker flows along the inner bank are present as secondary sub-ice channel.

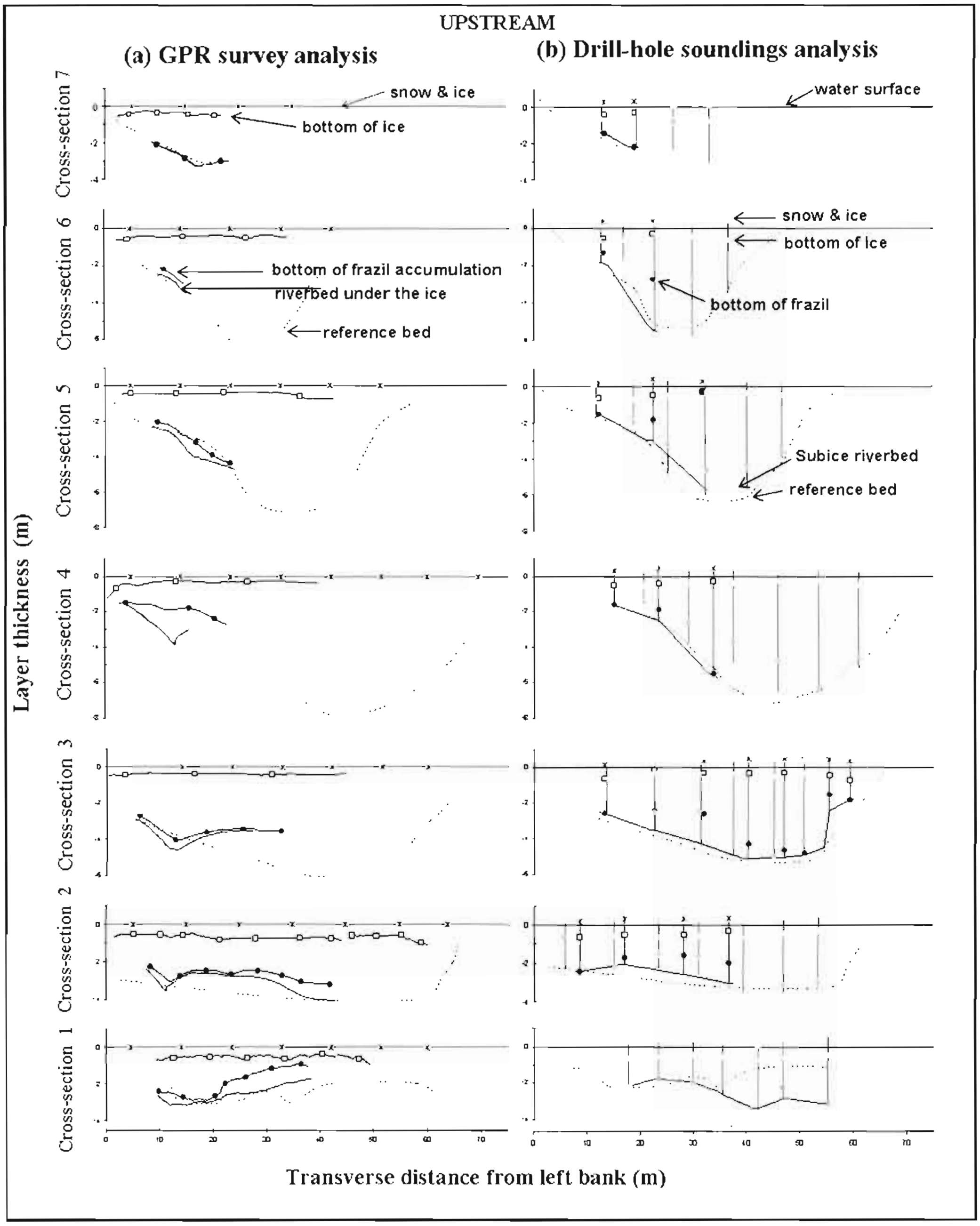


Figure 3.9 (a) GPR survey analysis cross-sections 01/29; (b) drill-hole sounding cross-sections 02/29 (black), 03/11 (gray). Cross-section thickness values should be regarded qualitatively. Symbols legend: ✳ snow-ice cover, □ bottom of ice layer,○ bottom of ice accumulation, - - sub-ice riverbed, — July2007 reference bed.

