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Sedimentary Geology 148 (2002) 425–447

Sedimentary Geology

www.elsevier.com/locate/sedgeo

Paleocurrent analysis of a deformed Devonian foreland basin in the Northern Appalachians, Maine, USA

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Received 11 January 2001; accepted 27 June 2001

Abstract

New paleocurrent data indicate that the widespread Late Silurian and Devonian flysch and molasse succession in Maine was deposited in an ancestral, migrating foreland basin adjacent to an advancing Acadian orogenic belt. The foreland-basin sequence spread across a varied Silurian paleogeography of deep basins and small islands—the vestiges of an intraoceanic arc complex that not long before had collided with the Laurentian passive margin during the Ordovician Taconic Orogeny. We report paleocurrents from 43 sites representing 12 stratigraphic units, the most robust and consistent results coming from three units: Madrid Formation (southwesterly paleoflow), Carrabassett Formation (northerly paleoflow), and Seboomook Group (westerly paleoflow). Deformation and regional metamorphism are sufficiently intense to test the limits of paleocurrent analysis requiring particular care in retrodeformation.

Keywords: Maine; Appalachian; Acadian; Paleocurrent analysis; Tectonics; Foreland basin

1. Introduction

Despite steady progress in unraveling details about the Acadian orogeny in New England, fundamental controversies persist about what it all means in terms of plate tectonics (Bradley, 1983; Hanson and Bradley, 1989; Robinson et al., 1998; Bradley et al., 2000). Most workers regard this Late Silurian to Devonian event as the result of collision between the Avalon terrane and the Laurentian (North American) margin (Fig. 1), which had already experienced an arc collision during the Ordovician. The problem is one of plate geometry: how many subduction zones existed, where were they located, and what was their vergence?

Strata that were deposited in the critical time immediately before and during Acadian deformation might be expected to hold clues to the plate configuration that led to orogenesis. Study of these rocks, however, has been hindered by structural complexity, limited exposure, obliteration of sedimentological details by deformation and metamorphism, poor paleontological control, and stratigraphy that many geologists have described—in conversation if not in print—as monotonous. In this paper, we present new paleocurrent data from three paleogeographic belts in central and northern Maine (Fig. 2). Our study lends quantitative support to the idea that the youngest part of the Acadian-deformed section was

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0037-0738/02/$ - see front matter
PII: S0037-0738(01)00161-0

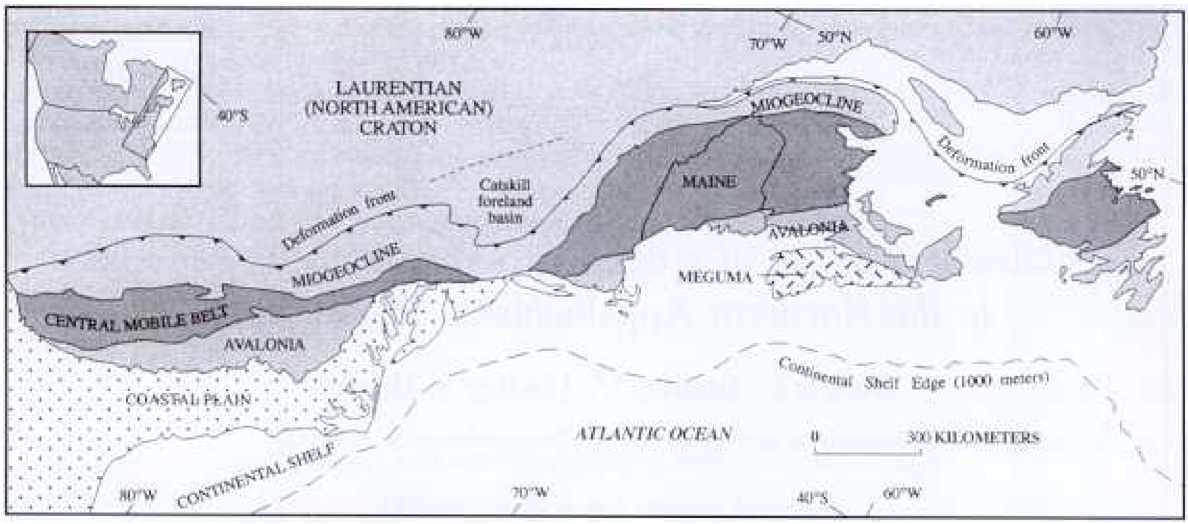


Fig. 1. Map of the Appalachian orogen showing selected tectonic features.

deposited in a migrating foreland basin along the Laurentian margin. Sediment transport during this interval was broadly toward the craton, across regional strike.

2. Regional setting

The study area is located midway along the length of the Appalachian orogen and 150 – 250 km behind the deformation front (Fig. 1). Pre-Silurian deformation in this region is attributed to two orogenies: Penobscottian (Cambrian) and Taconian (Ordovician); these events are too old to be of interest here. Our analysis focused on strata that were deposited after Taconian deformation and were penetratively deformed during the subsequent Acadian orogeny. Carboniferous to Permian orogenesis—commonly grouped under the catch-all "Alleghanian"—was important in some parts of the Appalachians but not in the study area.

Three main Silurian paleogeographic elements are recognized in the area of interest (Boucot, 1968; McKerrow and Ziegler, 1971; Roy, 1980) (Fig. 2). We will return to the paleogeography in greater detail later and only briefly mention the key elements here. The Central Maine basin, to the southeast, was an area of continuous deep-water sedimentation from at least Late Ordovician to Late Silurian or earliest Devonian (Ludman, 1976). We have long held that this belt represents the sedimentary off-scrapings of a closed ocean (Bradley, 1983), although this idea is controversial (Robinson et al., 1998; Bradley et al., 2000). To the northwest lay a relatively positive area ("anticlinorial belt" in Fig. 2) that approximately encompasses the Bronson Hill, Boundary Mountains, Lobster Mountains, Munsungun, and Pennington anticlinoria. This belt was the site of small islands flanked by siliciclastic aprons, shallow marine platforms that subsided at relatively slow rates, and local volcanic centers (Boucot, 1968). Still farther northwest lay another deep water area—the Connecticut Valley–Gaspe basin—where relatively deep-water sedimentation prevailed from Late Ordovician into Early Devonian time (Roy, 1989). Regional stratigraphic relations suggest that during Devonian times, the Silurian paleogeographic elements were buried beneath a thick siliciclastic succession derived from outboard sources (e.g., Bradley, 1983). The Devonian event is the main focus of the present paper.

3. Paleocurrent studies

The first paleocurrent determinations from Acadian-deformed rocks in Maine were published by Hall et al. (1976). They reported cratonward (west-directed) sediment transport from Acadian-derived turbi-

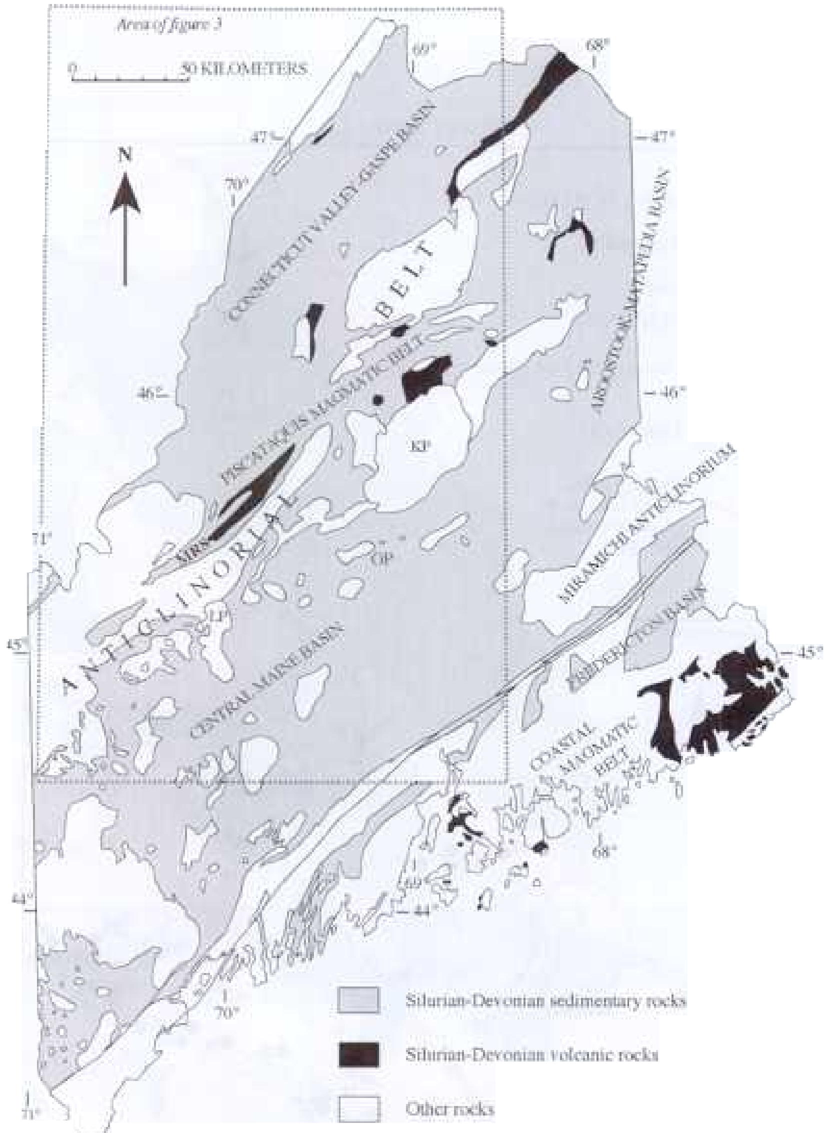
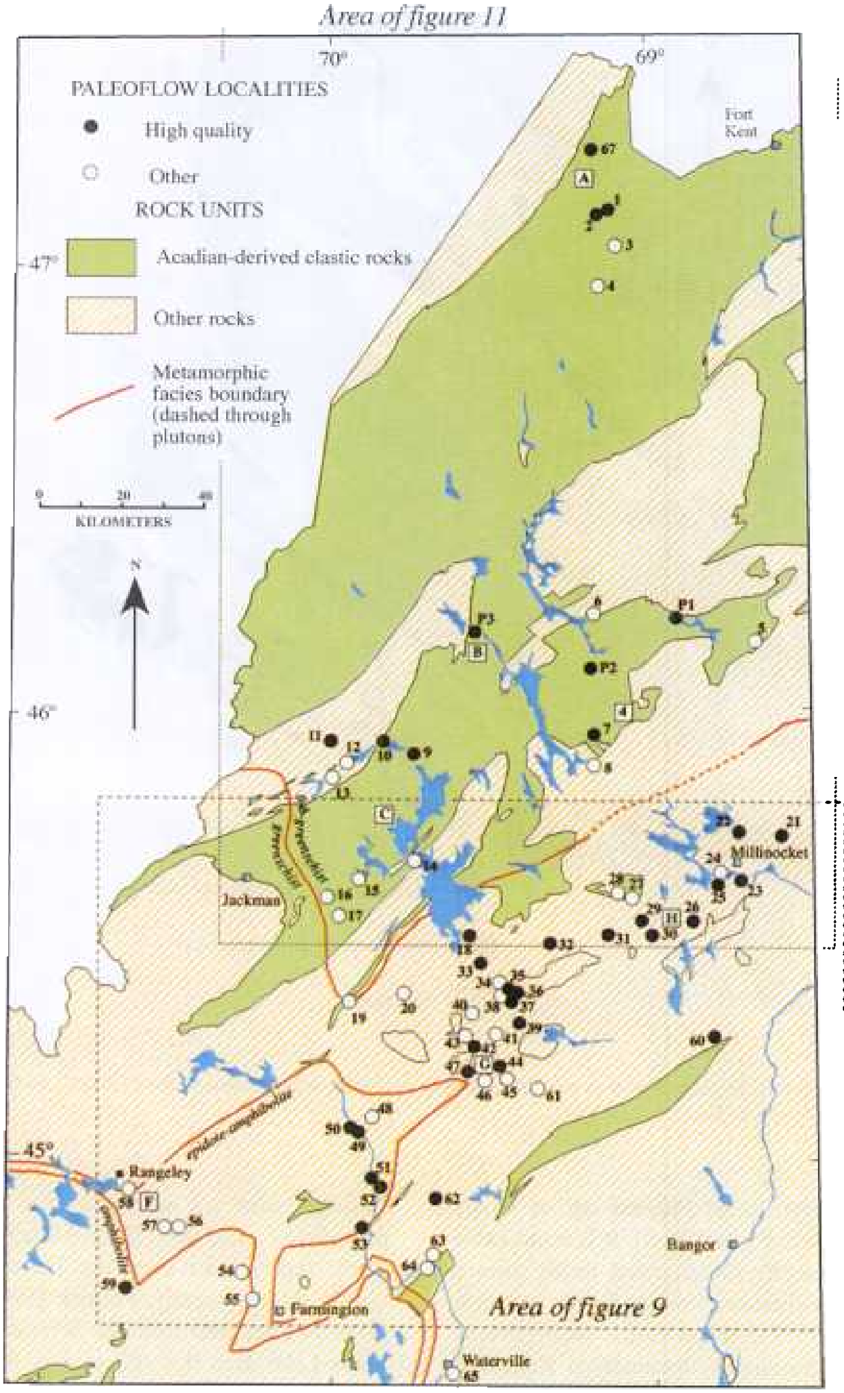


