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TURBIDITES, DEBRIS FLOWS, AND TYPE 1 MELANGE OF THE CARRABASSETT FORMATION, EAST BRANCH PLEASANT RIVER

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INTRODUCTION

The Kearsarge-Central Maine (KCM) Synclinorium is the largest and most controversial belt of metasedimentary rocks in New England. The principal controversy centers on the origin of the marine basin that received the now deformed sediments. Although ensialic rifting may explain some pre-Acadian basins in the New England Appalachians (e.g. Aroostook-Matapedia trough, Ludman and others 1993), a similar origin for the KCM basin cannot be readily assumed because it lacks preserved shallow-water facies along its southeastern margin. Bradley (1983) proposed that unsubducted crust of the Iapetus ocean formed the basin floor, while others (e.g. Ludman and others 1993, and Hopeck this volume) support basin formation on sialic crust by syn/post-Taconic extension. Although the events leading to the origin of the KCM basin[1] are unclear, events accompanying its destruction are better understood due to recent paleoflow and sedimentological studies of Devonian metasediments (Hall and others 1976, Hanson 1988 and 1989, and Hanson and Bradley 1989).

The onset of Acadian orogenesis in Central Maine was attended by the westerly progradation and migration of a fine-grained clastic wedge represented by the Madrid Formation and the Seboomook Group. The advent of easterly derived sediments foreshadowed the approach and subsequent convergence of the Taconic-modified Laurentian margin with one or more terranes to the east. The first documented influx of sediment from the east is represented by the Madrid Formation, which was shed from northeasterly sources and deposited along the basin axis (Hanson and Bradley, 1989, 1993). This first pulse may record the convergence and subsequent uplift of the Saint Croix and Miramichi belts to the northeast (West and others 1992). The second pulse, which arrived from the southeast, is documented by the Carrabassett Formation, the principal member of the Seboomook Group in central Maine and the focus of this trip. This northwesterly pulse of sediment may reflect the initial impingement of Avalon against the New York Promontory.

East Branch Pleasant River

In most areas of New England, clues revealing the Early Devonian depositional system were largely destroyed by Acadian deformation and metamorphism. Sedimentary features are either altered beyond recognition or are involved in complex structures that cannot be confidently retrodeformed. However, an extraordinary series of outcrops, where sedimentary features are remarkably preserved and well-exposed, is located along the East Branch Pleasant River. Along strike to the southeast, additional informative outcrops are exposed on Big Wilson Stream and Borestone Mountain. (Trip C4) Together, these important localities offer unique insights into the complexity of the Carrabassett Formation and the early Acadian sedimentary environment.

Carrabassett Formation

The Carrabassett Formation (Boone 1973) is a thick (>>5 km), fine-grained sedimentary assemblage of pelitic turbidites and other mass-flow deposits, with subordinate but highly visible packages of massive and medium-bedded turbiditic sandstone (Hanson 1988). Disrupted strata form more that 60% of the exposed formation and often occur in belts more than a kilometer in width. In some instances either gravity sliding or faulting can be identified as the cause of disruption. However, in many exposures the relative importance of sedimentary vs. tectonic disruption is obscure. Concealed by such ambiguities is the formation's true thickness and degree of imbrication by faulting.

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[1] KCM "basin" refer to the depositional trough into which the sediments of the KCM Synclinorium were deposited. The original basin was probably not geographically located at the present site of the Synclinorium. Destruction of the basin was ultimately accompanied by thrusting of the cover rocks, some undefined distance, over the adjacent basement.

Nevertheless, the abundance of disrupted strata is the hallmark of the Carrabassett Formation and therefore is an important clue to its formation.

The high volume of fine-grained turbiditic and disrupted strata composing the Carrabassett Formation suggests that it was deposited basinward from a rapidly prograding deltaic system in a tectonically active basin. The rarity of coarse sediment (>medium sand) and the occurrence of correspondingly fine-grained deltaic deposits (e.g. Tarratine, Tomhegan, Matagamon, and Jo-Mary Mt. Formations) indicate a source area of poorly consolidated fine-grain sediments. A high sedimentation rate for the Carrabassett Formation is supported by its abundant slump, debris, and pelitic turbidite deposits, minimal bioturbation, and low carbonate content. Such a high sediment influx could be explained either by 1) input from a fluvial system draining a large continental area, or 2) input from smaller fluvial systems draining a rapidly eroding and poorly consolidated source area. Our present understanding of the limited areal extent of the Avalonian source area, and knowledge of early Acadian deformation towards the east, both favor the latter explanation. Further, the geochemistry of the Carrabassett Formation supports an interpretation that incorporates sediment recycling (Hon and others 1992, Philippopoulov and others 1992). The following scenario best explains the tectono-stratigraphic environment of the Carrabassett Formation: Closure of the KCM basin was accompanied by deformation and concomitant uplift of marine sediments accreted to the Avalon margin to the southeast. Erosion of the rising hills and deposition of sediments into the adjacent foredeep of the KCM basin formed a cannibalistic system that migrated westward as convergence continued.

Events related to early plutonism could also have influenced resedimentation in the Carrabassett Formation. Instability within the northwestern portion of the basin might have been driven by faulting related to intrusion of the Moxie Pluton. (See Trip C5, Gabis and others.) Early intrusive activity is another event, albeit poorly understood, that could have contributed to the formation of chaotic strata, and would also explain aberrant northeasterly paleoflows recorded in the Carrabassett Formation north of Greenville (Hanson and Bradley unpublished data).

In summary the depositional model for the Carrabassett Formation includes a series of slope and mud-rich debris-lobe and fan deposits formed in association with the inner slope of a northwest-migrating foredeep (Hanson 1988, 1989, Hanson and Bradley 1989, 1993, 1994). The unstable slope, which was locally interrupted by slope basins, descended into a muddy turbiditic fan system. Slope instability resulted from high sediment influx, earthquakes accompanying basin closure, and over-steepening of the inner slope by thrusting. Debris flows carried large quantities of mud-rich sediments down slope and into the foredeep. Extrusion of muds along emerging thrusts may have provided additional debris deposits. (See discussion on disrupted facies.) Tremendous quantities of fine-grained sediment were discharged along the basin margin through deltas; the remains of which are preserved locally on Jo-Mary Mountain (prodelta) and to the northwest in the Moose River, Roach River, and Traveler Mt. synclinoria. Although the overall trend of deposition during the Late Silurian and Early Devonian was from east to west, the Carrabassett Formation was deposited from the south to the north. Shifting source areas from northeast (Madrid Fm.) to south-southeast (Carrabassett Fm.) back to northeast (North East Carry Fm.) may reflect discrete terrane accretionary events or alternating impingement of Avalon on the Saint Lawrence and New York promontories (c.f. Catskill delta complex, Ettensohn 1985).

Sedimentary Facies of the Carrabassett Formation

The depositional model described above is the product of detailed facies mapping (Hanson, 1988). Sedimentary units, called *lithofacies* (e.g. graded pelite, massive sandstone, etc.), reflect specific depositional conditions. Packages of related lithofacies form *facies associations*, that reveal the subenvironments within the prevailing depositional system. The facies model outlined in Table 1 for the Carrabassett Formation, and the vertical stacking of facies that it attempts to explain, is based on the assumption that facies associations are largely related to fluctuations of the local depositional environment within a migrating and prograding depositional system. In all probability tectonic pulses resulting in changes of relative sealevel were also important controls in facies distribution. (See Mutti, 1985.) For example, a sandstone-rich turbidite assemblage overlying pelitic slope sediments (e.g. Gauntlet Falls) could be interpreted as either a slope basin deposit or a influx of sand related to a temporary drop in relative sea level.