3.4.3 Dynamic bed-rods

Dynamic bed-rods (DBR) were deployed over the pool section to obtain dynamic measurements of erosional or depositional activity. The DBR were designed to reveal the time of occurrence of riverbed deformation in ice covered conditions. Over the monitoring period, flow-sensors #1, 3, 5 and 7 (respectively located at the entrance slope, the recirculation plateau, the exit slope, and main channel) maintained constant activity throughout the winter period. The clay-cliff DBR flow-sensor #2 responded to flow turbulence only once, on December 7th (fig. 3.10b). These flow-sensor motion responses are interpreted as near-bed free-flowing conditions, unaffected by frazil ice or by sediment deposition. The exit slope flow-sensors #4 and #6, were both responsive to flow upon installation, they also both recorded no-motion activity in ice covered conditions. Flow sensor #4's inactivity began in late November, with the formation of a partial cover while FS #6 became inactive on December 7th. They both became active again following thermal break-up on April 14th. This exit-slope flow-sensors inactivity could be the result of FS burial, FS ice jamming or of low flow conditions.

On December 7th, flow sensors #2 and #6 (located along the main channel) and buried ground sensors #7 and #8 (located in the main channel) simultaneously responded to the progression of the ice cover over the pool. These daily responses indicate strong and turbulent main channel flow conditions, causing short term erosional activity in the main channel. Similar DBR responses were registered following the very strong and rapid discharge increase of November 15th.

3.4.4 Bed deformation

Localized geomorphologic changes were observed in the main channel. The *in situ* measurements of the DBR-rods height above bed, showed the loss of 5 cm of marine clay near the entrance-slope at DBR #1 location. Remnant landslide deposits from the clay-cliff were observed during a scuba dive in 2008 near DBR#2. Unfortunately, DBR#2's sensors did not register the event nor did *in situ* measurements. The 2007 and 2008 bathymetric surveys comparison showed very heterogeneous and localized bedform changes (erosion at the cobble-bed entrance slope, at pool center and towards the inner bank, and deposition on the exit-slope and in pool-center). A higher sampling density would be needed to further analyze these annual changes.

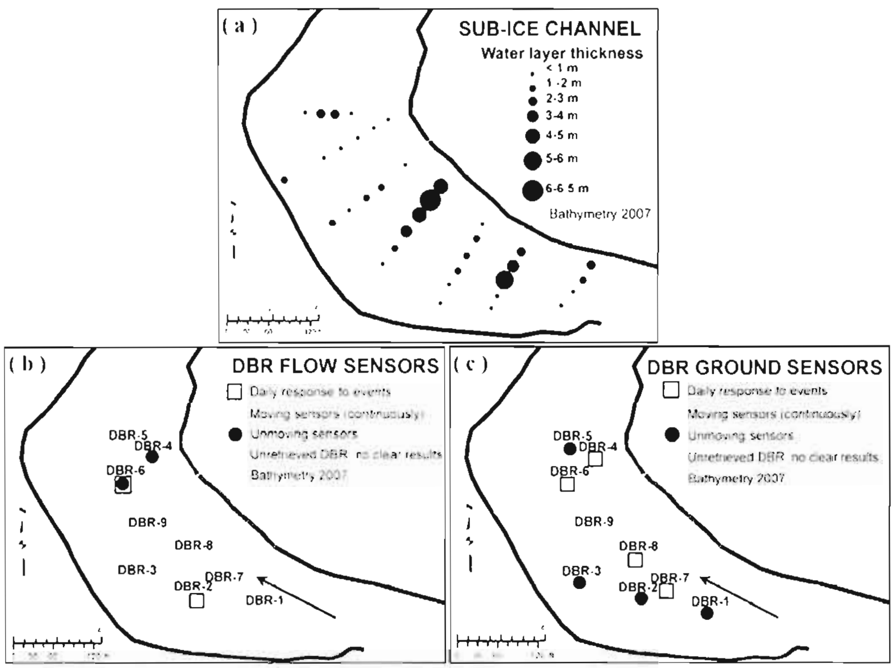


Figure 3.10 Frazil-pool result-maps (a) Sub-ice flow distribution from sounding measurements. (b) DBR flow-sensors responses (c) DBR ground-sensors responses.

3.5 Discussion

3.5.1 Dynamics of frazil ice production and accumulation

To better understand cold river dynamics, it is crucial to consider frazil ice behavior. Knowing where and when frazil ice is produced implies identifying the length of frazil producing river segments, the spatiality of river ice coverage, and continuous record of hydroclimatic conditions within river segments. These allow the estimation of quantities of frazil ice produced per segment and allows the understanding of the variability in production over time. Also essential to our knowledge is to understand how far and under what hydroclimatic conditions frazil ice travels downstream. Documenting the dynamics of frazil sinks in relation to their ice-holding capacity and the diversity of frazil-sink morphology along a river system is necessary to comprehend the consequences of undercover frazil accumulation.