Fig. 2. Map of Maine showing Silurian–Devonian paleogeographic belts. Metamorphic isograds (from Guidotti, in Osberg et al., 1985) are Acadian (Devonian) in the area of Fig. 3.

dites and deltaic deposits of Pragian age at Grand Lake Matagamon (loc. P1 in Fig. 3). It is remarkable how often this sketchily documented result from a single area has been cited as evidence for the transport direction of the entire Devonian clastic succession in New England. A thoroughly documented follow-up study by Pollock et al. (1988) confirmed the initial findings, and we can now say that the westerly paleoflows at Grand Lake Matagamon are indeed fairly representative of the Devonian Seboomook Group. More recently, Pollock (1994) reported paleocurrent directions from the Seboomook Group in the Telos Lake area (loc. P2 in Fig. 3), and included a synthesis of Devonian paleocurrent directions in


Area of figure 11
70°
69°
PALEOFLOW LOCALITIES
High quality
Other
ROCK UNITS
Acadian-derived clastic rocks
Other rocks
Metamorphic facies boundary (dashed through plutons)
KILOMETERS
N
47°
46°
45°
Fort Kent
Jackman
Millinocket
Rangeley
Bangor
Farmington
Waterville
Area of figure 9

Maine, based largely on his own previously unpublished measurements. A few paleocurrent directions have also been reported by Roy et al. (1983).

We began measuring paleocurrent data from Silurian and especially Devonian rocks in Maine in the early 1980s, largely in the Central Maine basin (Fig. 2). Although most of the paleocurrent data reported here are new, some were reported earlier by Hanson (1983, 1988), Bradley (1987), Hanson and Bradley (1989, 1993, 1994), Bradley and Hanson (1989), and Kusky et al. (1994). The directions quoted here may differ slightly from our previously published directions from the same locations, either owing to an improved structural correction or additional paleocurrent data.

Table 1 summarizes the more reliable paleocurrent determinations, which were found by the methods described in the next section. Sites are identified by locality number in Fig. 3; where two or more outcrops were studied within a 1-km radius, they were given a locality number and letter (e.g., loc. 12b). Fig. 2 also shows a number of locations where we obtained some paleocurrent data that we deemed unreliable owing to complex structure, scattered paleocurrent directions, or a shortage of measurable paleocurrent indicators. A tabulation of paleocurrent data from the less reliable outcrops is available through the Data Repository of *Sedimentary Geology*.

Although outcrop in interior Maine is generally poor, large exposures amenable to paleocurrent studies are found along many rivers and in mountainous areas. Paleocurrent indicators are fairly well preserved throughout most of the area of Fig. 3, but metamorphic grade increases to the southwest, where most sedimentary structures have been badly strained (Fig. 4d), to the point that they may be wholly unreliable (Fig. 4g). We selected sites for detailed sedimentological and paleocurrent analysis on the basis of quality of exposure, presence of appropriate sedimentary structures, and simple structure. Of several hundred outcrops that were examined, fewer than 50 yielded satisfactory results. Even the best sites expose only a few tens of meters of section, representing but a small fraction of formations that typically are hundreds to thousands of meters thick. At each site, we measured the section bed-by-bed, worked out the local structure, and measured one paleocurrent indicator per bed. We tried to measure at least 20 paleocurrent indicators per section. Some locations simply did not yield this many paleocurrent measurements but are nonetheless reported. Most of the rocks are turbidites, containing, in order of decreasing abundance, cross-laminae (ripple foresets) (Fig. 4c), flutes (Fig. 4b), and parting lineations (Fig. 4e). Convolute fold axial planes and ripple backsets are present in some cross-laminated beds. In the non-turbiditic deltaic deposits, trough and planar cross-bedding and symmetrical ripple crests are present (Fig. 4a). We also analyzed a number of slump folds for paleoslope direction (Bradley and Hanson, 1998).

4. Analytical methods and potential sources of error

The Silurian and Devonian strata of Maine test the limits of paleocurrent analysis. Accordingly, we here discuss the methods used and the various pitfalls. Table 1 gives any special information pertinent to each site.

4.1. Discrimination between sedimentary and tectonic structures

The most basic concern in any orogenic belt is to discriminate between sedimentary and tectonic structures. This presented no problems in the chlorite-grade rocks of most of our study area, except rarely in silt turbidites where cleavage and cross-laminae are subparallel. Curiously, some the best-preserved paleocurrent indicators were found in andalusite- or cordierite-grade rocks in the narrow contact aureoles of plutons that intrude the low-grade terrane. Crook (1979) described similar preservation in a contact aureole in Australia. At higher grades of regional metamorphism (southwest part of

Fig. 3. Map of northern Maine showing locations where paleocurrents were measured. Black dots represent localities that yielded higher-quality paleocurrent data (Table 1). White dots represent spotty or lower-quality data. Letters A–H in squares refer to general locations of stratigraphic sections.

Table 1
Paleocurrent directions from Silurian and Devonian clastic strata in Maine

No.	Rock unit	Location	Paleoflow direction	Directional indicators	Notes and references
	Seboomook Gr., St. John River Fm.	Walker Brook			This study; see Bradley and Bradley (1994) for discussion of early thrust-faulting at this location
2	Seboomook Gr., St. John River Fm.	Big Rapids	256°	29 cross laminae	This study
67	Seboomook Gr., Hafey Pond Fm.	Oxbow Brook	010°	15 cross laminae	This study
P2	Matagamon Sandstone	Telos Brook quadrangle	350°	176 cross beds from all deltaic facies	Pollock (1994)
P1	Matagamon Sandstone	Grand Lake Matagamon area	285°	230 cross beds from delta-plain facies	Pollock et al., 1988
P3	Seboomook Gr., undivided	Caucomogomoc Lake area	~270°	cross laminae; no details given	Pollock (1994)
P2	Seboomook Gr., undivided	Telos Brook quadrangle	281°	33 cross laminae	Pollock (1994)
7c	Seboomook Gr., undivided	Harrington Lake, outcrop 3	~260°	55 cross laminae	This study; Kusky et al. (1994)
7d	Seboomook Gr., undivided	Harrington Lake, outcrop 4	~290° and ~250°	44 cross laminae	This study; Kusky et al. (1994)
P1	Seboomook Gr., undivided	Grand Lake Matagamon area	~285°	Flutes; no details given	Hall et al. (1976)
6a	Seboomook Gr., Northeast Carry Fm.	Seboomook Lake-above dam	187°	30 cross laminae	Bradley, 1987
6b	Seboomook Gr., Northeast Carry Fm.	Seboomook Lake-below dam	210°	30 cross laminae	This study
10	Seboomook Gr., Ironbound Mt Fm.	N. end Seboomook Lake	277°	15 cross laminae	This study
28	Jo-Mary Mt Fm.	Johnston Pond	~190°	14 cross laminae	This study
42	Carrabassett Fm.	Barrows Falls	343°	14 cross laminae	This study
33	Carrabassett Fm.	Wilson Pond penstock	~330° and ~290°	14 cross laminae	This study
37 and 38	Carrabassett Fm.	Big Wilson and Little Wilson Streams, composite	~340°	36 cross laminae 38 flutes	Hanson, 1994a,b, subject to 65° ccw strike rotation
39	Carrabassett Fm.	Tobey Falls	~305° and ~025°	25 cross laminae	This study
36a	Carrabassett Fm.	Borestone Mtn., Visitor's center	~010° and ~065°	30 cross laminae	This study, subject to 55° ccw strike rotation
30	Carrabassett Fm.	Horseshoe Falls	~350°	20 cross laminae	This study
44	Carrabassett Fm.	Upper Abbott	~180° but scattered	14 cross laminae	This study
29	Carrabassett Fm.	Gauntlet Falls	~340°	33 cross laminae	This study
25	Carrabassett Fm.	Norcross	~355°	19 cross laminae	This study
31	Carrabassett Fm.	Saddleback Mtn, west flank	~230°	30 cross laminae	This study
32a	Carrabassett Fm.	Gulf Hagas Brook	~085° and ~185°	42 cross laminae	This study
32b	Carrabassett Fm.	Gulf Hagas Brook	~150° and ~185°	14 cross laminae	This study
18	Forks Fm.	Moosehead Isle Estates	~075°	10 cross laminae 11 ripple crests	This study