Lithofacies (See Appendix for description of classification.)

Disrupted facies. This facies category includes strata broken by syndepositional mass flow and/or premetamorphic faulting, and is volumetrically the most dominant facies in the Carrabassett Formation. Chaotic units[2] are a major component of the Seboomook Group, indicating that rapid sedimentation and active slopes prevailed throughout its depositional history. However, the Carrabassett Formation is unique in that it also contains abundant F2-3 disrupted facies that exhibit clast fabrics composed of stretched and broken bedding. These units of broken formation occur in broad belts and were related earlier (Hanson 1988, Hanson and Bradley 1989) to the Type 1 melange of Cowan (1983). Some of these units contain apparent phacoidal (scaly?) cleavages, which may be relict premetamorphic tectonic features. The abundance of F2-3 disrupted facies in association with other chaotic strata suggests that mass-movement and faulting were contemporaneous, if not mutually controlled events.

A modern analog might be found in Indonesia where fault-related mass-flow deposits were observed by Breen and others (1986) along the lower slope of the Sunda Arc accretionary prism. Here muds, extruded upward along thrust faults, form linear mud ridges that are later over-ridden by thrust sheets. A similar process may be evoked for some disrupted strata in the Carrabassett Formation; for example, as the toe of the slope encountered and over-rode the basement of the Piscataquis magmatic belt, buttressing would have aided in the upward deflection and emergence of thrust faults, and the formation of mud ridges.

Another explanation for facies F2-3 disrupted beds is that their phacoidal cleavages were produced by the overprinting of regional cleavage on chaotic beds (F2-4). The principle argument against this interpretation is the relative lack of facies F2-3 in the Seboomook Group of the Connecticut Valley Gaspe Synclinorium; where chaotic strata are nearly as abundant and regional deformation is as intense.

Pelitic and sandstone-rich facies. The Carrabassett Formation contains abundant pelitic turbidite, and subordinate medium- to fine-grained, sandstone-rich turbidite and mass-flow deposits. Pelitic turbidites and hemipelagites(?) blanketed the interchannel regions of both slope and fan realms (discussed below) and were the primary strata from which most chaotic beds were derived. Packages of sandstone-rich turbidites range in thickness from a few meters to a few hundred meters and are interpreted as; 1) slope-channel or gully deposits (Stop 1a), and 2) fan-channel and levee deposits, (Stops 2). Sandstone in the former association, typical of slope a setting, are laterally discontinuous and massive; beds are several meters thick and show evidence of fluidized flow. The latter association occurs in a fan environment, and is composed of stacked thinning-upward sandstone sequences that contain individual sandstone beds ranging from less than two meters at the base to a centimeter or less at the top.

Sedimentary Facies Associations

The depositional system of the Carrabassett Formation contained two interfingering realms that produced a complex internal stratigraphy; 1) an inner, slope-dominated realm, and 2) an outer, fan-dominated realm. Each realm is distinguished by a particular set of facies associations, many of which are laterally discontinuous. Although slope associations of inner realm may predictably overlie fan associations of the outer realm, the vertical sequence of associations within each realm typically varies both along and across strike. To further complicate the stratigraphy, fan facies, similar to those of the outer realm, may be interpreted as slope basin deposits of the inner realm.

The inner (slope-dominated) realm contains pelitic turbidites cut by narrow channels that are filled with thick, massive sandstones (facies B1) and debrites (facies F2-4). Chaotic deposits are abundant. Levee and interchannel deposits are relatively sand poor because flow stripping (Piper and Normark, 1983), coupled with steep channel walls, inhibited overbank sand deposition. Sediment remobilization by mass-wasting, as evidence by abundant chaotic deposits, produced a series overlapping debris lobes that blanketed the lower slope environment.

The slope deposits graded basinward into the less chaotic and more organized fan deposits of the outer realm. Filled fan channels typically contain amalgamated facies-B2 sandstones that grade upward and laterally into medium-

[2] The term *chaotic* applies to facies in this category that were disrupted by sedimentary rather than tectonic processes.

Inner Realm (Slope) Associations

Type localities=Borestone Mt, East Branch Pleasant River (north of Gauntlet Falls)

1. *Slope channel/gully*: Characterized primarily by massive sandstones and debrites (dominant lithofacies: B2, B1, F2-4, and M)
2. *Interchannel and levee deposits*
   a. *Unstable slope and debris fans:* Thick chaotic units with interbedded fine-grained (sand-poor) turbidites (dominant lithofacies: F2-1, F2-3, L, E, and D2)
   b. *Stable slope or slope basin:* Fine-grained turbidites and hemipelagite (lithofacies: L, E, and M); E and L dominate levee deposits.
3. *Tectono-sedimentary facies(lower slope?):* (lithofacies: F2-3)

Outer Realm (Fan) Associations*

Type locality=Big and Little Wilson Streams, Elliotsville

1. *Channel deposits:* Characterized primarily by thinning and fining upward sandstone-rich turbidite sequences up to 300 m thick (lithofacies: B1, C, D1, F2-4, D2 and E)
2. *Interchannel and levee deposits:* (lithofacies: D, E, L, M and F2)

*This association may also represent an inner realm slope-basin deposit, or a Type 2 turbidite system (Mutti 1985) form during a relative sea level low, perhaps reflecting an uplift pulse.

Table 1. Facies associations of the Carrabassett Formation. Most common lithofacies are listed in parentheses and described in the Appendix to this guide.

and thin-bedded sandstone-rich levee deposits. Interchannel regions are characterized by thick accumulations of mud turbidites and subordinate hemipelagites. Chaotic deposits in the outer realm are almost exclusively debrites (facies F2-4), associated with channel sandstones. (See trip C4.)

Internal stratigraphy

For reason discussed above, the Carrabassett Formation lacks a regionally predictable internal stratigraphy. Although all facies associations are generally present in a single map area, on a regional scale their relative position and abundance vary both vertically and laterally. (Compare Figure 3 of this trip with Figure 2 in Trip C4.) The lack of stratigraphic order is an inherent characteristic of the depositional environment; fan deposition that occurs in both foredeep and slope basins, fluctuations in sea level, shifting channels, ephemeral gullies, and recurrent submarine slides all resulted in an irregular stacking of facies associations. Further complications arise from the presence of intraformational (premetamorphic?) faults, which are particularly hard to distinguish from (and in) chaotic assemblages.