Such river-length approach has been applied in a case study of frazil-ice related flooding in the Moira River (Beltaos *et al.*, 2007). The authors focused on the important role of a lacking ice cover in a high slope reach on the supply of large quantities of frazil ice to the river mouth. A similar seasonal frazil generating reach is also present in the Mitis River. Such clear causality emphasizes the importance of geomorphic analysis (sinuosity, slope and channel pattern) in the assessment of cold river dynamics.

The freeze-up survey revealed the presence of numerous small openings or polyneas in ice covered riffles. The frazil generating potential of river polyneas has not yet been extensively documented.

The infilling of frazil sinks relies on the successful transport of ice to low flow areas. At the pool site, frazil ice runs were apparent only at temperatures of 0.8°C or below and did not correlate with regional atmospheric conditions. This is indicative that sub-ice frazil accumulation predictions need to include water temperature measurements as much as frazil crystallization predictions.

The infilling of the pool site happened within 10 weeks of full cover formation and the frazil layer and ice cover physical attributes have not changed considerably during the following winter months. In accordance with the concept of jam equilibrium, the frazil-pool undercover accumulation had reached equilibrium shortly following the jam's formation, possibly on December 14th when frazil ice runs confirmed the renewed availability of frazil ice for transport to downstream reaches. This implies the future challenge of working in younger and thinner ice cover conditions to investigate the dynamics of frazil accumulation, before a state of equilibrium is reached.

3.5.2 Evolution of ice cover and frazil layer

In 1975, Michel and Drouin wrote the following observation: «The growth of a suspended frazil layer is limited by hydraulic and thermal variables resulting in a unique winter hydraulic geometry». Most frazil jam case studies enclose sub-ice hydraulic geometries based on cross-sections and longitudinal profile. (Sui *et al.*, 2002; Lawson *et al.*, 1986, Michel and Drouin, 1975; Kuuskoski, 1972; Gold and Williams, 1963). This study has identified the unique flow patterns of an entire reach encompassing a suspended accumulation of frazil, based on multi-date investigations at several cross-sections. One common findings of our result with other studies is that frazil deposits can occupy more than 60% of cross-sectional areas. Another common finding is that sub-ice channels morphology often reflects open-channel flow processes, suggesting that currents in sub-ice channels erodes the impeding frazil material where flow normally takes place. Having uncovered the sub-ice flow dynamics of the frazil-pool, further studies will be able to address the matter of sub-ice flow velocities.

The noted lateral shifting of the channel at the exit-slope is not uncommon. Frazil accumulations are known to migrate in upstream, downstream and lateral directions (Sui *et al.*, 2002; Shen and Wang, 1995; Lawson *et al.*, 1986). It could also result from local bed deformation. Unfortunately the DBR were not deployed in this area because of the shallow depths and thus could not confirm any morphological changes. As presented by Shen (1995), frazil granules are subject to shear stress like bed particles. Induced deformation by frazil accumulation should be found where sediment size resistance to shear stress is lower than frazil resistance. Based on this, the lateral displacement of the channel over the exit-slope could be the result of the lateral erosion of small-size sediments indirectly-caused by a more resistant frazil ice accumulation. The exit-slope riverbed is mainly composed of sand and silt, a common sediment population for exit-slopes. This remains to be documented.

3.5.3 Morphological changes and implication for pool dynamics

The dynamic bed-rods results show a clear relation between riverbed deformation and the progression of the ice cover over the frazil-pool. On December 7th, numerous responses were triggered from DBR flow-sensors and ground-sensors along the main channel. These daily responses are interpreted as short term erosional activity followed by sediment deposition, suggesting a rapid and strong flow increase in the main channel. Such flow dynamics could be explained by either the passage of a backflow wave preceding the ice cover progression towards the pool or to a pressure flow effect in response to the formation of a new boundary layer (Prowse and Gridley, 1993). It has to be noted that the recorded scouring was not enduring and could not be related to frazil accumulation.

Two exit-pool flow sensors, #4 and #6, showed no-motion responses over the ice cover period. This suggests that frazil ice rapidly accumulated at the exit-slope when a partial cover was available and propagated upstream with the ice cover. This was also observed on the Yellow river, where the jam had propagated, with the ice cover, from downstream to upstream (Sui *et al.*, 2002).