Table 1 (*continued*)

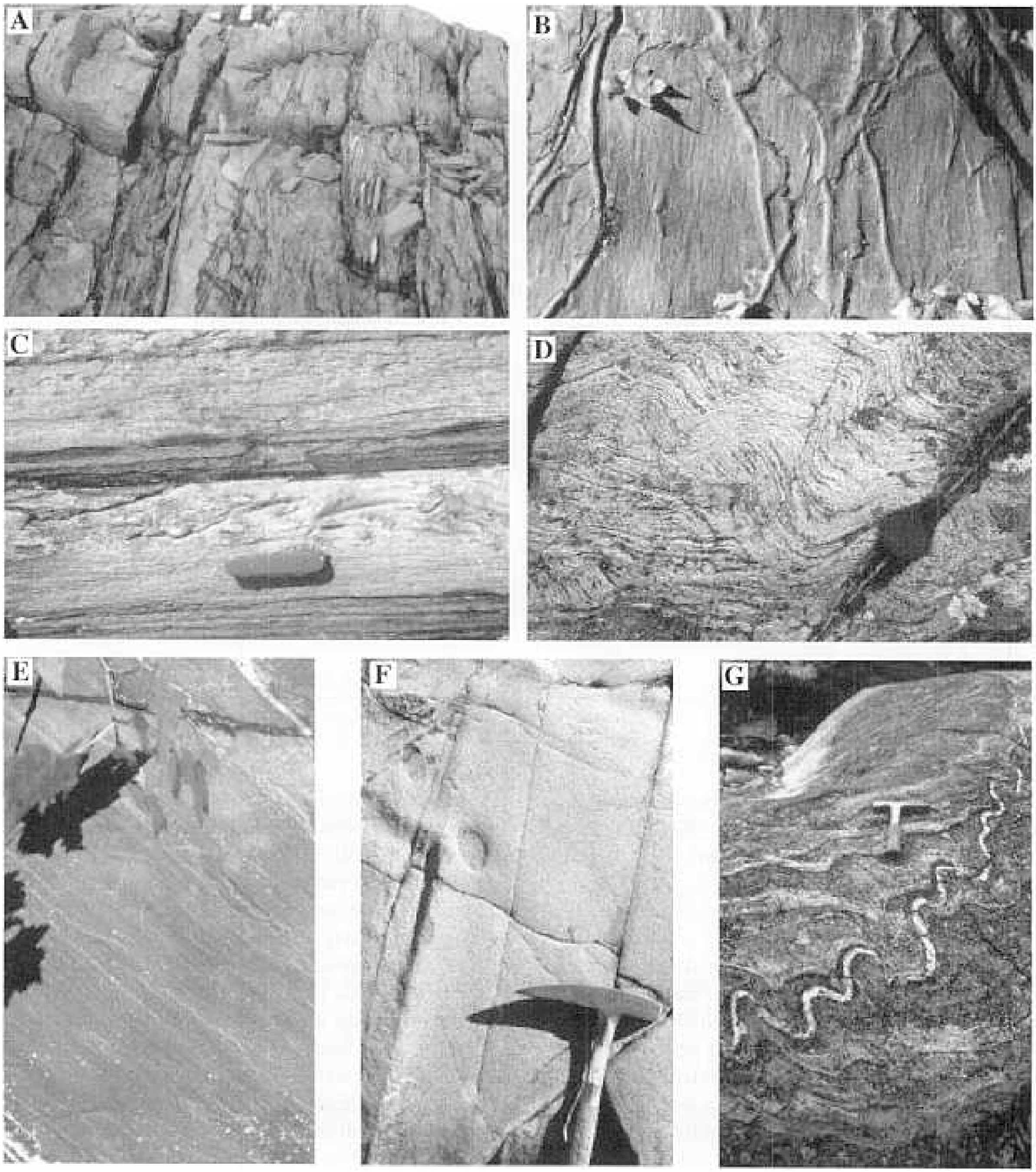
		Location	Paleoflow direction	Directional indicators	Notes and references
51	Madrid Fm.	Arnolds Landing			This study; result is strain corrected. Supercedes results in Bradley and Hanson, 1989
48	Madrid Fm.	Austin Stream	~250° and ~180°	11 cross laminae	This study
47	Madrid Fm.	Thorn Brook	247°	18 cross laminae	This study
52	Madrid Fm.	Fall Brook	~165° and ~240°	31 cross laminae	This study Supercedes results in Bradley and Hanson, 1989
53a	Madrid Fm.	N. Anson, gorge (outcrop A)	224°	6 cross laminae	This study
53c	Madrid Fm.	N. Anson (C) telephone lines	253°	5 cross laminae	This study
53d	Madrid Fm.	N. Anson (D) RR tracks	280°	6 cross laminae	This study
	Madrid Fm.	Quakish Lake (Stone Dam)	246°	19 cross laminae	This study
50	Smalls Falls	Houston Falls	148°	9 cross laminae	Double tilt correction; single tilt yields 139
49	Smalls Falls	Wyman Dam	152°	32 cross laminae	Double tilt correction; single tilt yields 204
22a	Allsbury Fm.	Millinocket Stream, outcrop 1	~345°	11 cross laminae	This study
22b	Allsbury Fm.	Millinocket Stream, outcrop 2	~255°	19 cross laminae	This study
22c	Allsbury Fm.	Millinocket Stream, outcrop 3	~225°	8 cross laminae	This study
26	Allsbury Fm.	West Seboeis Stream	~195° and ~270°	17 cross laminae	This study
	Allsbury Fm.	Grindstone-upper rapids	191°	14 cross laminae	This study
62	Sangerville Fm.	Athens	~230°	26 cross laminae	This study

Fig. 2), few directional indicators survived, and even when they did (e.g., Fig. 4D), the proper retrodeformation path was unknown.

4.2. Retrodeformation procedures

Exactly reversing the effects of deformation is a routine matter for dipping strata in structurally simple settings where fold plunge is negligible, and where there is no evidence for significant penetrative strain or vertical-axis rotations. The standard "single-tilt" restoration is appropriate in such cases; many textbooks illustrate the method (e.g., Potter and Pettijohn, 1977). This is the method used on most of the data in the present paper.

Where fold plunge is more than a few tens of degrees, the usual approach to structural correction is the "double-tilt" restoration (Ramsay, 1961): fold plunge is first removed, then bedding is unfolded to horizontal. We are wary of the validity of this procedure in our study area, and quote paleocurrent directions thus obtained with qualification. A case study of the Smalls Falls Formation at Wyman Dam (loc. 49) illustrates some of the problems. Thirty-two cross-laminae from Bouma T_c divisions from the limb of a plunging fold were rotated using both the double-tilt and single-tilt methods (Fig. 5). These methods yield vector mean paleocurrent directions that differ by 41°; certain individual results differ by as much as 53°. In fact, neither the single-tilt nor the double-tilt procedure seems particularly appropriate for the area. The single-tilt correction implies a single regional folding about horizontal axes, whereas the double-tilt correction implies that regional folding occurred first and was followed by a second period of folding or tilt, at right angles to the first. In this part of central Maine, there are no outcrop-scale structures, nor are there regional-scale fold interference patterns, that provide



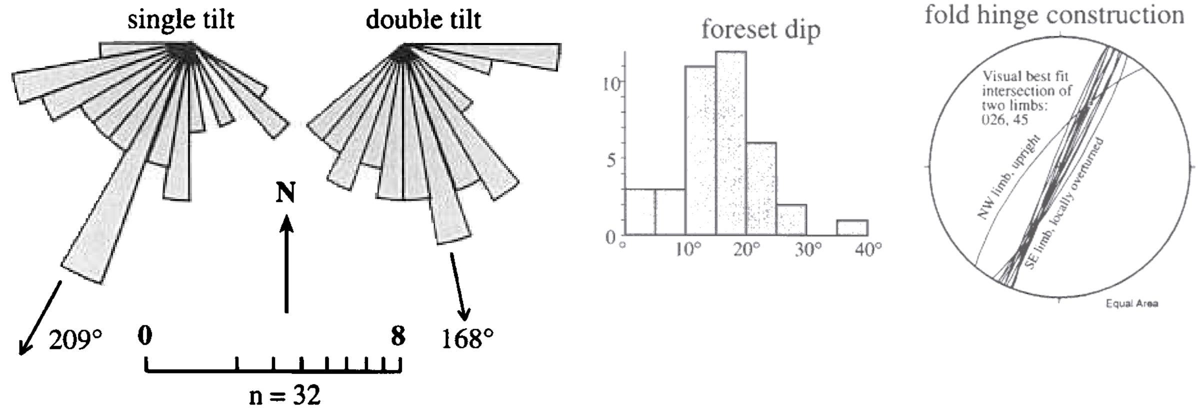


Fig. 5. Paleocurrent rose diagrams comparing the results of different tilt-correction methods on data from the Smalls Falls Formation at Wyman Dam (loc. 49). Six cross-laminae with foreset dips less than 10° were purged from the data set.

evidence for any such second deformation event. Instead, mesoscale folds display a variety of plunges that appear to have been imposed during a single progressive event. Accordingly, it may be more appropriate to unfold the rocks by removing plunge and bedding dip simultaneously and incrementally. This tedious process would yield paleocurrent directions somewhere between the values determined by the single-and double-tilt methods. Code to automate this process would be valuable.

Paleocurrent analysis in the vicinity of the Onawa pluton (locs. 34–38) illustrates another retrodeformation problem. The 405-Ma (early Emsian) pluton (Bradley et al., 2000) truncates regional-scale Acadian folds in the Carrabassett Formation, which have negligible plunge. In its aureole, contact-metamorphic minerals overprint a pre-existing cleavage (van Heteren and Kusky, 1994; Bradley et al., 2000). The map pattern clearly shows that bedding strike is deflected around the pluton, by about 55–65° clockwise in the area of our paleocurrent studies. This suggests a two-stage restoration procedure for paleocurrents: (1) remove strike deflection by an appropriate counter-clockwise rotation about a vertical axis, then (2) single-tilt correction. Steps 1 and 2 are commutative, and the opposite order is actually more convenient on the stereonet.

