Structural Geology of Parautochthonous and Allochthonous Terranes of the Penokean Orogeny in Upper Michigan—Comparisons with Northern Appalachian Tectonics

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Chapter Q

Structural Geology of Parautochthonous and Allochthonous Terranes of the Penokean Orogeny in Upper Michigan—Comparisons with Northern Appalachian Tectonics

By WILLIAM J. GREGG

U.S. GEOLOGICAL SURVEY BULLETIN 1904

CONTRIBUTIONS TO PRECAMBRIAN GEOLOGY OF LAKE SUPERIOR REGION

P.K. SIMS and L.M.H. CARTER, Editors

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Structural Geology of Parautochthonous and Allochthonous Terranes of the Penokean Orogeny in Upper Michigan—Comparisons with Northern Appalachian Tectonics

By William J. Gregg[1]

Abstract

The Precambrian terrane of Michigan's Upper Peninsula consists largely of Archean massifs and mantled gneiss domes surrounded by weakly to highly strained belts of Proterozoic metasedimentary rocks deformed during the Penokean orogeny, about 1.85 Ga. The rocks in these Early Proterozoic belts consist of a thin lower sequence of shelf and shallow-water sediments, overlain by a thick upper sequence of slate, siltstone, and discontinuous graywacke units, known as the upper slate member of the Michigamme Formation. The Baraga belt, in northern Baraga County, typifies the structural setting of the upper slate member. The belt can be divided into a weakly deformed terrane, the Huron River parautochthon, to the north, and a polyphase-deformed terrane, the allochthonous Falls River slice, to the south. The change in structural style and finite strain between the two terranes occurs abruptly along the east-west-trending Falls River thrust, which appears to correlate with a similar, previously recognized boundary in Carlton County, Minnesota.

In the parautochthonous terrane the rocks contain only one style group of folds and associated cleavage. Thrust faults typically deform early first generation structures such as S_1 cleavage, but do not occur in association with B_2 structures. In the allochthonous terrane, these same faults are overprinted by B_2 folds and S_2 crenulation cleavage, suggesting a late B_1 timing for the thrusting in both terranes.

The overall structural setting in northern Michigan, as well as in east-central Minnesota, appears to comprise a series of Early Proterozoic (Penokean) northward-verging folds that have been transported northward along imbricate thrusts. No large-scale overturned fold limbs have been observed in either area; therefore a "thrust nappe" setting appears more appropriate than a "fold nappe."

In northern (Upper) Michigan this structural setting is part of a regional terrane assemblage that is strikingly similar to that of the northern Appalachians. Both regions contain a parautochthonous and allochthonous foreland, followed laterally by a detached basement massif, a tightly appressed metasedimentary synform, and a compressional mantled gneiss dome–synform complex. The two regions differ in that the deformation of both basement and cover rocks is of higher intensity in the Appalachians.

In northern Michigan, curved metamorphic isograds may be related to relative structural position in the gneiss dome–synform terrane, with the highest metamorphic grade being in the deep synformal hinges. The gneiss domes may have been rotated from an initial east-west trend by late Penokean faults produced by north-northeast-directed tectonic indentation by an irregular projection on the north edge of the Wisconsin magmatic arc terranes.

The tectonic interpretations presented in this report suggest that at the beginning of the Proterozoic, the southern complex and possibly part of the northern complex of the Marquette district existed as a surface of low relief, forming a single large sedimentary basin extending over all the structural terranes, rather than forming a series of small basins in tectonic troughs. This single large basin received uniform sedimentation in a subsiding continental-shelf–shallow-deltaic environment facing open waters to the south. The relatively weak imprint of the collisional event in Upper Michigan might be a result of the smaller size of the colliding magmatic arc terranes compared to that in the Appalachians.

INTRODUCTION

The Precambrian terrane of Upper Michigan consists largely of Archean massifs and mantled gneiss domes (fig. 1) surrounded by weakly to highly strained belts of Proterozoic metasediments deformed during the Penokean

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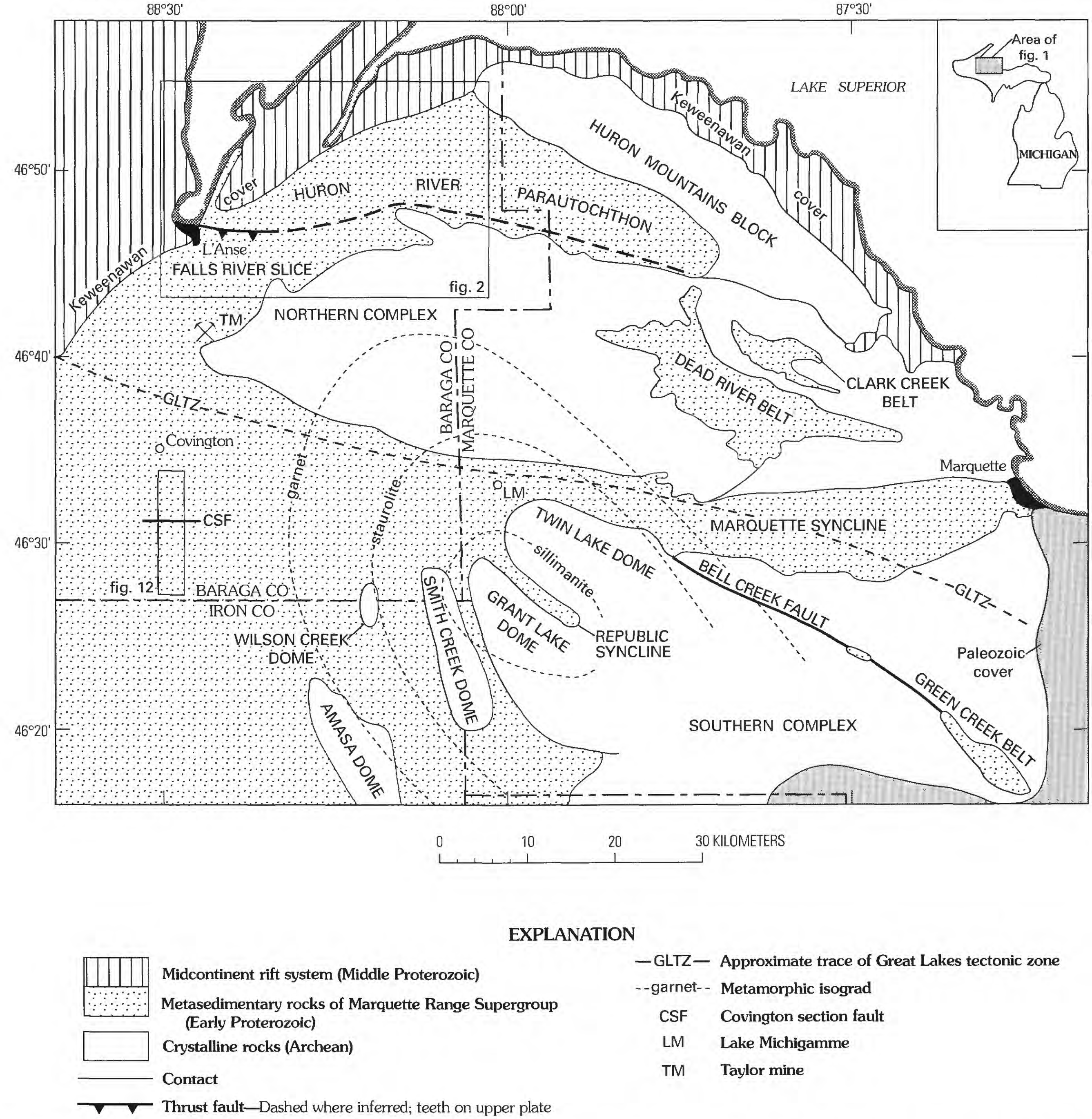

Figure 1. Tectonic setting of the Early Proterozoic of Upper Michigan. Modified from Klasner (1978) and Cannon (1977, 1986).

