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# From outwash to coastal systems in the Portneuf–Forestville deltaic complex (Québec North Shore): anatomy of a forced regressive deglacial sequence

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## ABSTRACT

Deglacial sequences typically include backstepping grounding zone wedges and prevailing glaciomarine depositional facies. However, in coastal domains, deglacial sequences are dominated by depositional systems ranging from turbiditic to fluvial facies. Such deglacial sequences are strongly impacted by glacio-isostatic rebound, the rate and amplitude of which commonly outpaces those of post-glacial eustatic sea-level rise. This results in a sustained relative sea-level fall covering the entire depositional time interval. This paper examines a Late Quaternary, forced regressive, deglacial sequence located on the North Shore of the St. Lawrence Estuary (Portneuf Peninsula, Québec, Canada) and aims to decipher the main controls that governed its stratigraphic architecture. The forced regressive deglacial sequence forms a thick (>100 m) and extensive (>100 km<sup>2</sup>) multiphased deltaic complex emplaced after the retreat of the Laurentide Ice Sheet margin from the study area *ca* 12 500 years ago. The sedimentary succession is composed of ice-contact, glaciomarine, turbiditic, deltaic, fluvial and coastal depositional units. A four-stage development is recognized: (i) an early ice-contact stage (esker, glaciomarine mud and outwash fan); (ii) an in-valley progradational stage (fjord head or moraine-dammed lacustrine deltas) fed by glaciogenics; (iii) an open-coast deltaic progradation, when proglacial depositional systems expanded beyond the valley outlets and merged together; and (iv) a final stage of river entrenchment and shallow marine reworking that affected the previously emplaced deltaic complex. Most of the sedimentary volume (10 to 15 km<sup>3</sup>) was emplaced during the three-first stages over a *ca* 2 kyr interval. In spite of sustained high rates of relative sea-level fall (50 to 30 mm·year<sup>-1</sup>), delta plain accretion occurred up to the end of the proglacial open-coast progradational stage. River entrenchment only occurred later, after a significant decrease in the relative sea-level fall rates (<30 mm·year<sup>-1</sup>), and was concurrent with the formation and preservation of extensive coastal deposits (raised beaches, spit platform and barrier sands). The turnaround from delta plain accretion to river entrenchment and coastal erosion is interpreted as a consequence of the retreat of the ice margin from the river drainage basins that led to the

drastic drop of sediment supply and the abrupt decrease in progradation rates. The main internal stratigraphic discontinuity within the forced regressive deglacial sequence does not reflect changes in relative sea-level variations.

**Keywords** Deglaciation, forced regressive delta, Holocene, St. Lawrence Estuary, Wisconsinan.

#### 4 INTRODUCTION

Deciphering the cyclical patterns of ice sheet advance and retreat from the sedimentary record is crucial for understanding the behaviour of past ice sheets. The processes of deglaciation are commonly inferred from sedimentary successions that include ice-contact deposits (e.g. backstepping morainal banks) overlain by more ice-distal glaciomarine depositional suites (Boulton, 1990; Visser, 1997; Syvitski & Lee, 1997; Lønne *et al.*, 2001; Jakobsson *et al.*, 2011; Carling, 2013; Batchelor & Dowdeswell, 2015). Yet, a wider spectrum of depositional systems, mainly including turbiditic, deltaic, lacustrine, coastal, tidal or fluvial deposits, constitute the bulk of many other deglacial sequences related to continent-scale ice sheet recessions, as evidenced in post-Late Glacial Maximum (LGM) successions (e.g. Syvitski & Hein, 1991; Hansen, 2004; Fraser *et al.*, 2005; Corner, 2006; Eilertsen *et al.*, 2011; Nutz *et al.*, 2014, 2015; Normandeau *et al.*, 2015) and in the deep time glacial record (e.g. Proust & Deynoux, 1990; Ghienne, 2003; Le Heron *et al.*, 2006, 2011, 2013; Isbell *et al.*, 2008; Loi *et al.*, 2010; Girard *et al.*, 2015). The nature and mode of deposition of such deglacial sedimentary sequences are controlled by complex interactions between the history of ice margin retreat (e.g. slow, punctuated, rapid retreat) and evolution of the related sediment entry points, the inherited basin physiography and paraglacial reworking processes, among others (Syvitski & Farrow, 1983; Brookfield & Martini, 1999; Lønne *et al.*, 2001; Powell & Cooper, 2002; Lajeunesse & Allard, 2002; Occhietti, 2007; Storms *et al.*, 2012; Hein *et al.*, 2014). Across large segments of formerly glaciated inner shelves, rates (up to 10 cm·year<sup>-1</sup>) and amplitudes (>100 m) of the glacio-isostatic rebound significantly outpace those of the eustatic sea-level rise throughout the entire deglacial period (Boulton, 1990). It therefore results in the uplift of deglacial sedimentary

successions, which were themselves deposited under conditions of relative sea-level (RSL) fall (e.g. Hart & Long, 1996; Lajeunesse & Allard, 2002; Dionne & Occhietti, 1996; Dionne *et al.*, 2004; Eilertsen *et al.*, 2011; Nutz *et al.*, 2015). The imprints of such an overall forced regressive context have to be formally understood. Forbes & Syvitski (1994) and Normandeau *et al.* (2015) stressed that the inland extent of river drainage basins is of prime importance in the stratigraphic development of deltas tied to deglacial sequences.

Depositional systems built after the LGM offer widely datable sedimentary archives and demonstrative connections having a well-preserved morphosedimentary record. The present study takes advantage of an uplifted, Pleistocene to Holocene (ca 12 500 to 4000 year BP) deglacial sedimentary complex built on the Québec North Shore and includes ice-contact, in-valley and open-coast deltaic and shallow marine deposits (St. Lawrence Estuary, Fig. 1). The outstanding exposure (80 m high and 7 km long, Fig. 2) enables the following: (i) detailed documentation of internal stratigraphic architectures and the related depositional facies; (ii) proposal of a detailed chronostratigraphic framework; and (iii) deciphering of the relative influence and signatures of forcing factors that controlled the deposition patterns throughout the forced regressive deglacial development. It is postulated here that the resulting scheme may help in unravelling the deep time deglaciation record, for which no dating at a resolution higher than 100 kyr can be expected, in spite of glacial tempos most likely comparable to that of the Late Cenozoic (Pazos, 2002; Fielding *et al.*, 2008b; Ghienne *et al.*, 2014). This study will also show that no river incision occurred during most of the deglacial sequence, corroborating some recent models depicting sediment accumulation during RSL fall conditions (Swenson & Muto, 2007; Blum *et al.*, 2013; Prince & Burgess, 2013).

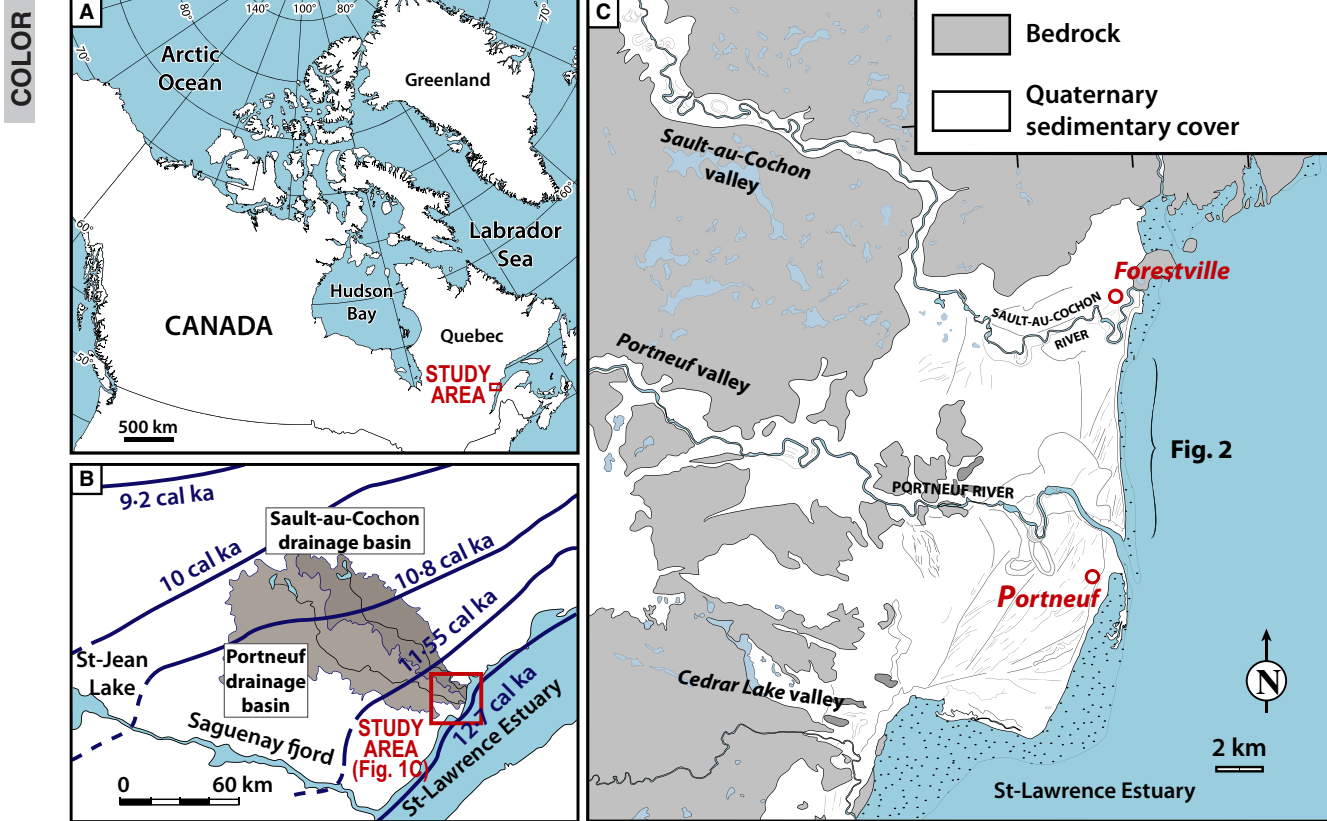


Fig. 1. (A) Location of the study area (B) in the St. Lawrence Estuary. (B) The drainage basins of the Portneuf (light grey) and Sault-au-Cochon (dark grey) rivers, relative to the reconstructed, post-LGM, retreating ice fronts of the Laurentian Ice Sheet (Occhiatti *et al.*, 2011). (C) Outline of the Portneuf Peninsula (a detailed version of this map is given in Figure S1).

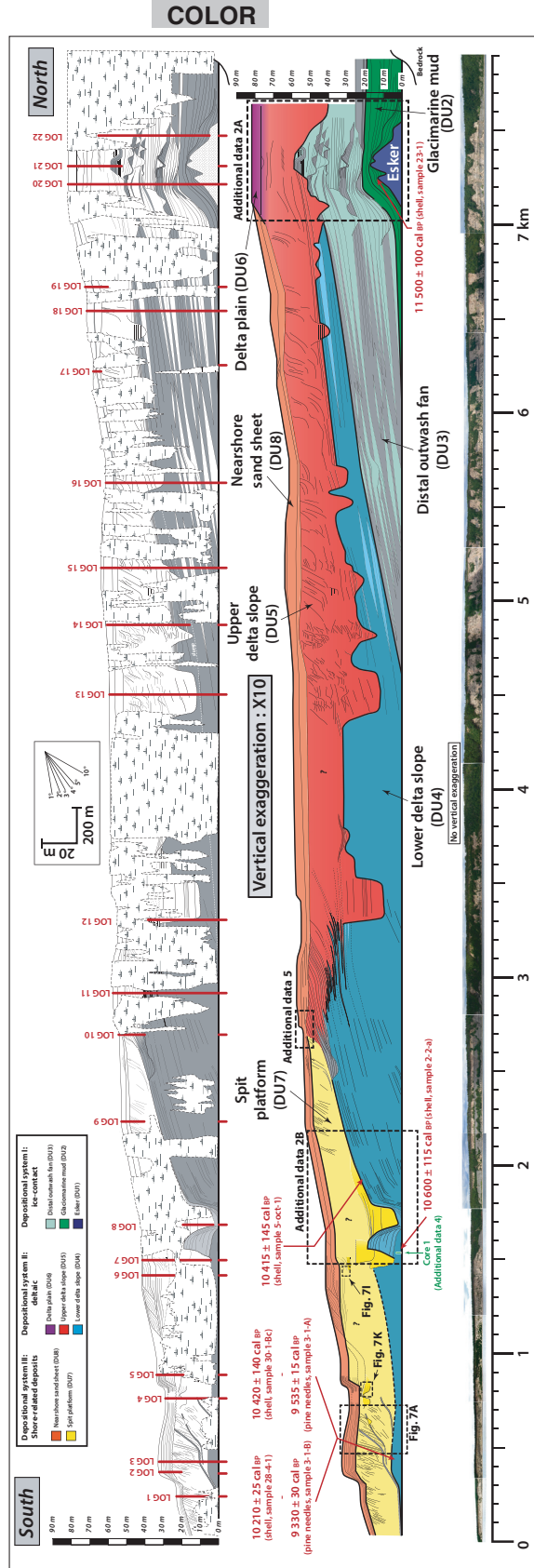
## REGIONAL SETTING

### Geological setting

The study area comprised the Portneuf Peninsula and its hinterland, which together cover *ca* 200 km<sup>2</sup> on the northern coast of the St. Lawrence Estuary (Québec North Shore, Fig. 1). To the west, Proterozoic basement rocks of the Grenville Province (Thériault *et al.*, 2012) crop out across a gently sloped plateau area at *ca* 300 m elevation. Seawards, only bedrock foothills emerge from the coastal Late Wisconsinan to Holocene sedimentary wedge, the upper surface of which runs from the *ca* 140 m.a.s.l. (above sea-level) elevation down to the present-day shoreline. This surface displays relatively well-preserved fluvial and marine terraces (Dionne *et al.*, 2004). The sediment wedge thickness is highly variable depending on the underlying bedrock topography. Coastal erosion and river incision have severely cross-cut the sedimentary

wedge, and exposures reaching 90 m in height are distributed along riverbanks and sea cliffs (Fig. 2). Offshore, a shallow coastal shelf extends down to depths of 100 m. At *ca* 8.5 km off the present-day shoreline, a south-west/north-east-oriented break-in-slope makes an abrupt transition with the St. Lawrence rift system and trough (Tremblay *et al.*, 2003, 2013; Duchesne *et al.*, 2007, 2010), where the basement top lies at a depth of 700 m.

Onshore, bedrock terrain frequently displays preserved glacial striations, grooves and crescentic chattermarks (Figure S1); bedrock morphologies being thus viewed as inherited relief forms. In addition, deep, steep-flanked and narrow structural valleys are deeply cut into basement rocks, which focus the flow of the Portneuf and Sault-au-Cochon rivers. The depth of the bedrock incisions, which may locally include a basal sedimentary infill underneath the present-day river bed, is unknown. At times, the width and depth of the Portneuf and Sault-au-Cochon



**Fig. 2.** The 7 km long, south–north-oriented cliff extending from the Portneuf River mouth to Forestville (location in Fig. 1). From the base to the top: photomosaic with no vertical exaggeration, interpreted stratigraphic architecture and line drawing (sand-dominated deposits in white; mud-dominated deposits in grey). Undocumented segments correspond to vegetation-covered areas. Colours in the interpreted stratigraphic architecture are tied to depositional units DU1 to DU8. Logs 1 to 22 are given in Fig. 3. See Table 1 for details about <sup>14</sup>C ages.

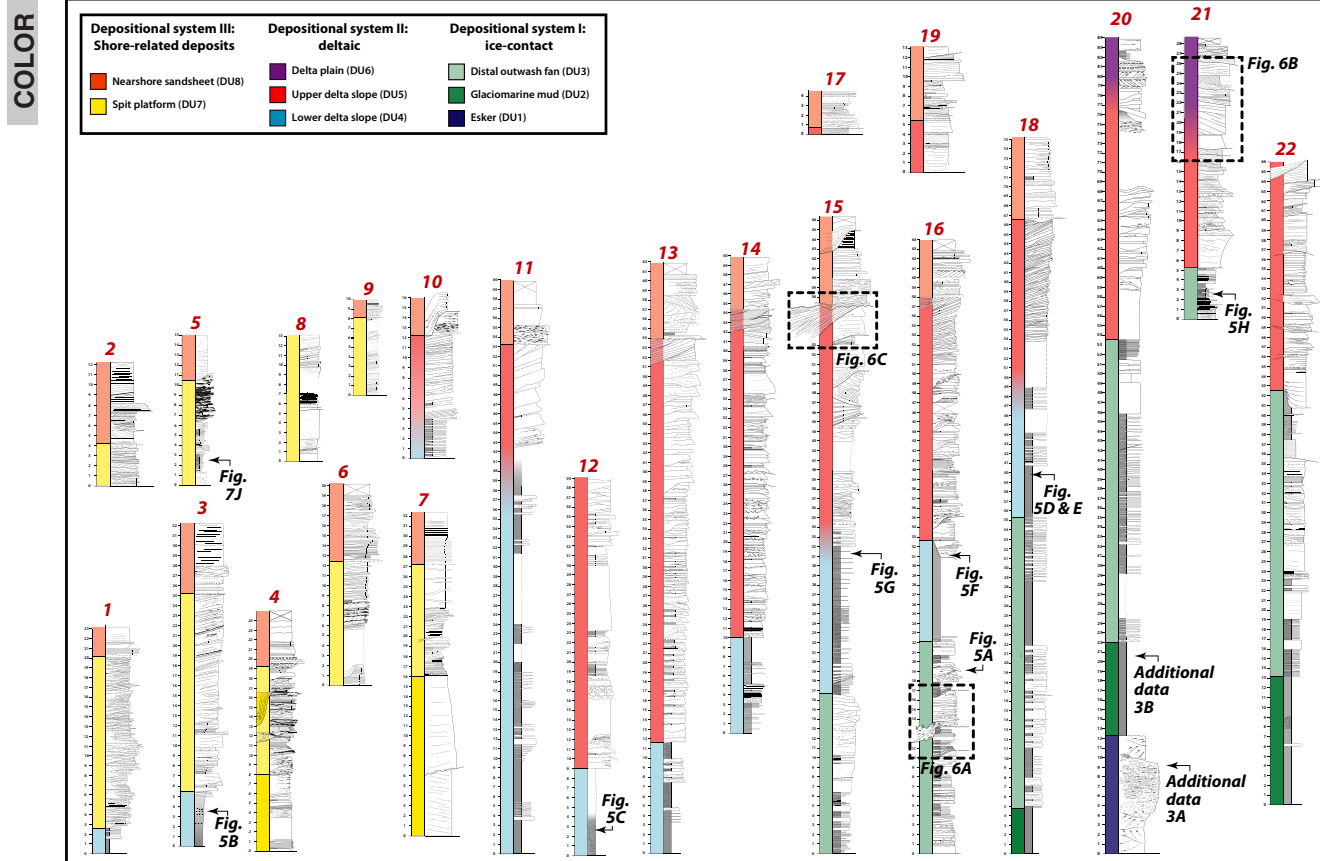


Fig. 3. Sedimentological logs distributed along the cliff (locations in Fig. 2).

valleys are *ca* 1 km and 100 to 150 m, and 0.5 to 0.8 km and 80 m, respectively. Drainage areas extend 120 km to the north-west (Fig. 1B). Bedrock thresholds suggest that the structural valley depths may not be significantly greater than the observed flank heights with the possible exception of inner gorges (Lajeunesse, 2014). Marine sediments preserved in terraces along the lower reaches of the valleys lying below 140 m.a.s.l. (marine limit) indicate that they were initially inundated at an early time of deglaciation and were temporarily shallow fjords.