Local metamorphism and structure

The East Branch Pleasant River flows southeast, exposing the metasediments on northwestern limb of the Kearsarge-Central Maine Synclinorium. The Carrabassett Formation underlies most of the river and the adjacent highlands and locally lies within the contact aureole of the Katahdin and Moxie plutons (Fig. 1). Regional deformation and metamorphism was followed by contact metamorphism (and a subsequent retrograde event) that accompanied the emplacement of Moxie and Katahdin plutons, which are exposed to the northwest. The width of the contact aureole (>6 km) suggests that contact between the plutonic zone and the metasediments dips gently toward

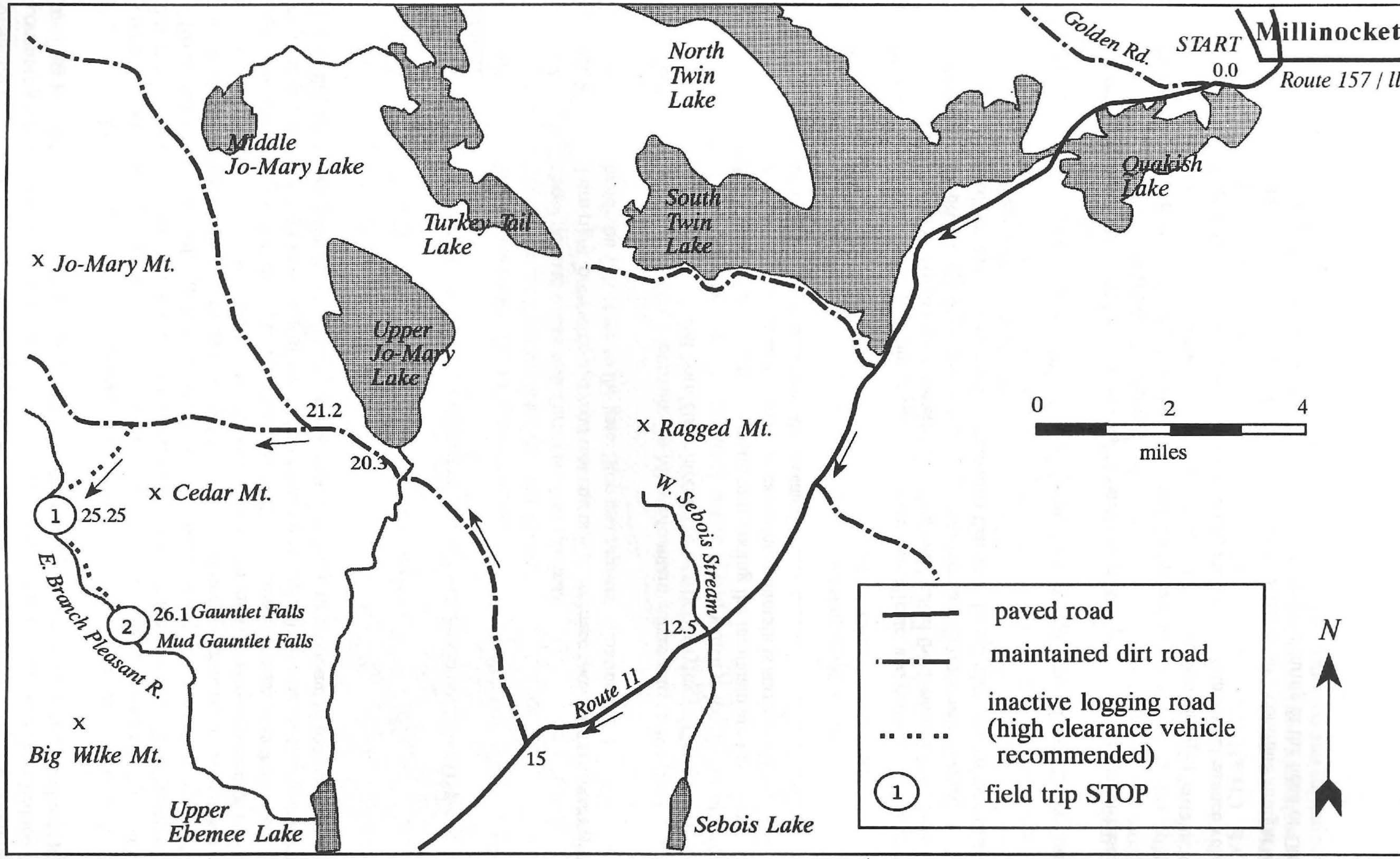

Figure 1. Location map of field trip stops.

the south. Contact metamorphism annealed slaty cleavages formed by regional metamorphism and enhanced the visibility of sedimentary structures. Pelitic beds recrystallized forming medium- to coarse-grained assemblages containing biotite, cordierite, andalusite, muscovite, and quartz in varying amounts. Interbedded hornfelsic sandstones are finer grained, quartzose-feldspathic assembleges that typically lack cordierite and andalusite porphyroblasts.

Locally, deformation accompanying regional metamorphism resulted in broad (>2 km), doubly plunging open folds that typically plunge less than 10°. Isoclinal folds or nappes are not present. Bedding dips commonly range between 45° and 80°, however dips less than 25° are common. (The flat top of Jo-Mary Mountain reflects near-horizontal bedding.) Overturned beds occur only in slump folds or along faults (Stop 2b). Whether the occurrence of gently dipping strata is related to depositional unconformities, or location along fold axes, or both is still unresolved. The first and last of these relationships would imply upper-slope and/or basin deposition on a rising and deforming thrust complex. The stratigraphy (Fig. 3) along the East Branch Pleasant River, from the old railroad trestle (Stop 1a) to the axis of the Wilkie Mountains (Fig. 2) syncline does not disallow this possibility. This highly schematic column was drawn from exposures along the river and in the adjacent highlands.

ROAD LOG

Road log begins at the Junction of Route 11 (Brownville Road) and the Golden Road (Fig. 1)

Mileage	
0.0	Head south on Route 11 towards Brownville.
3.0	Cross over the West Branch Penobscot River.
6.7	Cross railroad tracks near Partridge Cove Marina.
15.00	Turn right onto wide dirt road that enters K.I. Jo-Mary Mountain recreational area. After a broad bend you will come to the gate. Stop and register. Continue.
18.80	View of Jo-Mary Mountain. Jo-Mary Mountain comes into view from the northwest. The mountain is resistant hornfels composed of nearly flat-lying (10°±) pelitic and silty turbidites of the Jo-Mary Mountain formation (Hanson, 1988), which is locally the uppermost member of the Seboomook Group. The lowlands surrounding the mountain are underlain by less resistant migmatite, granite, and diorite. Because the mountain is surrounded and underlain by plutonic rock and migmatite, it is structurally isolated.
19.8	Gravel pit
20.30	Rusty-weathering, thin-bedded sandstone, calcsilicate, and pelite is exposed in banks and sometimes in low road pavements. This unit, called the Johnston Pond formation (Hanson, 1988), locally separates the Jo-Mary Mountain and Carrabassett Formations. It is a relatively shallow-water, nonturbiditic formation that may be correlative with the Hildreths Formation of Osberg and others (1968).
20.66	The road to Jo-Mary Campground joins from the right (northeast). Do not turn. Continue straight.
21.20	Road splits. Take lefthand fork. Continue straight, passing three gravel-based logging roads which enter from left (south).
23.90	Road splits. Take lefthand fork towards Jo-Mary Pond.
24.70	Road from Cedar Mountain joins from the south. Continue straight.
25.25	Turn left (south) on road to Gauntlet Falls. Participant with low-clearance vehicles may want to hitch a ride.
26.10	Park in clearing on west side of road. STOP. After Stop 1 continue straight to Gauntlet Falls (mileage 27.8) and STOP 2. At stop 2 Park in large clearing at the head of Gauntlet Falls.