orogeny, about 1.85 Ga (Cannon and Klasner, 1975; Foose, 1980; Klasner and others, 1982; Cannon, 1986; Sims and others, 1989). The rocks in the Proterozoic belts consist of a thin lower sequence of shelf and shallow-water sedimentary rocks near the Archean contacts, overlain by a thick upper sequence of slate, siltstone, and discontinuous graywacke units (table 1). In the western part of the area, the lower sequence consists of the Goodrich Quartzite and the lower slate member and Bijiki Iron-formation Member of the Michigamme Formation, whereas in the eastern part additional units of the Menominee Group are present below the Goodrich (Cannon and Gair, 1970; Cannon, 1974, 1977, 1986). In both areas the thick sequence of slate, siltstone, and graywacke above these lower units is known as the upper slate member of the Michigamme Formation. The volume of the upper slate member far exceeds that of all

Table 1. Correlation chart of Early Proterozoic rocks in northern Michigan

[Modified from Klasner (1978). Wavy line, unconformity; diagonal line pattern, hiatus]

	Huron River parautochthon			Western Marquette syncline		
Early Proterozoic Marquette Range Supergroup (part)	Baraga Group	Michigamme Formation	upper slate member	Baraga Group	Michigamme Formation	upper slate member
			Bijiki Iron-formation Member			Bijiki Iron-formation Member
			lower slate member			lower slate member
						Clarksburg Volcanics Member
						Greenwood Iron-formation Member
		Goodrich Quartzite			Goodrich Quartzite	
				Menominee Group	Negaunee Iron-formation	
					Siamo Slate	
					Ajibik Quartzite	
Archean gneisses, granites, and metabasites						

the lower units combined (Cannon, 1986); however, the discontinuous nature of the psammitic units within it makes stratigraphic mapping difficult.

Work by the U.S. Geological Survey over the past 50 years has outlined the structural style and orogenic sequence of the Penokean in this region (Cannon, 1973). Few attempts have been made at detailed structural analysis, especially in the voluminous upper slate member, but that which has been done shows a widespread variation in the complexity of overprinting relationships, amount of finite strain, and degree of metamorphism.

The greatest amount of deformation (Klasner, 1978) occurs in the western Marquette syncline near the Twin Lake dome of the southern complex (fig. 1), where the rocks consist of staurolite-bearing phyllites and schists (Cannon and Klasner, 1972; 1976). Klasner (1978) showed that the mesoscopic structures in this area consist of at least four generations of foliation and three tectonic lineations. West of this area, in Baraga County, the degree of deformation decreases and the rocks are garnet-bearing slate, phyllite, and siltstone. Farther to the north, in the Baraga belt, the grade of metamorphism in the upper slate member is below the biotite isograd (James, 1955) and the rocks are much less deformed (Cannon, 1986). In some locations the rocks contain only an incipient slaty cleavage and very low amplitude first-generation folds (Van Roosendaal, 1985). In the more highly deformed areas, particularly near the western Marquette syncline, the scarcity of detailed information on the geometry of Penokean deformation in these rocks has placed severe constraints on tectonic modeling, and has rendered the recognition of Penokean structures in Archean rocks all but impossible.

The work outlined in this report is an extension of Klasner and Cannon's pioneering efforts, with an increased emphasis on microscopic and mesoscopic foliation studies, statistical geometric analysis, and finite strain measurements. Letter designations for style group elements follow Turner and Weiss (1963). The mapping, typically at scales of 1:720 or greater, was initiated in the less deformed rocks of northern Baraga County. This allowed a systematic identification of structural subfabrics and mesoscopic style groups that has subsequently been extended into other, more highly deformed areas to the south and east.

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I thank my former graduate students, Dan Van Roosendaal, Kevin Sikkila, Gary Dyke, and Dave Saja, for the highly detailed and rigorous structural analyses they conducted along several of the river sections in Baraga County. Thoughtful reviews by David B. Rowley, David L. Southwick, and Paul F. Williams were very helpful and greatly appreciated. I especially thank Paul Sims and

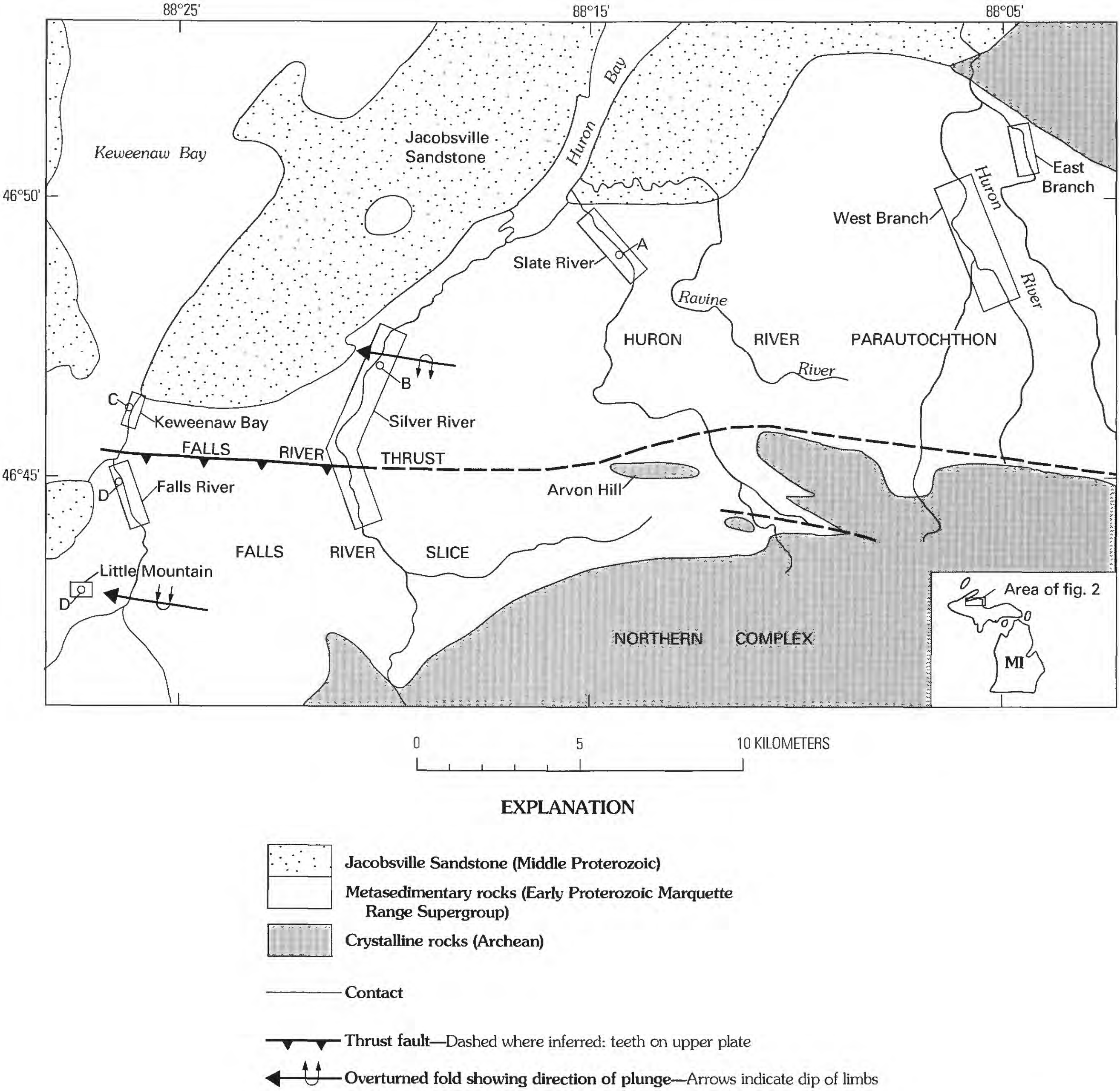


Figure 2. Macroscopic structural setting of the Baraga belt. The Falls River thrust separates the belt into structurally dissimilar parautochthonous and allochthonous terranes. Boxes outline named areas discussed in text; capital letters refer to locations of sections in figure 3. Archean contacts modified from Cannon (1977).

Lorna Carter for their patience and thoroughness while editing this report.

STRUCTURAL SETTING IN NORTHERN BARAGA COUNTY

The main area of study in northern Baraga County consists of the east-northeast-trending belt of Early Proterozoic rocks (fig. 2) previously referred to as the Baraga basin. The area is not a basin in the usual sense, because neither bedding nor cleavage dips inward from the northern and southern boundaries of the belt, and no thickening of sedimentary units toward its geographic center has been recognized. The northwest boundary of the belt is an angular unconformity; the Early Proterozoic rocks are covered by the Keweenawan (Middle Proterozoic) Jacobsville Sandstone. To the northeast and south, the belt is in tectonic contact with the Archean Huron Mountains and northern complex rocks, respectively.

