Sedimentology and Stratigraphy of the Cap Barre through Saint Yvon Section, Cloridorme Formation, Gaspe Peninsula, Quebec, Canada

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INTRODUCTION

The Cap Barre through Saint Yvon section of the Cloridorme Formation is located on the northern coast of the Gaspe Peninsula, Quebec, Canada. The Cloridorme Formation is of late Middle Ordovician, or Caradocian, age, and is more than 7.0km thick (Enos, 1969). The Cloridorme is a siliciclastic foreland basin flysch sequence. These sedimentary rocks were laid down during the closing of the proto-Atlantic, or Iapetus Ocean, as part of a westward-prograding wedge shed from an approaching system of tectonic lands and volcanoes (see fig. 1) (Hiscott, Pickering and Beeden, 1986).

The purpose of this project is to examine the sedimentology and stratigraphy of the Cap Barre through St. Yvon section of the Cloridorme Formation. The sandstones and shales of this formation have been divided into three lithofacies (A, B, and C) based on bedding thickness and relative proportions of shale and sandstone.

FACIES DESCRIPTIONS

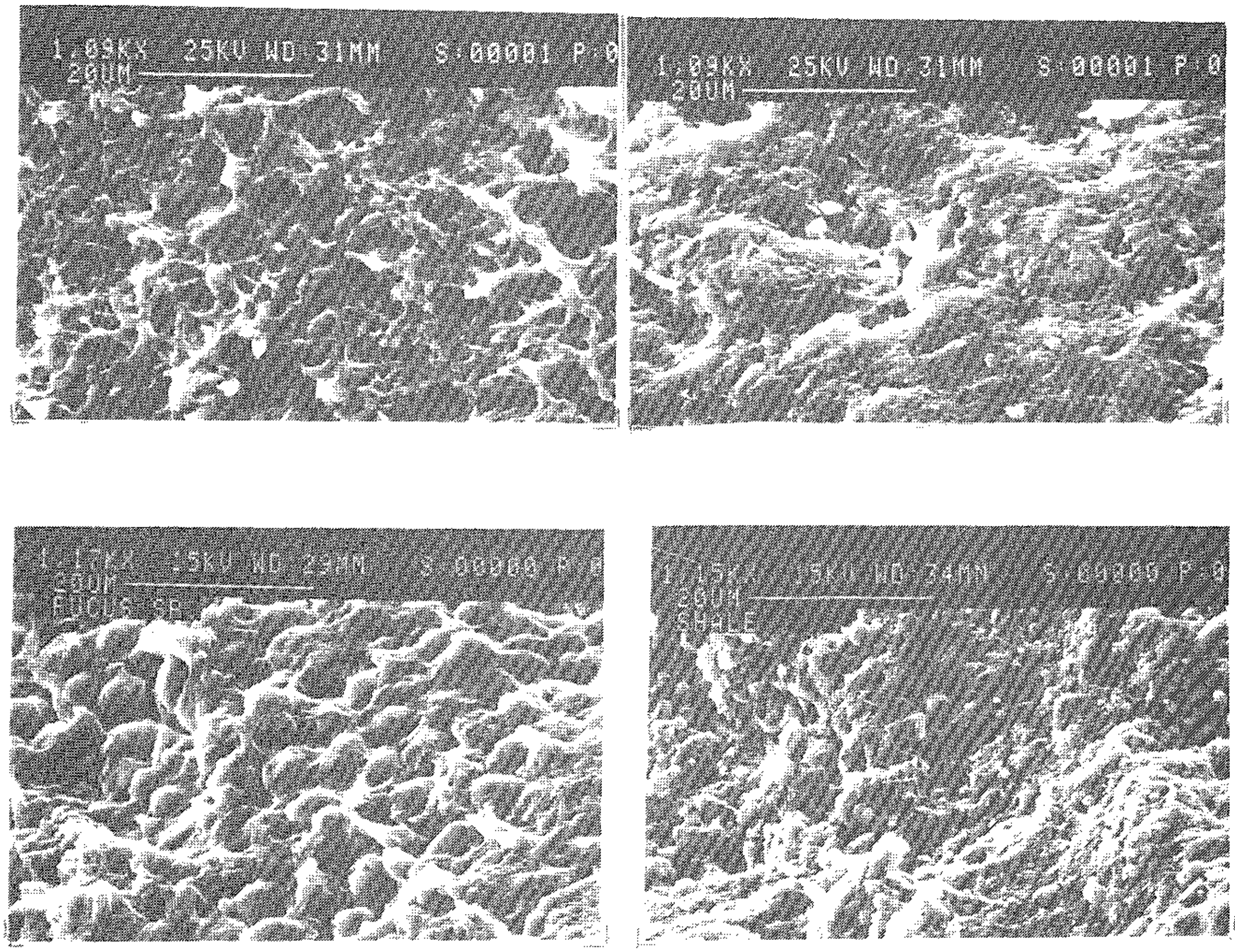
Facies A: Thick sandstone and shale couplets. The lower part of these couplets are made up of medium to thick beds (20cm—1.2m thick) of orange-weathering, calcite-cemented, very fine to fine sandstone. The upper part of these couplets consist of black fissile shale beds, from 50cm—4.5m thick. The contacts between sandstone beds and shale beds are usually sharp but in some cases are gradational.