### Deglacial setting

The development of the Portneuf–Forestville deltaic complex dates back to the Laurentide Ice Sheet (LIS) retreat following the LGM. An initial and rapid glaciomarine phase of ice margin retreat is interpreted to have occurred across the whole of the St. Lawrence Gulf and Estuary up to the Late Wisconsinan (Shaw *et al.*, 2006). Then, the ice-marginal zones temporarily stabilized along the Canadian Shield highlands

bordering the Québec North Shore (Dyke *et al.*, 2002; Shaw *et al.*, 2002, 2006; Occhietti, 2007; Occhietti *et al.*, 2011; Margold *et al.*, 2015; Lajeunesse, in press). After this stage, ablation-dominated continental ice fronts characterized the recession of the Québec Ice Dome up to its final withdrawal dated at *ca* 6 kyr (Clark *et al.*, 2000; Storrar *et al.*, 2014). In the St. Lawrence Estuary and Gulf, the resulting depositional sequence consisting of a discontinuous subglacial till covered with ice-proximal to ice-distal glaciomarine deposits is ubiquitously draped by 20 to 50 m of post-glacial hemipelagic muds (Syvitski & Praeg, 1989; Josenhans & Lehman, 1999; Duchesne *et al.*, 2007, 2010; St-Onge *et al.*, 2008). The post-glacial marine invasion accompanying the retreat of the LIS margin flooded a glacio-isostatically depressed strip of land (Goldthwait Sea, Occhietti *et al.*, 2004), over which proglacial and paraglacial coastal landforms and deposits developed (Dionne & Occhietti, 1996; Hart & Long, 1996; Dionne *et al.*, 2004). The coeval glacio-isostatic rebound led to a regional and continuous RSL fall in spite of the concurrent

1 global glacio-eustatic rise (Peltier & Fairbanks,  
 2 2006). As a consequence, a significant part of  
 3 the growth of the Late Wisconsinan to Holocene  
 4 sedimentary wedges occurred within a forced  
 5 regressive context (Hart & Long, 1996), under  
 6 high rates of sea-level fall ( $>2 \text{ cm}\cdot\text{year}^{-1}$ , Shaw  
 7 *et al.*, 2002; Tarasov *et al.*, 2012). According to  
 8 regional ice-front reconstructions, the Portneuf  
 9 area was free of ice shortly after 12.7 kyr cal BP  
 10 (Occhietti *et al.*, 2011; Fig. 1B) and the marine  
 11 limit – the elevation of the highest marine sedi-  
 12 ments deposited in the Goldthwait Sea – is here  
 13 estimated to 140 m.a.s.l. (Dionne *et al.*, 2004).

14 The resulting deltaic complexes have been  
 15 investigated by onshore or offshore studies in  
 16 several areas of the St Lawrence Estuary and  
 17 Gulf (e.g. Natashquan: Long *et al.*, 1989; Sala &  
 18 Long, 1989; Sept-Îles: Dredge, 1983; Norman-  
 19 deau *et al.*, 2013; Boyer-Villemaire *et al.*, 2013;  
 20 Betsiamites and Manicouagan-Outardes: Ber-  
 21 natchez, 2003; Hart & Long, 1996; Portneuf:  
 22 Dionne *et al.*, 2004; Tadoussac: Dionne, 1996;  
 23 Dionne & Occhietti, 1996). A recent overview of  
 24 the subaqueous expressions of the deltaic sys-  
 25 tems in the St. Lawrence Estuary is given by  
 26 Normandeau *et al.* (2015) who conclude that at  
 27 present day, they are essentially inactive.

### 30 Present-day shorelines

31 Development of the Portneuf Peninsula since the  
 32 Mid-Holocene has been dominated by relatively  
 33 low rates of RSL fall ( $<1 \text{ cm}\cdot\text{year}^{-1}$ , Shaw *et al.*,  
 34 2002). The southward-oriented longshore drift  
 35 nourishes southward-prograding shorelines with  
 36 a correlative migration of barrier islands and  
 37 back barrier marshes developed since 4000 years  
 38 cal BP (Dionne *et al.*, 2004; Fig. 1C and Fig-  
 39 ure S1). In contrast, the northern Portneuf Penin-  
 40 sula is today experiencing active shoreline  
 41 retreat (Bernatchez & Dubois, 2004). As a conse-  
 42 quence, the sedimentary complex is intersected  
 43 in its northern portion by a  $>7 \text{ km}$  long, remark-  
 44 ably linear, active cliff extending between  
 45 Pointe-des-Fortins and Patte-de-Lièvre Island  
 46

(Figure S1), the main data provider for this study  
 (Fig. 2).

### METHODS

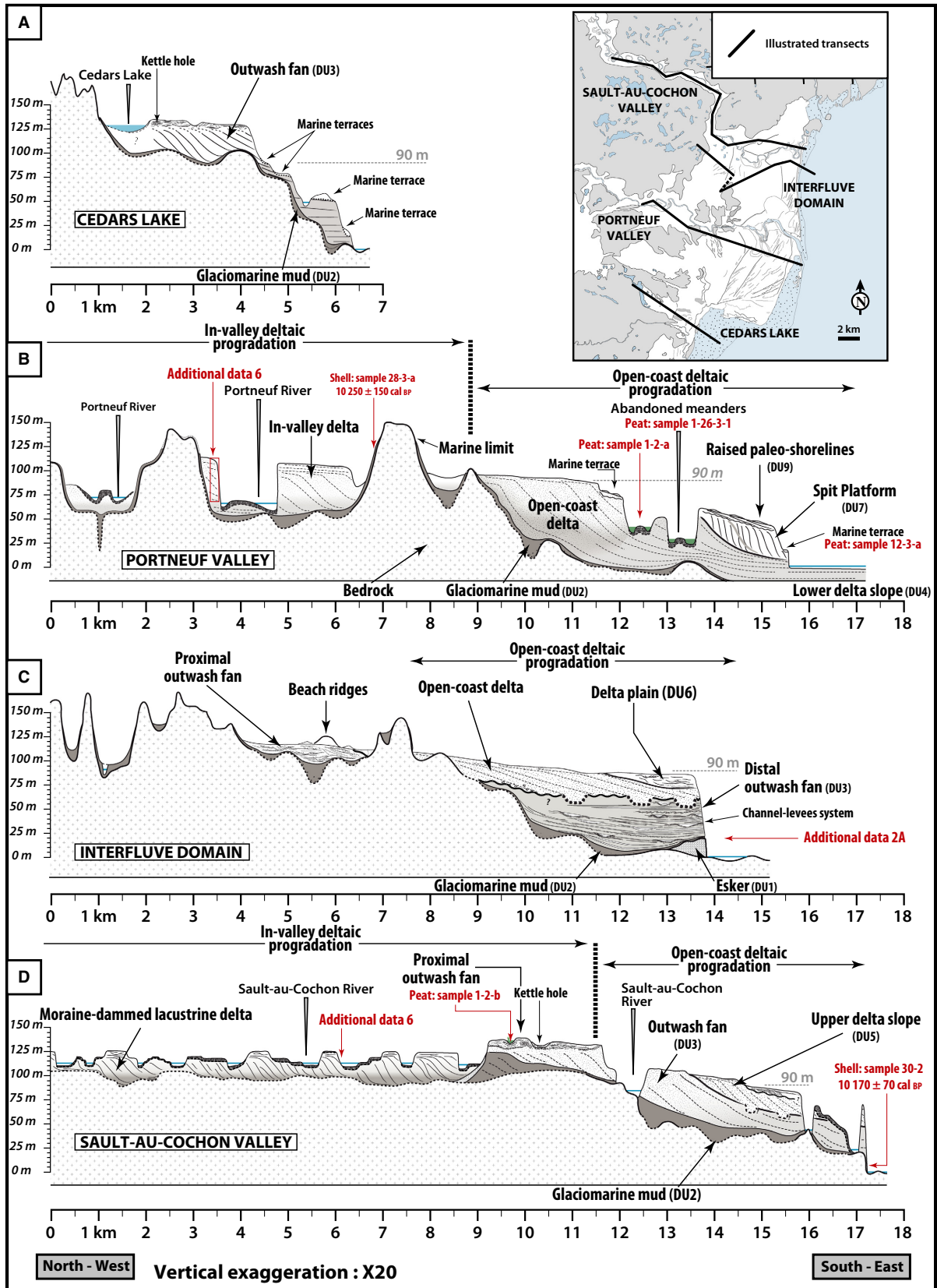
The geomorphological context is provided by  
 maps of surficial deposits (Fig. 2 of Dionne *et al.*,  
 2004; Cousineau *et al.*, 2014). The nature, eleva-  
 tion and pattern of landforms were characterized  
 from aerial photographs (1 : 40 000, available at  
 the Ministry of Natural Resources and Wildlife,  
 Government of Québec) and satellite images (Goo-  
 gle Earth<sup>TM</sup>) combined with the SRTM3 DEM  
 (grid resolution: 93 m, available on the USGS  
 website). Geographic information system (GIS)  
 software was used to account for the effects of gla-  
 cio-isostatic flexure and rebound (see Leverington  
*et al.*, 2002) on the basis of isobase curves taken  
 from Shaw *et al.* (2002) and Tarasov *et al.* (2012).

The internal architecture of the deltaic com-  
 plex was analysed from a panoramic photomo-  
 saic of the cliff taken from a boat at a distance  
 of *ca* 400 m (Fig. 2 and Figure S2). Regularly  
 spaced sedimentary logs at the 1 : 100 scale  
 were performed including sedimentary facies  
 description (Fig. 3), with inspection of bounding  
 surfaces observed in the photomosaic. Addi-  
 tional photomosaics and sections were made  
 inland along riverbanks. The inland sections  
 characterize the inner (and older) portions of the  
 deltaic complex (Figs 1 and 4, Figure S1). Six  
 handmade 30 to 40 cm long cores were recov-  
 ered in mud facies for the identification of inter-  
 nal structures by X-ray imagery.

Thirteen radiocarbon age determinations were  
 performed at the  $^{14}\text{C}$  measurement laboratory of  
 the CNRS (Gif-sur-Yvette, Artemis program/  
 CNRS-INSU) and at the Poznan Radiocarbon  
 Laboratory (Table 1). Data calibration was per-  
 formed using the IntCal09 procedure (Reimer  
*et al.*, 2009). An estimate of the reservoir effect  
 in the St. Lawrence Estuary at the time of the  
 Goldthwait Sea derives from measuring  $^{14}\text{C}$  ages  
 of shells and conifer needles in nearby horizons.

48 **Fig. 4.** Synoptic transects from the hinterland to the current shoreline, integrating sedimentological logs and mor-  
 49 phosedimentary features. (A) In the Cedars Lake valley, the outwash fan (apex at 140 m.a.s.l.) is reworked by mar-  
 50 ine terraces. (B) The Portneuf Valley transect illustrates the relationship between the basement and the sediment  
 51 wedge. The delta was initiated as an in-valley fjord head delta before prograding as an open-coast delta. (C) The  
 52 transect in the interfluvial domain highlights the architecture from the proximal to distal outwash fan, superim-  
 53 posed by the proglacial delta. Both are exposed along the cliff (Figure S2A). (D) The Sault-au-Cochon transect  
 54 includes the in-valley moraine-dammed lacustrine delta, an antecedent outwash fan positioned at the valley outlet  
 55 (Figure S1) and the subsequent proglacial deltaic system.

COLOR





**Table 1.** Radiocarbon dates used for the reconstruction of the temporal framework. The time calibration was performed using the IntCal09 curve (Reimer *et al.*, 2009). See also Dionne *et al.* (2004) for ages younger than 4000 year BP.

Sample number	Nature	Location	Depositional system	<sup>14</sup> C age	Calibrated age	Significance
23-Jan	Marine shell fragments ( <i>Mya arenaria</i> )	Glaciomarine mud draping the esker (20 m.a.s.l.)	DU2	10 010 ± 50 BP	11 500 ± 100 BP	First sediments deposited after the deglaciation
28-3-a	Marine shell ( <i>Mya arenaria</i> )	Glaciomarine mud draping bedrock 20 km inland (105 m.a.s.l.)	DU2	10 380 ± 40 BP	12 250 ± 150 BP	First sediments deposited after the deglaciation
30-2	Marine shell fragments	Glaciomarine mud draping bedrock near Forestville (5 m.a.s.l.)	DU3	9 735 ± 40 BP	11 170 ± 0 BP	First sediments deposited after the deglaciation
2-2-a	Marine shell fragments ( <i>Mya arenaria</i> )	Base of the channel levée (1 m.a.s.l.)	DU4	9 360 ± 50 BP	10 600 ± 115 BP	Onset of the channel-levée development
05-Oct-01	Marine shell fragments ( <i>Mya arenaria</i> )	Top of the channel levée (20 m.a.s.l.)	DU4	9 250 ± 50 BP	10 415 ± 145 BP	End of the channel-levée development
30-1-Bc	Marine shell ( <i>Buccinidae</i> )	Top of DU4 (2 m.a.s.l.)	DU4	9 260 ± 60 BP	10 420 ± 140 BP	End of the proglacial dynamic
3-1-B	Plant debris (Pine needle)	Top of DU4 (2 m.a.s.l.)	DU4	8 350 ± 40 BP	9 370 ± 100 BP	Lower delta slope deposits
3-1-a	Plant debris (Pine needle)	Top of DU4 (2 m.a.s.l.)	DU4	8 370 ± 40 BP	9 535 ± 15 BP	Lower delta slope deposits
28-4-1a	Marine shell fragments ( <i>Mya arenaria</i> )	Top of DU4 (1 m.a.s.l.)	DU4	9 030 ± 50 BP	10 210 ± 25 BP	Lower delta slope deposits
12-3-a	Peat	Backshore marshes (20 m.a.s.l.)	DU8	3 560 ± 30 BP	3 780 ± 20 BP	Backshore (lagoon) at RSL = 20 m
1-2-a	Peat	Abandoned meander peat accumulation (32 m.a.s.l.)		5 550 ± 35 BP	6 435 ± 15 BP	Abandoned meander and peat growth at RSL < 32 m
1-26-3-1	Peat	Abandoned meander peat accumulation (23 m.a.s.l.)		2 180 ± 30 BP	2 270 ± 20 BP	Abandoned meander and peat growth at RSL < 23 m

1 An estimate of  $800 \pm 100$  year age difference is  
 2 determined, which is similar to the reservoir age  
 3 of the North Atlantic during the Younger Dryas  
 4 Event (Bard *et al.*, 1994).

## 7 LANDFORMS

9 An outline of landforms in the Portneuf Penin-  
 10 sula is given below, offering valuable informa-  
 11 tion for the facies interpretation of the  
 12 underlying deposits observable within the cliff  
 13 (Fig. 2). A succession of downstepping deposi-  
 14 tional surfaces separated by break slopes is evi-  
 15 denced from 140 m (i.e. marine limit) down to  
 16 the current shoreline. A significant break-in-  
 17 slope is observed around 90 m.a.s.l., marking  
 18 the transition from upward slopes in the 0.1 to  
 19 0.4% range, to steeper slopes (1 to 2%) down-  
 20 ward. Four types of landforms are identified: flu-  
 21 vial terraces and abandoned meanders, delta  
 22 plain, ice-contact and coastal forms.