This report documents the foreland fold-thrust model proposed by Klasner, Ojakangas, and others (1988), Klasner, Sims, and others (1988), and Barovich and others (1989) in Upper Michigan, providing evidence for the division of the Baraga belt into the Huron River parautochthon to the north, and the allochthonous Falls River slice to the south, which are separated by the east-west-trending Falls River thrust. Evidence for this thrust comes from abrupt changes in structural complexity and finite strain between the adjacent terranes. The thrust can be identified along two north-trending sections in the belt, one along the Silver River and the other along the Falls River and Keweenaw Bay (fig. 2).

STRUCTURES IN THE HURON RIVER PARAUTOCHTHON

Folds

Rocks of the upper slate member of the Michigamme Formation throughout the Huron River parautochthon belong to only one structural style group. This style group is generally characterized by B_1 folds with S_1 cleavage, accompanied by small synkinematic low-angle thrust faults. The folds have shallow axial plunges to the northwest or southeast, and axial planes that dip 45° to 60° SW. (Van Roosendaal, 1985; Dyke, 1988; Gregg, 1989a).

Throughout the Huron River parautochthon, structures associated with B_1 deformation show an increase in intensity toward the south. In the northeast, along the Huron River (fig. 3), folds are gentle, with amplitudes on the order of 2 m, and occur only in close association with widely spaced synkinematic thrust faults. B_1 fold amplitudes develop progressively to the southwest (fig. 3*A–C*), with tight, overturned folds occurring along the Keweenaw Bay section (fig. 4). Throughout the parautochthon, B_1 folds are upward facing and display northward-trending Z-vergence. Only along the northern Silver River (fig. 2), within the hinge of the Silver Falls anticline, are the folds symmetrical over a large area. S-vergence folds are not generally present within the parautochthon.

Foliations

An incipient to moderately developed S_1 slaty cleavage is symmetrically disposed about the axial surfaces of B_1 folds, forming an L_1 bedding-cleavage intersection lineation parallel to the fold axes. In most rock types S_1 is a domainal slaty cleavage defined by anastomosing, continuous mica films of neomineralized white mica. Quartzofeldspathic microlithons between the S_1 mica films display bedding-parallel fine detrital layer silicates and opaque minerals. In rare examples, where bedding in the slate is defined by fine, competent laminations, S_1 is a weakly developed discrete crenulation cleavage. In the northern part of the parautochthon, S_1 is visible as a weak mica-film foliation in thin section but is not sufficiently well developed to produce a parting surface in outcrop. To the southwest, S_1 increases in intensity, and is readily visible on a mesoscopic scale.

Large chlorite–white mica aggregates (fig. 5) occur commonly within the microlithons in both slate and siltstone. These aggregates, which have (001) traces aligned along bedding in weakly deformed samples, have been shown to originate as prekinematic, possibly authigenic, chlorite porphyroblasts, that become interlayered with white mica when strained (Gregg, 1986). Both the aggregates (fig. 5*C*) and accompanying irregularly shaped opaque mineral grains (fig. 6) have strain shadows parallel to S_1, which increase in length with increasing strain, a feature that distinguishes S_1 from all other foliations in both parautochthonous and allochthonous rocks throughout the entire region. In addition to these strain shadows, long seams of opaque minerals occur along some S_1 mica films, where late displacements parallel to cleavage have taken place. The microstructure of S_1 is compatible with Beutner's (1980) studies, which indicate that evidence is lacking for a soft-sediment, gravity-sliding origin (Klasner, 1978) for any of the early structural features.

Strain Associated with B_1 Deformation

Reduction spots ranging in size from 0.1 mm to 1.0 cm occur throughout the Baraga belt in red silty shale containing fine white silty laminae. Along the East Branch Huron River, where mesoscopic cleavage is poorly developed, the spots are nearly circular in the plane of foliation, with R_{yz} ranging from 1.33 to 1.35 (fig. 7 and table 2). Along the West Branch Huron River the spots are also circular in the xy plane, but R_{yz} ranges from 1.57 to 1.65 (table 2). All these values are below the lower limit of the deformation field for slaty cleavage (fig. 7; Ramsay and Wood, 1973), and occur in rocks with lower fold limb rotation angles (Beutner, 1978) and lower finite strains (Cloos, 1947; Wood, 1973; Ramsay and Wood, 1973) than typically observed in slate belts.

A direct estimate of the finite strain in these weakly deformed rocks was made using fine iron-rich chert laminae in a few rare samples, which have been deformed by telescoping and rotation during cleavage development (fig. 8). Elongation strain was obtained by summing up the lengths of telescoped segments in the xz plane and comparing this initial length to the final length of the telescoped layer. A simple Mohr Circle for Finite Strain construction was employed on data from two beds oriented normal to and at 40° to the xy plane of the strain ellipse to obtain elongation strain values of ε_3=–0.28 and ε_1=+0.04 with R_{xz} of 1.46. These figures compare well with the results from reduction spot samples and suggest a shortening of 25–30 percent for rocks in the northeast end of the parautochthon. Strain is minimal in both the x and the y directions, an observation compatible with the reduction spot results from other samples and the fact that the beds were not telescoped in the xy plane. It seems likely, therefore, that the deformation was dominated at these low strains by volume-loss mechanisms.

Samples from Keweenaw Bay on the southwest end of the parautochthon (fig. 2) show obvious ellipticity in the xy plane, and have R_{yz} ranging from 1.62 to 1.79. These samples lie across the lower limit curve (fig. 7) and display well-developed parting surfaces in outcrop. The results are indicative of the higher strains associated with the increase in B_1-fold amplitudes and tightening of fold profiles to the west. A minor rotation of the long axis of the finite strain ellipse occurs from east to west in the parautochthon, and is associated with a decrease in the angle of dip of B_1-fold axial planes (fig. 3).

Faults

Numerous small thrust faults occur throughout the Huron River parautochthon (Van Roosendaal, 1985). Along the East Branch Huron River, where B_1 folding is weak, faults occur parallel to both bedding and S_1 cleavage. Some of the larger faults form a series of macroscopic-scale, postkinematic structural domains, marked by changes in the trend of L_1 and in the dip of B_1-fold axial surfaces and slaty cleavage (fig. 9). The magnitude of displacement along these thrusts is unknown, but the rotations of L_1 within the thrust slices are pronounced. Minor thrust faults within the domains are generally parallel to bedding or cleavage and in a few places are accompanied by flattening of B_1 axial surfaces to about 25°.

Pitchblende and secondary uranium minerals (metatyuyamunite and volborthite) occur in the thrusts as small discontinuous stringers and pods surrounded by brecciated quartz and calcite (Johnson, 1977). Along the West Branch Huron River, several old silver prospect pits are aligned along these faults.

In the central part of the autochthon, along the Silver River, faults are exclusively parallel or subparallel to bedding and have relatively little displacement (Dyke, 1988). The faults generally strike west-northwest as in other areas, and display white quartz veins with slickenlines plunging to the south-southwest. In rare cases, black slate bodies adjacent to these faults show a weak crenulation cleavage that overprints S_1; however, no mesoscopic B_2 folds are present.

Along Keweenaw Bay (fig. 4), at the southwest end of the parautochthon, where B_1 folds are very tight, faults are relatively uncommon. Large-scale rotations of L_1, such as those associated with faulting along the Huron River, have not been noted, and the displacements along most faults appear to be small.

The faults present in the Huron River parautochthon typically deform early first generation structures such as S_1 cleavage, but do not occur in association with widespread B_2 structures. In the allochthonous Falls River slice, however, faults identical in mineralogy and morphology to those in the parautochthon are clearly overprinted by numerous B_2 macrokinks and S_2 crenulation cleavage. These observations suggest a post-B_1 timing for the faults in the parautochthon as well as the allochthon.

