

Sedimentary Facies and Tectonic Interpretation of the Lower Devonian Carrabassett Formation, North-Central Maine

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ABSTRACT

The Lower Devonian Carrabassett Formation in north-central Maine is a complex assemblage of metasedimentary rocks composed of chaotic and fine-grained turbidite deposits. Detailed measured sections provide the basis for recognizing several distinctive lithofacies within the Carrabassett Formation, which are organized into locally mappable units referred to as facies associations. In order of decreasing abundance these are the (1) chaotic, (2) thin-bedded turbidite, (3) pelitic(?), (4) thick-bedded turbidite, and (5) massive sandstone facies associations. Facies associations in the study area are laterally discontinuous on the scale of kilometers or less, and hence it is not surprising that they should fail to correlate one-for-one with the four members of the Carrabassett Formation at its type area, 100 km to the southwest. The chaotic facies association, locally comparable to the type 1 mélange of Cowan (1985), was found to comprise over 50% of the Carrabassett Formation in a reconnaissance survey of nine 15' quadrangles.

The Carrabassett Formation, the youngest, widespread pre-Acadian unit in the low-grade sector of the Kearsarge-central Maine synclinorium, was being deposited during the demise of a long-lived, deep-water basin. Facies analysis, paleocurrent data, and regional relations together permit the interpretation that the Carrabassett Formation was deposited in a northwest-migrating inner trench slope during the initial stages of Acadian collision. Local slope basins were filled by mud-dominated debris aprons that were traversed by submarine channels. NW-directed thrust faulting continually produced and regenerated steep slopes, whose episodic failure resulted in olistostromes with both extensional and contractional soft-sediment deformation fabrics.

INTRODUCTION

The Lower Devonian Carrabassett Formation is the youngest widespread formation exposed in the Kearsarge-

central Maine synclinorium in Maine. It was originally broken out from the Seboomook Formation in the Little Bigelow Moun-

tain 15' quadrangle by Boone (1973), who described it as a succession of metapelites and metasandstones. In the type area, even the most basic sedimentological observations were hampered by intense Acadian deformation and metamorphism. Consequently, the Carrabassett Formation contributed few details to studies of Lower Devonian paleogeography, prior to the inception of Hanson's studies in the early 1980's (Hanson, 1983).

This paper describes results of detailed facies investigations of the Carrabassett Formation along strike to the northeast of the type area, where the effects of Acadian tectonism are far less intense, and where primary sedimentary features are well preserved. Here, the formation can be recognized as an assemblage of metamorphosed turbidites, pelites, and olistostromes, the latter comprising as much as 50 to 100% of exposed sections. Facies mapping, paleocurrent analysis, and regional relations together permit the interpretation that the Carrabassett Formation was deposited in a northwest-migrating slope and foredeep(?) environment during the initial stages of Acadian collision.

GEOLOGIC SETTING

The Carrabassett Formation is exposed in a northeast-trending belt that includes portions of the western limb of the Kearsarge-central Maine synclinorium (Lyons et al., 1982) and the eastern flank of the Piscataquis magmatic belt (Fig. 1). The Kearsarge-central Maine synclinorium is a 50 to 100 km-wide belt of Upper Ordovician(?) through Lower Devonian(?) marine strata that extends nearly 500 km from Connecticut to north-central Maine. An enduring controversy in Acadian tectonics centers on whether or not the Kearsarge-central Maine synclinorium was the site of a pre-Acadian ocean basin that closed by subduction prior to an Acadian collision. Direct evidence is lacking, because the polydeformed metasedimentary fill of the Kearsarge-central Maine synclinorium is nowhere known to be attached to depositional basement, nor are obducted basement rocks known to exist. What can be established is that deep-water (i.e. below wave base) conditions and rapid rates of sediment accumulation persisted from about Upper Ordovician to Lower Devonian times. Because it was deposited just before or during the transition from sedimentation to orogeny, the Carrabassett Formation is a key to deciphering early Acadian conditions.

The northwestern margin of the Kearsarge-central Maine synclinorium is flanked by pre-Silurian continental basement which is locally overlain or intruded by Upper Silurian through Medial Devonian volcanic and plutonic rocks: the Piscataquis volcanic belt of Rankin (1968) and the comagmatic Greenville plutonic belt of Hon (1980). The term Piscataquis magmatic belt is used here to include both volcanic and intrusive rocks. Geochemical studies suggest a subduction origin for at least the Devonian igneous rocks (e.g. Hon and Roy, 1981), but these results do not bear on the question of subduction polarity. One

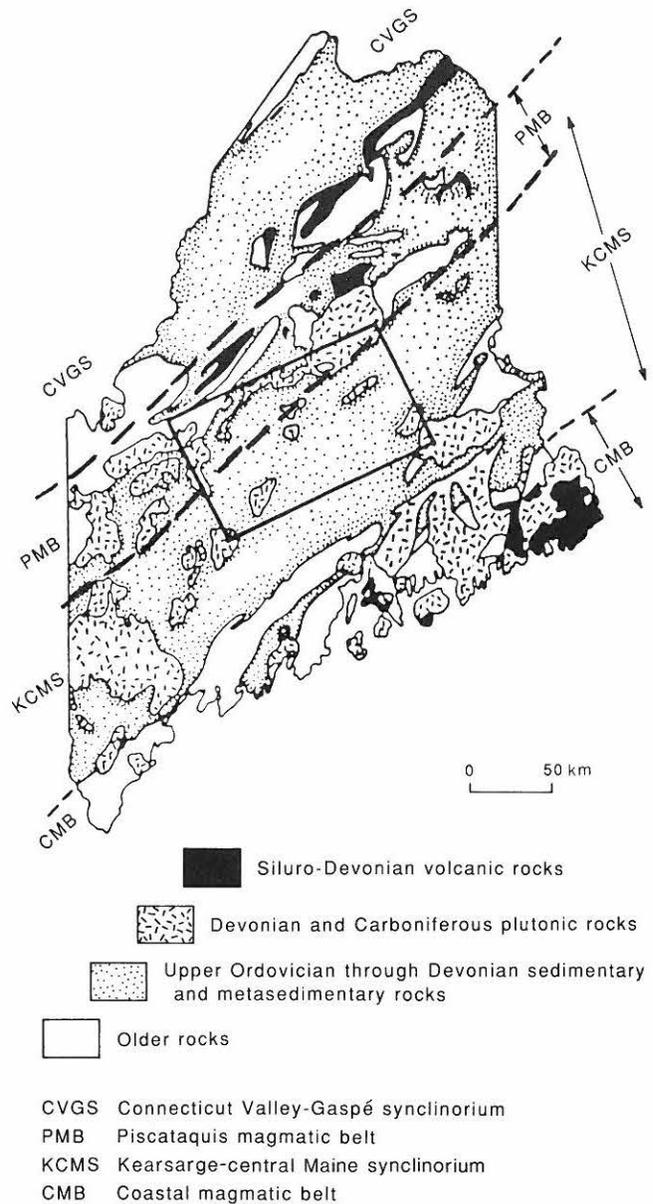


Figure 1. Generalized geologic map of Maine showing the distribution of Siluro-Devonian volcanic rocks, Devonian and Carboniferous plutonic rocks, Upper Ordovician through Devonian sedimentary and metasedimentary rocks, and older rocks. The coastal magmatic belt is roughly coextensive with the composite Avalon Terrane. The Kearsarge-central Maine synclinorium is roughly equivalent to the Merrimack synclinorium of Billings (1956), and the Merrimack trough as used by Bradley (1983; but cf. Lyons et al., 1982). Boxed area in central Maine shows location of Figure 2.

possibility is that a Piscataquis magmatic arc formed above a northwest-dipping subduction zone where oceanic basement of the Kearsarge-central Maine synclinorium was consumed (e.g. McKerrow and Ziegler, 1971; Hon and Roy, 1981; Bradley, 1983). Alternatively, Hon (pers. commun., 1987) has suggested that the arc formed above a southeast-dipping subduction zone

associated with destruction of oceanic crust in the Connecticut Valley-Gaspé synclinorium.

The Kearsarge-central Maine synclinorium is flanked on the southeast by Precambrian continental basement of the composite Avalon Terrane. Thick accumulations of Silurian and Lower Devonian volcanic rocks occur here in a belt that extends from Massachusetts to Nova Scotia. [Gates (1969) referred to this tract as the coastal volcanic belt; here we call it the coastal magmatic belt and include cogenetic intrusive rocks.] The coastal magmatic belt has been interpreted by several workers in terms of a southeast-dipping subduction zone along the southeastern margin of the Kearsarge-central Maine synclinorium (e.g. McKerrow and Ziegler, 1971; Osberg, 1978; Rodgers, 1981; Bradley, 1983). On the other hand, Gates and Moench (1981) have attributed the magmatism to lithospheric extension.

Results of the present study do not bear conclusively on the alternative tectonic interpretations of the Kearsarge-central Maine synclinorium, Piscataquis or coastal magmatic belts, nor on the larger controversy surrounding the nature of the Acadian orogeny as a whole. However, our results are consistent with an Acadian tectonic model developed by Bradley (1983) from ideas of McKerrow and Ziegler (1971). Tectonic implications are discussed in the concluding section.

AREA OF STUDY

This paper summarizes the results of reconnaissance (Fig. 2) and detailed (Fig. 3) mapping and section measuring in 1983-87 by Hanson (Hanson, 1988) and 1985-86 by Bradley. Hanson mapped the Jo-Mary Mountain and portions of the Norcross 15' quadrangles on contract with the Maine Geological Survey; related reconnaissance covered parts of the Millinocket, Sebec, Sebec Lake, and Greenville quadrangles. Key exposures were also studied (by Bradley and Hanson) in the Anson, Kingsbury, Rangeley, Phillips, and Little Bigelow Mountain quadrangles.

Most of the area of Figure 2 was subject to tight to isoclinal folding and greenschist-facies metamorphism during the Acadian orogeny, making detailed sedimentological studies difficult, but not impossible. In contrast, the area of Figure 3 lies within the contact aureole of the Moxie and Katahdin plutons (c.f. Fig. 2). Here the effects of regional deformation and metamorphism were substantially less (folds are open and tectonic foliations are subdued or absent), and primary sedimentary and soft-sediment deformation structures are beautifully preserved.

These observations suggest that contact metamorphism preceded regional deformation and metamorphism. Thin sections of spotted slates from the outer portion of the contact aureole (location 10, Fig. 3) support this order of events. Foliation, defined by the preferred orientation of micas, overgrows and is deformed around well-preserved andalusite and

retrograded cordierite porphyroblasts; there is no evidence of an older foliation in the porphyroblasts.

Thus, the hornfels offers a unique opportunity to study rocks which were frozen in time and not subjected to major Acadian overprinting. These rocks should be particularly useful in unraveling the sedimentary and tectonic environments that preceded the culmination of the Acadian orogeny.

STRATIGRAPHY

Most formations in the area of Figures 2 and 3 have been previously described in western and central Maine (Osberg et al., 1968; Boone 1973; Pankiwskyj et al., 1976; Moench et al., 1982). The oldest exposed strata (Fig. 4) are Lower Silurian sandstone and slates, which have been mapped as the Allsbury Formation (Ekren and Frischknecht, 1967; Neuman, 1967; Roy, 1981), and which have been correlated with the Sangerville Formation (Roy et al., 1983; Hill and Roy, 1985). The Allsbury Formation is overlain by an unnamed rusty-weathering unit of thin-bedded turbidites of metapelite and metasandstone. This unnamed assemblage has been assigned to the Smalls Falls Formation on the basis of stratigraphic position and lithology.

Feldspathic, locally calcareous, and sandstone-rich metaturbidites overlie the Smalls Falls Formation and are assigned to the Madrid Formation. Metamorphosed chaotic strata and fine-grained turbidites overlying the Madrid Formation are assigned to the Carrabassett Formation. This contact is clearly gradational at good exposures in the Anson and Kingsbury quadrangles (Fig. 2). However, on the northern slope of White Cap Mountain (Fig. 2), the Madrid Formation is abruptly overlain by a thick package of pelitic olistostrome marking the base of the Carrabassett Formation. While the present paper is mainly concerned with the Carrabassett Formation, the sedimentology of the Madrid Formation is also germane to the tectonic interpretation and is briefly discussed in the Appendix.

Rocks younger than the Carrabassett Formation, tentatively correlated with the Hildreths Formation (Osberg et al., 1968), and the Seboomook Formation of Boucot (1961; but see Boone, 1973), occur locally in the Jo-Mary Mountain quadrangle (Hanson and Sauchuk, 1986; Hanson, 1988).