### 24 Fluvial terraces and incised, abandoned 25 meanders

27 Extensive portions of the Portneuf and the  
 28 Sault-au-Cochon rivers are meandering and dis-  
 29 play abandoned meanders and oxbow lakes as  
 30 well as fluvial terraces (Fig. 1; Figure S1).  
 31 Incised, abandoned meanders and fluvial ter-  
 32 races are mainly observed at a short (<8 km) dis-  
 33 tance from the present-day river mouths, cutting  
 34 into the coastal sedimentary wedge. Four aban-  
 35 doned meanders are delineated along the lower  
 36 Portneuf River (Fig. 1; Figure S1). From the  
 37 older to the younger, they are positioned at  
 38 decreasing elevations (65 m, 32 m, 23 m and  
 39 15 m). Peat accumulation in abandoned mean-  
 40 ders at 32 m and 23 m.a.s.l. provided minimum  
 41 ages of  $6\,435 \pm 15$  and  $2\,270 \pm 20$  year cal BP  
 42 (Table 1) for their abandonment, respectively.  
 43 Abandoned meanders along the Sault-au-Cochon  
 44 River are staged from 60 to 20 m (Figure S1).

45 Today, the occurrence of local base levels  
 46 imposed by bedrock thresholds drastically  
 47 reduces river incision rates. However, because no  
 48 bedrock thresholds occur seaward of the aban-  
 49 doned meanders, the related river profiles were  
 50 adjusted to the regional base level (St. Lawrence  
 51 Estuary). As shown by meander development  
 52 along the lower reaches of some of present-day  
 53 rivers of the Québec North Shore, coastal mean-  
 54 ders have elevations very close to the concurrent  
 55 sea level. Palaeomeander elevations in the study

area thus provide valuable (misfit <5 m) estimates  
 of former sea levels at the time of incision.

### Delta plain landforms

Dionne *et al.* (2004) interpreted the flat sedimen-  
 tary plain located upslope of the 90 m elevation  
 contour as deltaic sand. Relic braided channel  
 morphologies in the 90 to 110 m elevation range  
 are preserved off the outlet of the Portneuf Valley  
 to the south of the current river course. Similar fea-  
 tures are also observed north of Forestville (Fig-  
 ure S1). Throughout the delta plain, braided  
 streams have significantly reworked any former  
 coastal landforms and related deposits (see below).

### Ice-contact landforms

Two bodies of ice-contact sediments are identi-  
 fied in the study area (Cousineau *et al.*, 2014).  
 The main body (Fig. 4; Figure S1) crops out at  
 the Sault-au-Cochon Valley outlet. It consists of a  
 large fan-shaped sediment accumulation reach-  
 ing 5 to 7 km in width. Its apex is located at the  
 narrowest point of the valley outlet. Radiating to  
 the south-east, its upper surface is even and sub-  
 horizontal near the apex that is positioned at  
 140 m.a.s.l. (i.e. marine limit). It progressively  
 slopes down, reaching 110 m.a.s.l. at 5 km to the  
 south. The upper surface of the ice-contact sedi-  
 ment body shows subdued relict channels and  
 subcircular to elongated, topographic depres-  
 sions, 20 to 50 m in diameter, occupied by lakes  
 or peat bogs (relict kettles). A subordinate body  
 of ice-contact sediments is identified in the  
 Cedars Lake valley (Figure S1).

The flat-topped and fan-shaped ice-contact  
 bodies were emplaced during the retreat of the  
 LIS margin (Cousineau *et al.*, 2014). The occur-  
 rence of kettles indicates an ice-marginal outwash  
 setting in front of glacier snouts debouching from  
 valleys. With their subaqueous, more ice-distal  
 counterparts cropping out downslope along the  
 current shore (see below, 'DU3' and Fig. 2), they  
 provide evidence of outwash fan systems larger  
 than the relatively restricted ice-contact deposits  
 observed in map view. It is worth noting that  
 comparable deposits do not exist off the outlet of  
 the Portneuf Valley.

### Coastal landforms

Two generations of coastal structures have been  
 observed. The first generation is observed above

90 m.a.s.l. and only includes sparse forms on the delta plain or bordering the ice-contact sediment bodies. Over the delta plain, a SSW–NNE-oriented, 3 km long and 50 m wide sand ridge is preserved below 110 m.a.s.l., halfway between the Portneuf and the Sault-au-Cochon rivers (Fig. 1; Figure S1). On both sides, it is bordered by subordinate sand ridges. Arcuate sand ridges are also identified over the original palaeodepositional profile of the ice-contact deposits, from its apex (140 m.a.s.l.) to nearly 110 m.a.s.l. (Figure S1). All of these sand ridges are interpreted as relict raised beach ridges, palaeocliffs, spits, cusps and tombolos reworking outwash or delta plain sands.

The second generation of coastal landforms is identified below the break-in-slope at 90 m.a.s.l. It consists of well-defined, extensive, linear landforms organized in a succession of flats and scarps. Scarps can be traced over the entire surface of the deltaic complex. The longest and highest scarp-oriented SSW–NNE extends along 17 km between Milles-Vaches Bay and Forestville (Figure S1). Its upper limit coincides with the 90 m break-in-slope and its height decreases northwards from 25 m to 5 m (Fig. 1; Figure S1). Four other, less-expressed, downstepping scarps are observed at lower elevations. Progressive change in their azimuth (SSW–NNE to SW–NE) causes truncation by the current cliff development, thereby offering exposed sections (Fig. 2). In flats between two successive scarps, sand ridges and intervening swales are 50 to 200 m in wavelength, 1 to 3 m in height. Beach ridges and sand spits were separated by back barrier marshes (Dionne *et al.*, 2004). The overall configuration reflects alternating phases of coastal erosion and beach accretion, for which it is difficult to assess the relative contribution of autocyclic processes – for example, nearby river mouth avulsion, Dominguez (1996) or cyclic formation patterns tied to finite sediment sources (Hein *et al.*, 2014) – or of allogenic forcing such as millennial climatic cycles (e.g. Sorrel *et al.*, 2012). Coastal landforms are intimately associated with an underlying sand sheet of nearshore deposits.

## STRATIGRAPHIC ARCHITECTURES AND DEPOSITIONAL FACIES

In the cliff exposure, eight depositional units (DU1 to DU8) characteristic of eight depositional environments are distinguished by their stratal

patterns and facies associations (Fig. 2). Facies associations encompass the 18 lithofacies summarized in Table 2. Information related to radiocarbon dating is summarized in Table 1. Sample locations are given in Figs 2 and 4. Depositional units are described from bottom to top, that is from the oldest and/or deepest to the youngest and/or shallowest (forced regressive setting). For sake of clarity, the eight depositional units are grouped into three depositional systems: (i) ice-contact (DU1 to DU3); (ii) open-coast deltaic (DU4 to DU6); and (iii) coastal (DU7 and DU8). In the following, notations such as ‘km 2’ or ‘km 2 to 3’ refer to positions along the cross-section shown in Figure 2 (origin at the southern tip of the cliff). An in-valley deltaic depositional system is described, which is not cropping out in the cliff but in the hinterland.

The entire region is characterized by continuous glacio-isostatic rebound with decreasing rates from the early phase of deglaciation up to the present day (Tarasov *et al.*, 2012; Peltier *et al.*, 2015). Thus, considering that the upper part of the cliff is characterized by coastal landforms, for example raised beaches, the minimal depositional palaeobathymetry for any underlying deposits is the height range up to the upright cliff top. The value is a minimum estimate because: (i) the erosional surface located at the base of the nearshore sand sheet inherently includes a hiatus (see below) and (ii) a depositional unit within the cliff is not strictly coeval with the beach deposits immediately above but relates to a shoreline positioned higher and landward.

### Depositional system I: Ice-contact deposits

#### *Depositional unit 1: Esker*

*Description.* Depositional unit 1 (DU1) rests directly on the striated bedrock cropping out on Patte-de-Lièvre Island (Fig. 2, km 7.0 to 7.5) where it forms an upwardly convex lenticular body of about 500 m in width and at least 18 m in thickness, having an irregular upper bounding surface (Fig. 2; Figure S2A). Depositional unit 1 lies immediately beneath distal glacial-marine muds (DU2).

Depositional unit 1 consists of poorly stratified sand and pebbles. Medium-grained to coarse-grained sand with faint oblique lamination (SGr2, Table 2) prevails. Well-rounded clast-supported pebbles (C1, Figure S3A) constitute metre-thick conglomeratic beds. A few decimetre-scale intraclasts of cohesive silty sands are observed.

**Table 2.** Lithofacies codes, description and interpretation.

Facies	Lithology	Sedimentary structures	Mean bed thickness	Interpretations	Overview
Mud					
Mu1	Mud or silty mud, granule-sized to cobble-sized outsized clasts	Massive or faintly laminated, mm-scale bioturbated laminae, relative abundance of shells ( <i>Macoma baltica</i> & <i>Mya arenaria</i> )	0.1 to 50 cm	Deposition from suspension fallout and/or from low-density gravity flows; outsized clasts (IRD)	
Mu2	Muddy silt	Well-bedded and well-laminated, mm-scale to cm-scale laminae, rare ripples	1 to 50 cm	Deposition from low-density gravity flows	
Mu3	Muddy silt	Contorted layers evidencing downslope shearing	1 cm to 1 m	Gravitational instabilities leading to creeping and/or slumping	
Fine-grained sand and silt					
SS1	Silt to very fine-grained sand	Well-laminated couplet of silt and very fine-grained sand, internal erosion at low angle, abundant bioturbation, contraction/dilatation cycles	1 to 10 mm	Turbulent deposition from low-density gravity flows; tidal or seasonal flow regime	
SS2	Very fine-grained to fine-grained sand and silt	Rhythmic climbing-ripple cross-lamination alternating sand and mud drape couplets, critical-to-supercritical angle of climb, occasional current reversal	1 to 10 mm	Tidal processes by reverse/bidirectional currents (Choi, 2010)	
SS3	Fine-grained to very fine-grained sand, silt drapes	Climbing ripples with increasing angle of climb, abundant micas, mud drapes	0.5 to 2 m	Deposition from tractive currents and suspension fallout, long-lived surging flow (Jobe <i>et al.</i> , 2012), seasonal or exceptional waning stage of flood events in a lake setting (Jopling & Walker, 1968)	
SS4	Fine-grained sand (well-sorted), subordinate muddy sand	Laminated to combined-flow ripples highlighted by heavy-mineral laminae, mud drapes	5 to 50 cm	Tractional deposition from bidirectional currents (wave action)	
SS5	Fine-grained to medium-grained sand (well-sorted), subordinate muddy sand	Large-wavelength (m-scale), HCS-like beds (planar to undulating)	30 to 50 cm	Deposition from high-energy oscillatory currents (storm waves)	

Table 2. (continued)

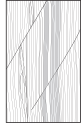


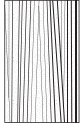
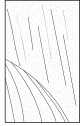
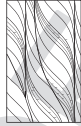
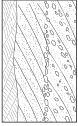




Facies	Lithology	Sedimentary structures	Mean bed thickness	Interpretations	Overview
SS6	Silt and very fine-grained sand	laminae), abundant heavy-mineral laminae Well-defined horizontal layering, intercalation of sub-critical climbing ripples, occurrence in place of syn-sedimentary deformation	<5 mm	Deposition from low-energy hyperpycnal flow and suspension fallout in a lake setting	
Sand S1	Fine-grained to medium-grained sand	Thin (<10 cm), normally graded and planar-laminated (Tb) and/or rippled (Tc) beds, occasional train of starved ripples	1 to 10 cm	Turbulent deposition from low-density turbidity currents (Talling et al., 2012)	
S2	Fine-grained to coarse-grained sand with occasional rip-up clasts and gravels	Thick (>10 cm), normally graded bed, basal massive (Ta) to horizontally laminated (Tb) with frequent angular rip-up clasts, capped by rippled sand (Tc) and mud (Te)	10 to 50 cm	Turbulent deposition from sediment-laden, high-density turbidity currents (Talling et al., 2012), possibly undergoing hydraulic jump (Postma et al., 2009).	
S3	Fine-grained to medium-grained sand, heavy-mineral rich (very well-sorted)	Evenly laminated sand, heavy and clear minerals alternation, thick (up to 15 cm) heavy-mineral placer with angular, large-scale sand intraclasts	10 to 50 cm	Tractional deposition in upper-flow regime in the beach swach zone (Clifton et al., 1971), sorting of heavy minerals by swach processes (Galloway et al., 2012)	
S4	Medium-grained to very coarse-grained sand (poorly sorted)	Horizontal to gently sloped layers, massive (Ta) and laminated (Tb) sand, erosion surfaces, forming large-scale undulated bedforms (10-20 m in wavelength)	1 to 5 m	Deposition from supercritical cyclic steps (Cartigny et al., 2011; Cartigny et al., 2014)	
Sand and gravels SGr1	Medium-grained to coarse-grained sand (well-sorted), subordinate granules	Sigmoidal cross-strata, laminae ubiquitously highlighted by heavy-mineral placers, reactivation surfaces, fugichnia burrows	10 to 50 cm	Tractional deposition in the lower-flow regime with an aggrading component; bioturbation and reactivation surface from tide influences	

Table 2. (continued)

Facies	Lithology	Sedimentary structures	Mean bed thickness	Interpretations	Overview
SGr2	Medium-grained sand with well-rounded gravels (well-sorted)	Horizontal to cross-stratified sand, well-rounded gravels and granules, angular mud clasts	10 to 50 cm	Tractional deposition in lower-flow regime forming dunes and bars	
SGr3	Medium-grained to very coarse-grained sand (poorly sorted)	Thick (m-scale) succession of massive to crudely laminated sand, sub-horizontal or upslope-dipping laminae, plug of steep-flanked channels	1 to 5 m	Deposition from supercritical flows, antidune or humpback dunes (Fielding, 2006; Cartigny <i>et al.</i> , 2014) in channelized setting as cut-and-fill processes	
SGr4	Coarse-grained sand and well-rounded gravels and pebbles (poorly sorted)	Horizontal to large-wavelength (<3 m) undulated layer of matrix-supported gravels and coarse-grained sand (dm-scale alternation)	20 cm to 2 m	Tractional deposition in upper-flow to supercritical regime forming upper-plane beds to antidunes (Fielding, 2006)	
Conglomerate					
C1	Clast-supported well-rounded pebbles and cobbles, coarse-graine matrix, m-scale sand intraclasts (poorly sorted)	Massive to faintly laminated, poorly sorted, clast-imbrications	10 cm to 2 m	Traction carpets of supercritical flows regime, possible occurrence of hydraulic jump (Russell & Arnott, 2003)	
C2	Clast-supported rip-up clasts and gravels with a sandy matrix (unsorted)	Clast-supported angular rip-up clasts and well-rounded gravels with medium-to coarse-grained sandy matrix	2 to 20 cm	En masse (freezing) deposition from sandy debrites (Talling <i>et al.</i> , 2012)	

1 Pebble imbrications and cross-stratal dips indi-  
 2 cate N150-oriented palaeoflows.  
 3

4 *Interpretation.* The stratigraphic position of  
 5 DU1 just above the glacially abraded bedrock  
 6 and beneath distal glaciomarine muds, as well as  
 7 its lenticular geometry, indicates that it origi-  
 8 nated as an esker. Conglomeratic deposits, faint  
 9 laminations and sand intraclasts together suggest  
 10 a high-energy depositional system tied to  
 11 hydraulic jump processes (Russell & Arnott,  
 12 2003; Cummings *et al.*, 2011).  
 13

#### 14 *Depositional Unit 2: Glaciomarine mud*

15 *Description.* Depositional unit 2 (DU2) is iden-  
 16 tified from km 6 to the northern end of the pro-  
 17 file. Depositional unit 2 overlies the esker, and  
 18 its minimal thickness is 4 m over the esker  
 19 crest. It thickens laterally to reach a maximal  
 20 thickness of 12 m. Mud beds show both onlap  
 21 onto the esker body and internal onlap rela-  
 22 tionships (Fig. 2; Figure S2A). The upper  
 23 bounding surface of DU2 is conformably over-  
 24 lain by DU3.

25 Depositional unit 2 consists of faintly lami-  
 26 nated and bioturbated muds (Mu1) with scattered  
 27 granule to cobble-sized limestones. The bedding  
 28 is, in places, highlighted by rare thin (<1 cm)  
 29 sand layers (S1). Reworked shells (*Macoma bal-*  
 30 *tica*) are relatively abundant (Table 1).  
 31

32 *Interpretation.* Limestones, representing ice-  
 33 rafted debris (IRD), and faint laminations, indi-  
 34 cating deposition by settling processes from  
 35 overlying turbid plumes, are indicative of glacio-  
 36 marine mud. Thickness changes over the inher-  
 37 ited esker topography as well as internal  
 38 unconformities suggest subordinate bottom cur-  
 39 rents. The rarity of large IRD, the intense biotur-  
 40 bation and the sparse presence of thin sand  
 41 layer interpreted as turbiditic beds suggest an  
 42 ice-distal depositional environment (Boulton,  
 43 1990; Ó Cofaigh & Dowdeswell, 2001).  
 44

#### *Depositional Unit 3: Outwash fan*

*Description.* Depositional unit 3 (DU3) is recog-  
 nized in the northern half of the cross-section,  
 where it forms a significant part of the cliff  
 (Fig. 2). It conformably overlies DU2 and corre-  
 sponds to a wedge-shaped body that reaches  
 40 m in thickness. To the north, the upper  
 bounding surface of DU3 is a truncation surface  
 (base of DU5), while southwards it corresponds  
 to a conformable contact below DU4 (Fig. 2). To  
 the north (km 6.8 to 7.6), the stratal pattern is  
 dominated by the occurrence of extensive  
 mud-rich heterolithic sheets (Figure S2A) inter-  
 stratified with numerous channel structures and  
 subordinate sand sheets. Southwards (km 4.5 to  
 6.8), channel structures are absent. Instead,  
 gently sloped heterolithic sheets alternate with  
 sand sheets. Both have an average slope of 1.05  
 to 1.4% (Fig. 2).