STRUCTURES IN THE FALLS RIVER SLICE

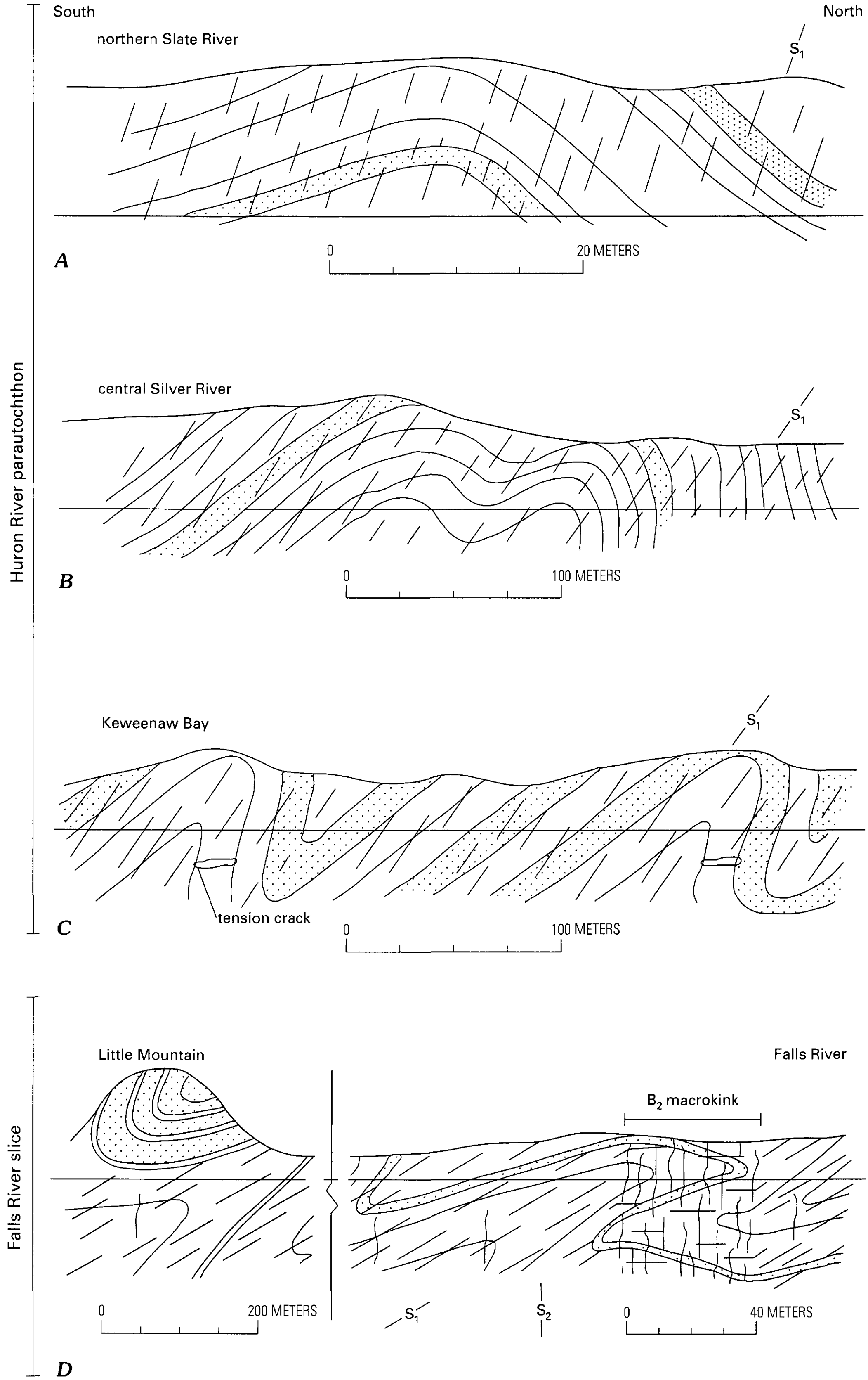
The deformational history of the slate and metagraywacke in the Michigamme Formation south of the Falls River thrust is remarkably different from that in the parautochthonous rocks to the north. Differences in the southern terrane include rotation of early first deformation structures to nearly recumbent positions, an increase in the finite strain associated with S_1, and the overprinting of first generation structures by second generation folds on all scales.

At the northern front of the allochthon, the Falls River thrust has been placed at the boundary between the contrasting structural terranes (Gregg, 1989a). In the western part of the Baraga belt, this boundary occurs between the Keweenaw Bay section and the Falls River (fig. 4) as well as along the Silver River (fig. 2), at a silver prospect about 100 m north of Dynamite Hill Road.

Folds

B_1 folds in the Falls River slice are tight to isoclinal, strongly asymmetric folds having a northward vergence (figs. 3*D* and 4; Sikkila, 1987; Sikkila and Gregg, 1987). The folds are generally overturned and commonly recumbent; axial surfaces and S_1 cleavage typically dip gently southward except where affected by second generation

Figure 3 (facing page). Typical B_1 fold profiles in Huron River parautochthon and allochthonous Falls River slice. Folds show increasing tightness toward Falls River thrust (*A* to *C*) and abrupt change in structural styles in allochthon (*D*). Locations of sections shown in figures 2 and 4. *D,* composite section of several localities on map in figure 4. Straight dashed line, foliation.



South
North
northern Slate River
S_1
A
0
20 METERS
Huron River parautochthon
central Silver River
S_1
B
0
100 METERS
Keweenaw Bay
S_1
tension crack
C
0
100 METERS
Falls River slice
Little Mountain
Falls River
B_2 macrokink
0
200 METERS
S_1
S_2
0
40 METERS
D

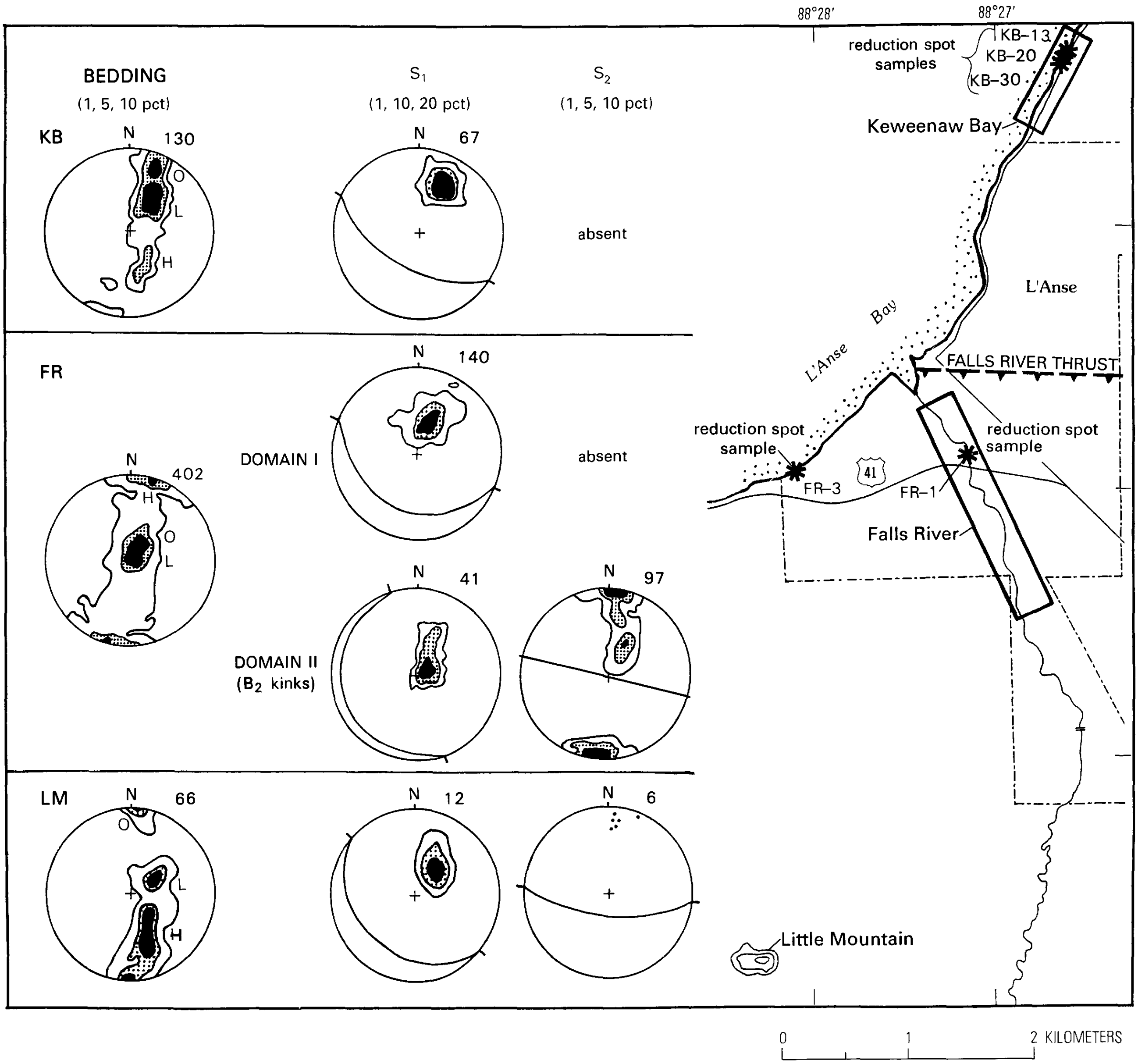


Figure 4. Equal-area hemispherical plots showing mesoscopic structures from Keweenaw Bay to the Falls River and Little Mountain. In stereonet plots: H, hinge region; L, long limb of fold; O, overturned limb; number, number of measurements. Both Falls River (FR) and Little Mountain (LM) are in the Falls River slice.