The Carrabassett Formation is generally assigned to the Lower Devonian, based on poor fossil control and stratigraphic position with respect to other units of known or inferred age. Espenshade and Boudette (1967) reported poorly preserved plant fragments from three localities in the Greenville quadrangle, and poorly preserved brachiopods (*Howellella* or *Acrospirifer*) from one locality in the Sebec Lake quadrangle. These fossils permit a Lower Devonian age, but do not require it. Also suggesting a Lower Devonian age, the Carrabassett Formation gradationally overlies the unfossiliferous Madrid Formation, which in turn overlies the fossiliferous (Ludlovian, Upper Silurian) Smalls Falls Formation (Ludman, 1976). Finally, the Carrabassett Formation resembles fossiliferous Lower

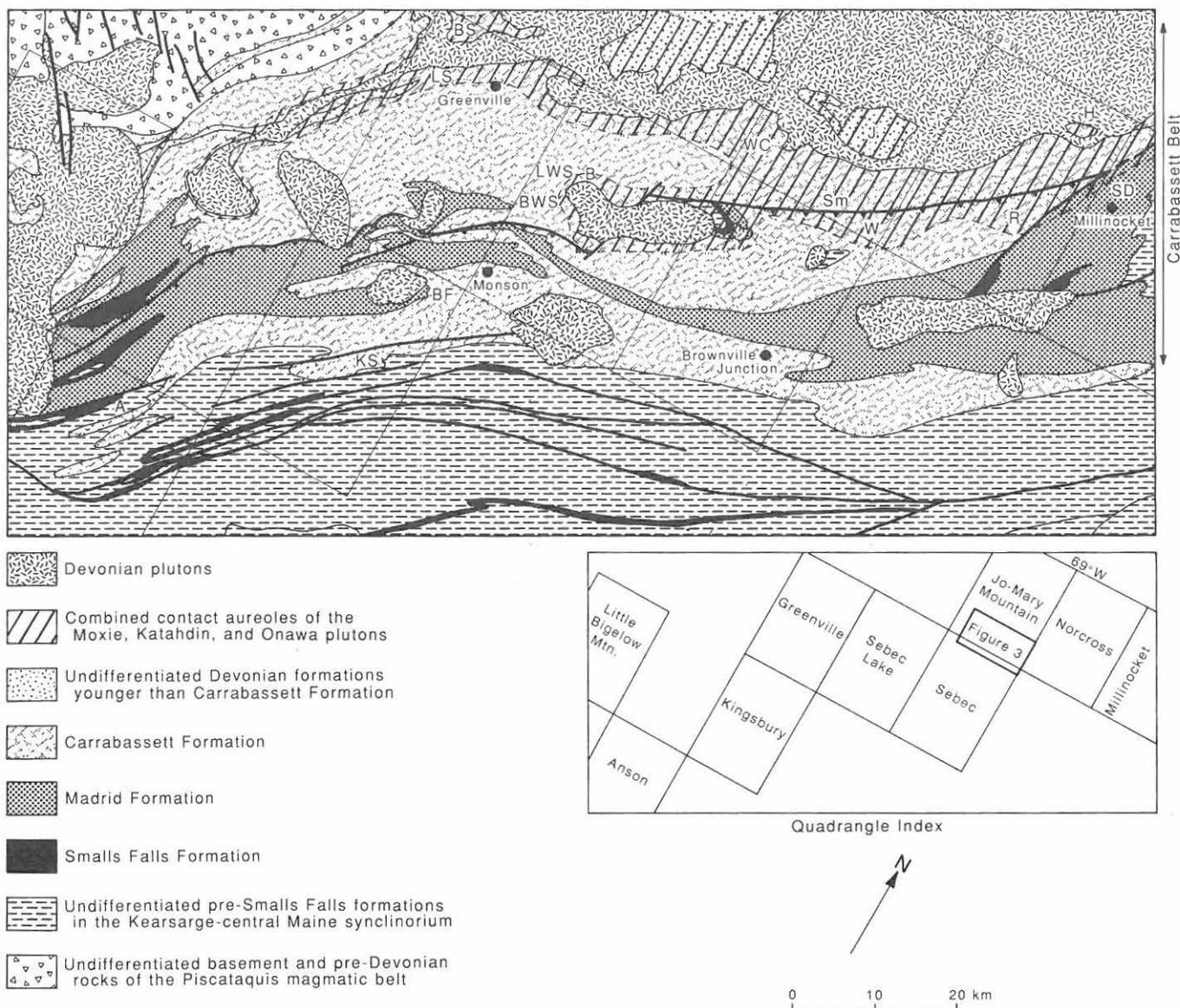


Figure 2. Geologic map of the reconnaissance study area (modified from Osberg et al., 1985). Letters indicate the following localities referred to in text: A = Arnolds Landing; BF = Barrows Falls on East Branch Piscataquis River; KS = Kingsbury Stream; BS = Big Squaw Mountain; LS = Little Squaw Mountain; B = Barren Mountain; LWS = Little Wilson Stream; BWS = Big Wilson Stream; WC = White Cap Mountain region, including Little Spruce Mountain; J = Jo-Mary Mountain; SM = Saddleback Mountain; W = Big and Little Wilkie Mountains; R = Ragged Mountain; H = Hammond Ridge; SD = Stone Dam on West Branch Penobscot River. The location of Figure 3 is shown on the quadrangle index.

Devonian strata of the Seboomook Group in the Moose River synclinorium.

The thickness of the Carrabassett Formation is impossible to measure accurately. In the area of Figure 3, we estimate its thickness at about 2000 m; Espenshade and Boudette estimated a thickness of about 1400 to 1750 m in the Greenville quadrangle.

SEDIMENTARY FACIES OF THE CARRABASSETT FORMATION

Classification Scheme

The term facies has a wide range of meanings and interpretations to sedimentologists (e.g. Moore, 1949; Reading, 1986;

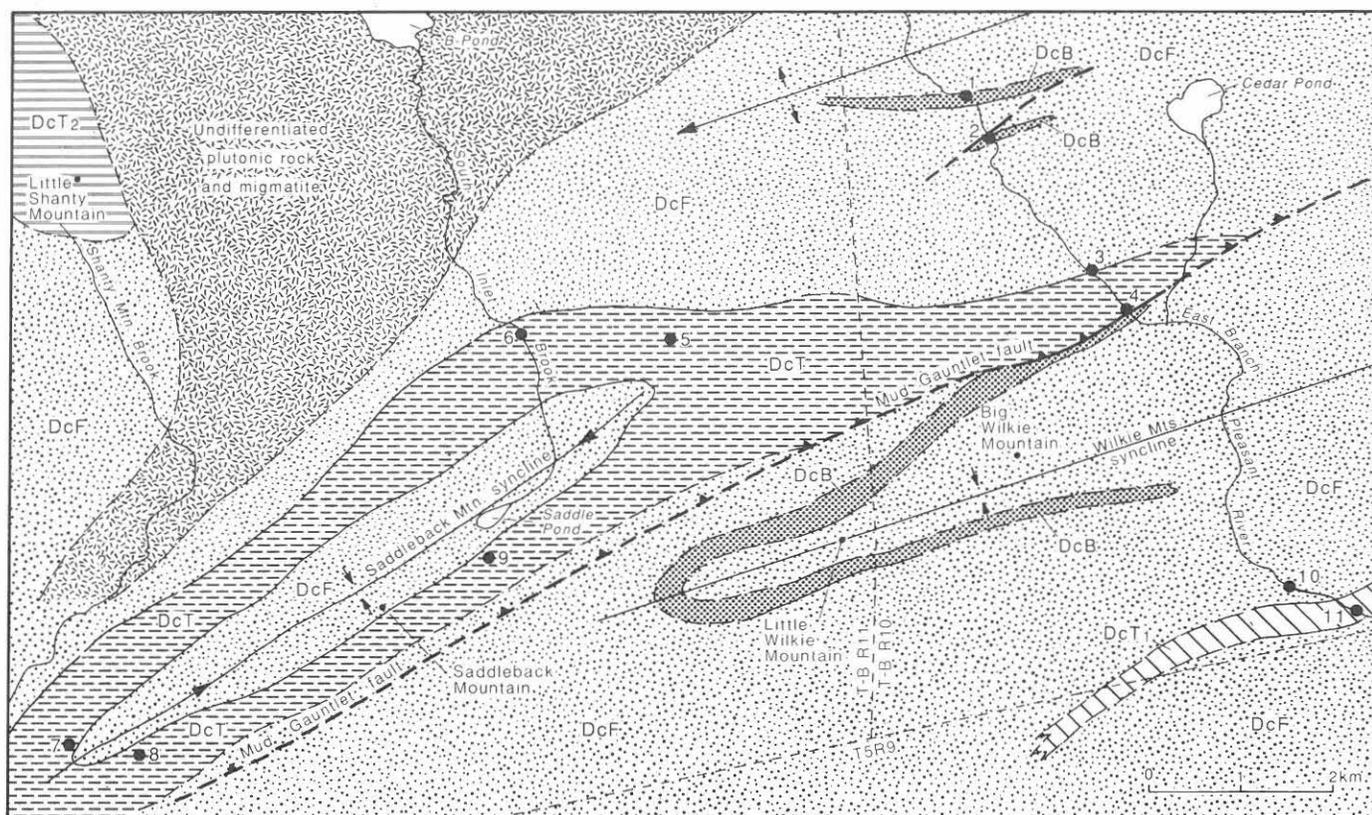


Figure 3. Geologic map of the East Branch Pleasant River and Saddleback Mountain, mapped by Hanson (1988). Numbers refer to locations discussed in text: 1 and 2 = exposures of the massive sandstone facies assemblage along the East Branch Pleasant River; 3 = Gauntlet Falls; 4 = Mud Gauntlet Falls; 5-9 = locally continuous exposures of the turbidite facies assemblage on Saddleback Mountain; 10 = spotted slates on the perimeter of the contact aureole; 11 = Horseshoe Falls. See Table 3 for explanation of geologic labels.

Anderton, 1985). A hierarchy of facies (Table 1) was developed by Hanson (1988) to facilitate mapping of the low and medium-grade, metamorphosed, turbidite-bearing formations in Maine. The primary focus of this paper is on the identification and interpretation of the lithofacies and descriptive facies associations of the Carrabassett Formation (Tables 2 and 3).

A *lithofacies*, at the base of the hierarchy, is a single bed or sequence of similar beds (Table 2) characterized by composition (e.g. sandstone-to-shale ratios), bed thickness and geometry, and sedimentary structures (e.g. presence of Bouma divisions). The lithofacies classification scheme used here is modified from Mutti and Ricci Lucchi (1972, 1975) and Walker and Mutti (1973), for turbidites and related deep-water sedimentary rocks. This lithofacies classification scheme should be equally applicable to the many other fine-grained, turbidite-dominated metasedimentary formations of the northern Appalachians. [The lithofacies classes are not to be confused with Bouma's (1962) divisions of individual turbidite beds based on sedimentary structures, which are denoted T_a through T_e.]

A *facies association* is a sedimentary package of juxtaposed lithofacies. Facies associations can be largely descriptive (e.g.

TABLE 1. HIERARCHY OF FACIES CLASSES (modified from Hanson, 1988). A more detailed description of lithofacies and facies associations found in the Carrabassett and adjacent formations is presented in Tables 2 and 3. Asterisks mark facies classes discussed in this paper.

Magnafacies: A regionally extensive, time transgressive package formed within a single tectonostratigraphic basin or system. Possibly correlative with group.

Facies assemblage: A lithologically distinct package of facies associations. Most assemblages are related to a single stratigraphic level and are therefore correlative with formations.

***Descriptive facies association:** Lithologically distinct, mappable packages composed of one or a number of associated lithofacies. Generally smaller than the scale of a formation and not restricted to a single stratigraphic position. May be correlative with member.

Facies categories and associated *lithofacies: A lithofacies is a descriptive term defined by the internal characteristics of an individual stratum. Categories are subjective groupings of facies used to facilitate discussion (e.g. turbidite and related facies).

***Subfacies:** Subdivision of facies based on small scale differences.

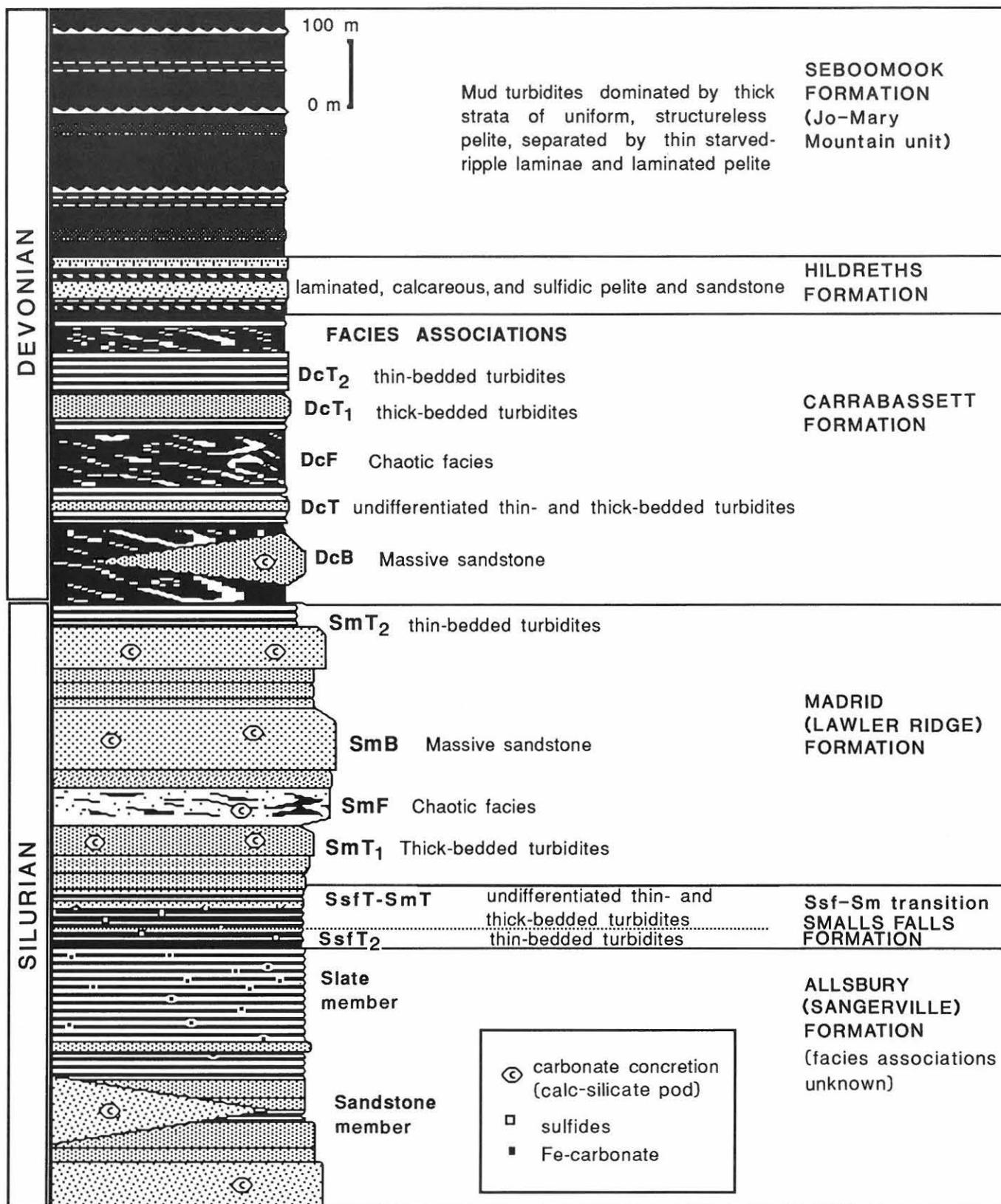


Figure 4. Generalized stratigraphic column of the Jo-Mary Mountain and Norcross 15' quadrangles (modified from Hanson, 1988).