Heterolithic sheets are mud-rich and well-stratified deposits including thin sand layers. Distinct internal facies distributions are identified in the lower and upper part of DU3. Well-laminated and occasionally bioturbated mud intervals (Mu2, 5 to 10 cm) and rippled sand layers (S1) prevail in the lower part (Fig. 5A). In the upper part, heterolithic sheets are coarser grained. The finest fraction consists of silt to very fine-grained sand facies (SS1). In places, rhythmites with contraction–dilation cycles are observed (Fig. 5F). Interstratified sand layers are more specifically characterized by trains of starved ripples (S1; Fig. 5H). Sand sheets consist of amalgamated normally graded sand beds (S2) with frequent liquefaction structures and rip-up clasts.

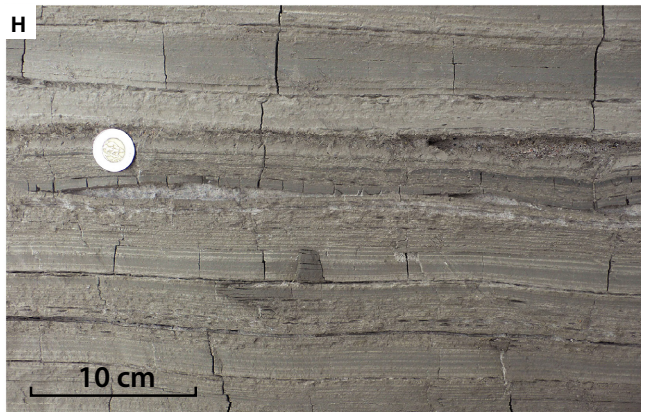
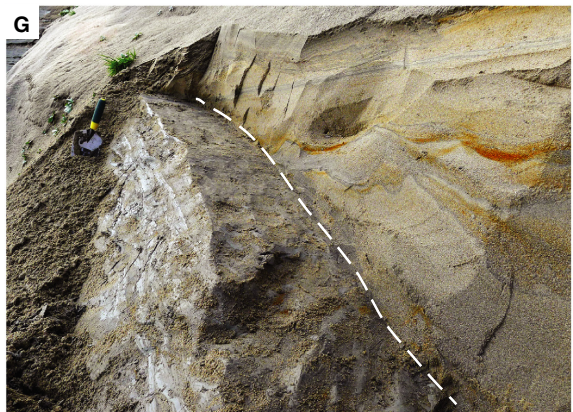
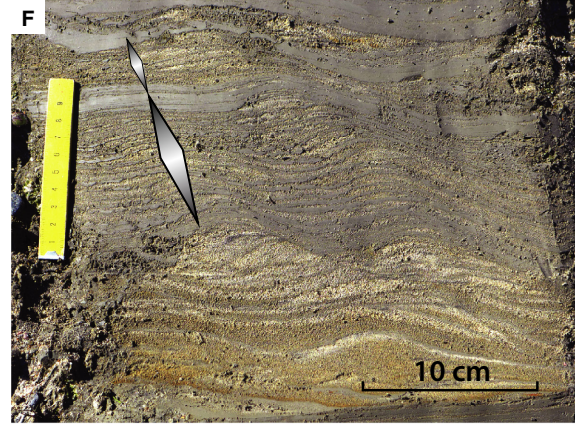
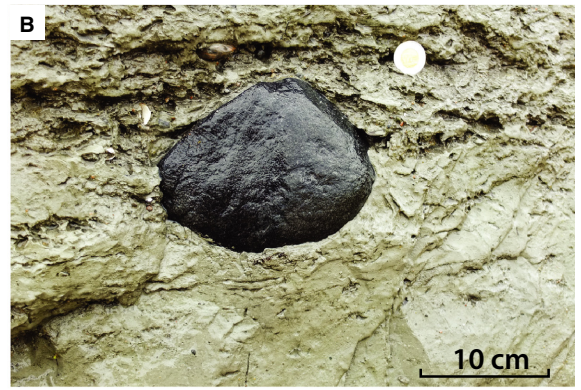
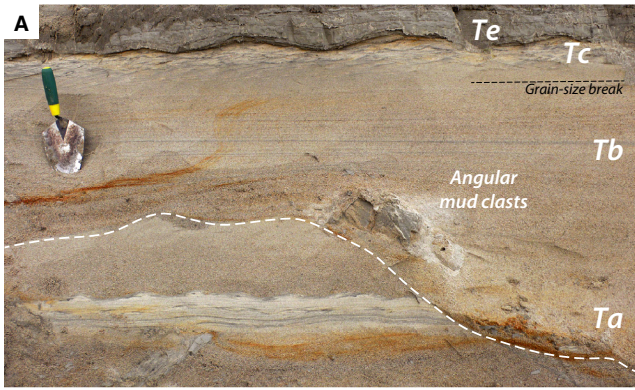
Channel structures are 10 to 200 m wide, up to 5 m deep incisions cutting in the heterolithic sheets. Each is characterized by a basal, 0.5 to 2.0 m thick, draping heterolithic unit (Mu2 or SS1), while the main infill corresponds to a sandy (S1 and S2) plug (Fig. 6A) directly onlapping the channel

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46 **Fig. 5.** Depositional facies from the distal outwash (DU3) to the upper delta slope (DU5, see location in Fig. 3).  
 47 (A) Normally graded turbiditic sand bed (facies S2) in the distal outwash fan (DU3) with the following subdivi-  
 48 sions above a basal erosive surface: T<sub>a</sub> (massive, with lithic and angular mud clasts); T<sub>b</sub> (subhorizontal lamination)  
 49 grain-size break; T<sub>c</sub> (ripples); and T<sub>e</sub> (mud drape) intervals. (B) Shell debris and a limestone (dropstone) in mas-  
 50 sive to laminated mud (Mu1) in the lower delta slope (DU4). (C) Laminated mud (Mu2) underlying a thick (>1 m)  
 51 contorted muddy layer (Mu3) that was used as marker bed in Fig. 2. (D) Lower delta slope (DU4) facies showing  
 52 alternating sand-sized (S1) and finer-grained deposits (Mu2, SS1). Details in the following images: (E) and (F)  
 53 rhythmic climbing ripples (RCRL structure) in SS1 facies; note double-mud couplets and lee-side accretion of the  
 54 sand fraction (E) and contraction–dilatation cycles (F). (G) Sharp transition (base of a U-shaped trough) between  
 55 the lower and upper delta slope (DU4 and DU5). (H) Well-laminated very fine-grained sand and silt including  
 couplets and contraction–dilatation cycles (SS1) in distal outwash fan deposits. Note the train of starved ripples.

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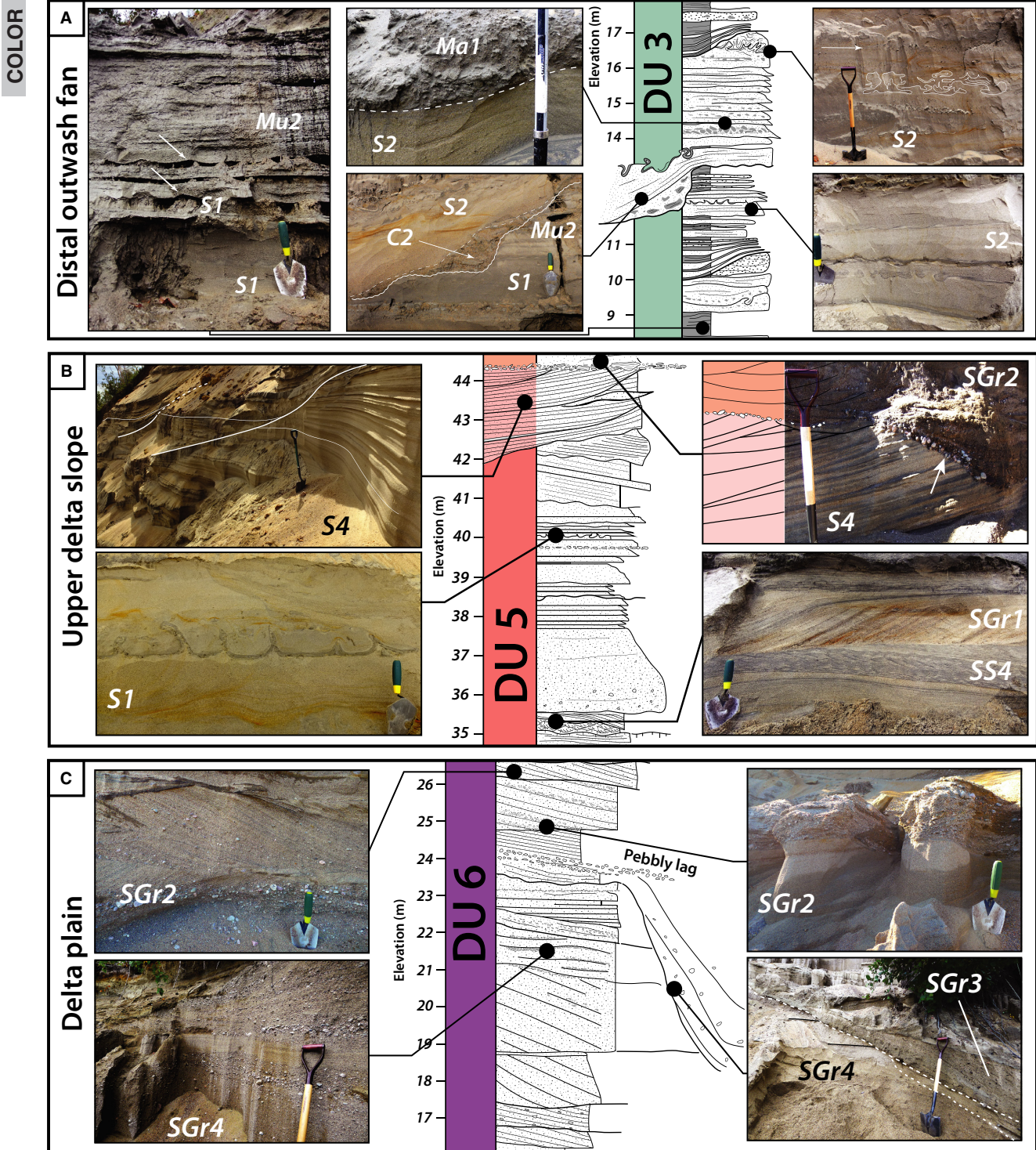


Fig. 6. Facies associations representative of the distal outwash fan (DU3) (A), the upper delta slope (DU5) (B) and the delta plain (DU6) (C). See Fig. 3 for stratigraphic position and Table 2 for facies codes.

flanks. Sandy channel plugs of some steep-flanked scours, 50 cm in depth and several metres in width, consist of normally graded sands showing in places a mixture of

immature rip-up clasts and granules (*Ma1*). In places, mud clasts are demonstrably derived from the immediate adjacent eroded mud beds (Figs 5A and 6A).

**1**

1 *Interpretation.* Along the subaqueous segment  
 2 of the outwash fan (i.e. the cliff exposure), and  
 3 in contrast to the underlying glaciomarine muds,  
 4 DU3 is notably devoid of IRD. It indicates the  
 5 demise of the glaciomarine depositional system  
 6 and suggests that the glaciomarine ice front  
 7 turned into a continental front (e.g. Lønne, 1995;  
 8 Lønne & Nemec, 2004). Depositional channel  
 9 and sheet geometries and depositional facies  
 10 together conform to a turbiditic system (Table 2).  
 11 The number and the limited size of the channels  
 12 suggest a network of relatively small channels  
 13 rather than a trunk channel. Sand sheets represent  
 14 lobe deposition, while heterolithic sheets  
 15 interfingered with either channel or lobe deposits  
 16 represent interchannel to interlobe deposits.  
 17 A channel distribution limited to the north and  
 18 prevailing gently sloped sand sheets downslope  
 19 to the south suggest a channel-lobe transition.  
 20 The topography inherited from the esker, an  
 21 expression of which was maintained during the  
 22 deposition of DU3 (Fig. 2), controlled at least in  
 23 part the position of this channel-lobe transition.  
 24 Autocyclic avulsions in the active channel-lobe  
 25 complexes probably resulted in the alternation  
 26 through time of heterolithic sheets and channels,  
 27 as well as sand and heterolithic sheets, upslope  
 28 and downslope, respectively. Similar high-density  
 29 turbiditic facies (S2) observed in both sand  
 30 sheets (lobes) and channels suggest that some of  
 31 the channels were plugged by backstepping lobe  
 32 deposits (e.g. Turmel *et al.*, 2015). Narrow scours  
 33 including mud clasts of very local origin suggest  
 34 hydraulic jump processes in agreement with  
 35 depositional conditions at the channel-lobe transi-  
 36 tions (Garcia & Parker, 1989; Kostic & Parker,  
 37 2006; Postma *et al.*, 2009; Postma & Cartigny,  
 38 2014). Heterolithic sheets were formed by the  
 39 incremental accretion of low-density turbidites  
 40 likely originating from overspilling and/or strip-  
 41 ping flows (e.g. Fildani *et al.*, 2006; Kane *et al.*,  
 42 2010).

23  
 43 The outwash fan deposits present in the cliff  
 44 are inferred to be the distal counterpart of the  
 45 fan-shaped depositional system recognized at  
 46 the outlet of the Sault-au-Cochon Valley at  
 47 140 m.a.s.l. (Cousineau *et al.*, 2014; Figure S1).  
 48 Here, ice-contact landforms (see above) and con-  
 49 glomerates (C1) including boulders and sand  
 50 intraclasts suggest flow deconfinement processes  
 51 at the outlet of a subglacial conduit over a sub-  
 52 aerial or shallow outwash plain (Cutler *et al.*,  
 53 2002; Russell *et al.*, 2009) overhanging the  
 54 downslope finer-grained subaqueous outwash  
 55

fan; depositional depths of the latter range from  
 140 to 80 m.

## Depositional system II: Open sea deltaic deposits

There is a demonstrably depositional continuum from DU4 to DU6. An upslope facies transition along clinof orm geometries is particularly well-expressed from DU4 to DU5 (Fig. 2; km 3.2 to 2.4), the latter being the proximal counterpart of the former. More upstream, a transition from DU5 to DU6 does exist, but is expressed less clearly. This depositional system is thus understood as a suite of depositional environments distributed along deltaic clinof orms running from the delta plain (DU6) down to the lower delta slope (DU4). The elevation difference between the lowest visible deposits of DU4 and the highest deposits of DU6 is 90 m (Fig. 2). It provides a minimal estimate for the vertical development of the deltaic clinof orms.

### *Depositional Unit 4: Lower delta slope*

*Description.* Depositional unit 4 (DU4) constitutes a major component of the cross-section extending throughout most of the cliff (Fig. 2), but it is a subvertical and poorly accessible portion of the cross-section (Figure S2). Depositional unit 4 consists of a relatively thick (<40 m) wedge of mud-rich heterolithic facies including some subordinate metre-thick sand units. Its lower bounding surface is conformable on the underlying distal outwash fan deposits. Depositional unit 4 pinches out to the north as it is truncated by DU5 (upper delta slope). Southwards, its upper bounding surface appears as a composite contact (Fig. 2): from km 7.0 to 3.2, U-shaped troughs (Fig. 5F) alternate with paraconformable portions; from km 3.2 to 2.4, lateral facies transitions develop along clinof orms (Figs 2 and 5G). A large-scale well-defined channel-levée structure is observed from km 1.4 to 2.1 (Fig. 2). 24

Heterolithic deposits consist of horizontally laminated mud (Mu2) and sand layers (S1) (Fig. 5D). Sand occurs as millimetre to centimetre-thick beds of continuous or starved ripple trains. From km 4 to 7, relatively sand-rich heterolithic beds include rhythmic climbing-ripple cross-lamination (RCRL, *sensu* Choi, 2010) in several metre-thick, uninterrupted, depositional successions (SS2, Fig. 5D and E). Occasional contorted intervals are observed (Mu3), the

1 thickest (1 m) being used as a marker bed  
 2 (Figs 2 and 5C). Interstratified sand units consist  
 3 of amalgamations of 10 to 30 cm thick normally  
 4 graded sand beds (S2).

5 The channel-levée structure, 20 m deep in the  
 6 thalweg axis (Fig. 2; Figure S2B), forms an  
 7 asymmetrical feature, including a well-defined  
 8 levée to the south that bears beds dipping  
 9 towards the channel axis (Figure S2b). The  
 10 opposite flank is formed by descending strata  
 11 constituting the bulk of DU4. In the levée depos-  
 12 its, the relative amount of sand – essentially  
 13 contained in starved ripple trains flowing away  
 14 from the channel axis – rapidly decreases away  
 15 from the thalweg. X-ray imagery (Figure S4)  
 16 shows inverse and normal grading in the verti-  
 17 cal stacking of millimetre to centimetre-scale  
 18 mud laminae.

19 At the southern tip of the cliff, the outcrop-  
 20 ping portion of DU4 is only 3 m thick. In the  
 21 lower portion, faintly laminated mud (Mu1) con-  
 22 tains abundant shells, plant debris (wood frag-  
 23 ments and pine needles) and some gravel to  
 24 pebble-sized limestones (Fig. 5B). The upper  
 25 portion shows intensively bioturbated deposits  
 26 including an increasing proportion of normally  
 27 graded sand (S1) interstratified with laminated  
 28 muds (Mu2).

29  
 30 *Interpretation.* Mud-dominated deposits, gently  
 31 sloped depositional geometries (dips  $<3^\circ$ )  
 32 beneath and beyond the sandier deposits of  
 33 DU5, indicate that DU4 encompasses lower  
 34 delta slope sediments. Older DU4 horizons at  
 35 the base of the cliff were penecontemporaneous  
 36 of a delta plain not exposed in the cross-sec-  
 37 tion – at an elevation comprised between 90 m  
 38 and 140 m.a.s.l. – while the upper horizons  
 39 were penecontemporaneous with delta plain  
 40 deposits constitutive of DU6. The minimal  
 41 depositional bathymetry was in the 30 to 90 m  
 42 depth range.