folding. First order B_1 folds are quite large, with long limb heights (Fleuty, 1964) ranging up to 150 m. In a few places short limb heights exceed 45 m, with axial-plane separations on the order of 15 m. Second order B_1 folds occur as parasitic folds on larger B_1 structures, and show roughly the same limb proportions on a hand specimen scale. The amplitudes of first order B_1 folds in the allochthon are commonly an order of magnitude greater than those in the parautochthonous rocks to the north.

B_1 folds have been overprinted on all scales by macroscopic B_2 kink folds (Sikkila, 1987), producing a variation in the orientation of B_1 structural elements within the kink band boundaries. Outside these boundaries, B_2 deformation is barely perceptible, and B_1 folds have axial planes that dip gently south, with horizontal to gently plunging axes trending west-northwest. Within a typical macroscopic kink zone (fig. 3*D*), B_1 folds are rotated so that axial planes and accompanying S_1 cleavage (fig. 4) are commonly recumbent or even gently north dipping. B_1 fold axes and associated L_1 bedding-cleavage intersection lineations may plunge as much as 20° within B_2 kink zones, and in rare cases they are doubly plunging. As in the parautochthonous rocks to the north, B_1 deformation is characterized as a system of constant-vergence asymmetric

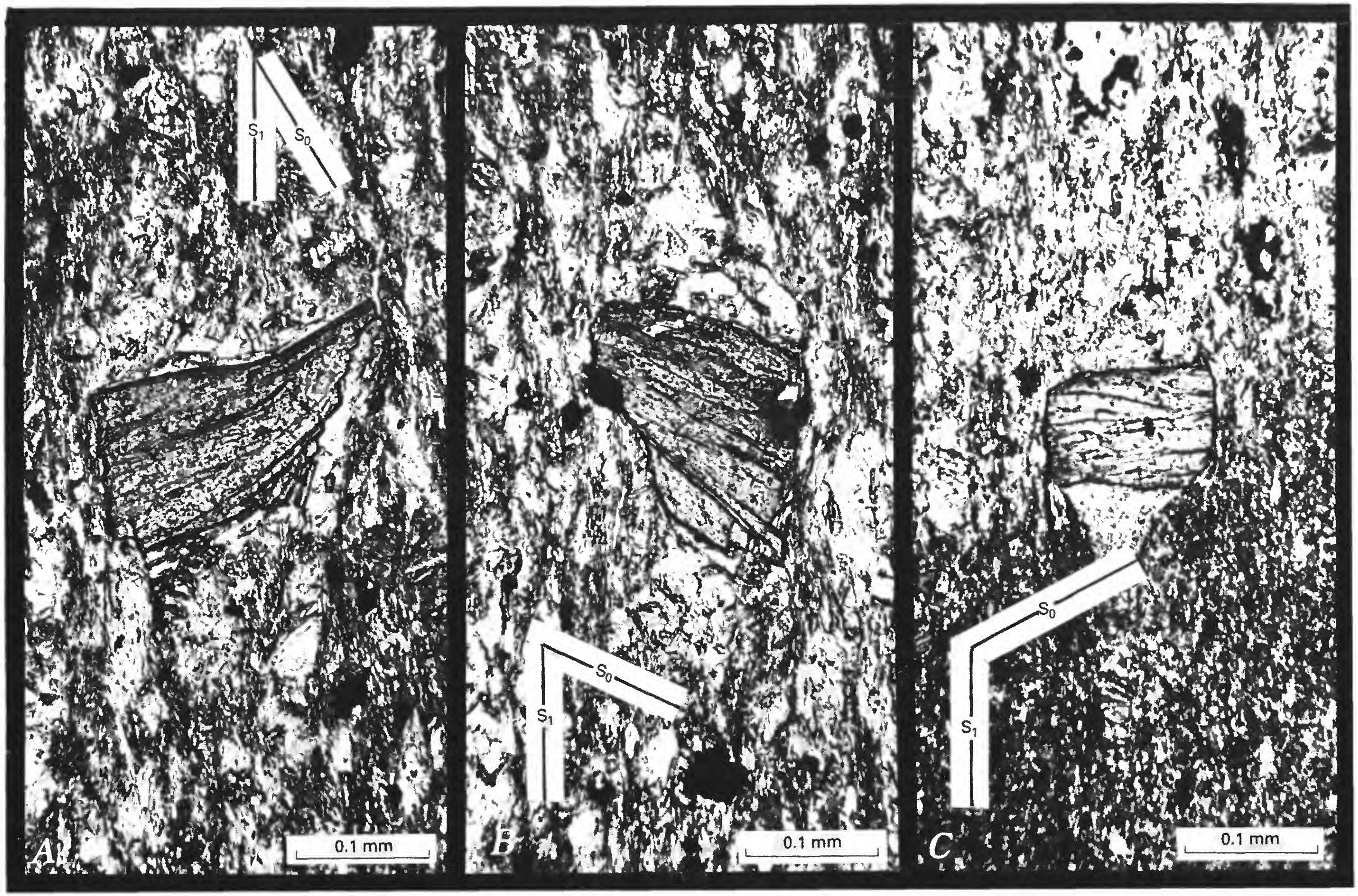


Figure 5. Photomicrographs of chlorite–white mica aggregates in slate of the Michigamme Formation of the Huron River parautochthon. *A*, *B*, Grains from the Silver River, showing rotation of (001) out of original bedding-parallel orientation (Gregg, 1986). *C*, Grain from the Huron River, showing development of strain shadows (dark areas) parallel to S_1 cleavage.