The sandstone beds contain the following sedimentary structures: convoluted bedding; psuedo-nodule layers; planar and undulatory bedding; flute casts, and in a few examples, climbing ripple lamination. Beds are well sorted and lack grading.

Facies B: Interbedded siltstone/sandstone and shale beds. Resistent-weathering beds consist of very thin to medium-bedded (0.5—40cm thick), orange-weathering, calcite-cemented siltstones to fine sandstones. [These sandstone beds were separated into two thickness populations, B and B', —see table 1] Shale beds have a similiar range of thickness and the overall ratio of siltstone/sandstone to shale is roughly 1 to 1. The contacts between the siltstone/sandstone and shale beds are extremely sharp.

Facies B sandstone beds display no grading, but in nearly all cases they contain climbing ripple cross-lamination. This cross-lamination shows various degrees of soft sediment deformation, usually in the form of convolute bedding. Sole marks and psuedo-nodules are rare in Facies B siltstone/sandstone beds.

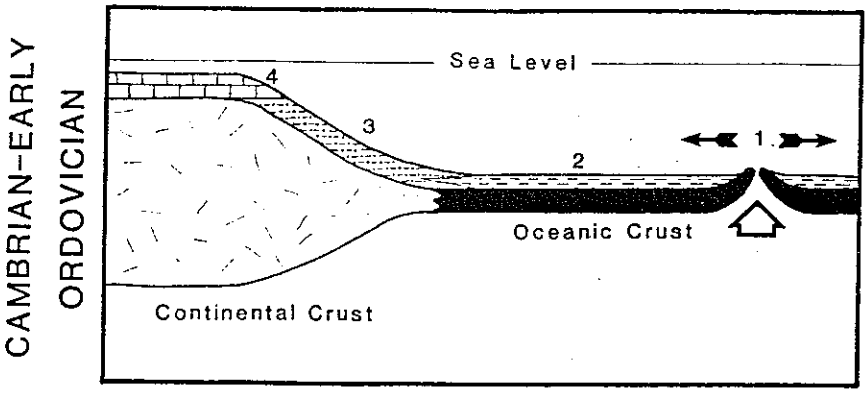
Facies C: Medium to thickly-bedded (25—90cm thick), massively-weathering, dark gray, siltstone to fine sandstone beds interbedded with thinly bedded (3-20cm thick) black shale. Many Facies C sandstone beds are