43 Deposition from low-density turbidity currents  
 44 (Mu2, S1) prevailed. Occasional subordinate  
 45 deposition from high-density turbidity currents  
 46 occurred that splayed from the upper delta slope  
 47 (DU5) over the lower delta slope forming tur-  
 48 biditic lobes. The RCRL suggests a tidal influ-  
 49 ence in spite of the relatively deep depositional  
 50 bathymetry. Fluviotidal interactions in mouth  
 51 environments (Gilbert, 1983; Cowan *et al.*, 1998;  
 52 <sup>25</sup> Ayranci *et al.*, 2012), the influence of internal  
 53 tides (Pomar *et al.*, 2012; see Saucier & Chassé,  
 54 2000; Bourgault *et al.*, 2014; Normandeau *et al.*,  
 55 2014 for internal tides in the St. Lawrence

Estuary) or daily discharges cycles in meltwater  
 flows may form structures very similar to the  
 RCRL, which are usually recognized in intertidal  
 environments (Choi, 2010).

In addition to the relatively unconfined tur-  
 bidites interpreted above, the channel-levée  
 structure provides evidence for the occurrence  
 of a confined sediment pathway. The position  
 and the asymmetry of the channel-levée suggest  
 a meandering channel (Babonneau *et al.*, 2010)  
 at the base of delta slope, near a transition with  
 prodelta environments. Inverse and normal grad-  
 ing in heterolithic levée deposits (Figure S2)  
 suggests river-derived hyperpycnal flows (Mul-  
 der & Syvitski, 1995). Levée accretion, including  
 sand layers without in-channel sand aggrada-  
 tion, indicates the bypass of sandy turbidity  
 flows. Beds dipping towards the thalweg axis  
 relate to inner levée, *sensu* Kane *et al.* (2007).  
 Radiocarbon dating (Fig. 2; Figure S2B) shows  
 that a significant portion of the channel-levée  
 structure was built in 100 to 200 years. The bas-  
 inward prolongation of such a channel is pre-  
 served as a buried channel on the Laurentian  
 Shelf (Normandeau *et al.*, 2015; their fig. 9).

To the south, faintly laminated mud (Mu1),  
 intensive bioturbation, limited sand deposition  
 and reduced aggradation rates (*ca* 0.6 cm-year<sup>-1</sup>  
 inferred from radiocarbon dating, Fig. 2) in dis-  
 tal DU5 deposits indicate prodelta environ-  
 ments. Whether limestones (Fig. 5B) were  
 derived from sea ice or from icebergs calved in  
 adjacent areas (e.g. the mouth of the Saguenay  
 Fjord, Fig. 1B) remains unclear, their apparent  
 relative abundance being preferentially inter-  
 preted as the result of reduced aggradation rates.

#### *Depositional Unit 5: Upper delta slope*

*Description.* Depositional Unit 5 (DU5) com-  
 poses the most important sand wedge of the  
 cross-section with a mean thickness of 30 m. Its  
 lower bounding surface on DU4 is marked by U-  
 shaped troughs up to 20 m in depth and 250 m  
 in width (e.g. km 3.4 and km 4.5), which alter-  
 nate with conformable segments (Fig. 2). A clear  
 facies transition (km 2.5 to 3.3) is observed from  
 DU4 to DU5. The upper bounding surface of  
 DU5 is a well-defined truncation surface corre-  
 sponding to the base of DU8 (nearshore sand  
 sheet).

Depositional unit 5 consists of offlapping  
 sand-rich clinothems. The infill of basal U-  
 shaped troughs consists of horizontal to gently  
 sloped beds ( $<1^\circ$ ) onlapping on trough flanks.  
 They correspond to the lower segments of the

1 clinothems (e.g. km 4.5). Upslope, clinofolds  
 2 locally include small-scale internal channel-like  
 3 incisions (10 to 20 m wide, 3 to 5 m deep).  
 4 Their dips progressively increase up to 6°,  
 5 reaching in places 17° (km 6.3). Small-scale  
 6 geometries similar in appearance to second-  
 7 order clinofolds, but that show internal intri-  
 8 cate onlapping patterns, are observed in the  
 9 uppermost part of the cliff (km 4.5 to 7.2;  
 10 Fig. 6B).

11 From the base to the top, a gradation in depo-  
 12 sitional facies is observed in DU5 (Fig. 3). The  
 13 infill of the basal U-shaped troughs consists of  
 14 vertically stacked, up to 50 cm thick, normally  
 15 graded sand beds (S2) and occasional debrite  
 16 beds (Ma1). Both beds include abundant angular  
 17 mud clasts and well-rounded lithic granules.  
 18 These beds generally show facies sequences lim-  
 19 ited to a superposition of massive and laminated  
 20 subdivisions (top cut-out turbidites). Upwards  
 21 and above the trough interfluvies, sand beds pro-  
 22 gressively thin and contain less abundant lithic  
 23 clasts; debrite beds became rare then absent.  
 24 From the middle of the cliff up to 15 m beneath  
 25 the basal truncation of DU8, sand beds coarsen  
 26 and thicken upwards, reaching one metre in  
 27 thickness. Meanwhile, mean stratal dips  
 28 increase upwards to clinofold dips (6 to 17°).  
 29 Sand beds contain an increasing upwards pro-  
 30 portion of well-rounded gravels, and their basal  
 31 surfaces show progressively better defined scour  
 32 structures. Bioturbated intervals of rippled fine-  
 33 grained sand (SS4) associated with sigmoidal  
 34 cross-strata (SGr1) are locally interbedded  
 35 (Fig. 6B). The uppermost part of DU5 includes  
 36 channelized backsets overlain by low-angle  
 37 cross-strata backstepping against southward-dip-  
 38 ping, regularly spaced (*ca* 10 m), concave-up  
 39 erosion surfaces with 'pseudo-clinofold' geome-  
 40 try. While backsets consist of massive to faintly  
 41 laminated gravelly sands (SGr3), backstepping  
 42 strata consist of a stack of massive to laminated  
 43 decimetre-thick sand beds including subordinate  
 44 layers of muddy sand (S4) (Figs 3 and 6B).

46 *Interpretation.* Based on well-defined clinofold  
 47 geometries (Fig. 2), DU5 is interpreted as being  
 48 upper delta slope deposits. Considering that  
 49 DU6 (delta plain) is more or less coeval with the  
 50 youngest DU5 and DU4 strata (see DU4), deposi-  
 51 tional bathymetries for DU5 are in the 20 to  
 52 60 m range, reaching 80 m regarding the lower-  
 53 most infill of the U-shaped troughs. As shown  
 54 by decreasing dip of strata from the top to the  
 55 base of DU5, the delta slope depositional profile

was concave-up, passing downslope asymptoti-  
 cally to the mud-dominated deposits of DU4.

Normally graded and laterally continuous  
 sand beds observed throughout DU5 are inter-  
 preted as turbiditic beds, essentially unconfined  
 above the U-shaped troughs. The infill of the U-  
 shaped troughs resulted from confined turbidity  
 flows that underwent hydraulic jump processes  
 as indicated by massive and/or faintly laminated  
 intervals (Postma *et al.*, 2009). The association  
 of rounded lithic and angular mud clasts indi-  
 cates that these deposits mixed far-travelled sed-  
 iments and clasts of local origin, the latter  
 resulting from the excavation of surrounding  
 beds. Upwards, and in interfluvies of U-shaped  
 troughs, deposition was tied to high-density tur-  
 bidity currents. A progressive increase in sedi-  
 ment concentration and flow competence is  
 indicated by the coarsening-upward trend, the  
 increasing content in lithic clasts and better  
 developed basal scours. The intervening biotur-  
 bated sigmoidal cross-strata were probably  
 emplaced during a short-lived interval of pre-  
 dominant tidal processes during deltaic accre-  
 tion.

The U-shaped troughs are interpreted as hav-  
 ing been formed at the break-in-slope marking  
 the transition between the upper and lower  
 delta slope. These U-shaped troughs may have  
 been formed by turbidity flows experiencing  
 hydraulic jumps due to slope reduction (e.g.  
 Massari & Parea, 1990; Breda *et al.*, 2007; Gobo  
*et al.*, 2014). Pseudo-clinofold geometries and  
 channelized backsets observed at the top of DU5  
 typify cyclic step deposits (Cartigny *et al.*, 2011;  
 Postma & Cartigny, 2014; Dietrich *et al.*, 2016).

In the upper delta slope, the unconfined tur-  
 biditic beds along the upper part of the clin-  
 othem contrast with deposits infilling U-shaped  
 troughs. Unconfined deposition is thought to  
 reflect background sedimentation when sands  
 were essentially deposited by sheet-like tur-  
 biditic flow events. Background sedimentation  
 was interrupted episodically by high-energy  
 events characterized by more dynamic events  
 forming channelized conduits along the delta  
 slope. The channelized conduits were possibly  
 connected downslope to a channel-levée struc-  
 ture similar to that described in DU4 and/or to  
 sandy depositional lobes identified in the lower  
 delta slope (see DU4).

#### *Depositional Unit 6: Delta plain*

*Description.* Depositional unit 6 (DU6) is iden-  
 tified at the top and northernmost part of the

1 cliff (km 7.0 to 7.5), immediately upstream of,  
 2 and above DU5 (Figs 2 and 4A; Figure S2A). It  
 3 consists of a 15 to 20 m thick wedge showing  
 4 subhorizontal bedding. The basal contact with  
 5 the underlying upper delta slope is transitional  
 6 in terms of its depositional facies. The upper  
 7 surface of DU6 defines the flat top of the cliff at  
 8 90 m.a.s.l. Depositional unit 6 extends to the  
 9 north under the forest cover. It pinches out  
 10 southwards over a short distance (km 7).  
 11 Although the contact with DU8 does not crop  
 12 out, a truncation surface at the base of the near-  
 13 shore sand sheet is most likely.

14 Two vertically superimposed coarse-grained  
 15 facies assemblages comprise DU6. The lower  
 16 facies, 5 m in thickness, consists of thick  
 17 (>1 m), erosion-based, subhorizontal beds of  
 18 gravelly sands (SGr4) (Fig. 6C). Bed contacts  
 19 and laminations show undulations, 5 to 10 m in  
 20 wavelength. Internal erosion surfaces define  
 21 poorly circumscribed steeply flanked scours  
 22 filled by massive to faintly laminated gravelly  
 23 sand (SGr3). The upper facies assemblage, trun-  
 24 cating the underlying one along a subhorizontal  
 25 surface underlined by pebbly lags, consists of a  
 26 fining-upward and thinning-upward succession  
 27 of coarse-grained, cross-stratified sand beds  
 28 (SGr2) (Fig. 3). Coarse-grained, planar cross-stratified  
 29 sands containing abundant well-rounded  
 30 gravels and pebbles pass upwards into lenses of  
 31 medium-grained planar cross-stratified sands  
 32 interbedded with thin (<20 cm) and laterally  
 33 discontinuous mud layers. Rare decimetre-scale,  
 34 angular mud clasts are scattered within cross-  
 35 stratified sand. Palaeocurrent measurements  
 36 indicate both westward and eastward flows.

37  
 38 *Interpretation.* The facies transition from the  
 39 underlying steeply sloped clinothems (DU5) to  
 40 subhorizontal stratal patterns indicates that DU6

represents the correlative delta plain, which was  
 separated from the delta slope by a topographic  
 break-in-slope (delta brink) reflected by the dras-  
 tic change in dip patterns. Large undulations in  
 the lower facies assemblage of DU6 strongly sug-  
 gest supercritical flows conditions (Fielding,  
 2006). In contrast, the upper facies assemblage  
 related to bedload processes formed in relatively  
 low-energy environments. A fluvial-dominated  
 setting is inferred for DU6 as relict braided  
 channels observed in plan view (Figure S1),  
 although not effectively connected to DU6, offer  
 the map view expression of such delta plain  
 environments. However, a tide-influenced set-  
 ting is also suggested by current reversals.

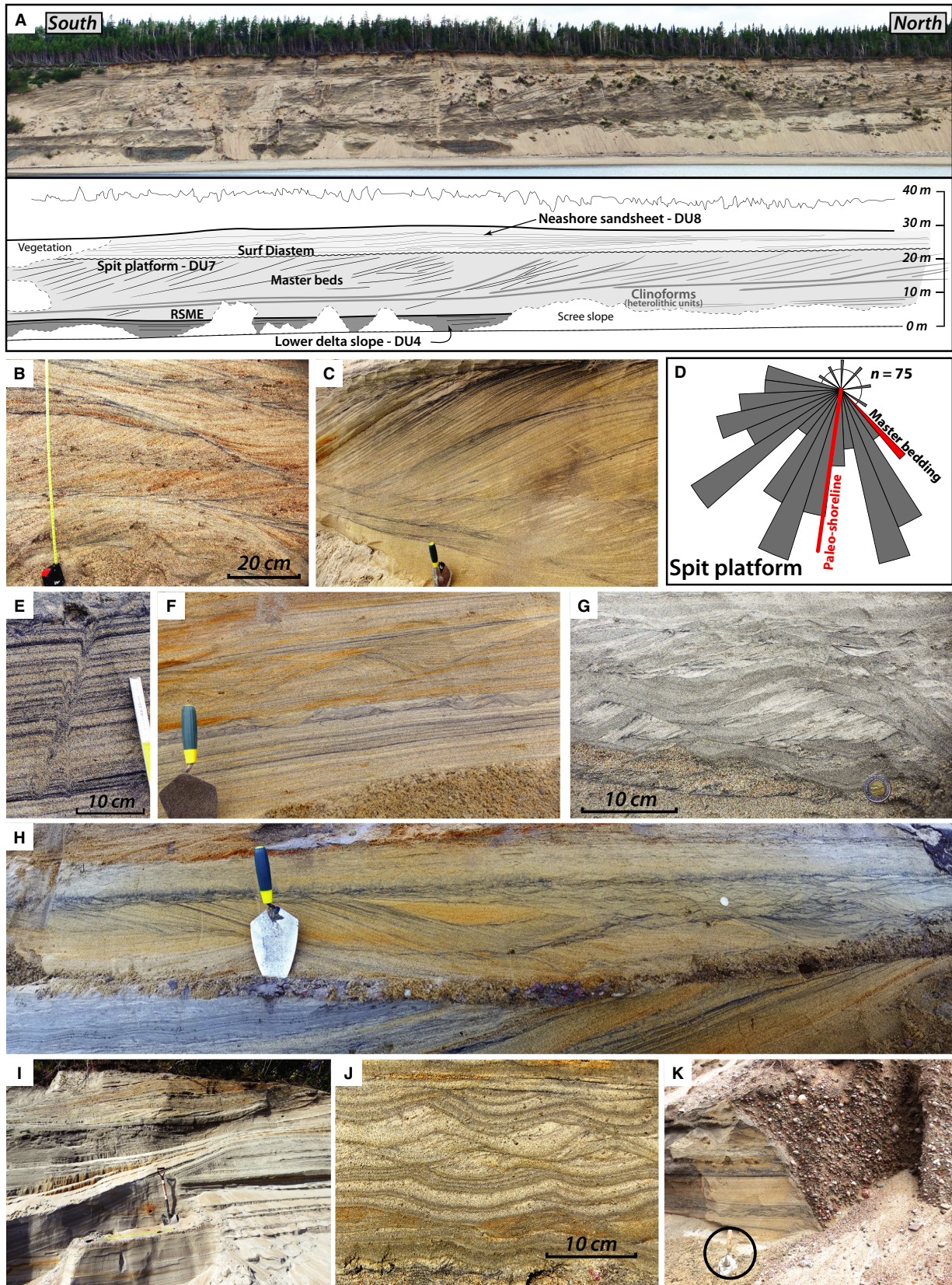
The vertically superimposed lower and upper  
 facies assemblages relate to spatially juxtaposed  
 depositional environments, which are ascribed  
 here to a subtidal and intertidal delta plain,  
 respectively. The change from a subcritical to  
 supercritical hydraulic regime from the inter-  
 tidal to subtidal delta plain requires a parameter  
 that allows the kinetic energy of the flow to  
 increase downstream in the absence of slope  
 increase. A 'tidal drawdown' process, acting as a  
 flush during ebb tide (Smith *et al.*, 1990; Diet-  
 rich *et al.*, 2016), may explain such a paradoxi-  
 cal transition. The channel structure filled by  
 SGr3 facies (Fig. 6C) probably breached the delta  
 brink, feeding directly the upper delta slope.

### Depositional system III: Coastal deposits

The third depositional system is comprised of  
 DU7 and DU8. Depositional unit 8 forms a later-  
 ally continuous sand sheet at the top of the cliff,  
 atop of which are the coastal landforms  
 described above (Fig. 1; Figure S1). Successively  
 truncating DU6, DU5 and DU7 from north to  
 south, it is a diachronous unit following step by

42  
 43 **Fig. 7.** Architecture and facies of the spit platform and mouth-related deposits (DU7). See Table 2 for facies  
 44 codes. (A) Photomosaic and line drawing (km 0.4 to 0.7; location in Fig. 3) highlighting clinofolds and master  
 45 beds, both dipping to the south. The basal surface of DU7 is ascribed to a diachronous erosion surface (RSME,  
 46 ravinement surface of marine erosion). The upper part of the cliff corresponds to the nearshore sand sheet (DU8,  
 47 see also Figure S5); its base is a surf diastem. (B) Bioturbated sigmoidal cross-strata with reactivation surfaces and  
 48 heavy-mineral laminae. (C) Cross-stratified sands (SGr2) constituting the master beds of the shoreface sand wedge.  
 49 (D) Palaeocurrents (cross-stratal dips) are almost perpendicular to the master bed dips and almost parallel to  
 50 inferred orientation of the palaeoshoreline. (E) Escape burrows (*fugichnia*) and heavy-mineral-rich laminae from  
 51 tidal bars illustrated in (B) (SGr1). (F) Well-defined, subhorizontal to undulating (hummocky-like) laminations in  
 52 sands (SS5) and rippled sands (SS4) in storm-related clinofolds. (G) Fine-grained and muddy sands in combined-  
 53 flow ripples (SS4). (H) Gravel lag at the interface (surf diastem) between DU7 and DU8. (I) Turbiditic sand beds  
 54 (S1) in aggrading channels and including and oxidized wood debris. (J) Pseudoclimbing ripples in heterolithic  
 55 facies including currents reversal (SS1). (K) Steep-flanked scour with a massive to faintly laminated coarse-grained  
 infill (SGr3) severely truncating underlying deposits (trowel for scale).