Unlike rigid rotations, penetrative deformation results in new angular relationships between bedding and sedimentary structures, which might lead to serious paleocurrent errors if ignored. Unfortunately, it is rarely possible to find good mesoscopic strain markers at outcrops that also preserve useful paleocurrent indicators. Consequently, the effects of strain are usually ignored, at the risk of introducing some unknown error. Orthographic strain analysis (De Paor, 1986) provides a simple way to unstrain directional indicators, which, to our knowledge, has not yet been widely applied by sedimentologists. Exposures at Arnold's Landing (loc. 51) provide the opportunity to document the effects of strain and to show, at least in this case, that the effect on paleocurrent directions is negligible for our purposes. Numerous turbidite cross-laminae of the Madrid Formation are exposed in an overturned homoclinal sequence. Fold plunge is negligible. Strain is triaxial. The plane of flattening

Fig. 4. Photographs of sedimentary and tectonic structures from Silurian and Devonian rocks, Maine. Locations are keyed to Fig. 1. (A) Trough-cross-bedded sandstone, Tomhegan Formation, Brassua Lake (loc. 15). (B) Burrows and small flute casts on a bedding surface, Carrabassett Formation, Big Wilson Stream (loc. 37). (C) Cross laminae, Seboomook Group, Harrington Lake (loc. 7). (D) Deformed cross-laminae in staurolite-grade rocks of the Seboomook Group, Bald Mountain (loc. 55). (E) Parting lineations from a Bouma T_b division, Carrabassett Formation, Big Wilson Stream (loc. 37). (F) Bedding surface showing strained concretions, Madrid Formation, Arnolds Landing (loc. 30). (G) Ptygmatically folded quartz vein cutting turbidites of the Seboomook Group, Bald Mountain (loc. 55). This photo, taken near (d), shows the intensity of deformation in rocks that still locally preserve cross-laminae—albeit ones that are of no value for paleocurrent analysis.

(XY), corresponds to a bedding-parallel foliation. The X direction pitches steeply, as is revealed by the preferred orientation of stretched sand grains, and by the long axes of calcareous concretions in sandstones (Fig. 4f). Axial ratios of four concretions exposed on bedding planes have a mean value of X:Y= 1:0.6. For the unstraining procedure, a conventional stereonet is first used to find the single-tilt-corrected orientation of two lines on bedding: X'' (the long axis of the bedding strain ellipse) and P'' (the paleoflow direction on bedding, still to be strain corrected) (Fig. 6a). Next, P'' and X'' are transferred to a new tracing-paper overlay on the orthonet with north at the top (Fig. 6b). As outlined by De Paor (1986), the bedding strain ellipse is constructed by tracing the appropriate great circle on the orthonet; a ruler is used to find the great circle such that X:Y= 1:0.6, where 1 is the radius of the orthonet. All that remains is to correct the paleocurrent indicator for strain. From the intersection between the strikeline of P'' and the strain ellipse, the paleoflow direction P is found by following the small circle straight to the edge of the net. Note that

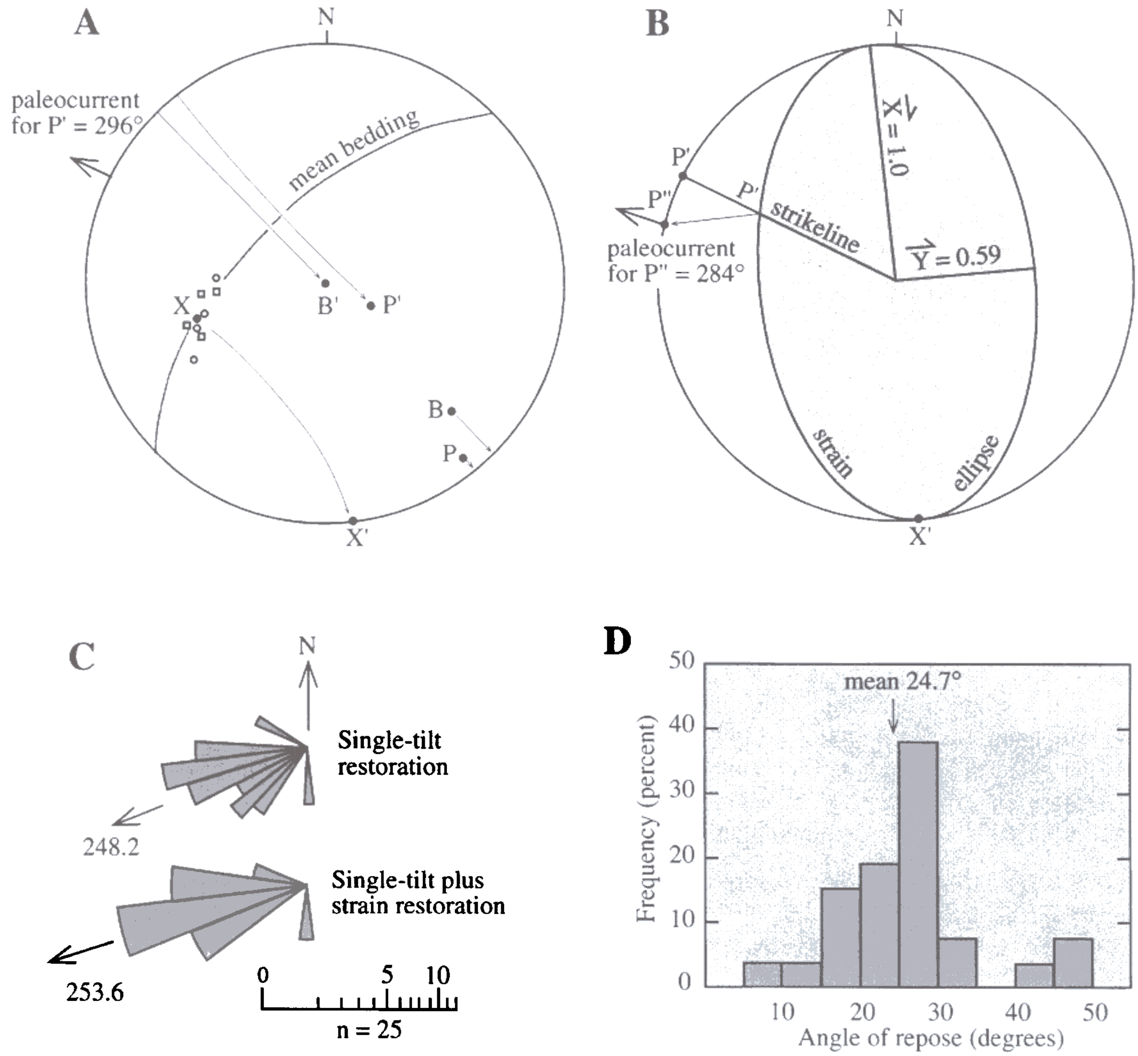


Fig. 6. Orthographic correction for strain based on an example from the Madrid Formation at Arnolds Landing (loc. 51).

the suggested order of operations (tilt correction, then strain correction) is less likely to replicate the deformation history than the reverse order—but either way, the operations are commutative, and the suggested order involves fewer steps. Rose diagrams comparing unstrained and strained paleocurrent directions from 26 cross-laminae are shown in Fig. 6c and d. The unstrained paleocurrent directions show an overall trend of 254°, which differs little from the trend of 249° prior to strain correction. Most paleocurrent directions shifted clockwise, but a few shifted counterclockwise, depending on initial orientation of P with respect to X. These results are encouraging because, as was noted, mesoscopic strain indicators are uncommon in the rocks we have studied.

4.3. Pavement exposures

Many of the best outcrops in Maine are glacially polished. Turbidite cross-laminae are commonly visible on such outcrops, but it is often difficult or even impossible to measure them accurately. Based on a case study at Seboomook Dam (loc. 9), Bradley (1987) showed that apparent paleocurrent directions measured from a set of 73 cross-laminae exposed on polished two-dimensional pavements gave essentially the same overall direction as a set of 30 cross-laminae exposed in three-dimensions from the same beds that were measured and analyzed in the normal way. As work on our regional paleocurrent survey progressed, however, we found that the need to resort to apparent directions rarely arose. With patience and a practiced eye, cross-laminae could almost always be measured in three dimensions.

4.4. Measurement errors

In any exposure, but particularly where three-dimensional exposures are marginal, some error in measurement is inevitable, both for sedimentary structures and bedding. A 2° error in measuring a cross-lamina with a 30° foreset dip, for example, could yield a paleocurrent error up to ±4°. For a 10° foreset dip, a 2° measurement error could yield a paleocurrent error as great as ±12°. The gentler the foreset dip, the greater the potential error in paleocurrent direction. Comparable paleocurrent errors also result from errors in bedding measurement. We took care to minimize any such errors. We calculated angles of repose for all foresets and discarded any cross-laminae with foreset dips less than 10°. While measuring many sedimentary structures on a given outcrop, we were careful *never* to round to the nearest 5° (many field geologists have acquired this seemingly harmless habit), and *always* to measure reference bedding at the slightest hint of a change.

An additional complication arises where the plunge of folds exceeds a few tens of degrees. Measurement errors involving fold data would result in yet another source of error in paleocurrent directions, compounding any errors like those described in the preceding paragraph. Moreover, one is faced with the problem of how to assess plunge. In the Wyman Dam case study, we stereographically estimated the trend and plunge of tectonic folds (Fig. 5) and used this value for all double-tilt restorations. Another reasonable approach would have been to determine a different local fold axis orientation corresponding to each of 11 bedding measurements. Yet another approach would have been to rely only on direct measurements of mesoscopic fold axes. Each of these approaches would yield paleocurrent measurements that might differ by as much as 10–20°.

4.5. Possible bias towards exposure of strike-parallel paleocurrents

At a given outcrop, the paleocurrent indicators that can be readily observed and measured may depend on the attitude of the surface of exposure, dip of bedding, bedding-cleavage intersection angle, and the orientation of paleoflow with respect to tectonic strike. Simply owing to the nature of the terrain, the majority of turbidite outcrops we studied are subhorizontal glacial pavements with steep bedding and cleavage. These conditions favor observations of cross-laminae produced by strike-parallel flow. Cross-laminae produced by flow normal to or oblique to strike, would, in contrast, be displayed as festoons on typical pavement exposures. Because such cross-laminae cannot be as readily measured or interpreted, the natural tendency is to skip these and move on to the next bed. Fortunately, despite the preponderance of outcrops favoring the detection of strike-parallel paleocurrents, our data do include many paleocurrents that are at a high angle to strike.

5. Paleocurrent data from Silurian and Devonian strata

Paleocurrent directions are presented below for twelve units, beginning with Silurian strata of the Central Maine basin, and working cratonward into the Devonian strata of the Moose River synclinorium and Connecticut Valley–Gaspe basin (Fig. 7). We also review paleocurrent data from these and correlative units published by other workers.