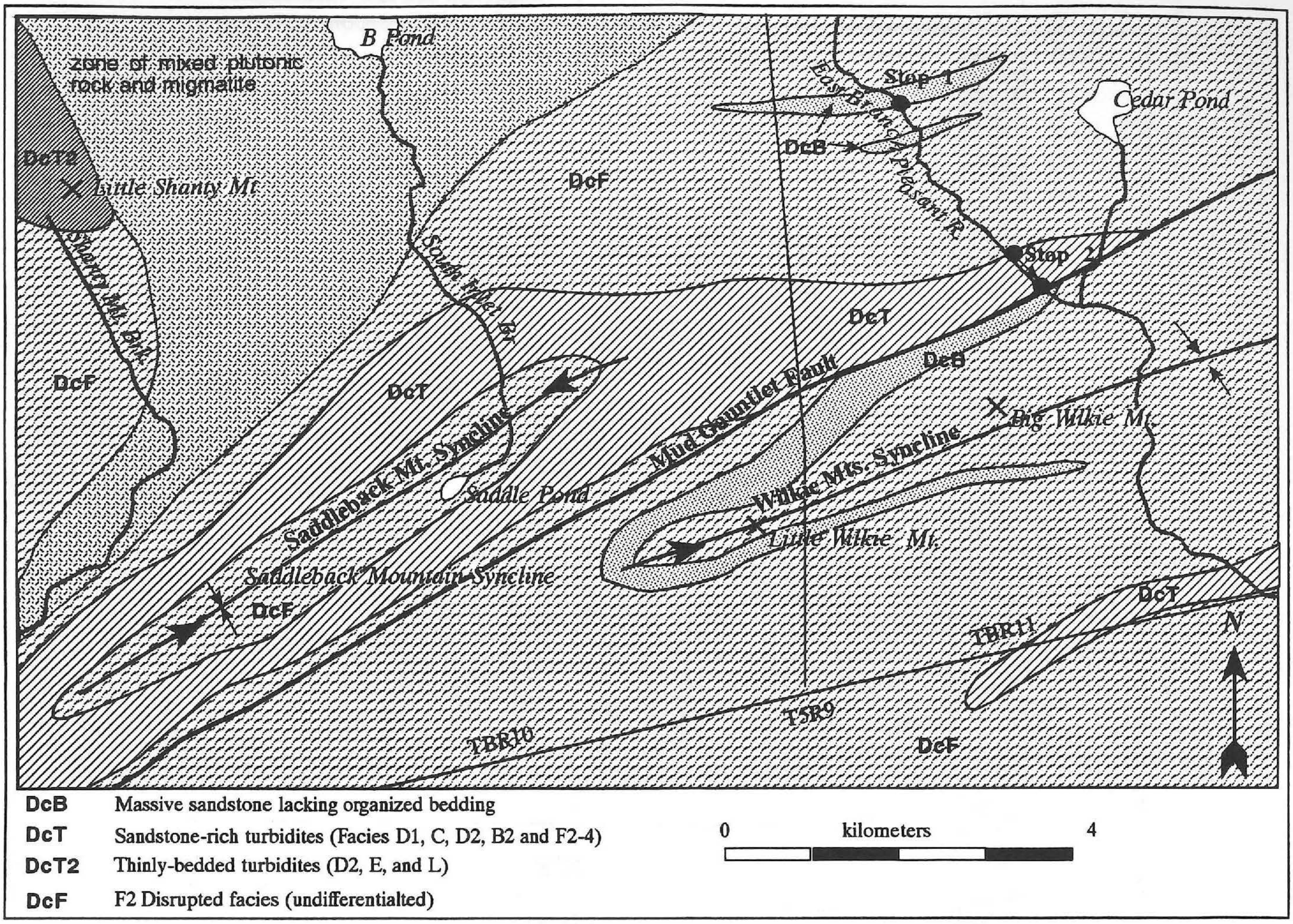

Figure 2. Geologic map of the East Branch Pleasant River region (After Hanson 1988). DcB and DcT are subdivisions of the sandstone unit (Dsc) on the Bedrock Map of Maine (Osberg et al. 1985).

STOP 1 EAST BRANCH PLEASANT RIVER--OVERVIEW. As shown in the generalized column in Figure 2, there is an overwhelming abundance of disrupted strata exposed along the East Branch Pleasant River. Most are chaotic beds derived from remobilized pelitic turbidites and hemipelagites. Bedded turbidites found between chaotic bodies are evidence of multiple slump and debris-flow events. Chaotic deposits form distinct hummocks that can be seen along the banks of the river and in the surrounding highlands.

There are probably more faults cutting this section than indicated in Figures 1 and 2. The thick chaotic zones camouflage faults within the formation. Also, because of the formation's complex stratigraphy, and general lack of unique, laterally-continuous marker beds, faults are not easily located or inferred where not exposed.

STOP 1a MASSIVE CHANNEL SANDSTONE. Exposed on the east bank, at the site of an old railroad trestle, is a massive metagraywacke. A tenth of a mile downstream (Stop 1b) the sandstone is abruptly overlain by chaotic strata. Upon close inspection you can see that the sandstone contains rare contorted pelitic laminae and deformed clasts of both pelitic and sandy

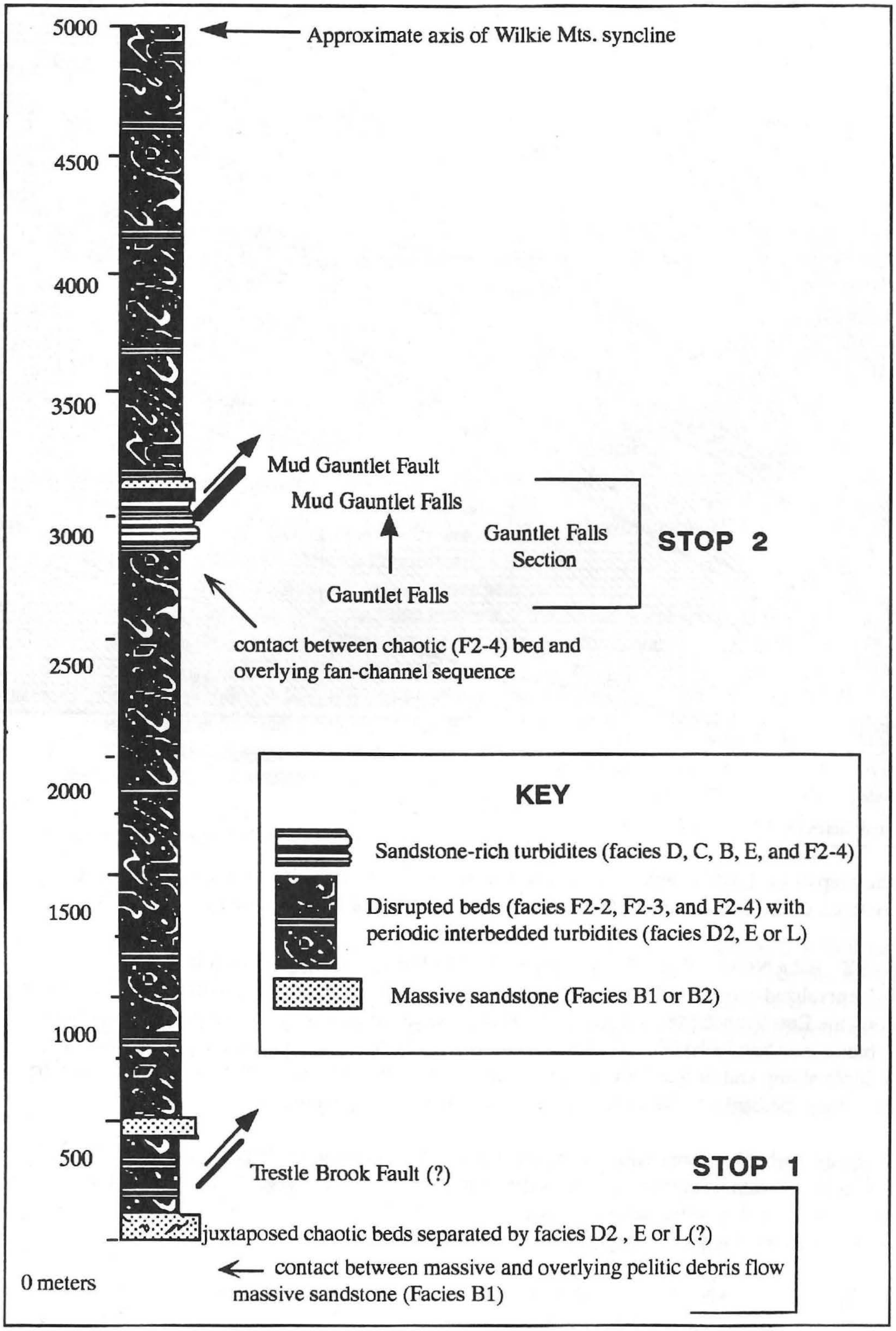

Figure 3. Stratigraphic column of the Carrabassett Formation. Compiled from exposures along the East Branch Pleasant River and along strike in the adjacent highlands.

composition. The bed was rapidly deposited, probably from a fluidized flow.