Lower Devonian Carrabassett Formation

TABLE 2. SANDSTONE AND SILTSTONE-RICH TURBIDITE AND RELATED LITHOFACIES

(adapted and modified from Mutti and Ricci Lucchi, 1972; Walker and Mutti, 1973; and Mutti and Ricci Lucchi, 1975; summarized by Nilsen, 1984). Additions and modifications include subfacies E₁ and E₂, and addition of the facies H, M, and L.

Facies A Conglomeratic facies

Facies B Arenaceous facies

Description Laterally pinching, thick-bedded, medium-fine to coarse-grained, massive or laminated sandstone.

Facies C Arenaceous-pelitic facies

Description Interbedded sandstone and pelite: Ta-c, Tab, Ta-de common. Sandstone-to-pelite ratio (>>9.5:1).

Facies D Pelitic-arenaceous facies I

Description Interbedded pelite and sandstone: Ta or Ta-Tb Bouma divisions missing.

Subfacies (Mutti and Ricci Lucchi, 1975)

D₁ sandstone(siltstone): pelite >1
D₂ sandstone(siltstone): pelite <1

Facies H (defined here) Uniformly graded, varve-like strata. Beds range from a few mm to over half a meter in thickness and occur in sets of 2 to 12 beds.

Facies E Pelitic-arenaceous facies II

Description Thin-bedded sandstone and pelite. Sandstones contain flat, sharp bases and wavy tops.

Subfacies (defined here)

E₁ Medium to coarse-grained sandstones; sandstone to pelite ratio is 1:1 or greater. Top and bottom contacts are sharp and ungraded.
E₂ Medium to fine-grained sandstone; sandstone-to-pelite ratio is variable; ripple tops may be graded and contain pelite laminae that thicken toward troughs.

PELITIC FACIES (defined here)

Facies M Uniform structureless (massive or graded) pelite.

Description Predominantly massive, structureless pelite. Small silty pseudonodules and discontinuous, wispy, silt laminae may be present.

Facies L Laminated pelite.

Description Pelite with parallel, convolute or rippled laminae of silt or fine sand.

DISRUPTED FACIES

Facies F Chaotic facies

Description Mildly to highly disrupted or mixed strata.

Subfacies (defined here)

F₁ Chaotic facies with extraformational clasts (e.g. quartz pebbles, greenstone, carbonate, etc.)
F₂ Intraformational chaotic facies (sandstone and pelite only)

Subdivisions of Subfacies F₂

F₂₋₁ Mildly extended chaotic facies
F₂₋₂ Moderately disrupted chaotic facies
F₂₋₃ Highly disrupted - foliated (by soft-sediment deformation) chaotic facies
F₂₋₄ Highly disrupted - nonfoliated chaotic facies

TABLE 3. DESCRIPTIVE FACIES ASSOCIATIONS OF THE CARRABASSETT FORMATION.

Sandstone (siltstone)-rich turbidite associations

B Massive sandstone association (facies B and/or C dominant)
T₁ Thick-bedded turbidite association (facies C, D, and F are volumetrically most abundant)
T₂ Thin-bedded turbidite association (facies D, E, L, and F are volumetrically most abundant)
T Undifferentiated turbidite association (C, D, E, L, and F present)

Other associations

F Chaotic assemblage (chaotic facies dominant)
P Pelitic assemblage (L and M dominant)

The above letter codes can be used in conjunction with formation codes (e.g. DcB, DcT, and DcF) in order to designate mappable units.

thick-bedded turbidite association) as emphasized here (Table 3), or interpretive (e.g. basin plain, outer-fan lobe, inner-fan channel, or slope association). Descriptive facies associations can be readily employed in reconnaissance mapping; however, the detail necessary to accurately map interpretive facies associations is commonly not obtainable in the Siluro-Devonian metasedimentary rocks of Maine. In the Jo-Mary Mountain quadrangle, certain descriptive facies associations in the Carrabassett Formation can be discussed in terms of more interpretive associations because such details as lateral continuity and vertical sequences can be locally mapped.

Lithofacies of the Carrabassett Formation

The lithofacies of Carrabassett Formation are summarized in Table 2 and are discussed below in detail. Lithofacies A, B, C, D, E, F, L, M, and H are recognized, and several are further subdivided. Facies A (arenaceous-conglomeratic facies) is abundant elsewhere in the Kearsarge-central Maine synclinorium (e.g. in the type area of the Rangeley Formation), but is extremely rare in the Carrabassett Formation, and is not described below.

Facies B (Arenaceous Facies). This facies is characterized by massive sandstones that cannot be readily described using the Bouma divisions. Walker and Mutti (1973) divided this facies into B₁ (massive sandstones with dish structures) and B₂ (without dish structures). In the Carrabassett Formation, facies B sandstones are structureless or contain parallel or convolute laminae; dish structures are relatively rare. Grading is common and interbedded pelite is generally absent or thin and discontinuous. Individual sandstone beds are commonly 30 to 100 cm thick, and average approximately 50 cm. Amalgamated sandstones may contain up to 5 beds and rarely exceed a few meters in total thickness. Sandstones are generally fine to medium-grained; coarse-grained sandstones are rare in the Carrabassett Formation. Beds typically thin laterally, and basal scour and tool marks are rare to locally abundant (Fig. 5).



Figure 5. Facies B exposed on Big Wilson Stream, Sebec quadrangle. The large bed in the foreground is about 1 meter thick and thins along strike. Large flutes are visible on the base of the overlying amalgamated sandstone (on left).

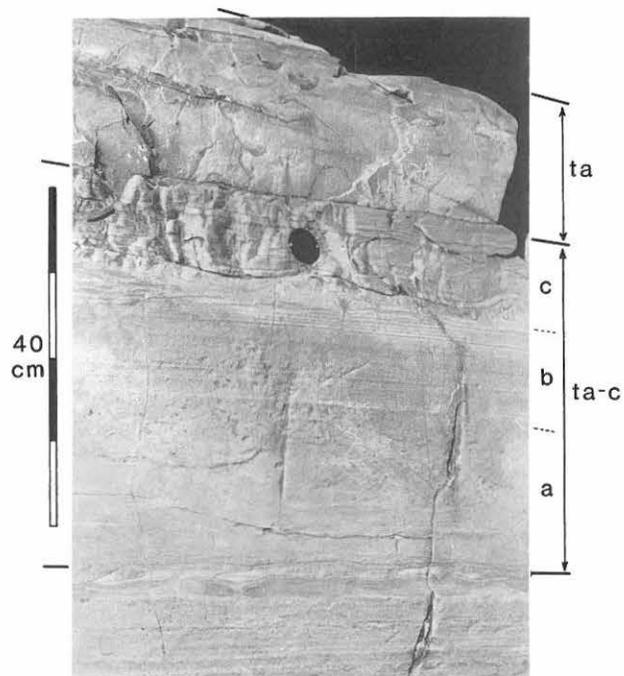


Figure 6. Facies C, containing two amalgamated sandstone beds. Bouma sequences T_{a-c} (lower bed) and T_a (upper bed), illustrated here, are sequences commonly seen in facies C.

Facies C (Arenaceous-Pelitic Facies). Facies C strata can readily be described in terms of the Bouma (1962) sequence. The complete sequence (e.g. T_a - T_e) and top-cut-out sequences (e.g. T_{a-b} , T_{b-c} , T_{a-c} , Fig. 6) are included in this facies. Bottoms of beds are generally sharp and flat, although flutes and tool marks may be present. Facies C beds are generally well graded from the base upward. Beds are laterally more continuous than



Figure 7. Facies D. (a) Facies D_1 shown here is composed of a series of T_{b-e} Bouma sequences. Note the high sandstone-to-pelite ratio. (b) Facies D_2 contains lower sandstone-to-pelite ratios (<1) and Bouma sequences T_{c-e} and T_{d-e} are more abundant.

facies B strata. Fine- to medium-grained sandstone is most common in the Carrabassett Formation.

Thick-bedded sandstones in poor exposures are best categorized as facies BC.

Facies D (Pelitic-Arenaceous Facies I). Facies D also can be described in terms of Bouma divisions, but lacks the massive sandstone (T_a) division. Mutti and Ricci Lucchi (1972) recognized three subdivisions of facies D on the basis of sandstone-to-pelite ratios, bedding thickness and occurrence of specific Bouma divisions (Fig. 7). Facies D_1 consists of Bouma T_{b-e} , T_{c-e} , and T_{de} divisions, and has a sandstone-to-pelite ratio greater than 1. Facies D_2 is comparable, but has a sandstone-to-pelite ratio less than 1. Facies D_3 has been reassigned in the present classification to pelitic facies L and M (see below).

Facies E (Pelitic-Arenaceous Facies II). Facies E is similar to facies D_2 , but differs in bedding geometry. Sandstone



Figure 8. Facies E₂ and B exposed on Big Wilson Stream. Facies E₂ (left) contains rippled sandstone beds with wavy tops and sharp, flat bases. Beds pinch out laterally.

beds are thin-bedded, tend to pinch out laterally, and are characterized by wavy ripples with sharp, flat bases; climbing ripples are locally present, but often difficult to recognize (Fig. 8). Facies E has been interpreted as an important constituent of channel margin associations (Mutti and Ricci Lucchi, 1972; Clifton, 1981). Although this facies was originally defined as poorly graded, coarse-grained, and sandstone rich, these distinguishing criteria have not been strictly adhered to in its identification (cf. Howell and Joyce, 1981; Graham and Ingersoll, 1981; Ingersoll, 1981; Ingersoll and Graham, 1981; Nilsen et al., 1981; Walker, 1985; and Nelson et al., 1977). Fine-grained and pelite-rich facies E strata also occur and reflect either a relative lack in the availability of coarser clastics, or overbank deposition from deeply incised channels that are not easily overtopped by coarse sediment. Facies E has therefore been divided here into two subfacies:

Facies E₁ corresponds to facies E as originally described by Mutti and Ricci Lucchi (1972). Beds are most commonly medium to coarse sandstone, but may include granule conglomerates. Grading is generally poor, tops of sandstone beds are sharp, and sandstone-to-pelite ratios are commonly high.

Facies E₂ is finer-grained. Beds are 2-30 cm thick, and vary from coarse siltstone through medium-grained sandstone (<1 to 10 cm thick), interbedded with thin (<1 cm) to relatively thick (10-30 cm) beds of pelite. Ripple-foresets commonly contain abundant pelite laminae which thicken toward the trough. Facies E₂ may have a greater lateral continuity than facies E₁, and a broader range of interpretations. Sandstone content is variable, ranging from approximately 10 to 80 percent. Facies E₁ is extremely rare in the field area; however facies E₂ is locally common and may be laterally associated with more mud-rich and sand-deficient turbidite flows.

Facies E, as defined by Mutti and Ricci Lucchi and discussed here, differs from the mud turbidite facies (E₁, E₂, E₃) of Piper (1978) and Stow and Piper (1984); the latter are pelitic facies G, discussed later.

Facies F (Chaotic Facies). The chaotic facies as used by Mutti and Ricci Lucchi (1972) refers primarily to resedimented strata such as debris flow, slump, and slide deposits. However, chaotic fabrics may also result from soft-sediment deformation produced by mud diapirism or penecontemporaneous faulting. Chaotic facies having a block-in-matrix texture that can clearly be assigned to sedimentary processes are referred to here as olistostrome. Facies F in the Carrabassett Formation is most commonly characterized by intraformational clasts of siltstone and/or sandstone, set in a pelitic matrix. Extraformational clasts have been observed, but are extremely rare.