5.1. Sangerville Formation (Ss)

The Sangerville Formation occupies the present axial region of the Central Maine basin (col. G in Fig. 7). Its age has been cited as late Llandovery through early Ludlow (Moench and Pankiwskyj, 1988) but the only tightly dated fossils are Wenlock

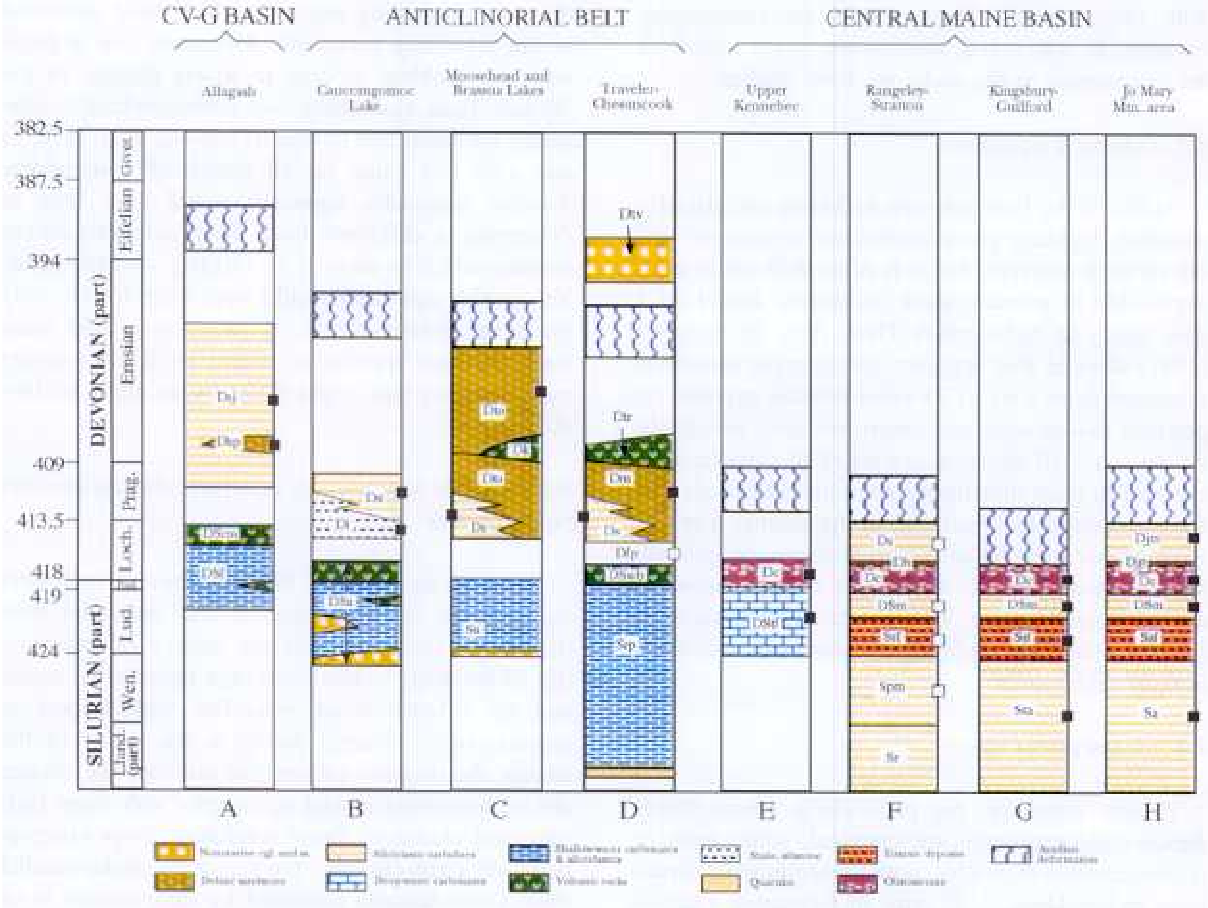


Fig. 7. Stratigraphic sections of Silurian and Devonian rocks considered in this study. Numerical ages of the various stage boundaries are from the time scale of Tucker et al. (1998). Abbreviations for rock units are as follows. Dc—Carrabassett Fm.; Dfp—Frost Pond Fm.; Dh—Hildreths Fm.; Dk—Kineo Rhyolite; Dm—Matagamon Ss.; Ds—Seboomook Gr. (formerly Fm.); Dta—Tarratine Fm.; Dto—Tomhegan Fm.; Dtr—Traveler Rhyolite; Dtv—Trout Valley Fm.; Du—Unnamed Devonian strata; DSm—Madrid Fm.; DStf—The Forks Fm.; DSwb—West Branch Volcanics; Spm—Perry Mtn. Fm.; SSr—Rangeley Fm.; Srp—Ripogenus Fm.; Ssa—Sangerville Fm.; Ssf—Smalls Falls Fm.; Su—Unnamed Silurian strata. Adapted from Bradley et al. (2000 and references therein) except as follows. Allagash column from Roy et al., 1991; Caucomgomoc column from Pollock (1985); Jo Mary Mountain column from Hanson and Bradley (1989). Black squares show approximate position of higher-quality paleocurrent data (Table 1). White squares show position of spotty or lower-quality data.

graptolites (Pankiwskyj et al., 1976). The Sangerville Formation consists of an estimated 1600 m of thin-to-thick bedded siliciclastic turbidites, metamorphosed deep-water limestone, metapelite, and metaconglomerate (Ludman, 1976). In most outcrops, Sangerville sandstones are strongly foliated and not suitable for paleocurrent work. We were able to obtain satisfactory paleocurrent measurements from only one locality

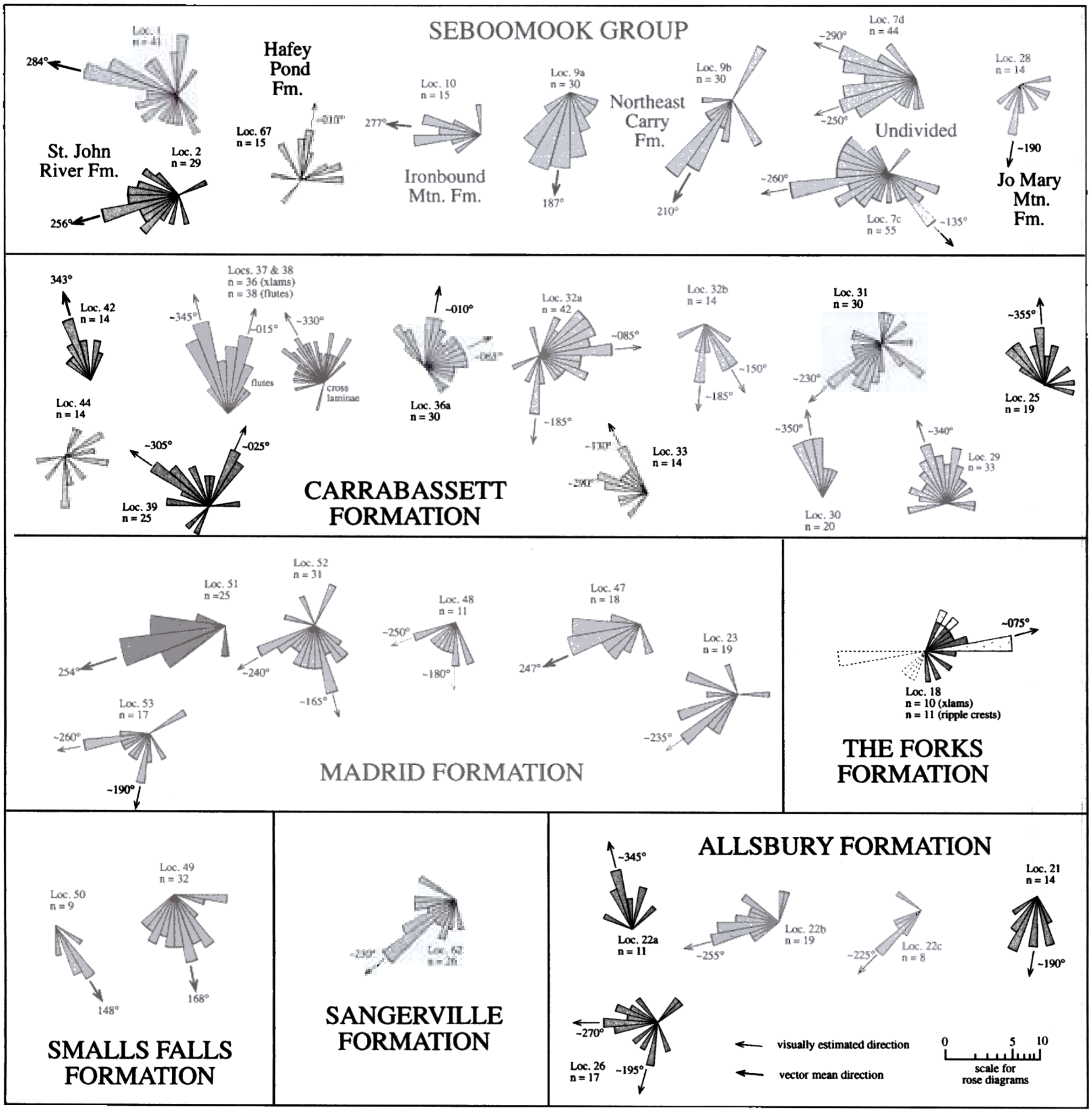


Fig. 8. Rose diagrams showing paleocurrent data from locations listed in Table 1. All rose diagrams are drawn to the same scale so the bigger ones represent more data. The area (rather than length) of each petal is proportional to the number of measurements. Arrows shown on paleocurrent maps (Figs. 9 and 11, below) are generalized from these rose diagrams.

(#62, Athens), and a few results from three other localities: Sangerville (loc. 61), Malbon's Mills (loc. 63), and Dover-Foxcroft (loc. 66).

At Athens, a series of outcrops along Wesserunsett Stream expose a few tens of meters of mixed siliciclastic and carbonate turbidites. The lower part of the section consists of coarse, gritty, noncalcareous feldspathic–lithic sandstone beds in Bouma T_{ab} sequences up to 2 m thick. These alternate with more calcareous thin-bedded, variably calcareous turbidites containing the Bouma T_{bcde} and T_{cde} divisions. In beds containing both the Bouma T_b and T_c divisions, the latter are markedly more calcareous. This shows that the same flows carried both siliciclastic and carbonate detritus and that the segregation into carbonate-rich and carbonate-poor layers was due to hydrodynamics and not, for example, due to interfingering of fans from different sources. We measured 26 turbidite cross-laminae at Athens which show a clear southwesterly flow direction (toward 232°) (Figs. 8 and 9). Axial planes of convolute synsedimentary folds are overturned toward 200° and thus imply a similar direction.