COLOR



step the falling sea level. In contrast, DU7 only characterizes the lowest and southernmost segment of the cross-section. It represents late deposits coeval with low RSL.

#### *Depositional Unit 7: Spit platform and mouth-related deposits*

*Description.* Depositional unit 7 (DU7) crops out at the southern tip of the cliff (km 0 to 2.7; Figs 2 and 7). It forms a well-individualized sand wedge pinching out northwards against DU5. The lower bounding surface is sharp but paraconformable onto prodelta deposits (lower part of DU4). It is truncated by the overlying nearshore sand sheet (DU8). The maximum observable thickness of DU7 is 35 m (Figs 2 and 7).

Depositional unit 7 mainly shows large-scale clinoform geometries and superimposed smaller-scale master beds (Figs 2 and 7A). Gently sloped (<2°) clinoforms that are underlined by 1 to 2 m thick, distinctively finer-grained deposits truncate steep-sloped (10 to 12°), coarse-grained master beds and downlap at the kilometre-scale onto DU4. Some channel structures locally depart from this well-organized clinoform/master bed system (Figs 2 and 7). They form either small-scale 'scour and fill' structures (km 0.8, km 1.5) or the plug of the large channel-levée structure described in DU4.

The bulk of DU7 consists of planar and trough cross-stratified, medium-grained to coarse-grained sands (SGr2, Fig. 7C) that form the steeply dipping master beds. In a few places, cross-strata correspond to metre-scale sigmoidal laminae sets bounded by reactivation surfaces highlighted by thin heavy-mineral laminae (SGr1, Fig. 7C) and burrows (mainly escape structures: *fugichnia*, Fig. 7E). Dips of cross-strata evidence a wide spectrum of palaeocurrents from the south-east to the north-west (Fig. 7D). A mean trend to the south-west, however, suggests palaeocurrents essentially transverse to the master beds (dipping up to 12° to the south-east). Draping the basal erosion surface defining each of the clinoforms, associated finer-grained deposits are made up of a specific heterolithic facies assemblage showing an alternation of fine-grained rippled sand and muddy drapes (SS4), with HCS-like undulating lamination highlighted by well-defined heavy-mineral-rich stripes (SS5, Fig. 7F).

Scour and fill structures show a variety of grain sizes and internal stratal patterns. At km 1.5, a large scour is filled by very thick (1 to

5 m) massive to faintly laminated sand beds (SGr3). They are characterized by abundant, decimetre-scale, angular rip-up clasts, which progressively decrease upwards in size and abundance. Above, well-stratified, fine-grained and normally graded sand beds (S1) are genetically linked with smaller-scale scours. Sand beds are 10 to 20 cm thick, include abundant wood debris (oxidized) and are separated by thin (<1 cm) mud intervals commonly associated with load and flame structures (Fig. 7I). In places, RCRL (SS2) is interfingered (Fig. 7J). Linked scour structures, 10 m wide and 3 m high, truncate the underlying beds but are, however, conformably draped by normally graded beds very similar to those that have been truncated. The basal erosion surface grades laterally into conformable contacts (Fig. 7I). Last, a third type of scour and fill structure is observed at km 0.7. Here, steep-flanked, very narrow scours (Fig. 2) are filled with pebbly sands, either massively or faintly laminated, including backsets (SGr3, Fig. 7K).

*Interpretation.* In a coastal setting, cross-strata associated with inclined master beds indicate either the lateral migration of a sandspit system or the frontal accretion of a delta. However, transverse palaeocurrent orientations rather suggest deposition from bars and dunes driven by diffracted longshore currents as evidenced by the distribution of palaeocurrents, conforming to a spit system (Fig. 7D). At the top of the cliff, coastal landforms and the absence of fluvial channels corroborate such an interpretation (Fig. 1; Figure S1). Coarse-grained, bioturbated, sigmoidal cross-strata suggest the influence of tidal processes down to the lower parts of the clinoforms (depositional depth >15 m). More specifically, DU7 is interpreted as a spit platform, *sensu* Meistrell (1972). A spit platform serves as a basement for a sandspit when the latter is prograding in a relatively deep (i.e. tens of metres) setting (Shaw & Forbes, 1992; Nielsen & Johannessen, 2009; Zecchin *et al.*, 2010). In the present case study, it can be argued that the spit platform inception occurred from the point at which the depositional depth, essentially inherited from the depositional profile of the former delta, increased significantly from a few metres to at least 30 m (Fig. 2, km 2.6). The spit itself (DU8) was then emplaced onto this platform.

The south-westwardly accretion of bars and dunes was recurrently interrupted by erosional

1 events resulting in the development of the larger-scale gently sloped clinofolds. Erosion surfaces cut into the shoreface environment, and the overlying occurrence of HCS-like laminations in the draping heterolithic deposits suggest storm events. The downlap surface of the clinofolds is understood as a regressive surface of marine erosion (Plint & Nummedal, 2000; Catuneanu, 2002; Swift *et al.*, 2003). In this interpretation, DU7 corresponds to the forced regressive shoreface sand wedge of Swift *et al.* (2003).

2 Subordinate deposition from gravity flows is identified in the infill of channels and scour and fill structures. Well-stratified and normally graded beds are interpreted as low-density gravity flow deposits derived from hyperpycnal currents. The abundance of terrestrial organic matter debris (Zavala *et al.*, 2010) and the lack of any observed slump scars (Plink-Björklund & Steel, 2004) together suggest riverine parent outflows. In places, RCRL-like structures (SS1) strongly suggest the influence of tidal processes. Narrow scour structures and the associated massive to faintly laminated deposits, including immature mud clasts and/or backsets, are interpreted as deposits related to hydraulic jump processes (Russell & Arnott, 2003; Postma & Cartigny, 2014). In this position, such processes may have eroded DU4 at the depth of 15 m, producing abundant angular mud clasts that were rapidly redeposited.

3 The restricted lateral extent of channel and scour infills as well as specific grain-size features suggests that gravity flow deposits in DU7 depart from the overall spit platform system. It is interesting that, in map view, channel and scours are located basinward of the large abandoned fluvial meander at 65 m.a.s.l. (Figure S1). Such a palaeogeographic location and depositional facies together suggest that related deposits were genetically linked with former river mouths. The interfingering of facies assemblages demonstrates that spit platform and mouth deposits were essentially coeval, the latter being deposited within localized incisions cut in the spit platform, with a subordinate contribution to the total sediment volume.

#### 49 *Depositional Unit 8: Nearshore sand sheet*

50 *Description.* Depositional Unit 8 (DU8) is a laterally extensive sand sheet atop the cliff (Figs 2 and 7A; Figure S5). It continuously runs from km 0 to 7.1 but is absent in the northernmost segment of the cliff at elevations higher than 85 m. Usually, its thickness is 6 to 8 m, but

reaches up to 15 m in some places. The basal bounding surface is a composite truncation surface that shows a suite of stair steps, each connecting with downstepping scarps identified in map view (Figure S1). Individual erosion surfaces are concave upwards, steep-sloped ( $>15^\circ$ ) in their upper part (Figure S2A). The tangential profile progressively flattens to the south eroding the underlying deposits. The erosion is highlighted by a lag deposit including basement gravels to rare cobbles (Fig. 7H). Thus, the basal bounding surface is the juxtaposition of the lower segment of successive, individual concave-upward erosional profiles. The upper bounding surface of DU8 is more or less parallel to its basal bounding surface reproducing the stair morphology.

26 Depositional unit 8 shows horizontal to sub-horizontal bedding. Intermediate, small-scale, gently dipping ( $<1^\circ$ ) clinofolds are identified in places, downlapping on the composite basal erosion surface. Three facies assemblages are generally distinguished from the base to the top of DU8. The lower facies assemblage consists of trough cross-stratified coarse-grained sand and gravel (SGr2, Fig. 7H). A bimodal palaeocurrent trend is evidenced showing north-eastward and south-westward orientations. In places, laminated to rippled fine-grained sand with heavy-mineral laminae and bioturbation (escape burrows, SS4) are present. In the middle part of DU8, the facies assemblage consists of well-sorted and well-laminated sand (S3) including metre-scale shallow scours. In places, heavy minerals are distinctively well-segregated, forming placers up to 20 cm in thickness (Figure S5C). Genetically associated with tangential erosion surfaces, decimetre-scale angular sand intraclasts show evidence of rotational slides (Figure S5B) and thin ( $<10$  cm) contorted beds. In the upper facies assemblage of DU8, heavy minerals are more diffusely distributed in evenly and monotonously laminated sands (Figure S3D).

*Interpretation.* Depositional unit 8 deposits are genetically associated with raised beach ridges and associated landforms and thus constitute a nearshore sand sheet. Truncation of progressively younger underlying deposits towards the south, together with decreasing elevation of DU8 deposits, is characteristic of diachronous sedimentation during forced regression and glacio-isostatic rebound deposits (Boulton, 1990; Dionne *et al.*, 2004; Fraser *et al.*, 2005;



Helland-Hansen & Hampson, 2009; Hein *et al.*, 2014).

More precisely, the lower facies assemblage represents surf zone deposits (Clifton *et al.*, 1971), emplaced above a surf diastem *sensu* Swift *et al.* (2003) (or wave erosion surface *sensu* Fraser *et al.*, 2005; their fig. 20), which is highlighted at this site by lag deposits. Sediment contribution to shore accretion was probably related to longshore bar migration that may have generated the gently dipping clinoforms. Note that the dips of related structures are only apparent. True dips are expected to be slightly steeper, considering that the cliff intersects the coastal landform suites at an angle of *ca* 30° (Figs 1 and 2). The middle facies assemblage represents foreshore sands including swash-backwash laminae (Clifton *et al.*, 1971) and placers (Gallaway *et al.*, 2012). Shallow scours are interpreted as ebb tide rip channels, incised within berms and/or longshore bars. At the top, the upper facies assemblage essentially represents aeolian deposits characterized by the absence of segregation of the heavy minerals.

Internal concave-upward erosional profiles are interpreted as coastal notches (Erikson *et al.*, 2007; Anthony, 2008), that is preserved bluffs subsequently left behind by the overall regressive trend. Oversteepening of the foreshore profiles resulted in gravity slides and/or in the formation of sand intraclasts, suggesting moisture cohesion or frozen sands (Runkel *et al.*, 2010). Although no evidence for the downslope continuation of the concave-upwards erosional profile has been clearly documented owing to subsequent surf diastem erosion, it is tentatively proposed that notches in the foreshore were formerly connected in shoreface environments to the large-scale clinoforms described in DU7 deposits (Fig. 7A). Erosion surfaces extending from the foreshore to the shoreface may relate to particularly severe storm events (Erikson *et al.*, 2007) or to more lengthy periods of enhanced coastal erosion.

### **In-valley depositional system**

Inland, exposures in the bedrock-confined Portneuf and Sault-au-Cochon valleys do not have the appreciable continuity of the cliff cross-section. Their integration in the stratigraphic framework is, however, crucial for the understanding of the overall deglacial sequence, because they allow the definition of a fourth, in-valley depositional system that records an earlier phase of

progradation relative to that documented along the cliff in depositional system II (Fig. 4). The fjord head delta and moraine-dammed lacustrine delta facies associations are recognized in the Portneuf and Sault-au-Cochon valleys, respectively.

### *Fjord head delta*

*Description.* In the Portneuf Valley, 15 km away from the cliff, the present-day river exposes a coarsening-upward, 50 m succession, at elevations ranging from 60 to 110 m (Figure S6). Here, three facies assemblages are delineated, successively dominated by mud (lower unit), sand (middle unit) and pebbles (upper unit, Figure S6A).

By comparison with facies associations documented in the cliff cross-section (Fig. 2), the overall succession appears as a variant of the DU4 to DU6 deltaic progradation. The lower unit, 7 m thick, is primarily composed of thin mud layers (Mu2) interstratified with thin (<30 cm) intervening sand beds (S1). Up to 1 m thick, contorted intervals are observed (Mu3, Figure S6F). The middle unit, 40 m thick, shows a coarsening-upward trend. Its lower portion is similar to the underlying lower unit but includes thick normally graded sand beds (S2) that contain abundant angular mud clasts. Ubiquitous rhythmical stratal patterns are observed (Figure S6D and E). Above, mud intervals disappear and turbiditic sand beds (S2) are stacked together. Bed interfaces ubiquitously show load casts and flame structures. Bioturbation also occurs in places. Syn-sedimentary downstepping extensional fractures are abundant and a several metre-long intraformational shear zone (subhorizontal décollement surface) was identified (Figure S4C). In the upper part of the middle unit, large-scale undulating and normally graded sand beds having a conspicuous rhythmical stratal pattern (S4, Figure S4B) are noted. The upper unit, which has a sharp, erosional contact, is characterized by cross-stratified, well-rounded pebbles.

*Interpretation.* The lower and middle units echo those deposits already described from the lower and upper delta slope in the sea cliff, respectively. The distinctive, coarse-grained upper unit is interpreted as a fluvial terrace. This inland succession charts the progradation of a fjord head delta prior to progradation in the open sea domain. The lower unit is dominated by low-density turbidity currents and settling from buoyant plumes. The contribution of high-density turbidity currents increases upward in

the middle unit reflecting the deltaic progradation. Here, the common occurrence of gravitational instabilities is apparent (step fractures and shear zone), in contrast to the delta slope setting of the cliff cross-section is interpreted as the signature of high rate of sediment accumulation.

#### *Moraine-dammed lacustrine delta*

**Description.** The moraine-dammed lacustrine delta facies association has been specifically identified within the Sault-au-Cochon Valley (Fig. 4; Figure S1). It was deposited upstream of and after the ice-contact outwash fan (DU3) that initially dammed the outlet of this valley and against which it onlaps.

A coarsening-upward succession is observed. At the base, subhorizontal well-laminated silty fine-grained sand (SS6, Figure S7A) grades upwards into in-phase climbing-ripple laminations (Figure S7B). Upwards, ubiquitous climbing current ripples (SS3) develop; they form gently inclined (1 to 2°) metre-thick master beds. At the bed scale, the climbing angle progressively increases upwards: subcritical at the base, then supercritical up to in-phase aggradation of a sinusoidal lamination. In some cases, ripple morphologies were subsequently draped by mud layers (Figure S7B). Syn-sedimentary extensional fractures have disrupted the bedding (Figure S7). The uppermost 5 m of the succession is made up of erosion-based gravelly to pebbly cross-stratified sand (SGr2).

**Interpretation.** The facies suite is typical of (glacio-) lacustrine deltas (e.g. Jopling & Walker, 1968; Nutz *et al.*, 2015); they are organized as well-laminated fine-grained bottomsets, climbing ripples in foresets and gravelly topsets. Suspension-charged river outflows arriving into the lake directly plunged owing to density contrasts between the flow and the ambient water body. They formed climbing-ripple assemblages that essentially recorded waning stage flow conditions as shown by the increasing angle of climb (Jobe *et al.*, 2012). The stratal pattern likely reflects seasonal meltwater discharge with mud drapes possibly representing low-inputs during winter conditions.