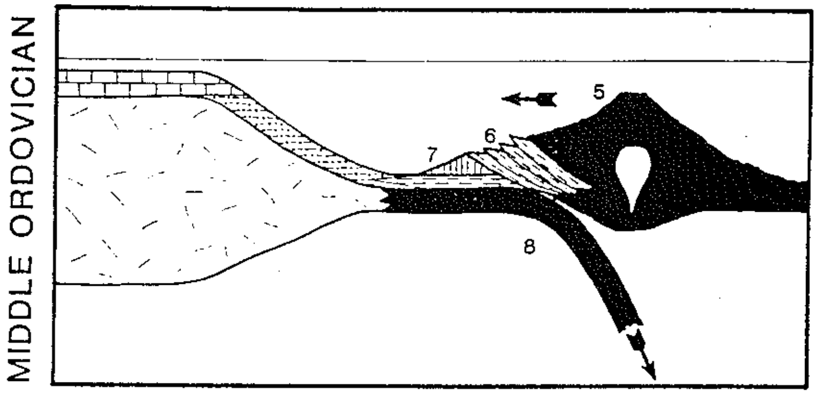
<u>Figure 2.</u>

From top to bottom:
Sample 1 F. vesiculosus scar
Sample 3 F. species scar

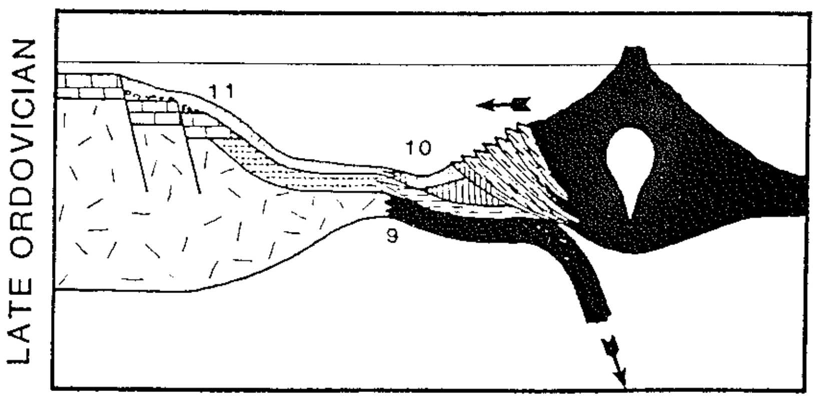
From top to bottom:
Sample 1 Argillite
Sample 3 "



1. Seafloor Spreading
2. Abbysal Plain Sediments
3. Slope Sediments
4. Shelf Sediments



5. Island-arc Magmatism
6. Accretionary Wedge
7. Submarine Fans, Tourelle Formation
8. Subduction



9. Loading/Downwarping

10. Foreland Basin, Cloridorme Formation

11. Subsidence

Figure 1. Evolution and plate tectonic setting of the foreland basin associated with the Cloridorme Formation. TOP: Cambrian and Early Ordovician. Atlantic-type margin and passive margin sedimentation during spreading of Iapetus Ocean. MIDDLE: Middle Ordovician. Subduction initiated with associated accretionary wedge. BOTTOM: Late Ordovician. Plate collision leading to downwarping of shelf and formation of elogate foreland basin.
Adapted from Hiscott, 1978.

amalgamated into thicker beds up to 2.5m thick. The amalgamation surfaces are undulatory and clearly erosional. The sandstone and shales beds of Facies C are commonly found in collections that contain no interbedding of the two other facies.

Rip-up clasts of black shale are common. These clasts are 3—30cm long and are usually ductilly deformed. The rip-up clasts are most commonly found at the top of Facies C sandstone beds; in a few cases they are found in the middle of beds.

DISCUSSION

The Cloridorme Formation is made up of interbedded turbidite sandstones and hemipelagic shales. Turbidites are the deposits of sediment gravity flows driven by density contrast between sediment and the surrounding water. These flows are triggered by a variety of forces: earthquakes; spontaneous liquefaction of rapidly deposited, fine-grained sediment with high pore pressures; and cyclic wave loading during large storms (Walker, 1984) (Middleton and Hampton, 1973). Tubidites are primarily associated with deep sea environments, specifically submarine fans (Walker, 1984).

The Middle Ordovician foreland basin was flanked by multiple submarine fans. The elongated geometry of the basin (perhaps 20km wide by over 100 kms long- see figure 1) constrained the turbidity currents to flow parallel to the axis of the basin (Hiscott, et al, 1986). In addition, the sedimentation rate in the basin was high, perhaps more than 400m/Myr. Thus, the deposits from from the many submarine fans overlapped and filled the basin. This overlapping of turbidites, some of which are continuous up to 2—3km (Enos, 1969), masks the progradation of single fans. Instead, a series of interbedded proximal and distal lobe deposits with similiar paleoflow directions (see figure 3) was laid down (Hiscott, et al. 1986). A lack of thickening or thinning upward sequences in sandstone beds in the studied section of Cloridorme Formation supports this multiple input, axial-trough flow hypothesis.