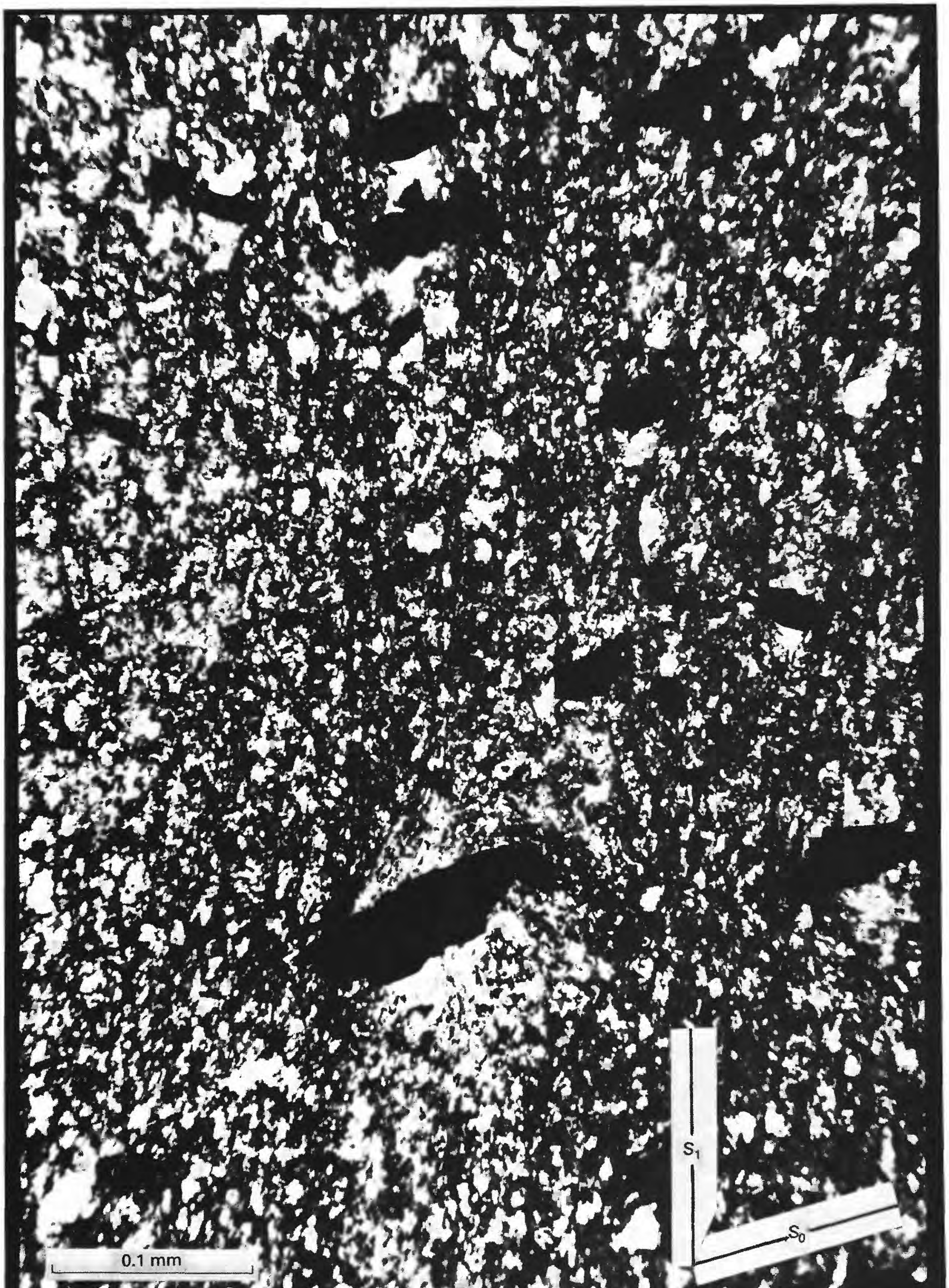


Figure 6. Photomicrograph showing microstructure of large opaque minerals in slate of the Michigamme Formation of the Huron River parautochthon. Opaques typically have long axes parallel to bedding and show strain shadows parallel to S_1. Specimen is from the Silver River (fig. 2).

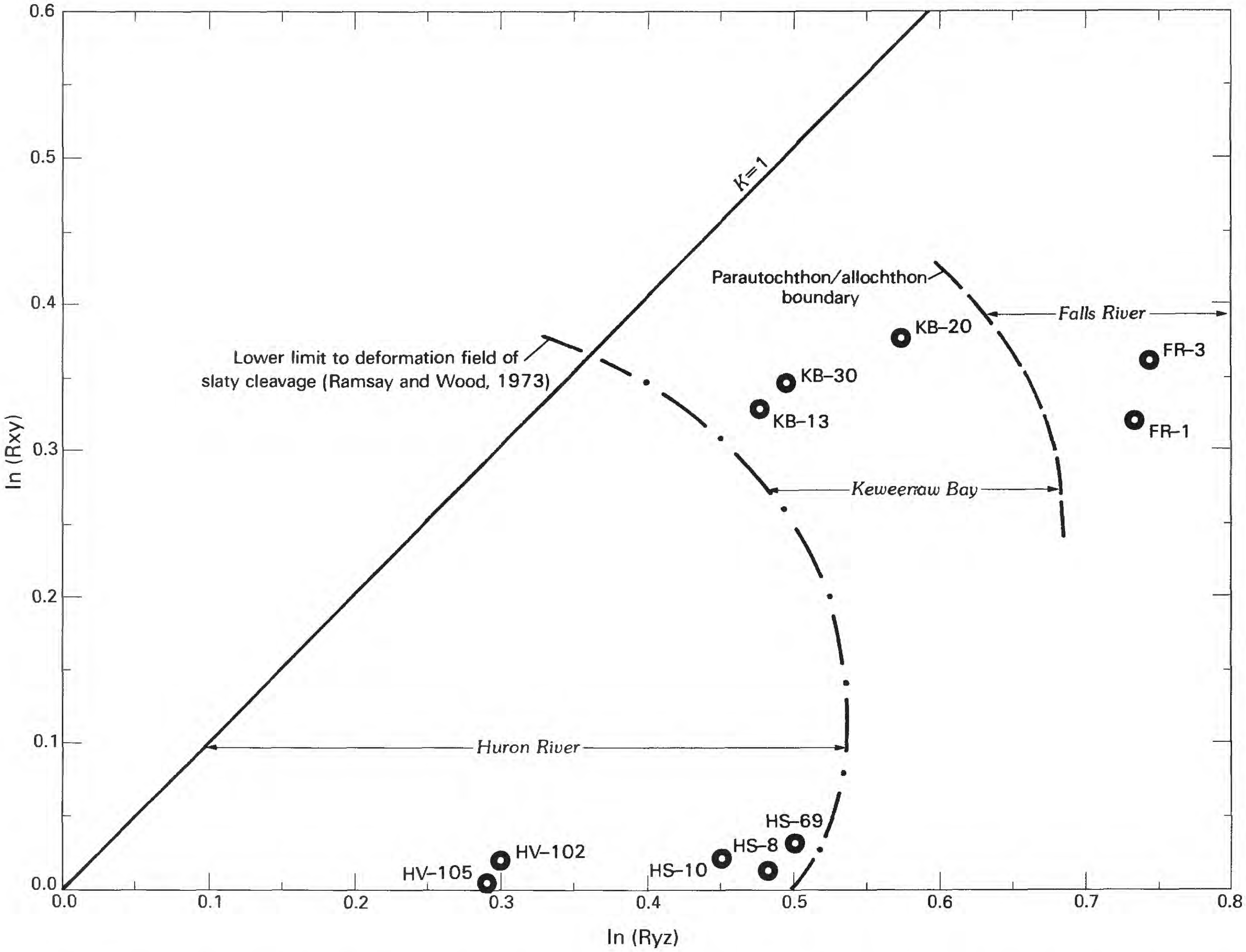


Figure 7. Modified Flinn diagram of reduction spot measurements from the upper slate member of the Michigamme Formation of the Baraga belt. Location of samples shown in figure 4. Samples from the parautochthon include: KB, Keweenaw Bay; HV, East Branch Huron River; HS, West Branch Huron River. Allochthonous rocks are represented by the Falls River samples (FR). R_{xy} and R_{yz} are the ratios of the axes in the xy and yz planes of the finite strain ellipsoid (table 2; Flinn, 1962; Ramsay, 1967, p. 138).

folds and associated thrusts, all of which were upward facing prior to B_2 deformation. There is no indication of the lower, downward-facing regional fold limb that is commonly associated with fold nappes.

B_2 folds are typically asymmetric flexures that deform S_1 cleavage, S_1-parallel thrust faults, and bedding. They are accompanied by an S_2 axial-planar crenulation cleavage (fig. 4) and an L_2 wrinkle lineation, both of which are readily distinguished from first generation structures. Along the Falls River, the intensity of B_2 folding increases systematically from open, gentle flexures in the north, to tight, short wavelength structures in the south. B_2 axial surfaces, which commonly appear as large-scale kink-band boundaries, are vertical and strike west-northwest in the north, whereas in the south they dip as low as 28° S., and strike about N. 39° W. This rotation of B_2 axial planes and S_2 cleavage is accompanied by a rare S_3 cleavage, which strikes N. 85° E. and dips 70° S. Mesoscopic B_3 folds have not been observed.

The size of macroscopic B_2 kink domains is on the order of 30–50 m. Within these domains, which are essentially defined by large-scale kink-band boundaries, smaller kink domains less than 1 m wide are present. On both scales it is typical for the L_2 wrinkle lineation to increase in amplitude as the boundary of the kink domain is approached from either direction; thus, the lineation and S_2 cleavage may actually diminish somewhat inside the rotated kink zone. Outside the kink zone, well away from the boundary, these structures are generally absent.