The degree of disruption, and the style of deformation associated with disruption (extensional versus contractional, or both), are highly variable and prove to be a useful basis for subdivision of facies F. The preliminary classification listed in Table 2 is based largely on the degree of stratal disruption. The term "fragment foliation" (Cowan, 1985) is used in reference to a premetamorphic foliation. In olistostromes, such foliation can be developed from layer-parallel extension (Cowan, 1982, 1985) induced by mass-sediment-gravity flow. At one extreme, competent beds of one or more facies described here (e.g. facies L and D, Table 2) are cut by minor extension fractures (F₂₋₁). In more extreme cases, bedding is still recognizable and individual horizons traceable, but strata are chaotically folded and may contain local boudinage (F₂₋₂). In still more extreme cases, bedding is entirely transposed, and thin sandstone beds form parallel bands of phacoidal clasts that are entirely surrounded by pelitic matrix (F₂₋₃). These clasts form a well defined fragment foliation. The term type 1 mélangé (Cowan, 1985) has been employed to describe the block-in-matrix character of subfacies F₂₋₃ and subfacies F₂₋₄ (Fig. 9).

Evidence of contractional deformation is common in facies F of the Carrabassett Formation. The fragment foliation of subfacies F₂₋₃, and the less deformed strata of subfacies F₂₋₂ are typically folded or cut by contractional faults. The folded fragment foliation is characteristically oriented at an angle to the undisturbed overlying strata. Complex fold interference patterns are commonly present and may be explained by differential motion down-slope or penecontemporaneous refolding by subsequent slumps or flows (Tobisch, 1984). Disharmonic, mesoscale slump folding in the Carrabassett Formation tends to produce distinctive hummocky outcrops (Fig. 10).

In rocks classified as facies F, deformation is largely inferred to result from sediment gravity flow, with extensional deformation occurring in the breakaway zone of the slump, and contractional deformation occurring at the toe. However, in many exposures, other mechanisms such as diapirism, syndepositional faulting, or tectonic deformation cannot be ruled out. Indeed, recent studies of accretionary prisms have suggested that processes of soft-sediment and tectonic deformation merge imperceptibly.

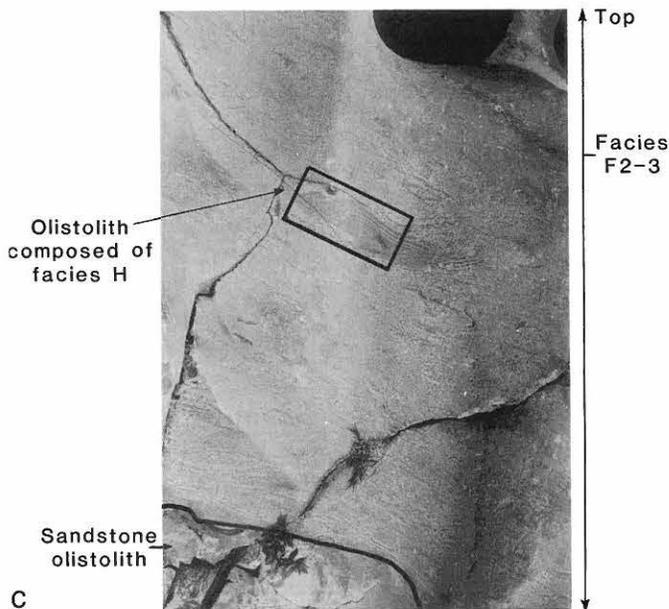
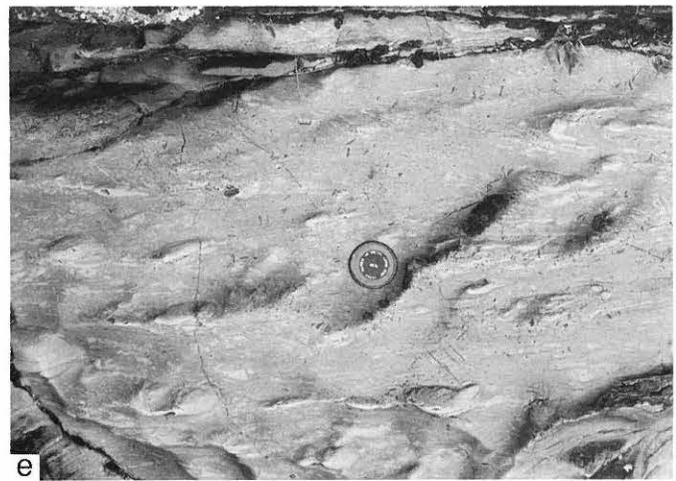
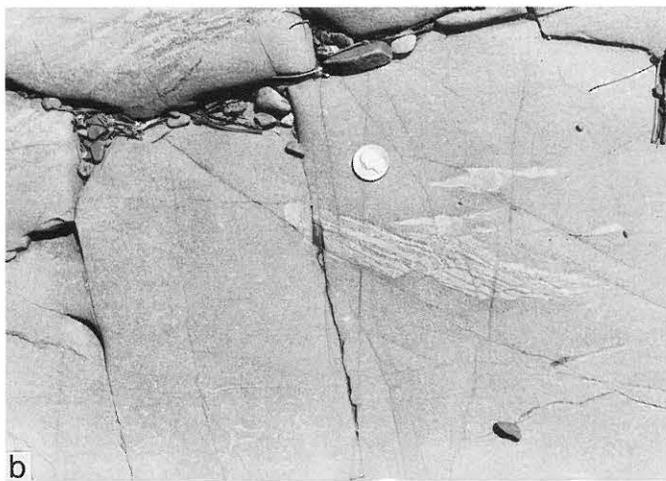
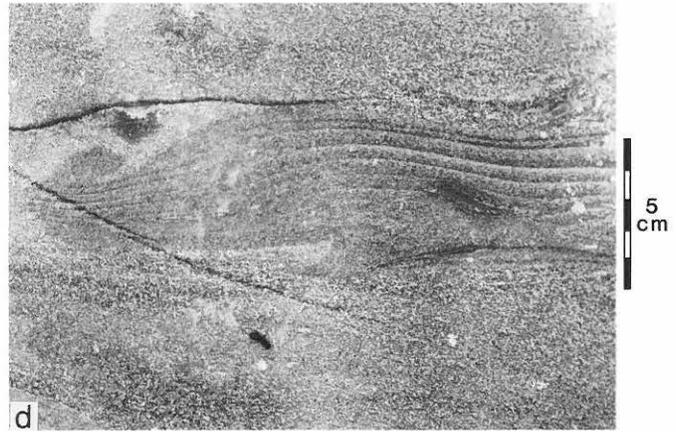
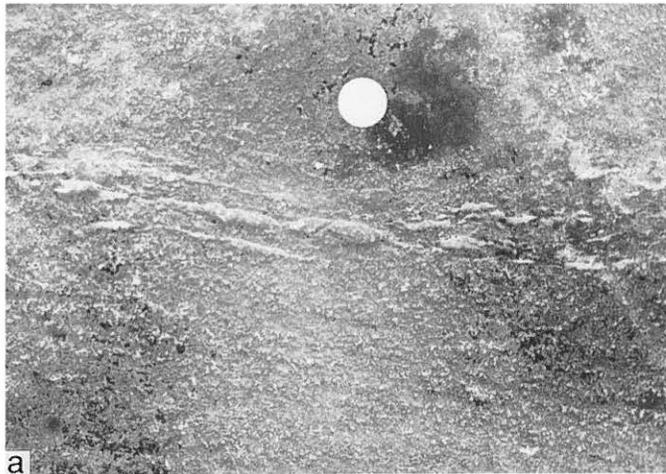


Figure 9. Chaotic facies of the Carrabassett Formation. Figures a-c are in hornfels exposed in the Jo-Mary Mountain and Norcross quadrangles. (a) Foliated chaotic facies. Sandstone beds are thoroughly broken, and phacoidal clasts form a distinct fragment foliation. (b) Chaotic facies lacking fragment foliation contains irregular blocks, which are isolated within a homogeneous, pelitic matrix. (c) Large clast (olistolith) of facies H in chaotic facies exposed at Gauntlet Falls (Fig. 3, loc. 3). (d) Enlargement of boxed area in central region of (c). (e) Olistostrome showing folded fragment foliation. This outcrop, located outside the hornfels zone (Barrows Falls, Kingsbury quadrangle), contains soft-sediment deformation fabrics overprinted by tectonic foliation (Fig. 2, BF).

tibly near the sediment-water interface, and that efforts to discriminate between the two in the ancient record may often prove futile. Where chaotic facies are stratabound within a succession of undisturbed turbidites, an interpretation based on submarine sliding is warranted. Similarly, a soft-sediment origin is likely for folds that are disharmonic, and whose axial surfaces are strongly transected by tectonic foliations.

Facies L and M (Massive and Laminated Pelitic Facies). Several elaborate classification schemes, with as few as two to

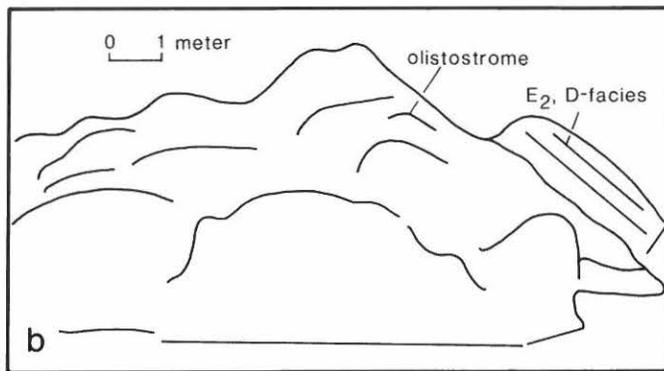
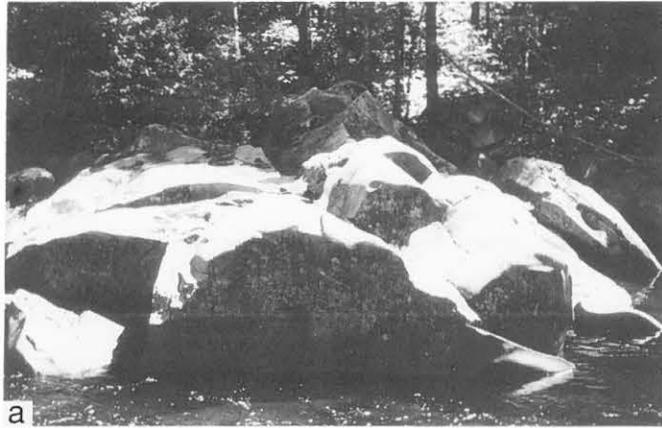


Figure 10. (a) Outcrop expression of facies F in hornfels along the East Branch Pleasant River. The folded fragment foliation influences the mechanical weathering properties of the hornfels, which forms rounded, hummocky ridges. (b) Interpretive sketch of outcrop in (a).

as many as eight subdivisions, have been proposed for fine-grained turbidites and hemipelagic sediments (see review by Stow and Piper, 1984). All authors agree that there is a characteristic succession of sedimentary structures produced by mud-rich turbidity currents and subsequent hemipelagic sedimentation (see also Banerjee, 1977; Lowe, 1982; and Carey and Roy, 1985). In the metasedimentary terrane of north-central Maine, two pelitic facies, (facies L and M) are generally recognizable (where present), despite significant variation in preservation of original sedimentary features.

Facies L (new name) is a laminated pelite characterized by a number of small scale sedimentary features, corresponding to divisions T₀ through T₅ of Stow and Shanmugan (1980) and facies E₁ of Piper (1978). These features include a variety of silt laminae, beginning with a relatively thick (up to 10 mm) silt layer with fading ripples (Fig. 11) and sequentially followed by lenticular, convolute, thin-irregular, thin-regular, indistinct, and wispy laminae. The entire sequence and most of the silt laminae display positive grading.

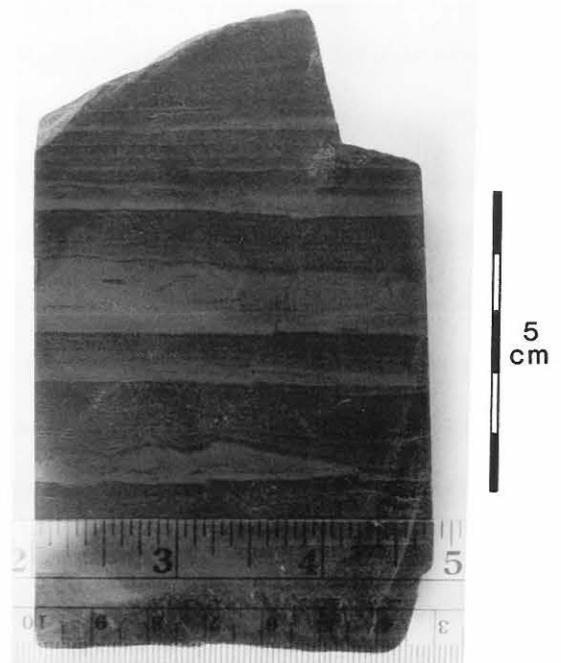


Figure 11. Facies L, a relatively sandstone-rich, laminated pelite. Basal siltstones and fined-sandstones contain ripples and pelite, and grade upward into pelite with siltstone laminae. At least four sequences, each inferred to have been deposited by a single turbidity flow, are shown here. The structures in each sequence correspond to the T₀-T₃ divisions of Stow and Shanmugan (1980) and are interpreted as being deposited from high-energy, fine-grained turbidity flows.