The documented frazil-pool is abnormally large and deep in comparison to other pools in the 24 km long river corridor. Its irregular dimensions, maximal width of 75 m and a maximal depth of 6.2 m (low flow), are hardly explainable when compared to average widths of 38-40 m and an average pool depth of 1.8 m at low flow (second deepest pool is 4.7 m deep). Irregular fluvial topography is common of semi-alluvial environments where rock and clay bed material causes constriction to the flow. The deep main channel of the frazil-pool is effectively subject to an underwater clay-constriction that might have enhanced the pool's depth. The pool is also deformed in width, by eroding into the Mitis terrace's silty intertidal deposits.

Analysis revealed an irregular, but stable, frazil ice layer of thicknesses ranging from 0 to 6.25 meters. Frazil ice occupied from 30 to 70% of its cross-sections, accommodating several sub-ice channels that extended to a maximal height of 2.5 meters. It remains a challenge to establish which component is causal in this complex relationship: bedform or iceform? Do irregular pools behave as frazil-sink allowing the infilling of a broadly proportional accumulation of frazil ice? Do frazil accumulations behave as constrictions to the flow inducing the lateral and vertical scouring of the frazil sink? Furthermore, which ice period induces the most ice-related morphological changes: freeze-up, ice cover progression, winter discharge increases or break-up? Facing the uncertainty of the origin of the scour of the documented frazil-pool, and knowing the extent of frazil ice presence into the pool, we hypothesize that frazil-pool maintenance and evolution is imputable to strong positive feed-back mechanisms between ice dynamics and fluvial dynamics.

Most well-known ice-related bedforms, like the *Bechevnick* introduced by Hamelin in 1979, result from the direct contact of ice. Deep pools are well known frazil-sinks in cold river systems and anomalous deep-pools have been intuitively associated to ice dynamics. The implication of complex interrelations seems to have restrained the identification of such bedfoms. Although it remains presently difficult to discriminate the role of ice dynamics from fluvial dynamics in anomalous frazil-pool morphology, we believe the appellation « frazil-pool » is evocative and could well describe these peculiar cold river morphologies.

The DBR has proven to be a useful tool to document the exact occurrence of erosional or depositional events beneath recurrent frazil accumulation. A similar tool, based on time-domain reflectometry (TDR), has been successfully used under ice conditions to continuously monitor scouring events (Zabilansky *et al.*, 2000). However, this wired-method requires an on-shore data collection system, and skilled operators for signal processing and data presentation. In comparison, the DBR is an autonomous round-the-clock monitoring system easily processed. Further research is expected to explore the capacity of the DBR to document XYZ acceleration data. The DBR could also be used, with the vertical arrangement of numerous in-flow sensors, to document the downward growth of the frazil layer in the pool when field work is impossible.

3.6 Conclusion

Knowledge of winter stream flow regime is required in cold catchments to evaluate the risk of damage and assess the vulnerability of aquatic habitat. Frazil ice is an important part of river ice dynamics and its sensitivity to climatic conditions makes it a valuable research subject. The objective of this study was to explore the geomorphic relation between a frazil-pool and frazil accumulation. The infilling dynamics of a frazil-pool is presented as it relates in space and time to hydroclimatic conditions and to morphologic features. Although the frazil-pool was continuously supplied in frazil ice, the frazil ice layer did not change much during the winter. The frazil ice layer thickness ranged from 0 to 6.25 meters and frazil ice occupied up to 70% of the cross-sectional area. The dynamic bed-rods effectively documented the temporality and spatiality of indirectly generated bed deformation to the formation of the ice cover. Frazil ice accumulation did not trigger any DBR response. However, the DBR successfully responded to the increase of flow associated to the ice cover progression over the pool. GPR signal analysis was successful, but signal radar strength is visibly an important factor in radar probing. In this paper, we have shown how frazil ice accumulations in a medium-size river is surprisingly important and should therefore be regarded as an intrinsic morphological component rather than as punctual localized ice deposits.

CHAPITRE IV

CONCLUSION

Ce mémoire veut souligner la complexité des interrelations entre la dynamique glacielle, la géomorphologie et la sédimentologie fluviale tout en proposant d'une part un modèle conceptuel (chapitre 2) et d'autre part des outils méthodologiques permettant de s'attaquer à la compréhension de cette complexité. À toute échelle, les caractéristiques physiques d'un système fluvial jouent un rôle important dans le régime des glaces fluviales et inversement, les glaces refaçonnent le modelé fluvial par contact direct ou par le biais d'interrelations complexes. Peu d'ouvrages décrivent les processus géomorphologiques associés aux rivières avec couvert de glace. Les avancées scientifiques dans ce domaine sont peu nombreuses en partie parce que les données terrain sont rares (Scrimgeour *et al.*, 1994; Shen and Wang, 1995 Sui *et al.*, 2006), mais également parce qu'il n'existe aucun cadre de travail universel de caractérisation fluvio-glacielle.