#### **DEVELOPMENT OF THE PORTNEUF-Forestville DELTAIC COMPLEX**

A four-stage development (Fig. 8; Table 3), which echoes the depositional systems detailed above

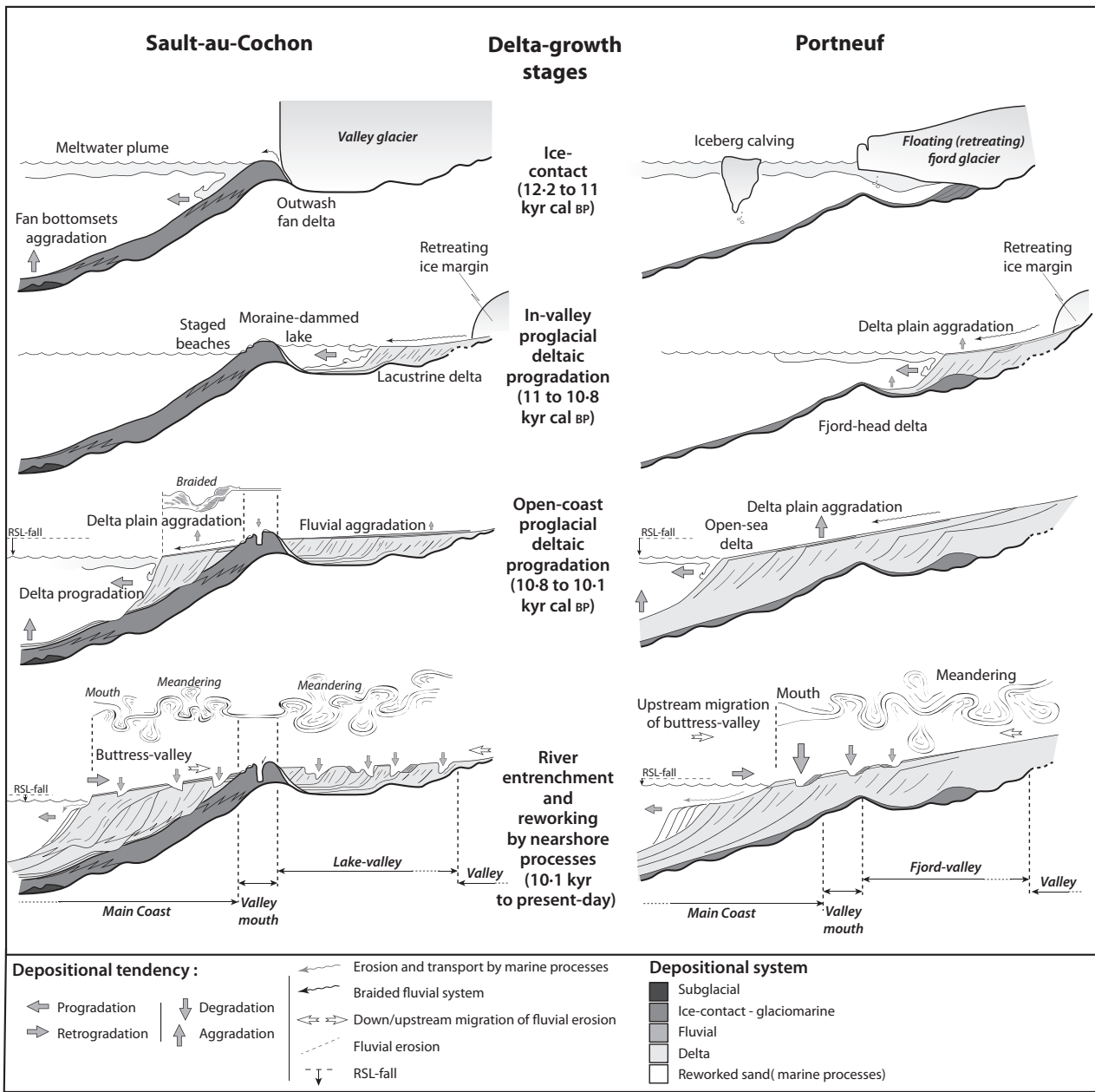
(depositional system I in the cliff, in-valley depositional system and depositional systems II and III), is proposed on the basis of landform assemblages (Fig. 1 and Figure S1), stratigraphic architectures (Figs 2 and 4), depositional facies (Figs 5 to 7) and radiocarbon dating (Table 1). Each stage is characterized by specific forcing parameters and sedimentation patterns (depositional rates, shoreline trajectories and relative intensity of marine reworking processes, Table 3).

#### **Temporal framework**

Combined with stratigraphic relationships, the internal coherency of the set of 12 radiocarbon dates (Table 1) allows for a robust temporal framework to be established. Although no radiocarbon date is directly available from the sandy upper delta slope deposits, ages are easily approximated by physical correlation with the lower delta slope (Fig. 2). From the age available in the glaciomarine mud (sample 30-2, 11 170 ± 70 year cal BP) to the older age available in the muddy lower delta slope (sample 2-2-a, 10 600 ± 115 year cal BP), a time span of ca 600 years accounts for the deposition of the outwash fan, the fjord head (Portneuf) and the moraine-dammed lacustrine (Sault-au-Cochon) deltas, and a notable part of the open-coast deltaic progradation (Fig. 2). This 600 year time span corresponds primarily to the period of in-valley and open-coast deltaic progradation that began with the development of the outwash fan; the latter most probably lasting a few hundreds of years in comparison with similar depositional settings (e.g. Nemeč *et al.*, 1999; Lønne *et al.*, 2001; Occhietti, 2007). A RSL curve is proposed here (Fig. 9) that represents the best fit between available radiocarbon dates and both observed and inferred stratigraphic relationships.

#### **Ice-contact outwash fan (12.2 to 11 kyr cal BP; shorelines from >140 m to 125 m.a.s.l.)**

After the disintegration of the LIS across the St. Lawrence Gulf and Estuary, the marine-based ice margin stabilized along the Canadian Shield and marine invasion extended over glacio-isostatically flexed lowlands (Goldthwait Sea). At that time, a fundamental dichotomy is noted between the successions observed at the outlets of the Sault-au-Cochon and Portneuf valleys. At the Sault-au-Cochon Valley outlet and beyond, down to the cliff (Fig. 4), an extensive ice-contact wedge was formed (depositional system I).



**Fig. 8.** The four stages of development of the Portneuf–Forestville deltaic complex, differentiating the distinctive stratigraphic evolution tied to the Portneuf (right) and Sault-au-Cochon (left) counterparts.

Here, the ice margin stabilized and an ice-contact depositional wedge developed at the still-stand position (Fig. 8). The ice-contact wedge initially developed downslope into a muddy glaciomarine turbiditic outwash system expanding down to the location of the present-day cliff (DU2; Figs 2 and 3). The upward disappearance of IRD and the onset of sand deposition characterize the transition from glaciomarine muds to the distal outwash fan, those two elements being the signature of a subaqueous to terrestrial

ice-front transition. The emergence of the fan resulted from a combination of sedimentary aggradation and RSL fall (e.g. Lønne *et al.*, 2001). At least one another outwash fan with a top surface lying at 140 m.a.s.l. is present at the outlet of the *Cedars Lake* valley (Figure S1). In contrast, at the Portneuf Valley outlet, the absence of any ice-contact deposits and the limited amount of glaciomarine muds during deglaciation implies that the ice front retreated significantly further up valley. At that time, the future Portneuf Peninsula was a

**Table 3.** Summary and semi-quantitative analysis of the characteristics of the four successive stages constituting the forced regressive deglacial sequence documented in the Portneuf Peninsula since 12.5 kyr cal BP.

Depositional system	Ice-contact	In-valley deltaic progradation	Open-coast deltaic progradation	Coastal sedimentary suites
Sedimentary structures	Deglacial	Proglacial (Ice-marginal retreat)	Proglacial (ablating ice margin)	Paraglacial (Ice margin retreated out from the drainage basins)
Depositional units	Outwash fan DU3 Glaciomarine mud DU2 Esker DU1	Fjord head and lacustrine deltas	Delta plain DU6 Upper delta slope DU5 Lower delta slope DU4	Nearshore sand sheet DU8 Spit platform DU7
Timing (duration)	12.5 to ~11 kyr cal BP (~1.5 kyr)	~11 to 10.8 (?) kyr cal BP (~0.2 kyr)	10.8 (?) to 10.1 kyr cal BP (~0.7 kyr)	10.1 kyr cal BP to today (>10 kyr)
RSL fall	140 to 125 m	125 to 110 m	110 to 90 m	90 to 0 m
Sedimentary volume	2 ± 0.5 km <sup>3</sup> (Sault-au-Cochon River outlet)	Fjord (Portneuf): <3 km <sup>3</sup> Lake (Sault-au-Cochon): 1.5 ± 0.5 km <sup>3</sup>	~5 km <sup>3</sup>	Unsignificant (sediment reworking)
Progradation length	4 to 10 km	~60 km	10 km	To the south: To the north: 5 km to 5 km (retrogradation)
Progradation rate	∅	~200 m·year <sup>-1</sup>	~15 m·year <sup>-1</sup>	~0.5 m·year <sup>-1</sup> to ~-0.5 m·year <sup>-1</sup>
Shoreline trajectory	∅	Accretionary Descending Regressive (Portneuf valley: 0.08% = 0.04°)	Accretionary Descending Regressive (0.2% = 0.115°)	Accretionary Descending Regressive (1.5% = 0.86°)

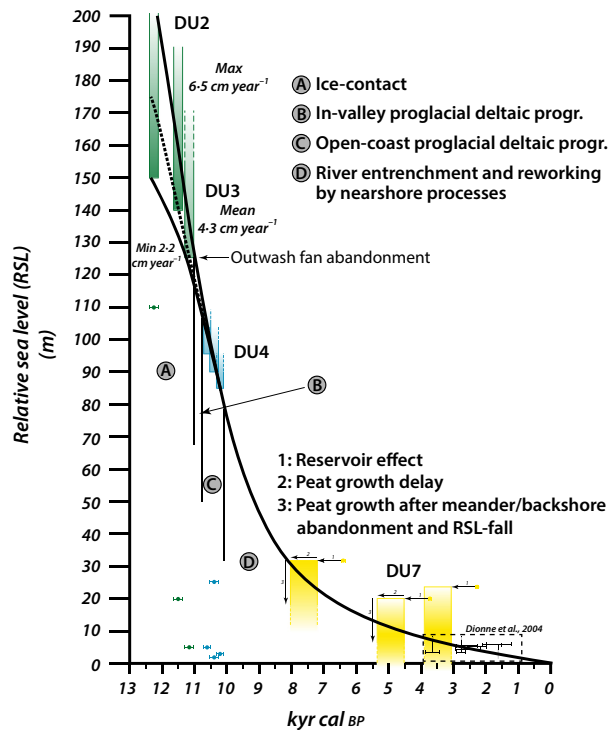
glaciomarine bay at the outlet of a fjord, with adjacent outwash fans.

The Sault-au-Cochon ice-contact outwash fan, which has a radius in excess of 15 km, stored gravely to sandy sediments from the present-day 140 m elevation (marine limit) down to beyond the current sea-level. Associated glaciomarine muds expand largely further to the east into the St. Lawrence Estuary where they correlate with depositional units 2 and 3 of the regional seismostratigraphy (Syvitski & Praeg, 1989; unit S2 in Duchesne *et al.*, 2010). In the Portneuf Peninsula, the volume estimate of outwash sediments accounting for known thicknesses and their lateral extent is 2 ± 0.5 km<sup>3</sup> (Table 3).

#### **In-valley proglacial deltaic progradation (11 to 10.8 kyr cal. BP; shorelines from 125 m to 110 m.a.s.l.)**

The second stage followed the retreat of the ice margin from its former stillstand at the outlet of the Sault-au-Cochon valley. Because no other nearby ice-marginal wedge has been identified up valley, the retreat of the ablating ice margin

in the Sault-au-Cochon Valley area was probably rapid (Fig. 8). The initial dichotomy between the Portneuf and Sault-au-Cochon valleys persisted as distinctive sedimentation styles characterize the infills of the two valleys: a moraine-dammed lacustrine delta in the Sault-au-Cochon Valley, a fjord head delta in the Portneuf Valley (in-valley depositional system). Beach ridges descending from 140 to 120 m along the basin-facing side of the abandoned and progressively emerging outwash fan shows that the marine shore was then temporarily sediment starved, owing to sediment retention in the moraine-dammed lacustrine delta. The stage of in-valley deltaic progradation is a very short period of time (*ca* 200 year) relative to the overall outwash to deltaic development (>2000 year; Table 3). The estimated volume of the sedimentary wedges is <3.0 km<sup>3</sup> and 1.5 km<sup>3</sup>, for the Portneuf and Sault-au-Cochon valleys, respectively (Table 3). The short time interval corresponding to this in-valley deltaic progradation implies very high progradation rates (>200 m·year<sup>-1</sup>), which are three to four times higher than those recorded in present-day fjord



**Fig. 9.** Reconstructed relative sea-level curve for the Portneuf Peninsula. The last two thousand years are partly derived from data published in Dionne *et al.* (2004). Colours relate to the depositional systems depicted in Fig. 2. Dots with error bars represent sampled and dated shell debris and organic matter (Table 1), and rectangles having a colour gradient reflect depositional units.

head settings (e.g. Cowan *et al.*, 1998, 50 to 70 m·year<sup>-1</sup>; Gilbert & Crookshanks, 2009, 50 m·year<sup>-1</sup>). High progradation rates resulted from the combined effects of effective RSL fall, high sediment discharge from the actively ablating LIS margin and reduced accommodation space owing to deposition in narrow bedrock-confined valleys (e.g. Eilertsen *et al.*, 2011). The shoreline trajectory of the fjord head delta was descending accretionary regressive, *sensu* Heland-Hansen & Hampson (2009). Although the downdip angle is poorly constrained, a maximal value can be assessed from the minimal gradient of the highest preserved fluvial terraces (*ca* 0.08%, Table 3), keeping in mind the differential glacio-isostatic flexure (Shaw *et al.*, 2002).

### Open-coast proglacial deltaic progradation (10.8 to 10.1 kyr cal BP; shorelines from 110 m to 90 m.a.s.l.)

The third stage is characterized by the development of a large and voluminous fluviodeltaic

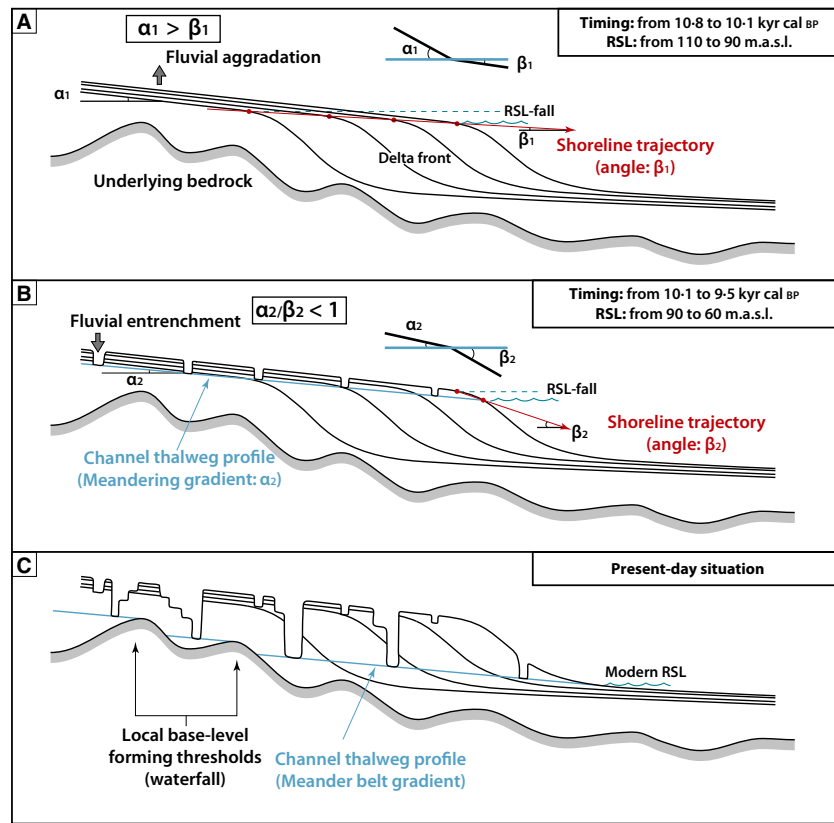
body (Depositional system II) after coalescence of the two deltaic systems now exiting the valley outlets (in-valley depositional system) at *ca* 110 m.a.s.l. and freely developing into an open-coast setting (Fig. 8). Regarding the Portneuf Valley, the open-coast deltaic progradation is the direct continuation of the fjord head delta. In contrast, a rejuvenation of the Sault-au-Cochon delta occurred after the infill of the moraine-dammed lake was completed; the antecedent lake outlet probably focusing the river mouth down to the regressive shorelines. Lower delta slope deposits (DU4) downlap the distal part of the outwash (DU3, Figs 2 and 4) and we then infer that topset beds of the rejuvenated Sault-au-Cochon delta overlapped against the depositional slope of the former proximal outwash fan (Figure S2A). The stratigraphic hiatus between DU3 and DU4, expressed by the virtual cessation of sand deposition (Fig. 2), corresponds to the time of lake infilling by the moraine-dammed lacustrine delta (Fig. 8). The channel structures found in the lower and upper delta slope (DU4 and DU5) were the feeder of basin floor fans located in the Laurentian Trough (Unit 4 in Syvitski & Praeg, 1989).

The open-coast progradation produced a *ca* 5 km<sup>3</sup> volume of sediment, considering an area of 100 km<sup>2</sup> (Figure S1) and a mean thickness of *ca* 50 m (Fig. 2). The progradation rate was about 15 m·year<sup>-1</sup>, showing an order of magnitude deceleration relative to the in-valley sedimentation (Table 3). This progradation rate reflects the significant increase in accommodation space provided by the depositional basin rather than a decrease in sediment supply (Fig. 10).

Interestingly, neither delta plain accretion nor fluvial entrenchment was noted during this stage, in spite of the forced regressive setting (Fig. 9). Further, the in-valley and open-coast deltaic accretions are characterized by similar facies, suggesting that sustained meltwater flows issuing from the ablating ice sheet maintained significant water and sediment discharge in the river catchment.

### River incision and reworking by nearshore processes (from 10.1 kyr cal BP to present-day; shorelines from 90 m to 0 m.a.s.l.)

The geomorphic and sedimentary records together indicate a major change when the RSL reached 90 m.a.s.l. (Fig. 2; Figure S1). The



**Fig. 10.** Conceptual model for delta plain accretion during RSL fall, as understood from the relationship between the fluvial gradient ( $\alpha$ ) and the shoreline trajectory angle ( $\beta$ ). (A) During the early proglacial deltaic progradation. (B) Fluvial entrenchment during the later paraglacial evolution. (C) Present-day situation illustrating the influence of local base levels imposed by bedrock sills (see text for details).

presence of beach ridges and associated deposits (Depositional system III) below 90 m elevation demonstrates the increased importance of marine processes relative to fluvial ones. The transition is interpreted as the result of a dramatic decrease in the fluvial input rather than being tied to an increase in wave energy (e.g. Swenson *et al.*, 2005). The present authors inferred that active coastal erosion began at that time in the northern Peninsula, nourishing spit progradations to the south. The braided fluvial network was no longer active and meandering rivers started to incise into the underlying deltaic sediment pile, contributing also to sediment reworking and spit accretion. The transitional boundary between the last deposits of DU4 and DU5 (Fig. 2, km 2.5 to 3.3), in contrast to scouring processes that occurred primitively further to the north, reflects the progressive decline in fluvial input that preceded the onset of predominance of coastal processes.