Bedrock exposures at the gravel pit in Sangerville (loc. 61) consist of less than 10 m of thin- and medium-bedded sandstone interbedded with gray slate. Sandstone beds, some of which contain calcareous concretions, range in thickness up to 75 cm. Seven cross-laminae from Bouma T_{bcde} and T_{cde} sequences show a scatter of southwesterly, westerly, and northwesterly paleocurrents. At Malbon's Mills (loc. 63), about 25 m of section are exposed at the old millrace on Wesserunsett Stream; the section is homoclinal and structurally uncomplicated. The rocks include a 13-m thick massive sandstone body and some thin-bedded turbidites beneath. A few cross-laminae are somewhat scattered but give an overall westerly direction. Along the Piscataquis River at Dover-Foxcroft (loc. 66), the Sangerville Formation consists of thinly interbedded fine sandstone and slate; some of the sandstone beds show cross-laminae but foreset dips were deemed too low-angle to give reliable paleocurrent directions.

5.2. Allsbury Formation (Sa)

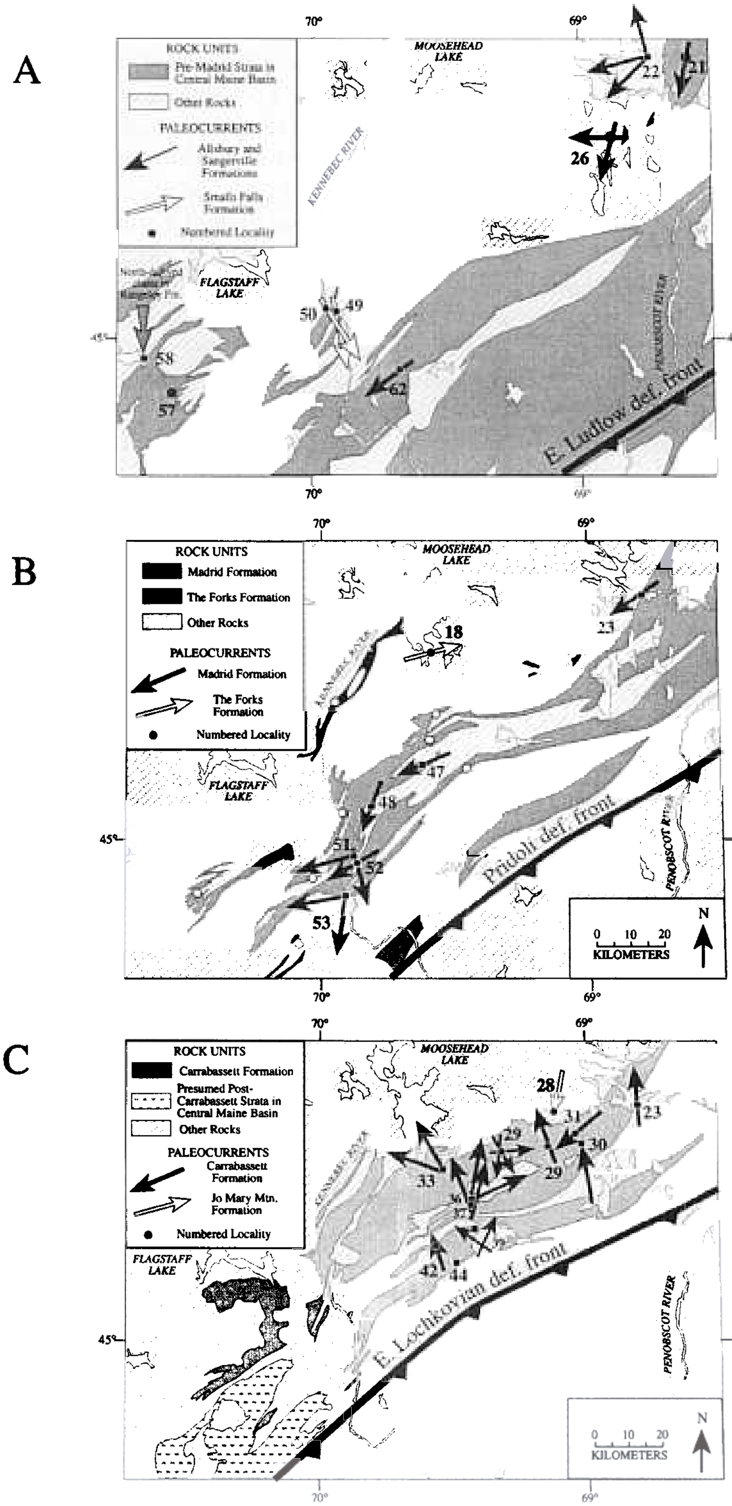
The Allsbury Formation is a deep-water formation that underlies much of the region where the Central Maine and Aroostook–Matapedia basins merge (near col. H in Fig. 7). It ranges from Llandovery to perhaps Wenlock in age (Roy 1981, p. 4). Roy (1981) divided it into a lower Sandstone Member and an upper Slate Member. The predominant rocks are turbiditic sandstone and conglomerate interbedded with gray and green slate, plus minor limestone. Neuman (1967, p. I-24) estimated the thickness of the Sandstone Member at approximately 1500–2300 m. We obtained satisfactory paleocurrent data from one locality in the Sandstone Member (loc. 21), where it consists of D- and B-facies siliciclastic turbidites deposited by southwest-directed currents (Figs. 8 and 9).

Four other outcrops that we assign to the Slate Member of the Allsbury Formation also yielded satisfactory paleocurrent data (Figs. 8 and 9). Along Millinocket Stream (loc. 22), three separate outcrops of thin-bedded turbidites interbedded with green slate yielded northerly, westerly, and southwesterly directions. These outcrops are in an area that has never been mapped in detail, but fall just outside the area mapped by Roy (1981), on strike with the Slate Member, which includes similar strata. At West Seboeis Stream (loc. 26), thin-bedded turbidites associated with gray slate and lenses of calcareous siltstone yielded southerly paleocurrents. This area, which also has never been mapped in detail, was assigned to the Carrabassett Formation on the Bedrock Geologic of Maine (Osberg et al., 1985). Calcareous rocks, however, are not known from other parts of the Carrabassett Formation, but are a typical lithology of the Allsbury Formation.

5.3. Smalls Falls Formation (Ssf

The Smalls Falls Formation is a distinctive rusty-weathering unit of thin-bedded (mainly D-facies) turbiditic metasandstone and metasiltstone beds that alternate with black, sulfidic metapelite. It is 700 m

Fig. 9. Non-palinspastic maps showing mean paleocurrent directions for selected rock units: (A) Sangerville, Allsbury, and Smalls Falls Formations; (B) Madrid Formation; (C) Carrabassett Formation. Successive positions of the deformation front, from Bradley et al. (2000), are interpolated between coeval pre-orogenic strata and syn- to post-orogenic plutons.



A
70°
69°
ROCK UNITS
Other Rocks
PALEOCURRENTS
Numbered Locality
MOOSEHEAD LAKE
KENNEBEC RIVER
FLAGSTAFF LAKE
PENOBSCOT RIVER
E. Ludlow def. front
22
21
26
50
49
58
57
62
45°
B
ROCK UNITS
Madrid Formation
The Forks Formation
Other Rocks
PALEOCURRENTS
Madrid Formation
The Forks Formation
Numbered Locality
MOOSEHEAD LAKE
FLAGSTAFF LAKE
PENOBSCOT RIVER
Pridoli def. front
18
23
47
48
51
52
53
0 10 20
KILOMETERS
N
C
ROCK UNITS
Carrabassett Formation
Presumed Post-Carrabassett Strata in Central Maine Basin
Other Rocks
PALEOCURRENTS
Carrabassett Formation
Jo Mary Mtn. Formation
Numbered Locality
MOOSEHEAD LAKE
KENNEBEC RIVER
FLAGSTAFF LAKE
PENOBSCOT RIVER
E. Lochkovian def. front
28
31
23
29
30
33
42
44
0 10 20
KILOMETERS
N

thick at its type locality in western Maine (Moench and Pankiwskyj, 1988) (col. F in Fig. 7). Its age has been cited as ranging from Wenlock through early Ludlow (Moench and Pankiwskyj, 1988) but the only tightly dated fossils are graptolites of early Ludlow age (Pankiwskyj et al., 1976). In the few large outcrops that exist of the Smalls Falls, sedimentological studies are hampered by rust coatings. At its type section (loc. 57), the Smalls Falls Formation consists of about 50% rusty metapelite, alternating with thin-bedded turbidites containing Bouma T_{cde} divisions, disharmonically folded horizons of probable soft-sediment deformation, and starved ripples. We were able to obtain meaningful paleocurrent data from only two places (Wyman Dam and Houston Brook; locs. 49 and 50, between cols. 6 and 7 in Fig. 7) which show approximately southerly and southeasterly paleocurrent directions (Figs. 8 and 9). The problems of structural correction at Wyman Dam were discussed earlier.

5.4. The Forks Formation (DStf)

The Forks Formation (Marvinney, 1984) underlies the Carrabassett Formation along the northwestern margin of the Central Maine Basin (col. E in Fig. 7). It consists of interbedded slate and calcareous siltstone, deposited below storm base. The Forks Formation is Wenlock to earliest Lochkovian in age (Bradley et al., 2000). The best exposures are a series of new roadcuts at Scammon Ridge (loc. 18), where a section that we assign to The Forks Formation is gradationally overlain by the Carrabassett Formation. This succession, which lies within the contact aureole of the Moxie pluton, is dominated by siltstone, much of it burrowed. Some calcareous rocks are present; turbidites are absent. Paleocurrents from cross-laminated siltstones reveal paleoflow toward the east (Figs. 8 and 9).