In 1983 I first correlated this sandstone with thick-bedded sandstones in the Upper Madrid Formation, primarily because the sandstone is massive, appears to contain calcsilicate, and lies beneath a massive, albeit chaotic, section of the Carrabassett Formation. These characteristics seemed to fit the prevailing lithologic and stratigraphic descriptions for the Madrid Formation. However, after defining the sedimentary facies of the Madrid Formation (Bradley and Hanson 1992) and after more detailed local mapping, I found some noticeable problems with my earlier interpretation. First, this chaotic sandflow exhibits none of the interbedded facies D turbidites or graded beds typically observed in the Madrid Formation; second, this sandstone is laterally discontinuous; and third, it is stratigraphically underlain and overlain by chaotic beds.

While viewing the sandstone, walk downstream to its contact (Stop 1b) with overlying chaotic beds.

STOP 1b. CONTACT BETWEEN SANDSTONE AND OVERLYING CHAOTIC PELITE. Here the massive sandstone is abruptly overlain by pelitic slump and debris flow deposits. The actual contact is not exposed but can be located to within a few meters. The overlying chaotic unit is readily recognized by its rounded outcrop form that mimics the slump fold structure.

Walk an additional 100 meters to the next exposure of chaotic strata (Stop 1c).

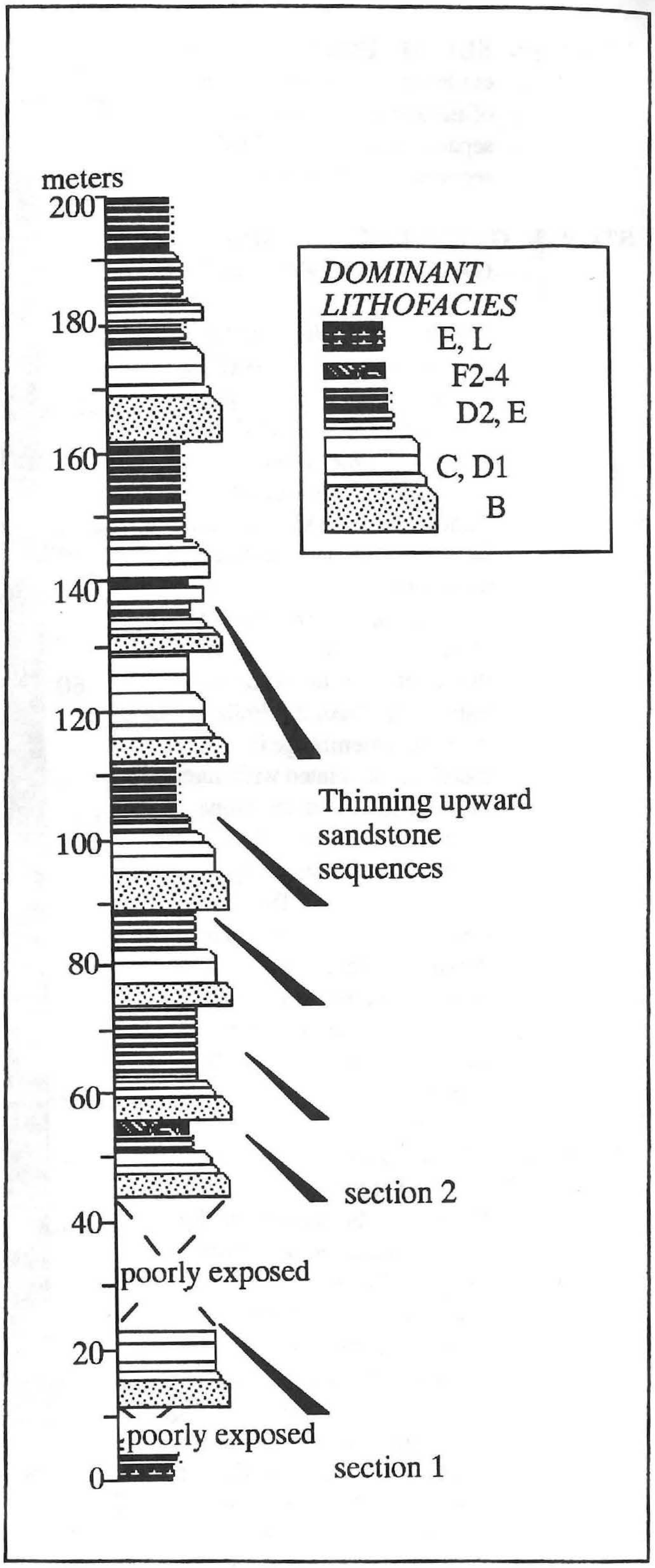

Figure 4. Schematic column of crude thinning-upward sandstone sequences in the Gauntlet Falls section.

STOP 1c. SLUMP FOLD. Look for evidence indicating that some of these chaotic units are separated by thinly-bedded sequences of turbidites.

STOP 2. GAUNTLET TO MUD GAUNTLET FALLS. A 200-meter thick, gently dipping turbidite assemblage is exposed between these two falls (Figs. 4 and 7). The section exposes several poorly organized fining-upward, channel-levee sequences, and is locally truncated by the Mud Gauntlet fault. However, on Saddleback Mountain the assemblage is stratigraphically overlain by chaotic strata, similar to those that underlie it at Gauntlet Falls. The Gauntlet Falls turbidite assemblage is therefore associated with inner-realm chaotic and sandstone facies associations. This stratigraphic relationship suggest that either these channel/levee deposits were laid down in a slope basin, or at a time when a relative lowering of sea level increased turbiditic sand deposition (c.f. Mutti 1985).