Relative sea-level was continuously falling during this phase. The shoreline trajectory, best represented by the surf diastem geometry (lower bounding surface of the nearshore sand sheet, Figs 2 and 7H) showed a biphased evolution. Initially, shorelines migrated over the inherited upper delta slope deposits (from km 7.0 to 2.6,

Fig. 2), and a severe truncation with shoreface deposits was observed here. The limited accommodation space resulted in a descending, non-accretionary regressive shoreline trajectory (Helland-Hansen & Hampson, 2009). From the point at which the shoreline crossed the inherited, inactive delta rim (*ca* km 2.7), the sudden depth increase led to the construction of a spit platform. As a consequence, the descending regressive shoreline trajectory became distinctively accretionary (Figs 2 and 10). Meanwhile, fluvial meanders formed on the delta plain and were successively abandoned and stepped from 65 to 15 m.a.s.l., below the rising, abandoned delta surface currently lying at elevations between 110 m and 90 m (Fig. 1; Figure S1). In places, total fluvial erosion depths are in excess of 50 m, denoting a fluvial entrenchment (Fig. 4).

## DISCUSSION

### Anatomy of the forced regressive deglacial sequence

This study demonstrates that the deltaic complex is temporally tied to a forced regressive

1 evolution. Volumetrically, most of the deposits  
 2 that constitute the deltaic complex relate to the  
 3 ice-contact, in-valley and open-coast progradat-  
 4 ion stages (Table 3). In other words, the forced  
 5 regressive deltaic complex represents the  
 6 deglaciation record. It displays the following  
 7 characteristics:

8  
 9 **1** Coarse-grained ice-contact sediment bodies  
 10 (the esker and the proximal outwash fan) are  
 11 volumetrically minor relative to the fine-grained,  
 12 ice-distal sediment bodies (glaciomarine muds  
 13 and distal outwash fan), which generally onlap  
 14 directly the basal glacial erosion surface.

15 **2** Two successive coarsening-upward deposi-  
 16 tional systems (excluding the esker) together  
 17 built a >50 to 100 m thick sediment wedge in a  
 18 restricted time interval. The lower system  
 19 resulted from outwash fan progradation (DU3),  
 20 the upper system corresponding to delta prograd-  
 21 ation (DU4 to DU6). In the cliff, they are super-  
 22 imposed but they may be laterally juxtaposed in  
 23 other case studies.

24 **3** The deltaic progradation of the deglacial  
 25 sequence was emplaced in a forced regressive  
 26 setting and then does not correspond to a post-  
 27 glacial system characterizing highstand condi-  
 28 tions. In domains with a lesser isostatic  
 29 rebound, similar sediment bodies are referred to  
 30 as 'lowstand delta' (Hein *et al.*, 2014).

31 **4** Delta plain accretion was maintained during  
 32 initial, rapid RSL fall but fluvial incision  
 33 occurred counter-intuitively later, when the rate  
 34 of RSL fall had already decreased significantly  
 35 (Fig. 9).

36 **5** Progradations (outwash and delta) left behind  
 37 a thick (>100 m) depositional wedge represent-  
 38 ing a relatively short time interval (*ca* 2000 year;  
 39 Figs 8 and 9). Such a succession would have  
 40 been easily misinterpreted as a typical third-  
 41 order depositional sequence in the absence of  
 42 any high-resolution time constraints and/or any  
 43 detailed stratigraphic architecture.

44 **6** The late depositional evolution was domi-  
 45 nated by shallow marine reworking (DU7 and  
 46 DU8) of the outwash and delta sediment  
 47 wedges, with no substantial fluvial input. In  
 48 other words, the deltaic complex is essentially  
 49 inactive today (see also Normandeau *et al.*,  
 50 2015).

51 This type of deglacial sequence differs from  
 52 other successions where glaciomarine deposi-  
 53 tional systems are predominant in terms of sedi-  
 54 ment volumes and related accumulation rates  
 55 (e.g. Boulton, 1990; Syvitski & Lee, 1997; Powell

& Cooper, 2002). Here, the ice-contact deposi-  
 tional system (outwash fan and glaciomarine  
 muds) contributed to only about 20% of the  
 overall volume of sediment deposited in the del-  
 taic complex. The short depositional time inter-  
 val denotes striking averaged accumulation rates  
 in the 25 to 50 m·kyr<sup>-1</sup> range, which is a conser-  
 vative estimate. Nevertheless, depositional suites  
 are made up of regular facies associations and  
 indeed no large mass flows or megabeds have  
 been recognized. It thus implies high preserva-  
 tion rates of the successive event beds and  
 minor intervening erosional or a non-deposi-  
 tional hiatus throughout the succession. Without  
 a framework based on radiocarbon dating, these  
 relationships would have remained unsus-  
 pected. Such deglacial sequences may have been  
 underestimated when deciphering the deep time  
 (de)glacial record. The short duration of deposi-  
 tion, negating the role of subsidence in the  
 depositional stratigraphic architecture of such  
 forced regressive deglacial sequences, would be  
 viewed as virtually instantaneous in the coarser  
 resolution deep time record (Girard *et al.*, 2015).  
 In the forced regressive context, accommodation  
 space is then only provided by inherited  
 topographies.

### Control of drainage basin extent on sedimentary architecture

The turnaround terminating the actively pro-  
 grading open-coast phase occurred when the  
 RSL reached the break-in-slope at the 90 m ele-  
 vation (Fig. 8). This transition has clear signa-  
 tures in terms of stratigraphic architecture and  
 depositional facies as shown by: (i) the onset of  
 the river entrenchment after a period of delta  
 plain accretion; (ii) the replacement of the  
 braided plains by entrenched, meandering  
 streams; (iii) the generalization of coastal forms  
 (spit, berms and back barrier marshes); and (iv)  
 the demise of turbiditic deposition over the  
 delta slope, concurrent with the onset of the spit  
 platform to the south and coastal erosion to the  
 north. These elements suggest that the system  
 evolved from a high-discharge fluvial-dominated  
 depositional system to a new state, within  
 which shallow marine depositional processes  
 largely prevail over river dynamics.

In glaciation settings, the abrupt decrease in  
 sediment supply is recurrently associated with  
 the retreat of ice sheet margins from river drain-  
 age basins (Church, 1972; Syvitski, 1989; Syvit-  
 ski & Hein, 1991; Forbes & Syvitski, 1994;

1 Corner, 2006; Eilertsen *et al.*, 2011; Nutz *et al.*,  
 2 2014), considering that glacier-related sediment  
 3 yields are greater by one or two order of magni-  
 4 tude relative to those in non-glaciated basins  
 5 (Church & Slaymaker, 1989; Syvitski & Lee,  
 6 1997). The sediment starvation hypothesis is in  
 7 good agreement with: (i) the transition from  
 8 delta plain accretion to river entrenchment in  
 9 such a context of RSL fall (e.g. Leeder & Stewart,  
 10 1996; Blum & Törnqvist, 2000; Klinger *et al.*,  
 11 2003); and (ii) the enhanced marine processes  
 12 on deltaic accretion patterns (Reading & Collin-  
 13 son, 1996; Swenson *et al.*, 2005). In the present  
 14 case study, the age of the transition from proglacial,  
 15 river-dominated deltaic progradation to the  
 16 generalization of coastal processes confidently  
 17 dates back to 10 kyr cal BP owing to seven  
 18 radiocarbon dates, in particular those recovered  
 19 from lower delta slope muds (Fig. 2; Table 1).  
 20 On the basis of ice sheet reconstructions (Occhi-  
 21 etti *et al.*, 2011) and considering the extent to  
 22 the north-west of the drainage areas of the Port-  
 23 neuf and Sault-au-Cochon rivers, this age pre-  
 24 cisely coincides with a LIS margin positioned  
 25 close to the uppermost watershed of the two riv-  
 26 ers (Fig. 1B). Thus, this transition is interpreted  
 27 here as the signature of the retreat of the LIS  
 28 margin out of the Portneuf and Sault-au-Cochon  
 29 drainage areas. It corresponds for the deltaic  
 30 complex to a proglacial-to-paraglacial transition.  
 31 The shutdown of the glaciogenic sediment sup-  
 32 ply transformed an actively prograding delta  
 33 into a depositional system essentially dominated  
 34 by shallow marine reworking processes (see also  
 35 **30** Forbes & Syvitski, 1994; Swenson *et al.*, 2005).  
 36 The reduction in river-derived sediment supply  
 37 caused either coastal erosion or progradation of  
 38 beach/barrier systems, depending on patterns of  
 39 longshore drift (Bernatchez & Dubois, 2004;  
 40 Korus & Fielding, 2015). It follows that one of  
 41 the main stratigraphic discontinuity and erosion  
 42 surfaces within the forced regressive deglacial  
 43 sequence derived from the relatively far-field  
 44 deglaciation development in the drainage basins  
 45 rather than to local changes, for instance the  
 46 interplay between RSL fall and inherited  
 47 topographies.

### 49 **Fluviodeltaic deposition during relative sea-** 50 **level fall: from accretion to river** 51 **entrenchment**

52 The well-preserved Portneuf–Forestville forced  
 53 regressive deglacial sequence offers a unique  
 54 opportunity to test recent models and concepts  
 55

dealing with possible sediment deposition dur-  
 ing RSL fall. Subsurface (McMurray &  
 Gawthorpe, 2000; Tesson *et al.*, 2000; Rabineau  
*et al.*, 2005) and outcrop (e.g. Bhattacharya &  
 Willis, 2001; Porebski & Steel, 2006; Fielding,  
 2015) case studies have usually indicated that  
 limited or no fluvial/delta plain deposits are  
 typically associated with falling stage system  
 tracts (FSST); however, it is rarely determined  
 whether such deposits are lacking because of  
 bypass or erosion during forced regression or  
 owing to subsequent transgressive ravinement  
 surfaces (Plint & Nummedal, 2000; Posamentier  
 & Morris, 2000; Catuneanu, 2002; Embry, 2010;  
 Santra *et al.*, 2013). Significant volumes of flu-  
 vial sediments can be nevertheless deposited  
 and preserved within the FSST. In particular,  
 subaerial deposition during RSL fall has been  
 long recognized to occur where the downstream  
 basin topography has a gradient lower than that  
 of the fluvial equilibrium profile (Schumm,  
 1993; Burgess & Allen, 1996; Browne & Naish,  
 2003; Zecchin, 2007; Holbrook & Bhattacharya,  
 2012; Blum *et al.*, 2013). Fluvial aggradation  
 during forced regression can also be understood  
 as the result of a high rate of sediment supply  
 associated with low diffusion rates in subaerial  
 environments; a situation implying that sedi-  
 ment supply exceeds the carrying capacity of  
 the rivers (Swenson & Muto, 2007; Prince & Bur-  
 gess, 2013; Fontana *et al.*, 2014).

This study has described a deltaic complex  
 that was entirely emplaced in a context of sus-  
 tained RSL fall (Fig. 9). Its evolution was two-  
 fold: (i) early proglacial dynamics (shorelines  
 above 90 m.a.s.l.), showing fluviodeltaic accre-  
 tion and no river incision, but coeval with high  
 rate of sea-level fall (3 to 5 cm·year<sup>-1</sup>) and a  
 high sediment supply; (ii) late paraglacial  
 dynamics (shorelines below 90 m.a.s.l.), charac-  
 terized by river entrenchment, lower rates of  
 RSL fall (<3 cm·year<sup>-1</sup>) and restricted sediment  
 supply. In both cases, the basin slope was  
 greater than the fluvial gradient. Fluvial aggra-  
 dation was possible, however, owing to high rates  
 of sediment supply and ‘neutralization’ (*sensu*  
 Leeder & Stewart, 1996) of the progressive water  
 depth increase during progradation of the delta.  
 The case study here is an illustration of some  
 model predictions that consider the massive  
 reduction in sediment supply as being poten-  
 tially responsible for river incision (Muto &  
 Swenson, 2006; Swenson & Muto, 2007). **31**

The signature of the proglacial-to-paraglacial  
 transition can also be illustrated in terms of



shoreline trajectory (Helland-Hansen & Hampson, 2009). In the proglacial setting, the fluvial gradient ( $\alpha_1$ ) is best expressed by the delta plain slope gradient between 110 m and 90 m.a.s.l. (*ca* 0.25%), considering that it represents a snapshot of the braided fluvial system immediately prior to its abandonment (Fig. 10; Figure S1). Such a relatively high gradient is, however, consistent with that of a present-day sandar (0.14 to 3.0%, Maizels, 2002). An estimate of the shoreline trajectory angle ( $\beta_1$ ) during the open-coast progradation is *ca* 0.2% (see above; Figs 2 and 4, Table 3). Hence, when the RSL fell from 110 to 90 m.a.s.l., the ratio  $\alpha_1/\beta_1$  was slightly more than 1, allowing a limited but definite topset aggradation (Fig. 10; see also Helland-Hansen & Hampson, 2009; their fig. 9G). The paraglacial fluvial entrenchment and related downstepping meanders (Fig. 4; Figure S1) conform to a more conventional forced regressive evolution. Here, the shoreline trajectory in the axis of the meandering river essentially reflected interactions between the RSL fall rates and the inherited basin slope (Helland-Hansen & Hampson, 2009). The gradient of the paraglacial meandering river ( $\alpha_2$ ), either estimated from DEM or inferred from present-day coastal meanders flowing into the St. Lawrence Estuary and Gulf, was significantly flatter (*ca* 0.05 to 0.8%) than that of the earlier braided streams. Below 90 m.a.s.l. and beyond the former delta brink, the new shoreline trajectory angle ( $\beta_2$ ) was scaled approximatively to the angle of the inherited deltaic depositional profile, which was relatively steep (>1%, Figs 2, 4 and 10). Thus, the paraglacial  $\alpha_2/\beta_2$  ratio was  $\ll 1$ , a condition leading straightforwardly to river incision in conditions of restricted sediment supply. Following the shutdown of the glaciogenic sediment supply, river entrenchment was initiated upvalley, from the head of the drainage basins, with subsequent downstream migration (buffer valley model of Holbrook & Bhattacharya, 2012). Concomitantly, processes of fluvial incision leading to stepped and abandoned coastal meanders adjusted to the shoreline originated from an upstream migrating incision. It corresponds to the buttress valley model of Holbrook & Bhattacharya (2012) (Fig. 8).

## CONCLUSIONS

In this study, a post-Late Glacial Maximum (LGM) forced regressive, deglacial sequence was

documented mainly on the basis of landforms, stratigraphic architectures and radiocarbon dating. Facies associations were described in order to highlight depositional processes along the deltaic profile, but related palaeobathymetries were primarily inferred from the overall deltaic architecture. This case study provides an outcrop analogue for currently submerged lowstand deltas in the eastern Canadian outer shelf for example, only illustrated by seismic dataset or shallow cores. It is envisioned here that forced regressive deltaic complexes, related stratigraphic architecture and associated depositional facies definitely have preservation potential if (i) relatively thick depositional proglacial wedges have been emplaced and (ii) they are located in areas experiencing a significant relative sea-level (RSL) rise after deglaciation where a post-glacial glacio-eustatic transgression succeeds to an initial glacio-isostatic rebound. This study demonstrates that the bulk of the sediment volume, originating from ice-contact and forced regressive delta dynamics, relates to very short-lived (<2000 year here) stages of the glacial retreat and associated sedimentation. Most of the paraglacial time is essentially associated with sediment reworking and redistribution, either in regressive or transgressive conditions. This study is significant for the understanding of both Quaternary and pre-Quaternary forced regressive, deglacial successions because it highlights a case of stratigraphic distortion primarily linked to glacially controlled patterns of sediment supply rather than to accommodation trends. Indeed, timescales in proglacial-to-paraglacial sediment deposition are one to three orders shorter than basinal subsidence patterns. Being disconnected from the global sea-level cycle, careful examination is required in the identification of glacio-isotatically forced regressive deltaic complexes that record a deglaciation. Based on the Portneuf–Forestville case study, their main features should be as follows:

- 1 Underlying glacially related sediments wedges, from thin till covers to thick, but spatially restricted, ice-contact outwash fan deposits.
- 2 A shallowing-upward tripartite succession from a mud-dominated lower delta slope, sand-prone turbiditic upper delta slope to proglacial braided delta plain deposits.
- 3 A shift from active deltaic accretion to a shallow marine reworking stage, which marks the retreat of the ice sheet margin from the drainage basin of the rivers.

4 Raised and stepped beach structures, staged from the marine limit down to the interglacial shoreline, which will probably be reworked during any ensuing transgression.

5 Fluvial entrenchment processes occurring late in the deglacial history and marking the onset of the paraglacial evolution in the delta plain.

A diagnostic criteria differentiating glacio-isostatically forced regressive deltas from other glaciation-related deltaic successions are the embedded record of a drastic reduction in sediment supply. It is in contrast with other forced regressive systems where sediment supply is increasing during the forced regression.

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## Supporting Information

Additional Supporting Information may be found in the online version of this article:

**Figure S1.** (A) Landform distribution over the Portneuf Peninsula. Bedrock outcrops in grey, ice-contact deposits in red (from Cousineau *et al.*, 2014); red arrows: glacial striations and/or crescent marks on bedrock; green lines: networks of braided channels; orange: coastal landforms, mainly raised beach ridges. (B) Panoramic picture of one of the active meanders of the Sault-au-Cochon river.

**Figure S2.** Details of two selected segments of the cliff (location in Fig. 2). (A) the northernmost segment (km 7.0 to 7.7), from esker (DU1) to delta plain (DU6) deposits. Note draping and overlapping geometries of glaciomarine mud (DU2) above the upwardly-convex