5.5. Madrid Formation (DSm)

The Madrid Formation, which gradationally overlies the Smalls Falls Formation, records both an end to anoxia and an influx of sand from a new direction. The Madrid Formation, as used here, includes the Fall Brook Formation of Pankiwskyj et al. (1976), the Brighton Formation of Ludman (1976), and the Lawler Ridge Formation of Roy (1981). The Madrid is unfossiliferous but has been assigned a Pridolian to early Lockhovian age (Osberg et al., 1985) on the basis of regional correlations and the well-constrained age of the Smalls Falls which it depositionally overlies. In its type area in western Maine (col. F in Fig. 7), the Madrid Formation consists of a lower calcareous member and an upper member of medium to thick-bedded quartz-plagioclase turbidites. The upper sandstone member thickens to the east (about 1500 m in the Skowhegan quadrangle, near col. G in Fig. 7), at the expense of the lower calcareous member, which pinches out. Our studies of the upper member (or "eastern facies") at localities 47, 48, 51, 52, and 53 (cols. G and H in Fig. 7) have revealed turbidite Facies B, C, and D (Bradley and Hanson, 1989). The best exposures are at Arnolds Landing (loc. 51) where the uppermost ~ 30 m of the Madrid Formation is exposed below a gradational contact with the Carrabassett Formation (Bradley and Hanson, 1989). The thickest sandstones here (1–2 m) are complete Bouma sequences; most of the section, however, is made up of turbidites beginning with the Bouma T_b or T_c divisions. A uniform, southwesterly transport direction is revealed by abundant turbidite cross-laminae (Figs. 8 and 9). The strain-correction case study discussed above was done at Arnolds Landing.

Quartz-rich turbidites in isolated exposures at Stone Dam (loc. 24) closely resemble Madrid turbidites at the above localities, and they also show southwesterly paleocurrents (Fig. 8). Although Roy (1981) mapped these rocks as Allsbury Formation, we suggest that an assignment to the Madrid Formation is more reasonable. Whichever the case, the regional paleocurrent direction is about the same for both units (Fig. 9).

5.6. Carrabassett Formation (Dc)

The youngest regionally extensive unit in the Central Maine basin is the Carrabassett Formation, which conformably overlies the Madrid Formation. It has been assigned an Early Devonian, probably Lockhovian age, but has not yet yielded age-diagnostic fossils. Thickness, although difficult to determine owing to structural complexity and an absence of marker units, is probably about 2000 m where the unit has been best studied, in the Jo Mary Mountain 15′ quadrangle (col. H in Fig. 7; locs. 29–31) (Hanson and Bradley, 1989). The Carrabassett Formation is a mud-rich turbidite succession with an internally complex stratigraphy (Hanson and Bradley, 1989; Hanson, 1988, 1994a,b). Disrupted strata of olistostromal origin comprise about 70% of the unit. Interbedded with the olistostromes are massive sandstones and thinner-bedded turbidites. We have suggested that slope, slope-basin, and base-of-slope environments are all represented in the Carrabassett Formation (Hanson and Bradley, 1989) (Fig. 10). Sandstones of the Carrabasett Formation consist largely of quartz and plagioclase and appear similar to sandstones of the Madrid Formation. Paleocurrent data are now abundant for the Carrabassett Formation. Data from 11 localities reveal a dominant flow direction toward the north, and secondary flow directions toward the east, northwest, and south (Figs. 8 and 9).

5.7. Ironbound Mountain Formation (Di)

The Seboomook Group (of Pollock, 1987; formerly Seboomook Formation) has been subdivided into two formations in the Moose River synclinorium (cols. B and C in Fig. 7). The Ironbound Mountain Formation, the older of the two, is predominantly slate (Marvinney, 1984). Marvinney (1984) assigned it an

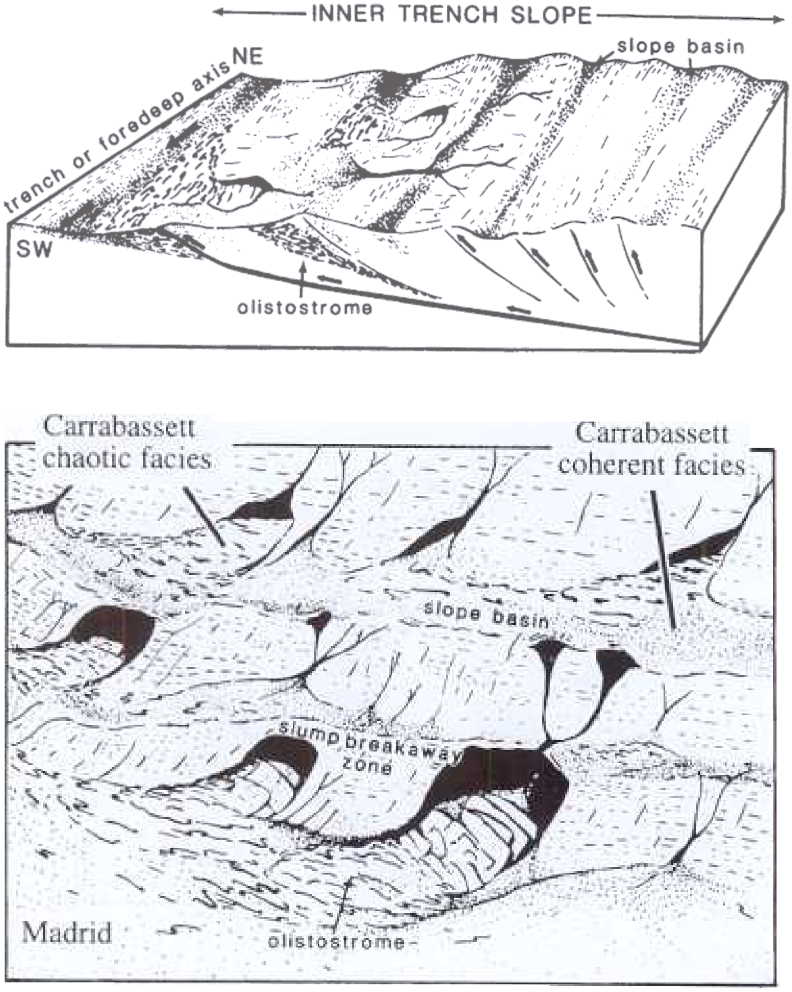


Fig. 10. Depositional model for the Carrabassett Formation, from Hanson and Bradley (1989).

age of "Late Silurian – Early Devonian", but fossil control is poor; he estimated its thickness at about 900 m. At locality 10, cross-laminated silt turbidites show west-directed paleocurrents (Figs. 8 and 11).

5.8. Northeast Carry Formation (Dn)

The upper part of the Seboomook Group in the Moose River synclinorium is assigned to the Northeast Carry Formation (col. B in Fig. 7). These strata, which include the type section of Boucot's (1961) Seboomook Formation, attain a maximum thickness of about 6100 m. Brachiopods suggest that it is mainly Pragian in age, although locally, the lower part is Lockhovian (Boucot and Heath, 1969; Bradley et al., 2000). The formation consists largely of thin- and medium-bedded turbidites which contain the Bouma T_{cde} sequence and are assigned to facies D

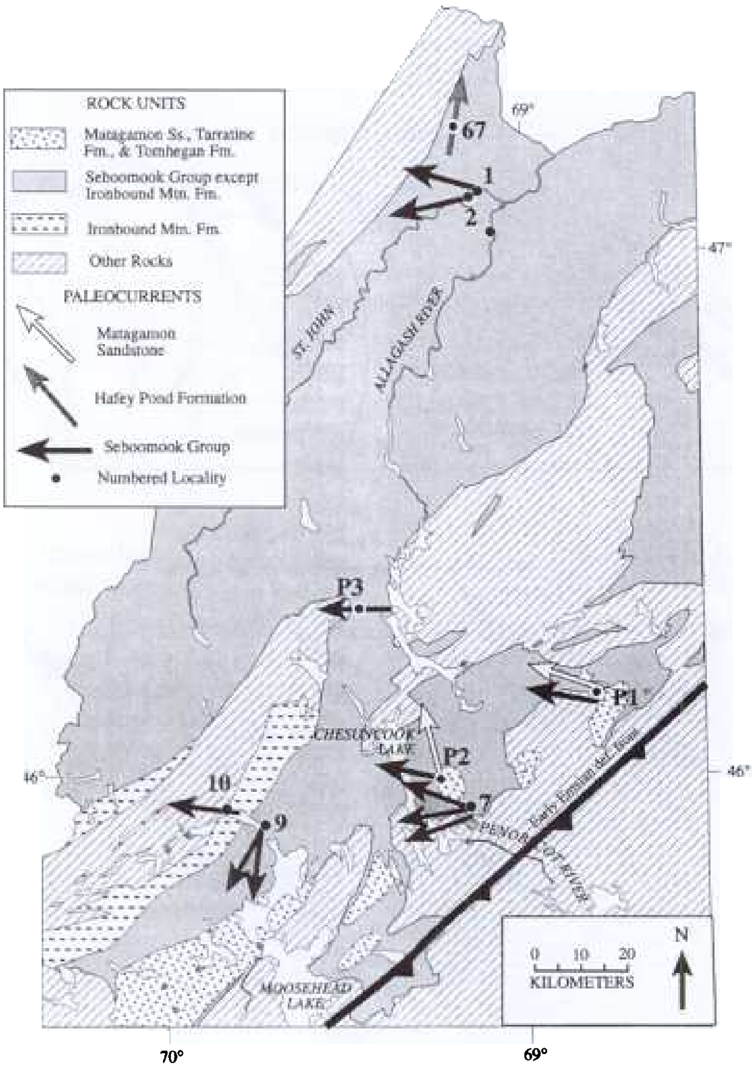


Fig. 11. Non-palinspastic map showing mean paleocurrent directions for the Seboomook Group, Matagamon Sandstone, and related units. Position of the deformation front, from Bradley et al. (2000), is interpolated between coeval pre-orogenic strata and syn- to post-orogenic plutons.

of Mutti and Ricchi-Lucci (1978). Abundant paleocurrents were measured at two widely separated outcrops at the type area at Seboomook Dam (loc. 9). Southwesterly paleoflows were obtained below the dam, and southerly paleoflows at low-water outcrops above the dam (Bradley, 1987).

5.9. Undivided Seboomook Group (Ds)

Much of Pollock's (1987) Seboomook Group has not yet been subdivided into formations. At widely separated outcrops at Harrington Lake (col. D in Fig. 7; loc. 7), the Seboomook Group shows west-directed paleoflows (Figs. 8 and 11). In the Grand Lake Matagamon area (loc. P1), Hall et al. (1976, p. 61) reported that an unstated number of flute casts indicate paleoflow toward 280° (Fig. 11). Pollock (1994) reported overall westerly paleoflows based on 33 measurements at various Seboomook outcrops in the Telos Brook area (loc. P2) (Fig. 11). He also showed westerly to northerly paleocurrents in the Caucomgomoc Lake area (loc. P3) (Fig. 11), apparently based on 12 paleocurrent measurements.