Facies M (new name): The laminated-pelite facies may grade into the massive-pelite facies, containing a sequence of graded and ungraded pelite (T₆ and T₇ of Stow and Shanmugan, 1980; E₂ and E₃ of Piper, 1978). Silt lenses and silt pseudonodules occur locally within the massive pelite. This facies is thought to be comparable to the unifites of Blanpied and Stanley (1981), although other interpretations are possible (e.g. hemipelagic deposits or homogenites of Cita et al., 1982).

Deposition of pelitic facies has been related to thick, low velocity, mud-rich turbidity currents and hemipelagic settling (Stow and Shanmugan, 1980; Piper 1972, 1978; Stow and Bowen, 1976, 1980). The laminated-pelite facies, deposited from relatively high-energy turbidity flows, has been interpreted as channel margin overbank deposits within a mud-rich and sand-starved system (Slatt and Thompson, 1985). Massive pelite, deposited either from the low energy tail of a turbidity flow or hemipelagic settling, is suggestive of basin-plain or interchannel sedimentation.

Facies H (Uniformly Graded Facies). Facies H (new name) is composed of uniformly graded, varve-like strata lacking visible parallel or ripple laminae (Fig. 12). Beds range from a few millimeters to over half a meter in thickness and commonly have sandstone-to-shale ratios greater than 6:1. This facies

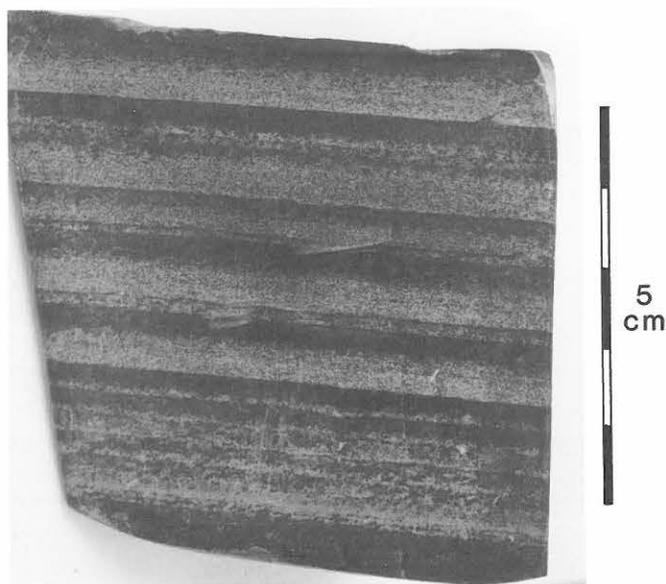


Figure 12. Facies H, consisting of fine-grained sandstone which grades rapidly to pelite. Laminae and other structures are conspicuously absent, and beds occur in clusters. The section shown here is 7 cm thick.

commonly occurs in sets of 2 to 12 beds that separate thicker sandstone beds (facies D₁, C or B).

The lack of laminae, high sandstone and siltstone content, and excellent grading suggest that these beds were deposited rapidly from suspension. Ponding of thin sand- or silt-rich turbidity flows may be responsible for the deposition of facies H; the sudden decrease in current motion would induce rapid settling and the development of a varve-like bed. Alternatively, these beds could be deposited from a single surging flow (e.g. Lowe, 1982). Thinning-upward sequences would be expected in such instances, reflecting the progressive decrease in surge intensity.

Facies Associations of the Carrabassett Formation

Exposures along the East Branch Pleasant River and adjacent highlands demonstrate that distinct, mappable sedimentary packages exist within the Carrabassett Formation. These packages are referred to here as descriptive facies associations. Boone (1973) recognized three lithologic packages within the Carrabassett Formation that presently have member status; the (1) massive metapelite member, (2) thinly layered schist member, and (3) quartzwacke member. These members can be related to one or more of the descriptive facies associations listed in Table 3. The facies associations of the Carrabassett Formation are discussed below in estimated order of abundance.

Chaotic Facies Association (DcF). Although facies F has been encountered in nearly all facies associations of the Carrabassett Formation, particularly large areas are dominated by the facies (e.g. Little Squaw Mountain, Barren Mountain, Big

Wilson Cliffs near Big Wilson Stream, White Cap Mountain, Little Spruce Mountain, Ragged Mountain, Big and Little Wilkie Mountains, and Hammond Ridge; Fig. 2). However, as much as 20 to 40 percent of the association may consist of interbeds composed predominantly of facies L, M, D₂ and E₂. Belts of chaotic facies, with interbedded turbidites, are traceable for tens of kilometers along strike across the area of Figure 2 and are not confined to any one part of the Carrabassett belt. Many areas presently mapped as the massive metapelite member (see Osberg et al., 1985) are underlain by the chaotic facies association. Hence, these deposits cannot be related to a single mass-wasting event, but instead to numerous smaller events which must have occurred repeatedly along the entire length of the basin.

Pelitic Facies Association (DcP). In the area of detailed mapping, a mappable pelitic facies association has not been encountered in the Carrabassett Formation. On close inspection, most exposures of apparently massive pelite were found to be chaotic strata. Although the chaotic strata are commonly composed of laminated pelites (e.g. Hammond Ridge) the designation of a distinct pelitic facies association has not been warranted because they have been resedimented. However, the south-westerly-trending belt of structureless pelite that extends through Brownville and Monson has not been studied in detail. This belt appears to contain a thick pelitic facies association that we have not encountered elsewhere in the Carrabassett Formation. Although rare in the Carrabassett Formation, thick pelitic facies associations undisturbed by gravity flow dominate the Seboomook Formation in the Jo-Mary Mountain quadrangle. These pelitic facies associations are discussed in detail by Hanson (1988).

The massive metapelite member of the Carrabassett Formation (Boone, 1973) may encompass two facies associations: (1) a pelite-rich chaotic facies association (cf. Vehrs, 1975); and (2) a pelitic facies association(?) comparable to that encountered in the Brownville-Monson slate belt.

Thin-Bedded Turbidite Facies Association (DcT₂). With a few exceptions (i.e. facies B and F) strata within thick-bedded (next heading) and thin-bedded turbidite associations can be described, bed-by-bed, in terms of the Bouma divisions.

Facies D, L, and E₂ are the principle components of the thin-bedded turbidite association. Beds containing T_a and T_b divisions are rare, and amalgamation of these beds even less common. Sandstone-to-shale ratios range between 4:6 and 8:2. Beds are composed of base-cut-out Bouma sequences that are generally less than 10 cm and average 5 cm in thickness. Interbeds of chaotic facies a few centimeters to several meters thick are not uncommon. The thin-bedded turbidite association may grade vertically and laterally into thick-bedded turbidite associations. However, the outcrop control is commonly too poor to define the nature of this transition. The thinly layered schist member of the Carrabassett Formation in the type area (Boone, 1973) is a thin-bedded turbidite facies association.

Interpretations of the thin-bedded turbidite association could include deposition within a channel margin or levee,

interchannel, or basin plain environment. In the Carrabassett Formation, the local abundance of facies E₂, occurrence of facies F, and association with thinning-upward sandstone packages having locally discordant contacts (e.g. Fig. 8), are together suggestive of deposition in channel-margin/levee and interchannel environments. Channel-margin/levee deposits grade laterally and vertically into interchannel deposits.

Thick-Bedded Turbidite Facies Association (DcT₁). Thick sandstone beds of facies B, C, and D₁ characterize this turbidite facies association. However, facies F and E₂ are common associated facies, and are abundant in thin-bedded sequences that are interbedded with the thicker sandstone packages. Sandstone thicknesses are usually 5-30 cm, but may range from 1 to over 100 cm. Some units mapped as the quartzwacke member of the Carrabassett Formation are thick-bedded turbidite associations.

Thick-bedded turbidite associations of the Carrabassett Formation are less than 200 m thick and characterized by thinning-and fining-upward sequences. The thick-bedded sandstones and associated thin-bedded facies can be related to channel axis and channel margin deposition respectively. In contrast, the thick-bedded turbidite associations encountered in the Madrid Formation appear to thicken upward and are more suggestive of outer-fan lobe deposits (see Appendix).

Massive Sandstone Facies Association (DcB). This facies association is composed of massive, amalgamated sandstone beds (facies B and C), with rare pelitic interbeds. Where exposed along the East Branch Pleasant River (Fig. 3, locations 1, 2, and 4), individual beds generally cannot be distinguished, and vertical sequences have not been identified. The massive metasandstone facies association is, in part, analogous to the quartzwacke member of Boone (1973).

The stratigraphic position of this facies association within the Carrabassett Formation is poorly understood in the study area. In many instances the base of the unit is either faulted or not exposed. The upper contact is always abruptly overlain by olistostrome (facies F). Along strike the unit thins and disappears, where it is replaced by olistostrome. The massive sandstone association may be interpreted as either (1) a channel deposit within the Carrabassett Formation, or (2) slivers of sandstone from the Madrid Formation transported along faults. Whereas poor exposure greatly limits interpretation, the laterally discontinuous geometry, abundance of massive sandstone, and association with chaotic facies is inconsistent with a slope channel origin.

PARTIAL MEASURED SECTIONS OF THE CARRABASSETT FORMATION

Outcrops on the East Branch Pleasant River and adjacent highlands form the most continuous and best exposed section of the Carrabassett Formation in the area of Figure 3. The turbidite section between Gauntlet and Mud Gauntlet Falls was originally identified as the Seboomook Formation (Hanson, 1983; Osberg

et al., 1985) because of its positioning on Carrabassett-like rocks and resemblance to rocks identified as Seboomook Formation by Boucot (1961; Northeast Carry Formation of Pollock, 1987) at Seboomook Lake, and by Boone (1973) in the Little Bigelow Mountain area. On Saddleback Mountain, these turbidites stratigraphically underlie, in apparent conformity, a thick package of chaotic facies very similar to that below the section at Gauntlet Falls. Similar observations elsewhere (e.g. Little Wilkie Mt., Horseshoe Falls - Fig. 3; Little Wilson Stream, and Kingsbury Stream - Fig. 2) suggest that many Seboomook-like rocks are simply turbidite associations within the Carrabassett Formation.

Gauntlet Falls Section, East Branch Pleasant River

The rocks exposed between Gauntlet Falls and Mud Gauntlet Falls (locations 3 and 4 in Fig. 3) are part of a turbidite association which is locally thick-bedded. The turbidites form a homoclinal sequence that dips gently 10°-30° to the south. A generalized column is presented in Figure 13, and details are shown in Figures 14 through 17.

The base of the sequence is draped over the hummocky surface of a large olistostrome, part of an underlying chaotic association with an estimated thickness of over 1000 meters. The original bedding in the olistostrome (facies F) has been thoroughly stretched and disrupted, forming an indistinct fragment foliation composed of irregular stringers and phacoidal clasts of sandstone and siltstone. The foliation is disharmonically folded and is visible only on water-worn or weathered surfaces where feldspathic silt and sand grains have been weathered to a chalky white. Where surfaces are fresh, the olistostrome appears massive, except for a few irregular blocks of bedded material surrounded by massive pelite. Directly overlying the olistostrome is a single discontinuous bed, 10-80 cm thick, of medium-fine sandstone (facies B). The sandstone, and in many areas the olistostrome itself, is overlain by a meter or more of medium-bedded pelite separated by thin, wavy trains of non-climbing ripples (facies E₂; Fig. 14). The ripples have linear to slightly sinuous, asymmetrical crests which are approximately 2 mm high and have a mean spacing of 7 cm. Abundant paleocurrent indicators range from 325° to 360°, with a mean of 345°.

The section rapidly coarsens upward, interrupted episodically by thin olistostromes. Sandstones are fine to medium grained. Thick, locally amalgamated sandstones commonly lack the upper T_{b-e} or T_{c-e} beds of the Bouma sequence (Fig. 15) and are assigned to facies BC. Facies B and BC packages are commonly interbedded with a series of non-laminated graded beds (facies H). The thick sandstone section, which is composed of a series of thinning upward cycles, passes up-section into a thin-bedded turbidite association (Figs. 16, 17) that is truncated by the Mud Gauntlet fault.