The interbedding of proximal and distal turbidites due to the elongate shape of the basin is useful as a model to explain the relationship between Facies A and B. These two facies appear much the same in terms of weathering, paleoflow direction and composition, but differences in bed thickness and sedimentary structures clearly separate the two (see table 1). Turbidite beds thin laterally and in the direction of flow. It is probably the case that a Facies B bed, traced in the upflow direction, would thicken into a Facies A bed. In addition, Facies A and B are randomly interbedded throughout the section as the model would suggest.

The abundance of soft sediment deformation in the thicker Facies A beds reflects the ability of large, quickly deposited sediment to trap more water than thinner beds during deposition and, consequently, be more prone to liquefaction. Post-depositional deformation features such as convoluted bedding and psuedo-nodules are more common in the thicker beds. Preservation of syndepositional features, such as climbing ripple cross-lamination are, not surprisingly, more common in the thinner Facies B beds. These thinner beds are below the "critical" thickness for liquefaction. Both Facies A and B sandstone beds are made up predominantly of well-sorted, very fine and fine sandstone with no grading. Sediment of this sort is particularly susceptible to liquefaction (Andressen and Bjerrum, 1967) (Middleton and Hampton, 1973).

Classical turbidites, defined by Bouma (1962), result from deposition from waning turbidity currents. The T_c division of the Bouma sequence contains ripple cross-laminated silt to fine sand sized sediment, that is commonly deformed into convoluted lamination. These features indicate deposition during lower flow regime conditions and subsequent liquefaction. Both Facies A and B sandstone beds are dominated by T_c Bouma divisions. This is consistent with the axial trough model of deposition. Basin plain turbidites commonly lack the T_a and T_b divisions and begin with the T_c division. The lack of the lower divisions is due to the distal positions relative to the distributive channels (Walker, 1984). In the case of the Cloridorme Formation, the mixing of multiple submarine fan deposits and the axial flow of this sediment created, in effect, a huge series of overlapping lobe fringe deposits (Hiscott, et al., 1986).

The Facies C beds are different in many respect from the two other facies in this section (see figure 2). Facies C sandstones and shales are found in collections of beds up to 10m thick with no interbedding of other facies. The Facies C beds are coarser and include more volcanic fragments than Facies A and B. Rip-up clasts are common in the Facies C beds but absent in the two other facies. Sedimentary structures are rare in the Facies C sandstones—nearly all beds are nonstratified. Flutes and other paleoflow indicators are not present in Facies C beds (see figure 2). It is unknown if these characteristics represent a different source area.

The effects of small changes in local sea level or local tectonics may account for the differences between Facies C and the other facies. Hiscott et al (1986), believe the high sedimentation rate during the deposition of the Cloridorme Formation "makes it impossible to assess the importance and effects of short-duration local...sea level changes". The exhaustive cataloging of beds for my study might provide evidence of the small-scale changes that Hiscott, et al, dismiss. The intermittent collections of Facies C beds may represent parasequences. Parasequences are cyclic pulses consisting of alternations sediment in response to rapid rises and stillstands of local sea level. These are commonly recognized only on the scale of outcrop (Vail, et at., 1977).

Alternatively, the Facies C beds may represent typical submarine fan deposits that were not reflected down the axis. Deflection may alter competency of the turbidity current, dropping out the larger grains and rip-up clasts that typify Facies C. This lack of reflection may be due to occasional changes in basin geometry or deposition, perhaps stemming from sea level fluctuations. On the other hand, perhaps Facies C beds are simply the intermittent deposits from a nearby submarine fan.

Preliminary petrographic study of the sandstone beds is consistent with the tectonic and depositional framework described above. Quartz grains are predominantly monocrystalline, have multiple inclusions and showed heavy undulatory extinction. The angularity of fossils and sedimentary rock fragment detritus clearly suggests rapid deposition. Feldspars are nearly absent, suggesting deposition was far from the craton. Using the diagnostic criteria of Dickenson and Suczek (1978) this suite of sediments plots in the recycled orogen province. Depositional environments for this setting include turbidity currents in forearc basins near unstable plate boundries.