Ce mémoire est structuré autour d'une réflexion entourant la relation frasil – formes du lit. L'approche préconisée vise d'abord à identifier clairement les composantes dépendantes et les liens directs, rétroactifs et complexes qui interviennent dans la dynamique fluviale hivernale. Dans un deuxième temps, une étude de cas documente en continu l'évolution temporelle complexe de la déformation du lit et l'évolution des formes glacielles dans un environnement favorable à l'accumulation de frasil sous couvert de glace.

Les résultats de ce mémoire émergent d'une approche originale, d'abord théorique, puis pratique, mais également méthodologique. La figure 2.3, présentée au second chapitre, présente un cadre conceptuel simple et flexible pour la caractérisation de la dynamique fluviale des rivières froides. Trois composantes glacielles sont mises en évidences et intégrées à la dynamique fluviale de Leeder (1983), dans un modèle conceptuel où les six composantes entrent en relation. Tout lien établi entre deux composantes est soutenu par une revue de littérature exhaustive. Cet outil de représentation novateur illustre efficacement la complexité de la relation documentée au chapitre 3 et peut aisément être intégré à d'autres recherches.

L'étude de cas présentée au chapitre 3 révèle la dépendance de la répartition spatiale du frasil à la géomorphologie fluviale et, inversement, elle montre l'ajustement morphologique d'une fosse à la dynamique glacielle. Cet essai s'organise à deux échelles spatiales. Dans un premier temps, une approche par corridor alluvial est essentielle à la compréhension de l'approvisionnement de la fosse en frasil car elle permet d'évaluer la spatialité et la temporalité de la production et du transport de frasil. Cette approche est complémentaire au suivi glacio-géomorphologique de la fosse qui se démarque par une approche méthodologique systématique et novatrice. En effet, il importait de documenter la déformation du chenal en continu plutôt qu'à postériori pour distinguer avec exactitude l'influence du glaciel. À cet effet, une matrice de bornes sédimentaires dynamique a été développée et déployée sur le lit de la fosse. Cet instrument autonome (sans-fil) a montré son efficacité pour mesurer l'érosion et la déposition de sédiments dans un chenal. Il était également essentiel au projet de documenter la constitution, la répartition et l'évolution mensuelle de l'accumulation suspendue de frasil. Des travaux sur champs de glace ont servi à sonder l'épaisseur du couvert de glace, de l'accumulation de frasil et la profondeur d'eau libre sous-glacielle. Deux techniques de représentation ont été utilisées : le sondage manuel ainsi que le relevé géophysique. Ce choix, initialement exploratoire, confirme la faisabilité de l'utilisation des relevés géophysiques pour la détection des interfaces glace/frasil et frasil/eau.

Les résultats de ce mémoire dépassent l'intérêt fondamental de caractériser la dynamique hivernale d'une fosse-à-frasil. Ils rejoignent des intérêts du domaine des sciences biologiques, de l'ingénierie, de la science de la glace et de la gestion du territoire. La modélisation conceptuelle est une composante indissociable du travail scientifique, car il procure un cadre d'analyse pour le questionnement, la discussion et la coordination d'efforts de recherche. La représentation de la dynamique fluvio-glacielle selon un assemblage de relations formant un tout permet de comprendre et de communiquer les mécanismes internes des rivières froides.

D'importants liens unissent la dynamique fluvio-glacielle aux taux d'oxygène dissous, au transport de polluant, à la végétation riveraine, aux flux nutritifs, à la protection et la destruction de l'habitat du poisson et des invertébrés et la survie du poisson (Morse et Hicks, 2005). Le manque de données terrain sur la dynamique morpho-sédimentaire hivernale, particulièrement durant les phases de transitions glacielles, est décrié par des spécialistes de tous les domaines (Beltaos, 2000; Shen, 2003; Morse et Hicks, 2005). L'obtention de mesures quantitatives hivernales, sous couvert de glace, apporte des données inédites sur les conditions glacielles hivernales d'une rivière bas-laurentienne de taille moyenne. Ces résultats pourront être mis à contribution dans la gestion des risques d'embâcles, dans l'évaluation de la sensibilité de l'habitat du poisson aux conditions hivernales et dans la gestion des corridors fluviaux des régions nordiques.

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