The Gauntlet Falls section is interpreted here as channel and channel-margin deposits. The facies E₂ strata at the base of the association may represent overbank sediments deposited when

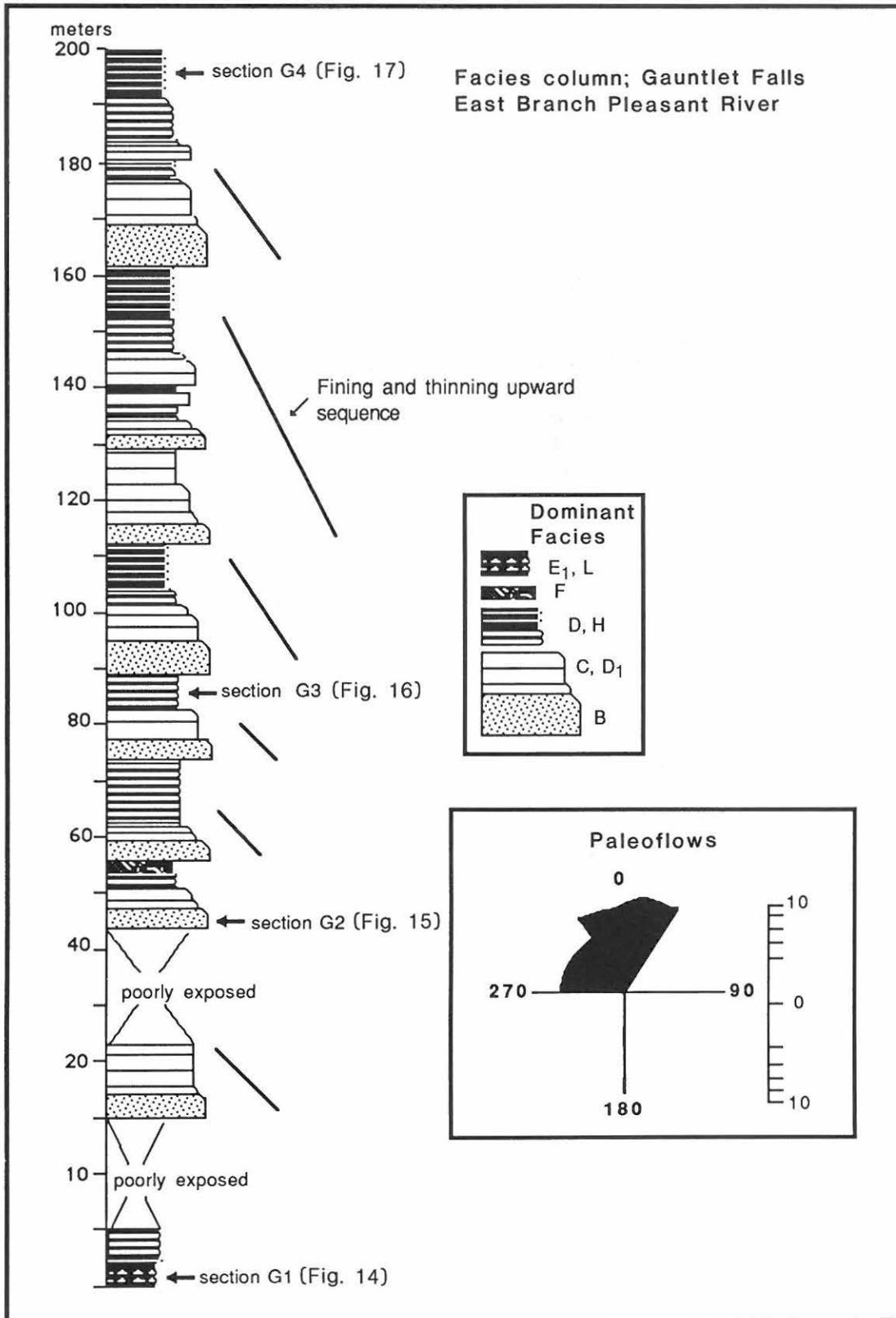


Figure 13. Generalized measured section from Gauntlet to Mud Gauntlet Falls, East Branch Pleasant River. Most of the section forms a thick-bedded turbidite facies assemblage, composed of thinning-upward megasequences. The uppermost part of the section is a thin-bedded turbidite facies assemblage which is truncated by the Mud Gauntlet fault. Note positions of detailed sections (Figs. 14-17, sections G1 - G4).

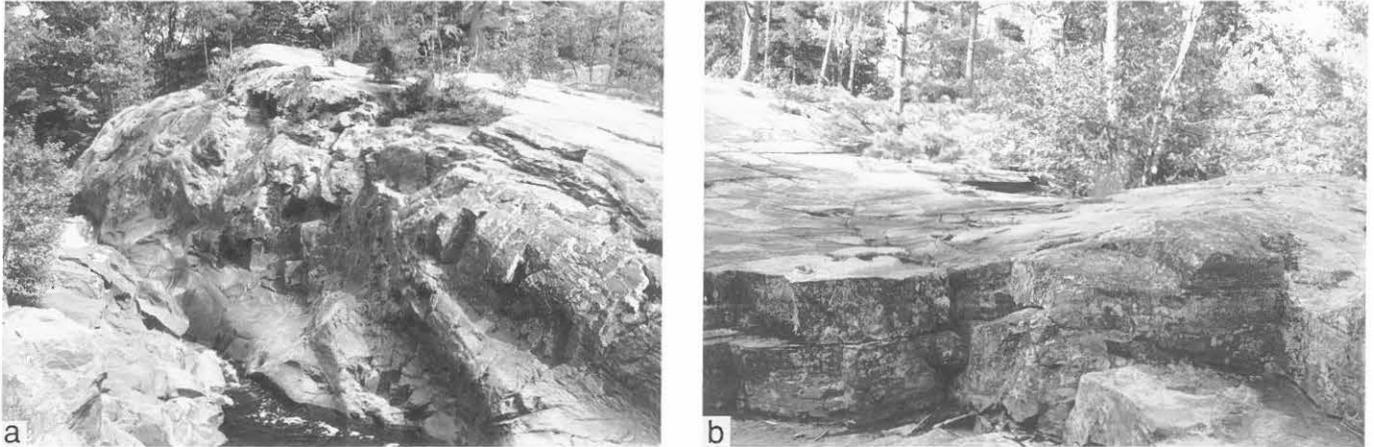


Figure 14. Base of Gauntlet Falls turbidite section. (a-b) A locally thin section of facies E₂ (Section G1) covers the irregular surface of olistostrome exposed at Gauntlet Falls (loc. 4, Fig. 3).

the channel axis was farther to the southeast. Location and subsequent abandonment of the channel in the Gauntlet Falls region is supported by the observed thinning- and fining-upward vertical sequences.

Saddleback Mountain

The turbidite facies association at Gauntlet Falls has been traced westward to the southern end of Saddleback Mountain. Measured sections (at locations 5-9 in Fig. 3; Hanson, 1988) from the northern and southern slopes of the mountain suggest that a lateral facies change from predominantly thick- to thin-bedded turbidites occurs along strike toward the west. In addition, interbedded olistostromes become thicker and more abundant. The thin- and thick-bedded turbidites on Saddleback Mountain and Gauntlet Falls are shown here as single turbidite association (DcT, Fig. 3).

PALEOCURRENT DATA

A number of workers have embraced the concept of a major reversal in sediment transport direction within the Kearsarge-central Maine synclinorium during the Late Silurian or Early Devonian. According to this hypothesis, the Ordovician and main Silurian part of the Kearsarge-central Maine synclinorium succession (e.g. Quimby, Greenvale Cove, Rangeley, Sangerville, and Perry Mountain Formations) were derived from the northwest. On the other hand, Devonian rocks in the Kearsarge-central Maine synclinorium are generally believed to have been derived from the opposite direction (i.e. from the southeast or east). It should be noted that the above interpretation is based on reconstructed lateral facies relationships, rather than quantitative paleocurrent data.

As part of a detailed paleoflow study still in progress, we have measured paleocurrent directions from several Carrabassett

and Madrid Formation exposures in the area of Figure 2. Predominantly west-directed paleoflows, ranging from 190° to 020°, establish a generally eastern source and generally westward sediment transport, albeit with significant local variation (Hanson, 1988).

These results suggest that the Madrid Formation belongs with the upper, east-derived succession (as was suggested by Roy et al., 1983), leaving the Smalls Falls Formation as the most probable transition unit. As shown in Figure 18, sedimentary structures are extraordinarily well preserved at several localities (e.g. East Branch Pleasant River and Big Wilson Stream).

INTERPRETATION

Depositional Setting of the Carrabassett Formation

In this and the following section, we present evidence and arguments bearing on our interpretation that the Carrabassett Formation was deposited during the earliest stages of an Acadian collision. We visualize deposition in an inner-trench slope environment, which may have featured a number of small, tectonically controlled sub-basins (Fig. 19). This preliminary model is supported by the following observations and reasoning.

The Carrabassett Formation in the Context of Other Siluro-Devonian Events. The Acadian orogeny brought about the end of a 40 Ma episode of deep-water sedimentation in the Kearsarge-central Maine synclinorium, during which an estimated 8 km of sediment accumulated in water depths below storm base. The orogeny coincided with the end of volcanism in the two magmatic belts which flank the Kearsarge-central Maine synclinorium and resulted in extreme horizontal shortening. These first-order observations are the basis for interpreting the Acadian orogeny as a result of collision following subduction, by one or another plate geometry. As the youngest widespread formation in the Kearsarge-central Maine

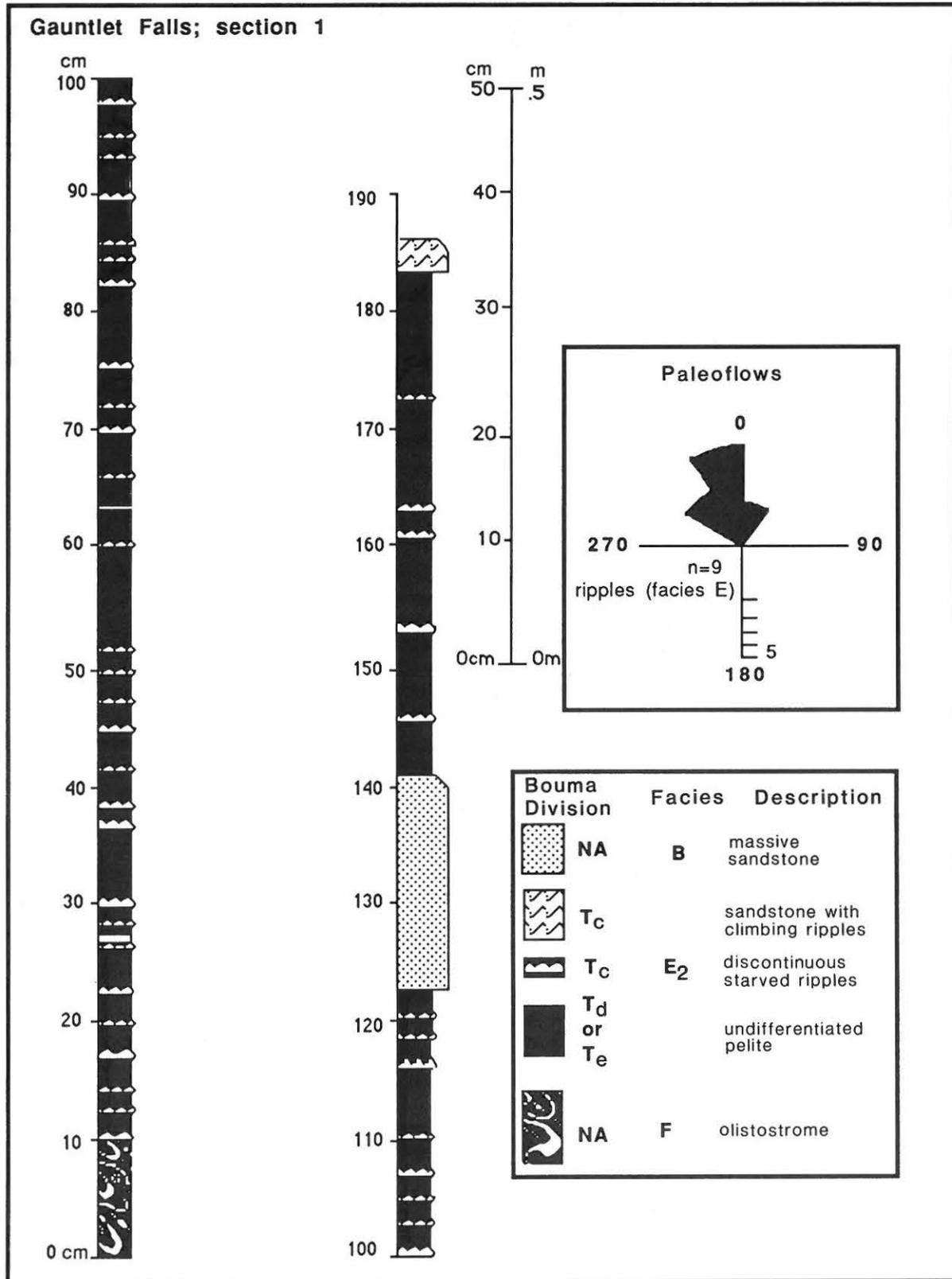


Figure 14 (Continued). (c) Partial column of facies E₂ strata overlying olistostrome in (a). These beds are interpreted as channel margin deposits.