Table 2. Reduction spot data for slate of the Michigamme Formation

[(*n*), number of samples; *, not available. Sample designations: FR, Falls River; KB, Keweenaw Bay; HV, East Branch Huron River (Van Roosendaal, 1985); HS, West Branch Huron River (Saja, 1991)]

Sample No.	R_{xy} (*n*)	R_{yz} (*n*)	$S_0 \times S_1$
FR–1	1.38±0.07 (82)	2.10±0.11 (117)	17
FR–3	1.44±0.05 (33)	2.11±0.09 (45)	21
KB–13	1.39±0.06 (38)	1.62±0.07 (29)	32
KB–20	1.46±0.04 (11)	1.79±0.09 (13)	14
KB–30	1.42±0.06 (14)	1.65±0.05 (10)	68
HV–102	1.02±0.03 (23)	1.35±0.09 (52)	37
HV–105	1.00 * *	1.33±0.08 (22)	*
HS–8	1.01±0.04 (73)	1.61±0.12 (70)	44
HS–10b	1.02±0.04 (178)	1.57±0.10 (100)	44
HS–69	1.03±0.04 (67)	1.65±0.14 (33)	72

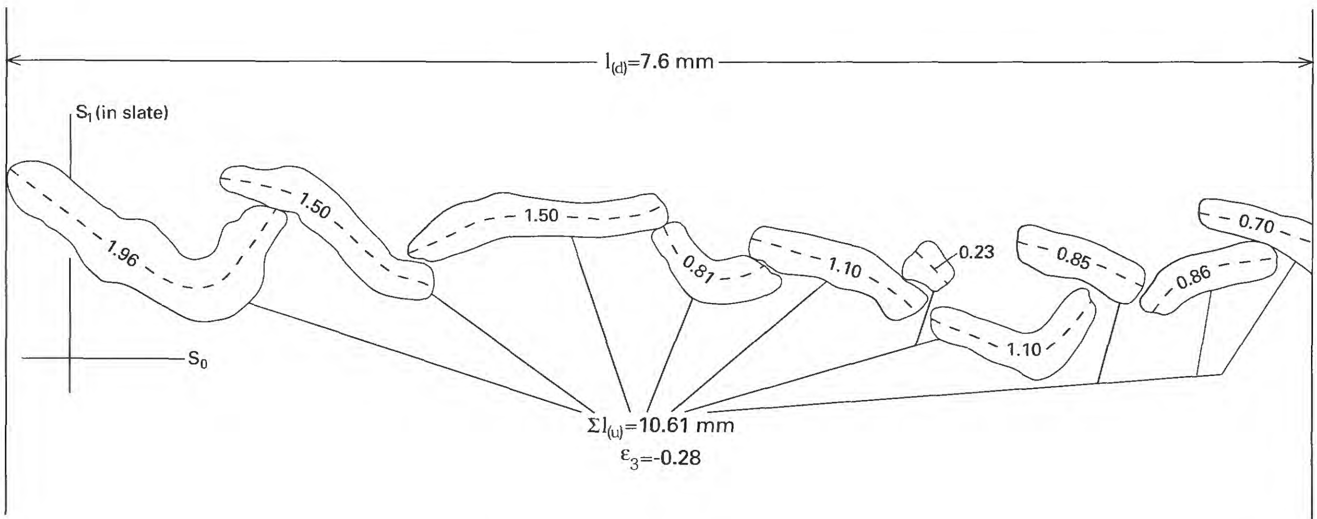


Figure 8. Telescoped iron-rich cherty bed in slate of Michigamme Formation. Finite strain in the **xz** plane was estimated by comparing length of summed segments ($\Sigma l_{(u)}$) to length of telescoped span ($l_{(d)}$) as shown. ε_3 is elongation strain value. Specimen is from the East Branch Huron River (fig. 2).

Foliations

S_1 varies in mesoscopic appearance from a well-developed penetrative slaty cleavage in pelite to an irregular domainal rough cleavage in graywacke (Sikkila, 1987). In the limbs of isoclinal B_1 folds, it is typically impossible to distinguish S_1 from S_0 without a thin section. On a microscopic scale, S_1 is defined by continuous, anastomosing mica films of oriented chlorite and white mica separated by quartzo-feldspathic domains. The width and degree of irregularity of the quartzo-feldspathic domains generally increase with sedimentary grain size. In thin sections of finer grained rocks, bedding can be readily distinguished within quartzo-feldspathic domains.

S_1 in the Falls River slice can be conclusively correlated with S_1 in the Huron River parautochthon not only

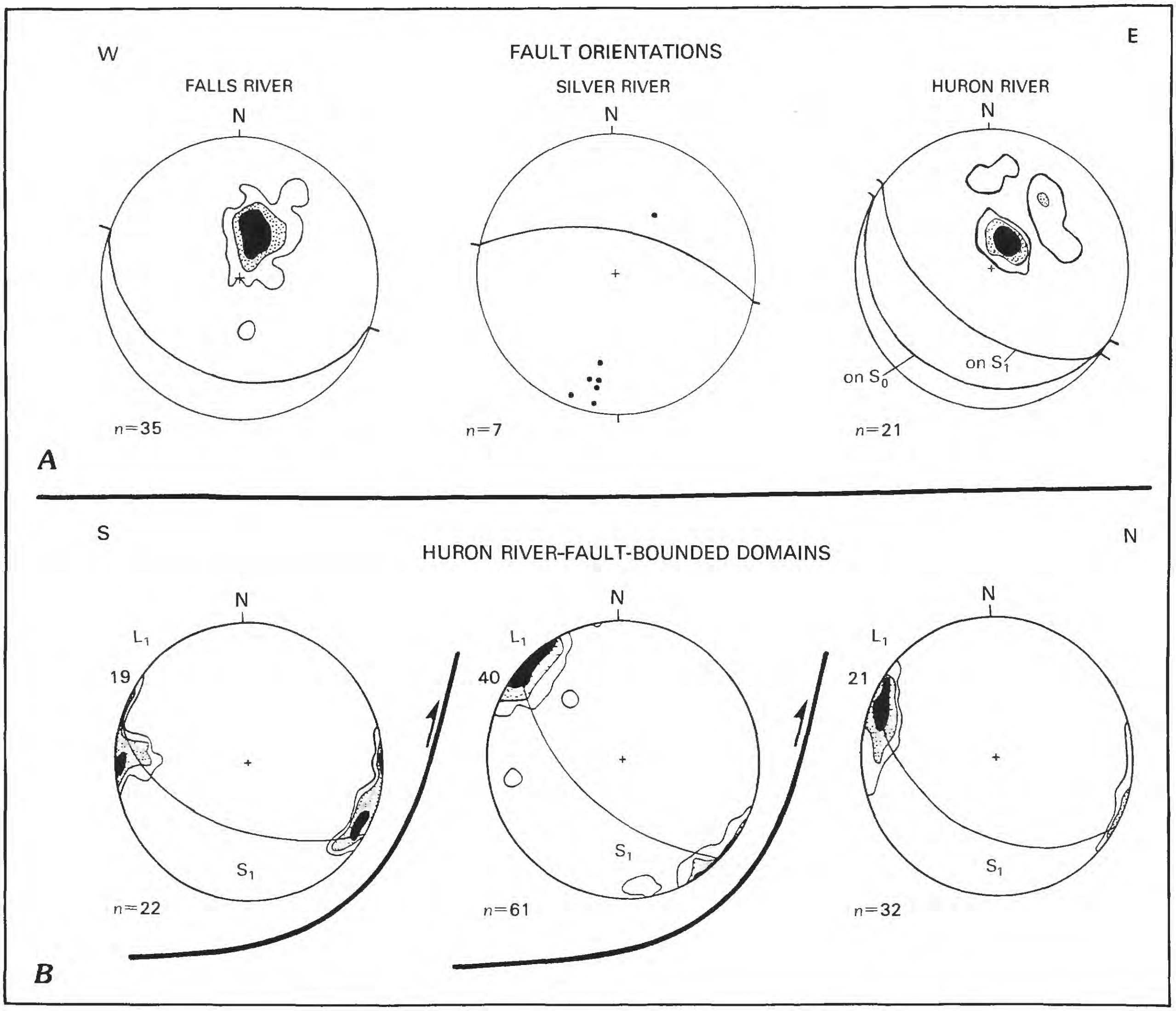


Figure 9. Equal-area hemispherical plots of faults and related features in the Baraga belt. *A*, Poles to fault planes in allochthonous rocks (Falls River) and parautochthonous rocks (Silver River and Huron River) in the Michigamme Formation. *B*, Rotation of L_1 bedding-cleavage intersection lineations across fault boundaries (diagrammatically shown in cross section) on the East Branch Huron River (fig. 2). Numbers, number of measurements; contours are 1 point, 10 percent, and 20 percent of 1 percent area for all plots.

because of its morphologic similarity, but also because of its relationship with the ubiquitous predeformational chlorite–white mica aggregates. In the allochthon, these aggregates display obvious prekinematic textures relative to S_1, including deformation of the aggregates and development of long strain shadows parallel to slaty cleavage (fig. 10*A*). As in the parautochthonous rocks, the strain shadows show an increase in length parallel to S_1 with increasing strain. S_1-parallel strain shadows are also present around large opaque mineral grains (fig. 10*A*, *B*), as in some of the parautochthonous rocks. In both cases the strain shadows are deformed by S_2 crenulation cleavage (fig. 10*B*).