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FACIES	Thickness	Convoluted Lamination	Climbing Ripple Lamination	Planar/ Undulatory Lamination	Flute Casts	Psuedo-nodules	Mud Drapes	Rip-ups	
								Top	Mid.
A (n:19)	R: 18–150cm x: 43.7 s: 36.1	79%	16%	68%	37%	53%	47%	–	–
B <20cm (n:50)	R: 1–15cm x: 4.7 s: 4.3	22%	42%	14%	12%	10%	4%	–	–
B′ >20cm (n:30)	R: 20–56cm x: 34.1 s: 24.9	70%	27%	40%	37%	20%	17%	–	–
C (n:54)	R: 3–85cm x: 26 s: 18.3	7%	7%	–	–	14%	24%	87%	3%

Table 1. Type and abundance of sedimentary structures, average bed thickness, and sample size for each of the facies in the study. Note: Facies B was split into two thickness populations, greater than or less than 20cm.

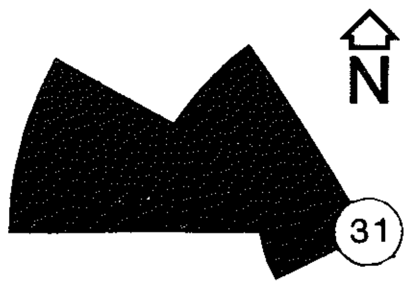

Figure 2. Flute mark paleocurrent rose diagram. Data is from all three facies in the study.

THE ORIGIN AND EVOLUTION OF PENOUILLE SPIT, FORILLON NATIONAL PARK, GASPÉ PENINSULA, QUEBEC

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INTRODUCTION

Penouille spit is located in Forillon National Park on the northern shore of Gaspé Bay (Fig. 1). The purpose of this study was to obtain information about the developmental history of the spit by studying the positions and orientations of beach ridges, sediment transport directions relative to the offshore bars, grain size distribution along each shore, the soil profile of a four-foot erosion scarp on the west side of the spit, bedding in two trenches dug on the south and west shores, and measurements of accretion and erosion made from comparison of 1989 and 1978 data.

GEOMORPHIC FEATURES

Seven north-south transects were surveyed with a tape and compass to locate the positions of the major ridges in the field (Fig. 2). These measurements were plotted on a map of Penouille made in 1978 (Allard and Germain, 1979) and compared with the ridges shown on a blow up of a 1970 aerial photograph. Each ridge represents a former shoreline of the spit.

Ten transects across the offshore bars on the south side of the spit recorded the relative locations and sizes of the bars close to low tide. These measurements and aerial photographs taken in 1989 were used to draw a contour map of the bars. The resultant pattern was compared to aerial photographs from 1948, 1961, 1966, 1970, and a map of the spit made in 1978 (Allard and Germain, 1979) to determine growth and movement of the bars. From 1970 to 1989 the bars grew to extend further west, parallel to the south shore. This growth is due to changes in topography and sediment transport along the south shore.

One change in topography was the development of a hook on the southwest corner of the spit (Fig. 2). The shape and orientation of this feature is indicative of the sediment transport directions. The path of transport runs from north to south along the west side of the spit, around the hook to the offshore bars. From there, sediment accretes onto the south shore. One source of sediment for this system is the eroding area marked by a scarp on the west side of the spit, which is continually being worked by the waves.

SPIT SEDIMENTS AND STRATIGRAPHY

Two trenches were dug to analyze the stratigraphy of the spit on the south and west shores (Fig. 3). At station C (Fig. 2), the trench showed the accretion of sediment onto the south side of the spit. The layers of sand in the trench were parallel to the surface indicating that layers of sediment were being deposited by the waves. The second trench dug on the west side between stations H and I showed erosional features and evidence of sediment that had been reworked by the waves. The layers within the trench were truncated by erosion of the foreshore. The bedding in these trenches proved the hypothesis that the west shore was being eroded back while the south shore was being built out.

A four foot erosional scarp on the west side of the spit was cleared off to reveal a soil profile . The horizons present in the profile included a dark organic horizon which would have taken a minimum of hundreds of years or as many as several thousand years to form (Birekland, 1974; Press and Siever, 1974).