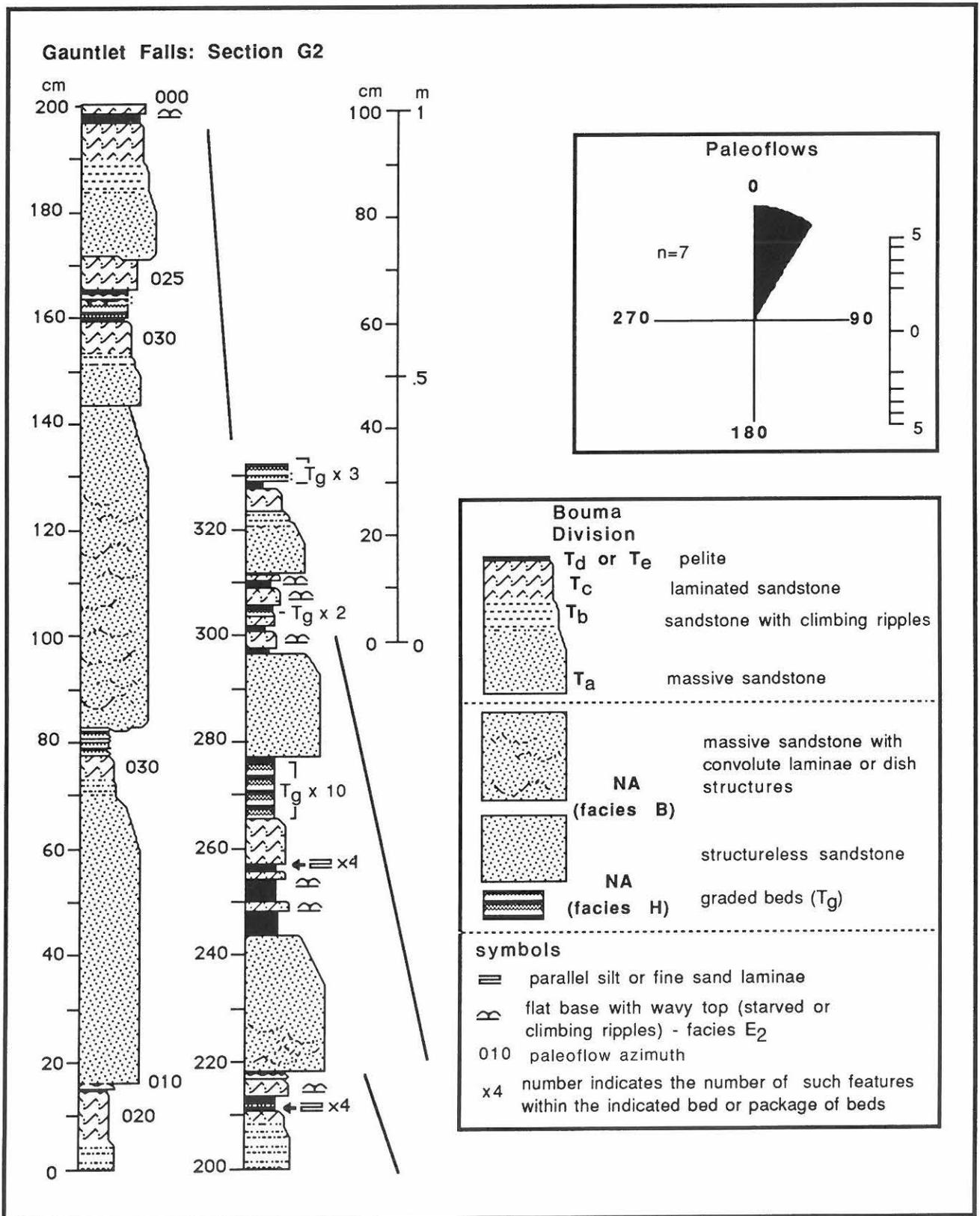


Figure 15. Section G2 between Gauntlet and Mud Gauntlet Falls. Two thinning upward megasequences are visible above the 80-cm marker. Facies B lies at the base of each cycle. Packages of facies C and D, separated by sets of facies H strata, occur sequentially up-section.

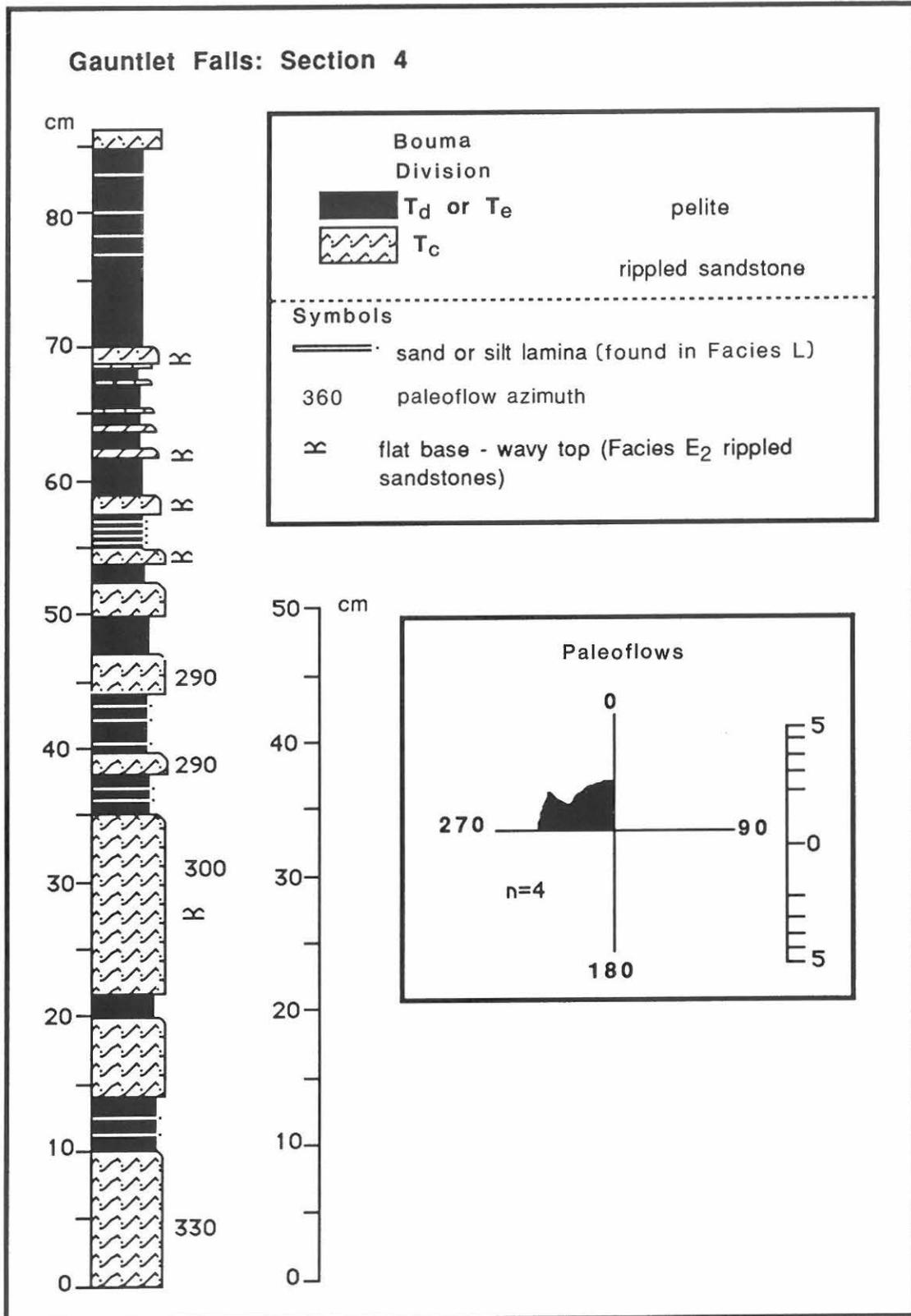


Figure 17. Section G4 between Gauntlet and Mud Gauntlet Falls. Facies D, E_2 , and L mark the upward transition into thin-bedded turbidites, resulting from channel migration and abandonment.

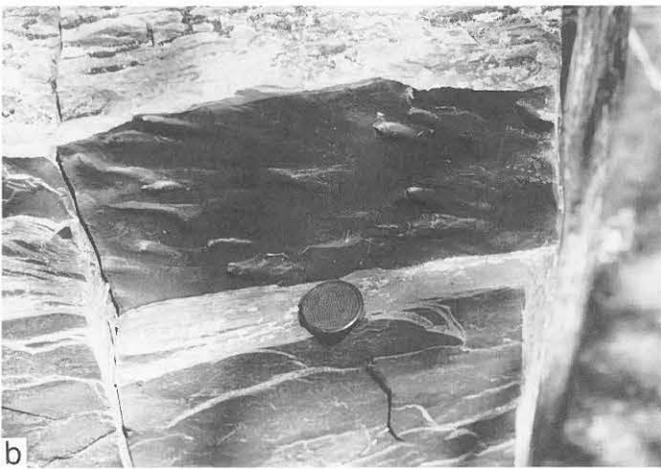
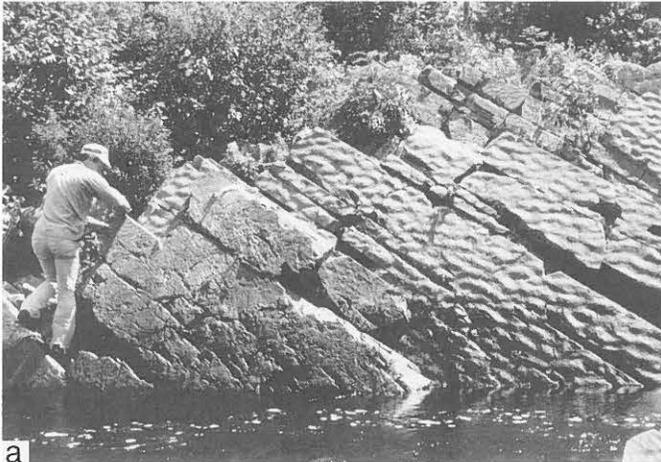


Figure 18. Well preserved directional indicators record a westerly sediment transport direction with significant local northerly and southerly variations. (a) Current-formed ripples at Horseshoe Falls, East Branch Pleasant River (Fig. 3, loc. 11). (b) Flute casts at Big Wilson Stream (Fig. 2, BWS).

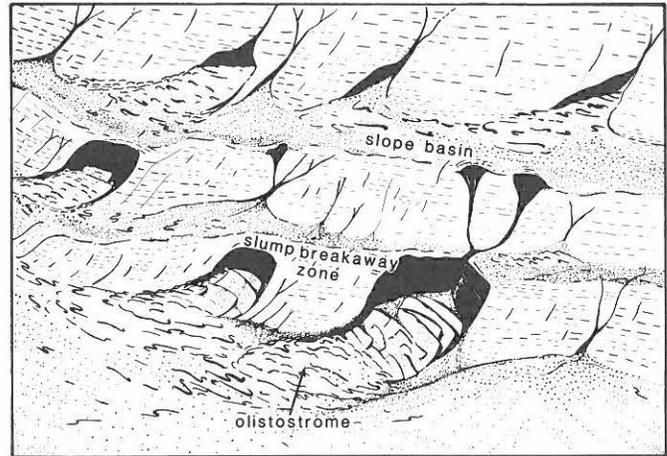
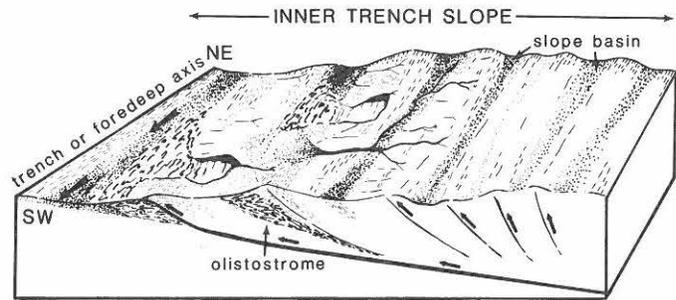


Figure 19. Depositional model for the Carrabassett and Madrid Formations. The Carrabassett Formation is interpreted as a package of inner trench slope and basin deposits. Pelitic and thin-bedded turbidite facies occupying slopes were subject to episodic failure. Thrusts bound local slope basins occupied by debris aprons and small elongate "fans". The entire depositional system was presumably transected by submarine canyons and channels. This depositional model can accommodate paleoflow directions with a dominant westerly component and strong northerly or southerly variations, as are observed in the Carrabassett Formation. The Madrid Formation is inferred to have been deposited along the axis of the associated trench (foredeep). Axial sediment transport was along strike toward the southwest.

synclorium, the Carrabassett Formation was deposited during the critical transition period when a regime of deep-water sedimentation gave way to a regime of tectonic deformation, metamorphism, crustal thickening, and uplift. Therefore, we believe that the Carrabassett Formation was deposited in a convergent tectonic setting.

Inferences Based on Sedimentary Facies. The combination of chaotic, thin-bedded turbidite, and pelitic facies associations is characteristic of lower slope environments at bathyal to abyssal depths. Within convergent plate boundary zones, two lower slope environments are plausible: (1) the inner trench slope of the accretionary prism and (2) the outer trench slope. In the case of the Carrabassett Formation, the latter setting is considered unlikely, owing to the scarcity of chaotic facies in such outer trench slope units as the Utica Shale of the Taconic

foredeep, New York (Bradley and Kusky, 1986). We therefore postulate that the Carrabassett Formation was deposited in an inner trench slope, that is, along an accretionary prism. Present-day inner trench slopes are characterized by local, thrust bounded slope basins and transverse submarine canyons, which are superimposed upon a broader slope (e.g. Underwood and Bachman, 1982, their Fig. 5). Facies associations of the Carrabassett Formation are comparable to those encountered in a number of modern inner trench slope and basin settings (e.g. Hellenic trench and Coriscan trough, Stanley and Maldonado, 1981; Stanley and Knight, 1979; and Stanley, 1980; inner slope of the Timor trough, Karig et al., 1987; Eastern Aleutian trench and slope, Piper et al., 1973).