5.10. Matagamon Sandstone (Dm)

The Matagamon Sandstone is a deltaic sandstone succession that prograded from east to west across the deeper-water Seboomook Group (col. D in Fig. 7) (Hall et al., 1976). Detailed studies by Pollock et al. (1988) at Grand Lake Matagamon indicate that delta front and delta plain environments are represented. The mean paleocurrent direction of 763 cross beds in all subenvironments is about 288° (Pollock, 1994). Pollock (1994) reported paleocurrent directions from two facies of the Matagamon in the Telos Brook area (loc. P2). The shoreface-estuarine facies showed westerly paleflows (Fig. 11). Paleoflows from the delta front facies are mainly toward the northwestern quadrant, but a subordinate flow direction was toward the northeast, which Pollock (1994) interpreted as the approximate trend of the shoreline.

5.11. St. John River Formation (Dsj)

In northwestern Maine, the axis of the Connecticut Valley–Gaspe basin is underlain by a thick succession of Devonian slates and turbidites assigned to the informally named St. John River Formation of the Seboomook Group (col. A in Fig. 7) (Roy et al., 1991). Emsian brachiopods have been reported from approximately equivalent rocks of the Temiscouata Formation, just across the Canadian border to the north (St. Peter and Boucot, 1981). The rocks consists of turbidite facies B, C, D, abundant contractional slumps (facies F), and mud turbidites. Paleocurrent data from two localities (Fig. 2, locs. 1 and 2) indicate sediment transport toward the west (Figs. 8 and 11). Hesse and Dalton (1995) reported paleocurrent measurements from the Temiscouata Formation at its type area, not far off the north end of Fig. 3. These are predominantly toward the northwest, with a secondary maximum toward the southeast. Hesse and Dalton (1995) considered this somewhat bimodal pattern to be the result of deposition on levees, and they speculated that the dominant flow of the turbidite system was toward the southwest, along the basin axis towards Maine.

5.12. Hafey Pond Formation (Dhp)

This informally defined subunit of the Seboomook Group is known only from reconnaissance studies in extreme northwestern Maine (col. A in Fig. 7) (Roy et al., 1991). It consists of thinly interbedded quartz-rich sandstone and slate, and occasional meter-thick beds of graywacke. A 17-m section at Oxbow Brook yielded 15 cross-laminae that indicate scattered flow directions, generally toward the north (Figs. 8 and 11).

6. Paleogeographic implications

6.1. Central Maine basin

The paleocurrent data presented here confirm and refine the hypothesis that an outboard-derived clastic succession spread across Maine in advance of the Acadian orogen (Bradley et al., 2000). In the Central Maine basin, the sequence of events is most clearly revealed in the deep-water turbidite succession along the basin's northwestern flank. Conglomerate clasts in the Llandovery-age Rangeley Formation were shed to the southeast from sources such as the Ordovician Attean pluton (Moench and Pankiwskyj, 1988). The Rangeley grades upward into the compositionally more mature, and finer-grained, Wenlock(?)-age Perry Mountain Formation, which seems to represent a continuation of the same paleogeography, but with the gradual degradation of the northwesterly source terrane, i.e., the Boundary Mountains anticlinorium (Moench and Pankiwskyj, 1988). Our meager paleocurrent data from the Smalls Falls Formation suggest that it was also shed toward the southeast.

The Madrid Formation, in contrast, was deposited by southwest-flowing turbidity currents, which flowed parallel to the basin's axis. The paleogeographic reversal was fully completed by the onset of deposition of the Carrabassett Formation, which consists largely of olistostromes interbedded with north-flowing and subordinate east-flowing turbidity currents. We speculate that the Carrabassett's complex pattern results from a combination of downslope flow, mainly toward the north, and alongslope flow in perched basins, toward the east (Fig. 10). A key point is that the slope formed in a place where, for tens of millions of years, many kilometers of turbidites had been accumulating in deep water. The sudden existence of the Carrabassett slope, followed shortly by an end to all sedimentation and then by Acadian orogenesis, suggests the paleogeographic setting shown in Fig. 10. We have suggested, accordingly, that the Carrabassett slope was the leading edge of the submarine Acadian orogenic wedge, and that the slope advanced over the Madrid foredeep axis (Hanson and Bradley, 1989; Bradley and Hanson, 1989). A similar Neogene succession of slope deposits *conformably* overlying ocean-floor turbidites has been described from the thickly sedimented Makran accretionary prism of Pakistan (Harms et al., 1984). Subduction – accretion need not be marked by an angular unconformity in the deep-marine record.

Circumstantial evidence suggests that foredeep sedimentation commenced slightly earlier in the more outboard, southeasterly part of the Central Maine basin. Acadian deformation had already taken place in southeasternmost Maine by the time the Pocomoonshine pluton was intruded at about 423 Ma (West et al., 1992) (early Ludlow according to the time scale of Tucker et al., 1998). Deformation probably took place during Wenlock time, i.e., after deposition of the Digdeguash Formation in the Llandovery, but before plutonism. At least some parts of Sangerville and Allsbury Formations were being deposited by axial, southwest-flowing turbidity currents during Wenlock time. These turbidites could represent the same tectonic environment as inferred for the Madrid Formation, but at an earlier time when the foredeep axis lay farther outboard.

6.2. Moose River and Traveler synclinoria

The onset of Seboomook Group deposition is one of the most obvious and significant changes in depositional regime in the history of the orogen, on a par with the Ordovician platform drowning associated with the Taconic collision (e.g., Bradley, 1989). Seboomook Group turbidites inundated a varied Early Devonian paleogeography that included shallow marine limestone (Beck Pond Limestone; Boucot and Heath, 1969), red shale (Frost Pond Shale of Griscom, 1976), and locally derived conglomerate (e.g., Hobbstown Formation; Boucot and Heath, 1969). Water depth and the rate of subsidence both increased dramatically with the deposition of the Seboomook Group flysch. Abundant paleocurrents show that the Seboomook Group was deposited by west-flowing turbidity currents. In the major synclines are preserved deltaic deposits of the Tarratine and Tomhegan Formations and Matagamon Sandstone, which prograded over the Seboomook. The molasse shows the same overall westerly paleocurrents as the underlying flysch. Seboomook deposition got underway in Lochkovian where the transition is tightly dated, but the bulk of the sediment evidently was deposited during Pragian time; molasse deposition continued into Emsian (Boucot and Heath, 1969). Recent U/Pb and $^{40}Ar/^{39}Ar$ age determinations on plutons that intrude the Central Maine basin show that the Acadian orogeny was already well underway by early Emsian, when the Matagamon and Tomhegan were still being deposited (Bradley et al., 2000).

6.3. Connecticut Valley–Gaspe basin

In the Connecticut Valley – Gaspe basin, the transition to Acadian flysch sedimention is most clearly revealed, and most accurately dated, along the basin's northwestern margin in Quebec. Near Lake Temiscouata (just north of the northern edge of Fig. 3), the Eifelian-age Touladi Limestone is gradationally overlain by flysch of the Temiscouata Formation. Deposition and deformation of the Temiscouata flysch in no older than the *costatus* Zone of the Eifelian, and is probably a bit younger (Bradley et al., 2000). Only a few tens of kilometers to the east of Touladi, the Temiscouata has yielded Emsian brachiopods (St. Peter and Boucot, 1981). This is further confirmation that the base of the flysch succession gets younger toward the craton. West-directed paleocurrents in the St. John River Formation, on strike with the Emsian part of the Temiscouata Formation, agree with paleocurrents from the older, Pragian part of the Seboomook Group farther southeast.

7. Closing remarks

How representative are the directional indicators that we report—mostly cross-laminae—of the depositional system as a whole? Many situations, such as levees, exist on submarine fans where the local flow direction may not be representative of the overall fan (e.g., Hesse and Dalton, 1995). For the units that yielded only a few paleocurrents, this possibility cannot be ignored. For the units that yielded a wealth of paleocurrent data, however, the paleocurrent maps show undeniable, broad trends that we believe are representative of the regional dispersal pattern. In light of all the possible sources of error and misinterpretation, the paleocurrent directions that we quote are probably accurate to within a few tens of degrees.

This study has shown that the Acadian-derived clastic succession spread cratonward across Maine during Late Silurian and especially Early Devonian times. The diachronous march of the foredeep in Maine is consistent with the broader regional pattern documented by Bradley et al. (2000) in the northern Appalachians and by Ettensohn (1987) in the central Appalachians. The existence of coeval, paired belts of flysch sedimentation in the foreland and of deformation in the orogen provides a sediment source for the vast "Devonian slate belt" of Maine and adjacent areas, previously a mystery (Boucot, 1970). Given the complex paleogeography of the rocks immediately underlying the Acadian flysch-molasse succession, and a basement of recently accreted terranes lacking the flexural rigidity of a craton, it should be no surprise that this particular foreland basin had a more complicated history and sediment-dispersal pattern than foredeeps associated with "simple" arc-passive margin collisions, such as the Taconic.

The Catskill clastic succession of New York and Pennsylvania (Fig. 1) has been regarded as "the" Acadian foreland basin. Our paleocurrent work in Maine supports an alternative view (Bradley et al., 2000): that the Acadian foreland basin originated hundreds of kilometers farther outboard, that it migrated across strike as orogeny progressed, and that the Catskill sequence merely marks its final position and youngest incarnation.

Our work, together with that of Pollock (e.g., 1994), has adequately revealed the sediment dispersal patterns of the Madrid Formation, the Carrabassett Formation, and the main outcrop belt of the Seboomook Group. Slump folds in the latter two units show the same regional pattern as the paleocurrents (Bradley and Hanson, 1998). Further paleocurrent analysis in these units would probably lead to diminishing returns. On the other hand, additional paleocurrent studies of the Rangeley, Perry Mountain, Smalls Falls, Sangerville, and Allsbury Formations would undoubtedly shed new light on the evolution of the Central Maine basin, which remains the key to unraveling the plate geometry that led to Acadian collision in the Silurian.

The Silurian and Devonian strata of Maine have been traced southward into New Hampshire and Massachusetts (Hatch et al., 1983), where the Acadian orogeny obliterated any fossils and paleocurrent indicators that presumably once existed. Our results therefore should be taken into account in ongoing efforts to unravel the pre-Acadian history of the high-grade metamorphic tract in central and southern New England.

Acknowledgements

Funding was provided by NSF Grant EAR-8803233, and a series of contracts and grants from the Maine Geological Survey. Scott Sauchuck developed most of our Macintosh software for paleocurrent analysis. Much of our fieldwork was done with the logistical support of Boston University's geology field camp. Bradley's salary during manuscript preparation was covered by a USGS Gilbert Fellowship. We benefited from advice and field trips with Dee

Caldwell, Phil Osberg, Gary Boone, Dave Roy, Steve Pollock, Allan Ludman, and Bob Moench, and from geochronological collaboration with Dan Lux and Bob Tucker. We thank reviewers Peter Cawood, Timothy Kusky, and Dee Caldwell for their efforts to clarify the manuscript.

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