STOP 2a. GAUNTLET FALLS. Exposed at the head of Gauntlet Falls is the broad hummocky surface of a large debris flow. Draping the debris flow is a discontinuous massive sandstone, overlain by a sequence of interbedded pelites and thin wavy sandstones (Fig. 5). Pelitic beds are 1-15 cm thick. Sandstone and siltstone beds are .5-3 cm thick, and contain ripples with flat bases and wavy ungraded tops (facies E). The ripples have linear, to slightly sinuous, asymmetrical

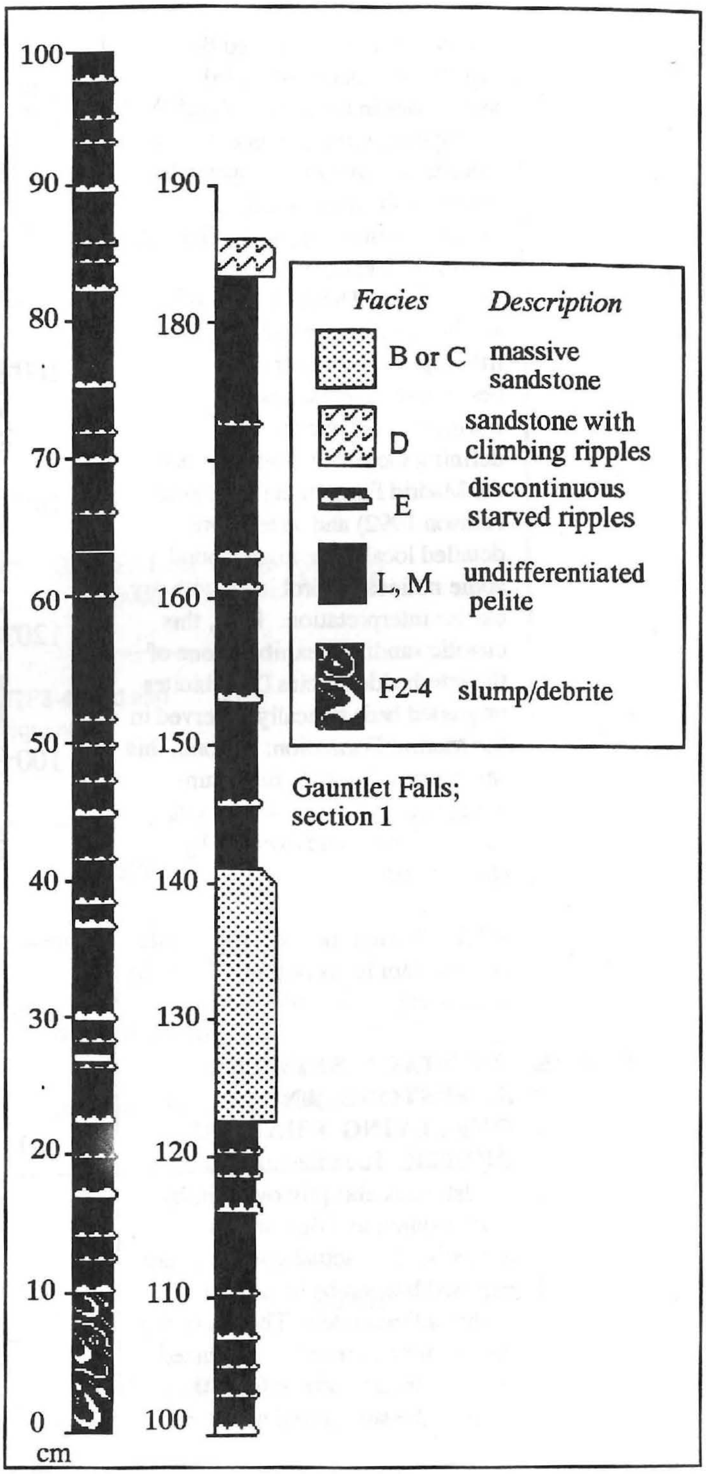

Figure 5. Detail of pelites and thin-bedded sandstones overlying debris flow at Gauntlet Falls.

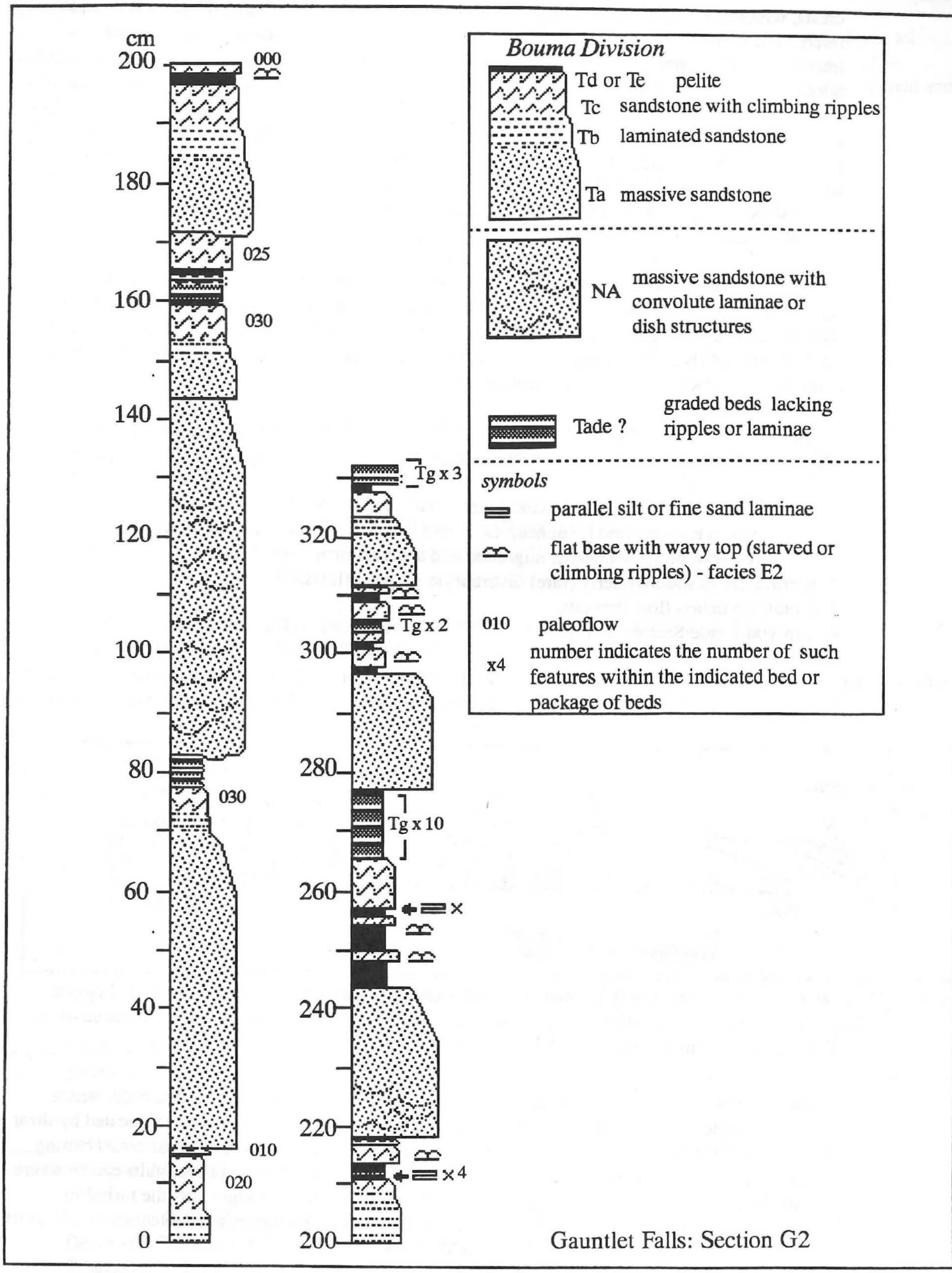

Figure 6. Detailed section of thinning-upward sequence within the Gauntlet Falls Section.

crests, which are approximately 2 mm. high with a mean spacing of 7 cm. These wavy ripples are interpreted as overbank sediments, deposited along a channel margin or in some proximal interchannel environment (e.g. levee). Ripples exposed on bedding-plane surfaces were formed by northward flows, ranging from N15E to N35W with a mean of N15°W. Beds dip only 0° to 20°.