The bathymetric profiles of accretionary prisms are continually steepened by contractional deformation. In the Car-

rabassett Formation, evidence for layer-parallel extension can be most readily interpreted in terms of slope failure, and repeated interbedding of individual chaotic and turbidite facies requires numerous mass-movement events. These could be related to slope instability triggered by sedimentation or by earthquakes, to oversteepening caused by thrusting, and/or to mass movement of material extruded along emergent thrust faults (cf. Breen et al., 1986)

Chaotic and thin-bedded turbidite facies assemblages in the Carrabassett Formation are interpreted as interchannel slope deposits, deposited on slopes that were episodically failing as described above. Thick-bedded, thinning-upward turbidite assemblages in the Carrabassett Formation are clearly channel deposits. Many of these channel deposits, such as the Gauntlet Falls section, traversed the slope, while others may have developed on elongate "fans" occupying the axes of narrow slope basins developed between thrusts (Fig. 19). The lack of coarse channel facies supports the concept of a broad, low-gradient, multibasinal slope or a newly emergent mud-rich source area. The highly disrupted chaotic facies and associated laminated pelite facies were deposited in lower slope, slope-basin, and base of slope settings.

Clearly identifiable basin-plain assemblages dominated by facies D (Mutti and Ricci Lucchi, 1972; Walker and Mutti, 1973) appear to be lacking in the Carrabassett Formation, as are sand-dominated, thickening-upward outer fan (Mutti and Ricci Lucchi, 1972) or suprafan lobe deposits (Normark, 1970, 1978; Walker and Mutti, 1973; Walker, 1978). Nevertheless, the pelitic facies association may be a good candidate for basin-plain-type sedimentation (cf. Blanpied and Stanley, 1981; Stanley, 1980, 1983; Cita et al., 1982; and Kastens and Cita, 1981). Deposited in a sand-starved trench fed only by mud-rich, low-velocity turbidity currents, these pelites would conformably overly sandstones of the Madrid Formation.

We believe the Madrid Formation (see Appendix) was deposited in a trench environment. Carbonate (calc-silicate) concretions, considered characteristic of the Madrid Formation and found primarily in thick sandstone beds, probably represent original carbonate sand deposited rapidly by turbidity currents. Rapid burial will easily result in preservation of carbonate material below the carbonate compensation depth (e.g. Middle American trench, Moore et al., 1982). Therefore, the occurrence of carbonate material, especially in a formation of thick sandstones like the Madrid Formation, does not necessarily reflect shallow water deposition. The slope-sediments of the Carrabassett Formation are thought to be in part diachronous with the Madrid Formation. However, the pelitic turbidites and olistostromes ultimately dominated the basin once sand supply to the trench was terminated.

Paleocurrents. Paleocurrent data are too sparse and scattered to paint a clear picture, but the absence of any Carrabassett or Madrid Formation paleocurrents with a significant easterly component (none between 020° and 190°) is probably meaningful. Within the inferred lower slope setting of the Carrabassett

Formation, the available data would best support a westerly paleoslope. In the Madrid Formation, southwesterly paleoflow at Arnolds Landing is consistent with the inferred depositional setting along a trench axis, that is, sediment was funneled down the axis of a northeast-southwest trending, tectonically controlled basin. The limited data from the Carrabassett and Madrid Formations are consistent with the broader regional pattern for the entire syn-Acadian clastic succession from Gaspé to Virginia, whereby sediment transport was generally westerly (Ettensohn, 1987; Bradley, 1987a; Hanson, 1988).

Implications for the Tectonics of the Kearsarge-Central Maine Synclinorium and Acadian Orogeny

Tectonic interpretations of the Acadian orogeny have been discussed in several recent reviews (e.g. Rodgers, 1970; Osberg, 1978; Robinson and Hall, 1979; Bradley, 1983), and a full treatment is beyond the scope of this paper. Bradley (1983) argued that a plate geometry analogous to the present-day Molucca Sea, involving two subduction zones in the Kearsarge-central Maine synclinorium, could best satisfy the first order observations mentioned in the previous section. However, a major weakness of Bradley's (1983) model was that it did not readily account for the simple stratigraphy that evidently characterizes the Kearsarge-central Maine synclinorium (e.g. Moench et al., 1982; Hatch et al., 1983). Figure 20 is a provisional attempt to incorporate the stratigraphy of the northwestern Kearsarge-central Maine synclinorium, as follows. According to this interpretation, the lower part of the succession (Quimby, Greenvale Cove, Rangeley, and Perry Mountain Formations) was deposited along the Taconic-modified margin of North America, in a fore-arc basin that was part of a post-Taconic northwest-dipping subduction system (Fig. 20a). The upper part of the succession (Smalls Falls, Madrid, and Carrabassett Formations) was deposited in outer trench slope, trench axis, and inner trench slope environments, respectively (Fig. 20b). At the onset of collision (Fig. 20b), this migrating trench or foredeep system evolved from Trench 2, and in a later position, it was the basin in which the vast expanse of Seboomook Group olistostromes and pelites of northwestern Maine were deposited (Bradley, 1987b). Our interpretation of the Smalls Falls, Madrid, and Carrabassett Formations implies that these units have diachronous contacts that shifted across strike in concert with plate motions, at rates of tens of kilometers per million years. The coarsening and thickening upward sequence which characterizes the transition between Smalls Falls Formation and the Madrid Formation is similar to that associated with ancient trench systems (Lash, 1985).

Whether or not this particular scenario is valid, an important corollary of all models involving Siluro-Devonian subduction in the Kearsarge-central Maine synclinorium is that some deformation which is traditionally regarded as "Acadian" and Devonian is in fact older. We visualize a continuum of deformation from Silurian to Devonian, beginning with subduction-related shor-

Figure 20 (at right). Tectonic model for the Acadian orogeny, modified from figures by Bradley (1983) to account for the stratigraphy of the Kearsarge-central Maine synclinorium. KCMS = Kearsarge-central Maine synclinorium; GMA = Green Mountain anticlinorium; CVGS = Connecticut Valley - Gaspé synclinorium; PMB = Piscataquis magmatic belt; CMB = Coastal magmatic belt. (a) Hypothetical paleogeography during mid-Silurian, shortly before onset of collision between opposed accretionary prisms. The column on the left shows the lower formations of the KCMS, which are interpreted to have been deposited in a fore-arc basin. Oq = Quimby Fm.; Sg = Greenvale Cove Fm.; Sr = Rangeley Fm.; Sp = Perry Mountain Fm. (b) Hypothetical paleogeography during Late Silurian to Early Devonian, at a very early stage of collision. The mid-Silurian fore-arc basin is inferred to have evolved into a foredeep, which is the direct descendant of Trench 2 in the mid-Silurian model. Three time-transgressive depositional environments (outer trench slope, trench axis, and inner trench slope) are visualized, corresponding to the Smalls Falls (Ssf), Madrid (DSm), and Carrabassett (Dc) Formations, respectively.

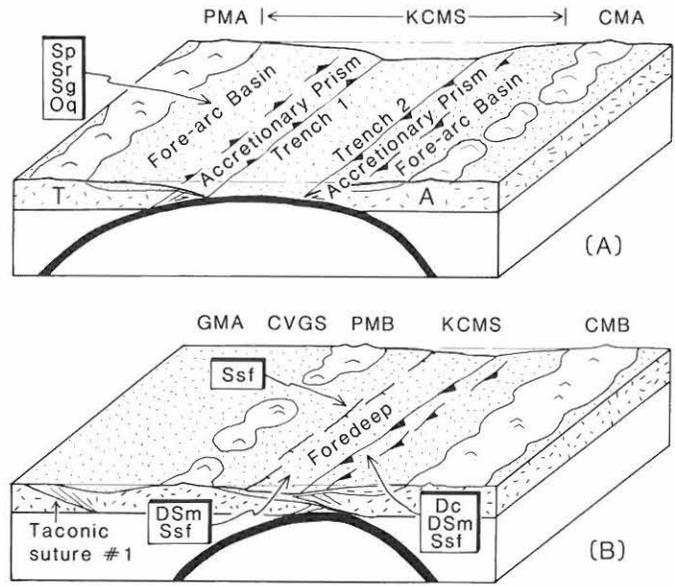
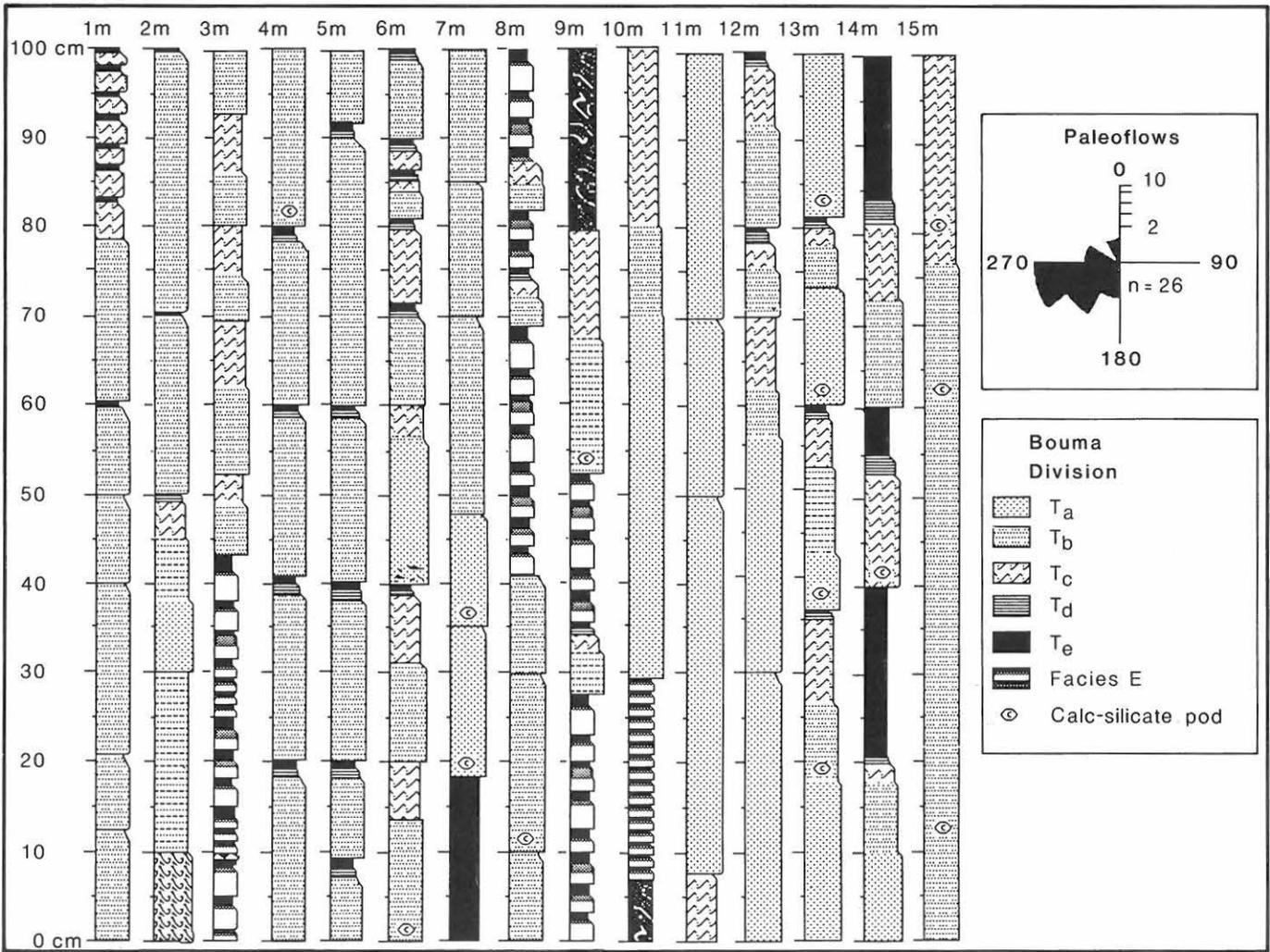


Figure 21 (below). Detailed measured section of upper Madrid Formation at Arnold's Landing (Fig. 2, loc. A).



tening and ending with collision-related shortening. Along strike in the Scottish Southern Uplands, deformation is now largely recognized as the consequence of pre-Caledonian subduction rather than Caledonian collision (Leggett, 1980; Leggett et al., 1982; McKerrow and Cocks, 1977).

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APPENDIX: MADRID FORMATION SEDIMENTOLOGY

Because of the gradational contact between the two units, the Carrabassett Formation cannot be properly understood without also considering the Madrid Formation. Quadrangle mappers recognize the Madrid Formation as feldspathic, locally calcareous metasandstone, while the Carrabassett Formation is characterized by metapelite. The results of the present study show that Madrid-like rocks are locally abundant within the Carrabassett Formation. Both formations contain well-graded T_a - T_{dc} sequences (facies C); however, amalgamated sandstones (facies B and BC) are much more common in the Madrid Formation. Thick-bedded turbidite facies associations of the Madrid Formation were examined at three localities: at the type section at Madrid Village, at Arnold's Landing on the Kennebec River (Fig. 2, A; Fig. 21), and Stone Dam on the West Branch Penobscot River (Fig. 2, SD).

Unlike the Carrabassett Formation, vertical sections of the Madrid Formation are characterized by coarsening and thickening upward sandstones. These characteristics are suggestive of outer fan lobe deposits. Bradley and Hanson (1989) described the Madrid Formation in greater detail.

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