Above the facies E sequence the section is not well exposed, but appears to exhibits a fairly rapid thickening-upward trend, interrupted episodically by debris flow deposits. The 200-meter sequence of turbidites is an fan-like deposit formed by one or more channels that meandered along the base of the slope, or within a local slope basin. Sandstones are quartzose-feldspathic and typically fine- to medium-grained.

Walk south towards Gauntlet Falls. Take time to look at the beds composing the section. Contact metamorphism recrystallized pelitic beds and laminae forming coarser mineral assemblages that are easily distinguished from their more quartzose sandy interbeds. The contrast results in an enhancement of sedimentary features that is amazingly sharp. Bouma sequences in turbidite assemblages and stratabound debris flow deposits are obvious.

As you walk downstream to the Mud Gauntlet fault, about 500 meters, try to observed the following features exhibited in this turbidite package:

1. Thick-bedded B, C, and D facies sandstones crudely organized into fining upward sequences and replaced by thinner D, E, and L facies. This cyclic sedimentation is the result of periodic channel migration and abandonment.
2. A northerly to southwesterly (rare) distribution of paleoflows.
3. Stratiform debris flow deposits.
4. Can you locate Section 2 (Fig. 6)

STOP 2-b: MUD GAUNTLET FAULT. The Mud Gauntlet fault is a 50-meter wide, reverse fault that truncates the Gauntlet Falls turbidite section (Fig. 7), and repeats the underlying massive sandstone

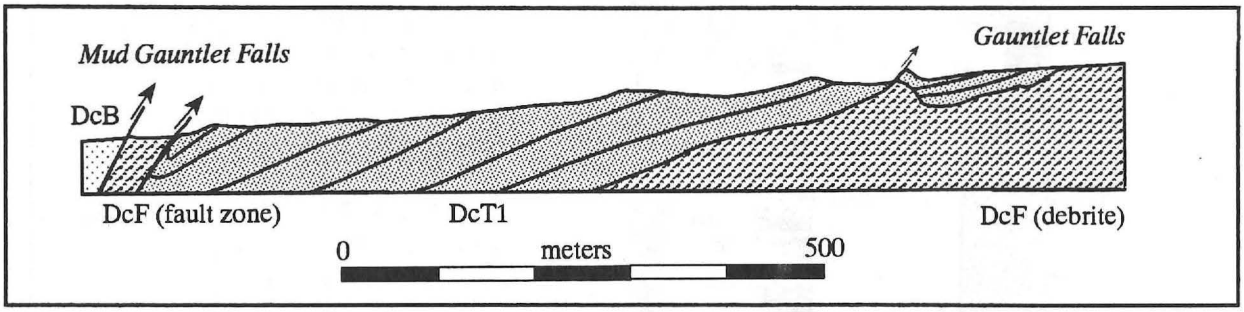

Figure 7. Cross-section of the Carrabassett Formation between Gauntlet and Mud Gauntlet falls (After Hanson 1988). Strata are labeled as follows: DcF (disrupted beds), DcT (thin and thick-bedded sandstone-rich turbidites), and DcB (massive sandstone).

(Stop 1). The fault zone contains slivers of tectonized chaotic strata and turbidite beds, which have been stretched and broken. Turbidite beds and fragment foliation are locally truncated by shear planes. Small faults cutting the turbidite strata are clearly visible. However, shear zones cutting the pelite-rich chaotic facies are much less obvious, demonstrating how elusive faults can be where chaotic facies dominant. Along strike to the northeast, the fault no longer cuts the turbidite section but extends into a broad belt of chaotic facies where its no longer traceable.

End of Field Trip

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APPENDIX

Sedimentary Lithofacies

Lithofacies categories

The lithofacies of the Carrabassett and Madrid Formations fall into three principle categories, outlined in Figure 1: a) sandstone/siltstone-rich turbiditic facies (Table 1); b) pelitic-rich turbiditic facies (Table 2); and (c) disrupted facies (Table 3). A sequence of related lithofacies combine to form a *facies association* that characterized a particular depositional environment.

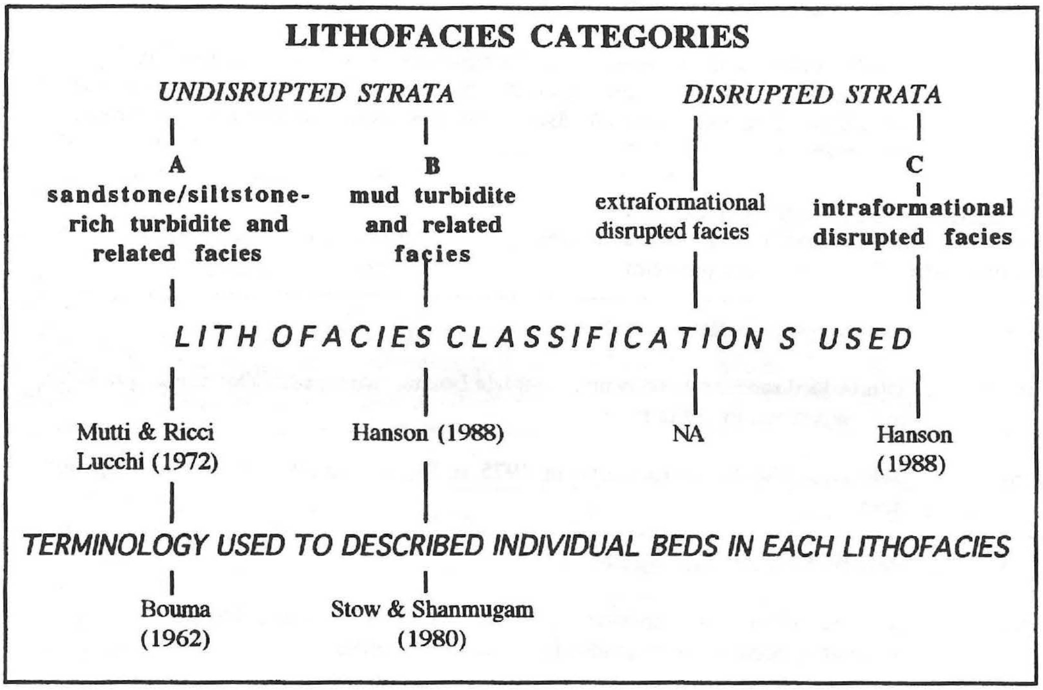

Figure 1. Facies categories and classifications applied to the Carrabassett and Madrid Formations.

Sandstone/siltstone-rich turbidite facies category. This facies category includes relatively sandstone-rich facies, typically containing greater than 40% sandstone. The sands were deposited by turbidity currents, grain flows, or some other related mechanism.

Mud turbidite and related facies category. This category includes pelitic strata deposited by mud turbidites and hemipelagic settling. They may also occur as distal deposits and overbank deposits of more sand-rich turbidity currents.

Disrupted facies categories. The intraformational disrupted-facies category includes all strata characterized by premetamorphic or synsedimentary disruption, such as submarine debris flow, slump or slide deposits, beds disrupted by diapirism, and strata disrupted by premetamorphic or synsedimentary faulting. Intraformational disrupted units are composed of strata that were derived locally from within the same basin. Extraformational disrupted units, which are not found in the Carrabassett or Madrid Formations, contain exotic blocks that have been transported into the basin from some other terrane.

Facies A	**Arenaceous-Conglomeratic Facies**
Description	Conglomerates, pebbly sandstones and medium- to very coarse-grained sandstone; Bouma sequence not applicable.
Comments	*Absent from the Carrabassett Formation*, but common in the Rangeley Formation.

Facies B	**Arenaceous Facies**
Description	Laterally pinching, thick-bedded, medium-fine to coarse-grained massive or laminated sandstones. Dewatering structures are common along with flutes and load casts.
Comments	Present, but not a major facies of the Carrabassett Formation. Sandstones in the Carrabassett Formation are medium-fine to medium grained and generally lack dish structures. Commonly indistinguishable from facies C unless lateral continuity is observable.

Subfacies defined for the Carrabassett include:
B 1: crudely laminated to massive organized sandstones
B 2: massive unorganized sandstones (chaotic)

Facies C	**Arenaceous-Pelitic Facies**
Description	Interbedded sandstone and pelite: complete Bouma sequences (Tabc) often represented; high sandstone to pelite ratio.
Comments	Subfacies of Mutti and Ricci Lucchi, 1975; or Walker and Mutti, 1973 are not applied here.

Facies D	**Pelitic-Arenaceous Facies I**
Description	Interbedded pelite and sandstone (siltstone); Ta or Ta-Tb Bouma divisions missing; moderately high to low sandstone (siltstone) to pelite ratios.
Comments	Tops and bases of sandstones are planar

Subfacies defined here for the Carrabassett Formation:
D 1: Tc sandstone typically less than 5 cm
D 2: Tbc and Tcd sandstones 10 - 50 cm common

Facies E	**Pelitic-Arenaceous Facies II**
Description	Lenticular beds of sandstone characterized by flat, sharp bases and wavy tops interbedded with pelite.
Comments:	Facies E in the Carrabassett Formation are composed of medium fine-grained sandstone. Sandstone (siltstone) to pelite ratio is variable. Ripple tops may be graded and contain pelite laminae which thicken toward the troughs. Interpreted as overbank deposits. This facies forms a continuum with facies in the Pelitic Facies category.

Table 1. Sandstone and siltstone-rich turbidite and related lithofacies. (Modified from Mutti and Ricci Lucchi 1972.)

Lithofacies classification

Sandstone-rich turbidites facies. The lithofacies classification used for these strata was adapted from Mutti and Ricci Lucchi (1972) and is outlined in table 1. With the exception of Facies A and B, the sandstone-rich turbidite facies (Table 1) are primarily classified by the Bouma sequences composing the sandstones. Bouma (1962) noted that the sedimentary structures with a turbidite bed that was deposited during a single event, occur in a predictable pattern or sequence. The complete sequence is shown in Figure 2. The Bouma sequences present reflect the strength and sand content of the turbidity currents, and the distance traveled. Turbidite beds are described using *T* (standing for turbidite) followed by the small letters (a,b,c,d and e) indicating the divisions present in the bed.

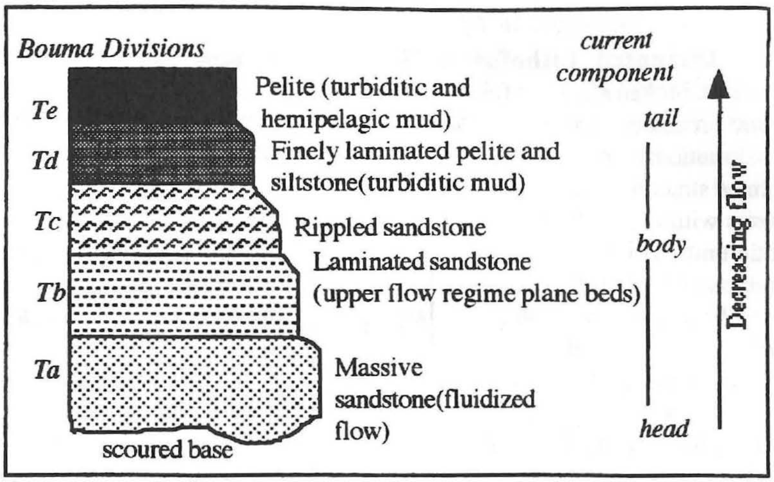

Figure 1. Complete Bouma sequence with description of Bouma divisions (Bouma, 1962)

Pelitic Lithofacies (Table 2). Pelitic strata are subdivided into two general facies (Hanson, 1988); laminated pelite (L), and massive or uniform structureless pelite (M). Beds and laminae composing these facies are described (Hanson 1988, Hanson and Bradley, 1989) using the terminology of Stow and Shanmugam (1980). Poor preservation of sedimentary features in most pelitic rocks of the Carrabassett Formation has hindered attempts to use the original facies classification of Stow and Shanmugam (1980). Most pelitic strata in the Carrabassett Formation contain silt laminae spaced a few millimeters to as much as 30 cm apart. In overbank and interchannel deposits the thickness and density of laminae reflect the distance from channel margins. There also appears to be a continuum between these facies and the more sandstone-rich facies D2 and E. Some thick structureless pelitic units were probably muddy debris-flow deposits and are often difficult to distinguish from non-disrupted massive pelite. Contourites have not been recognized in the Carrabassett Formation; the observed sedimentary structures and lack of extensive bioturbation suggest that most pelitic beds were deposited from gravity flows.

PELITIC FACIES (Hanson 1988)		**SEDIMENTARY STRUCTURES** (Stow and Shanmugam 1980)	
M	*Massive pelitic facies*	**P**	Bioturbated pelagic or hemipelagic pelite
		T7	Ungraded mud
		T6	Graded mud with or without silt lenses
L	*Laminated pelitic facies*	**T5**	Pelite with wispy, convolute silt laminae
		T4	Pelite with indistinct and discontinous silt laminae
		T3	Pelite with parallel silt laminae
		T2	Irregular to lenticular rippled silt layer in pelite
		T1	Convolute laminae in relatively thick pelite
		T0	Basal lenticular siltstone or fine-grained sandstone with climbing ripples

Table 2. Pelitic lithofacies of the Carrabassett Formation. These facies are related to the mud-turbidite sequence described by Stow and Shanmugam (column 1980).

Disrupted Lithofacies (Table 3). Disrupted facies are composed of either chaotically mixed lithologies, broken strata, or folded beds that were deformed or transported by syndepositional or premetamorphic-tectonic processes. facies F1, which is absent from these formations, contains clasts or blocks that are extraformational (not originating from within the basin). Intraformational chaotic facies (F2) are composed of deformed strata derived from deposits within the basin. All chaotic units within the Carrabassett Formation were derived from the mass-wasting or deformation of local strata. Subdivision of intraformational facies is based on the intensity of disruption. facies F2-1, -2 and -4 are common in the Carrabassett Formation and are the result of mass-wasting. Facies F2-3 is more problematic and may be related to premetamorphic faulting that accompanied early closure of the basin. (See Introduction.) Other processes, such as diapirism may have from chaotic units but would be difficult to identify.

F1	***Chaotic deposits with exotic clasts***
F2	***Chaotic deposits with intraformational clasts***
F2-1	***Extensional chaotic facies.*** Strata are cut by minor normal faults.
F2-2	***Moderately-disrupted chaotic facies.*** Strata are faulted and folded, but topping direction and original lithofacies is still recognizable.
F2-3	***Highly disrupted, foliated-chaotic facies.*** Similar to F2-4 but with the appearance of a fragment foliation or phacoidal cleavage.
F2-4	***Highly disrupted,Non-foliated chaotic facies.*** original bedding is unrecognizable.

Table 3. Classification of disrupted lithofacies.