S_2 typically consists of a zonal to discrete crenulation cleavage that overprints both S_1 and bedding (fig. 10). Where macroscopic B_2 folds are gentle, this cleavage is well developed only within the hinge regions of the folds. Where B_2 folds are tight and numerous, S_2 is pervasive and predominates in mesoscopic and microscopic appearance over other foliations.

S_2 morphology also varies with rock type. In pelitic rocks near gentle B_2 flexures, S_2 may be accompanied by thin, widely spaced mica films in the short limbs of B_2 microcrenulations. These films may contain thin seams of opaque minerals in rocks that initially contained abundant disseminated opaques. Where B_2 deformation is intense,

Figure 10. Photomicrographs of microstructures in slate of the Michigamme Formation. *A*, Typical chlorite–white mica aggregates and large opaque porphyroblasts in allochthonous rocks along the Falls River. Strain shadows occur parallel to S_1 slaty cleavage as in the parautochthonous rocks to the north. *B*, Deformation of strain shadows and porphyroblasts by B_2 crenulations along the Falls River.

S_2 is more tightly spaced, the angle of the short limb is higher, and the width of the domain is much smaller. In the pelites, this is commonly accompanied by a change in morphology to a more anastomosing appearance. In siltstones, S_2 is more strongly defined, since most of the chlorite in the rock lies along the microfold limbs, which have been subjected to solution-transfer removal of quartz. In graywackes, S_2 is poorly defined or absent.

S_3 foliation is present only in the south area of the Falls River section (fig. 4), near a large Keweenawan diabase dike, where B_2 folds and S_2 cleavage have been subsequently deformed. In this area, originally east west striking, nearly vertical S_2 crenulation cleavage has been rotated more than 70° about an east-northeast-trending horizontal axis. In the field, a steeply dipping S_3 cleavage occurs as a set of closely spaced planes, 2 to 4 cm apart, which overprints all earlier foliations. On a microscopic scale, S_3 is characterized by a series of simple, discontinuous fractures filled with dusty opaque material. No reorientation or recrystallization of layer silicates is associated with this foliation. This foliation has not been observed in any other area, and mesoscopic B_3 folds have not been found.

Strain Associated with B_1 Deformation

Reduction spots are relatively rare in the Falls River slice because of the scarcity of the characteristic red-banded pelites and psammites that contain them. Those spots that were measured are from samples that contained no S_2 cleavage and were not located within B_2 macrokink boundaries; however, some overprint of strain associated with B_2 deformation cannot be ruled out. As shown by Holst (1985), the superimposed B_2 strain is in a direction that would tend to decrease the ellipticity of the reduction spots, since the long axis of the ellipsoid lies in the plane of the nearly recumbent slaty cleavage and plunges southwest. The data can thus be viewed as a minimum strain associated with B_1 deformation in the Falls River slice. The measurements obtained from typical red silty slate (table 2) show R_{xy} ranging from 1.38 to 1.44 and R_{yz} of about 2.10 (fig. 7).

The wide distribution of reduction spots allows a comparison of finite strain throughout the parautochthon and across the Falls River thrust as well (fig. 11). Along the Keweenaw Bay and Falls River sections (fig. 4), the change in finite strain does not seem particularly abrupt, being roughly comparable to those observed on local faults in the Cambrian slate belt of Wales (Wood and Oertel, 1980). However, the change in finite strain is accompanied by a large rotation of the ellipse axis due to overprinting by B_2 structure. The strain ellipse along Keweenaw Bay has a relatively constant orientation, plunging about 60° SW., whereas the strain ellipse in the Falls River slice plunges from horizontal to gently northward in B_2 kink hinges to gently southward outside kink zones. This rotation amounts to a large post-B_1 strain that cannot be evaluated at the mesoscopic scale.

The reduction spot measurements show an increase in ellipticity from northeast to southwest in the parautochthon, toward the Falls River thrust. It is tempting to draw a smooth curve through the data (fig. 11), suggesting a relatively uniform strain distribution in the parautochthon. More detailed studies in progress, however, suggest that this may not be the case (Saja, 1991).

Faults

A system of S_1-parallel faults from a few centimeters to 1 m thick is present in the allochthon, typically appearing near large B_1 fold hinges. Most of the faults contain white quartz and slate fragments, but some contain abundant cataclastic slate fragments surrounded by carbonate. The vein material, which is also present in joints located near the faults, ranges in thickness from a few centimeters to 1 m. The mineralogy of minor faults is consistent with that of the mylonites from the Falls River thrust (fig. 12).

The rotation of cleavage in the slate fragments within these smaller thrusts shows them to be post-B_1 structures, as is the case with similarly oriented and mineralized faults in the parautochthonous rocks to the north. This suggests an association between the small thrusts and the Falls River thrust, which shows a much larger rotation of S_1 cleavage and is therefore also a post-B_1 structure. This timing relationship suggests that the Falls River thrust may have more or less the same orientation as the smaller structures, that is, it strikes roughly east-west and dips about 30°–40° S., flattening to the south.

The timing of the thrusting relative to the B_2 fold event is apparent from both field and macroscopic observations.

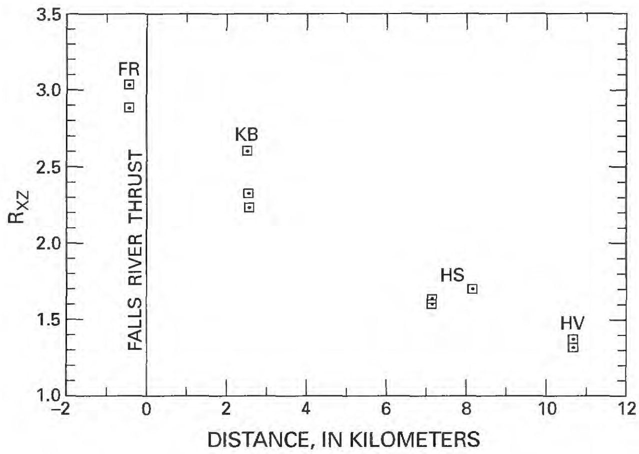


Figure 11. Strain distribution shown by reduction spots in slate of the Michigamme Formation. Sample localities shown in figure 4, except Huron River, general location shown in figure 2. Samples from the Huron River parautochthon and Falls River slice show an increase in finite strain toward the Falls River thrust. FR, Falls River; KB, Keweenaw Bay; HS, West Branch Huron River; HV, East Branch Huron River. Data calculated from values in table 2.

It is obvious that B_2 folding did not occur before the thrusting, because the steeply inclined B_2-fold axial surfaces are in no place cut by faults. On the other hand, in areas where B_2 macrokinks occur, it is the faults which are overturned and have severely convoluted and crenulated surfaces. The most reasonable suggestion is that the faulting is more or less synchronous with B_2 macrokink development, and that the folds represent the plastic deformation associated with ramping during the overthrusting event. This fits the macroscopic picture well, because only the transported rocks south of the Falls River thrust contain B_2 structures. If B_2 folding had occurred after the overthrusting, one would expect the B_2 macrokinks to overprint the thrust boundary and to be found in the parautochthonous rocks as well.

IMPLICATIONS FOR PENOKEAN TECTONICS IN MICHIGAN AND MINNESOTA

Nappe Structures

Detailed structural analysis (Sikkila, 1987; Dyke, 1988; Gregg, 1989a) has shown that the rocks of northern Baraga County can be divided into a northern weakly deformed parautochthonous structural belt and a southern polyphase-deformed allochthonous belt, separated by the east-west-striking, southward-dipping Falls River thrust

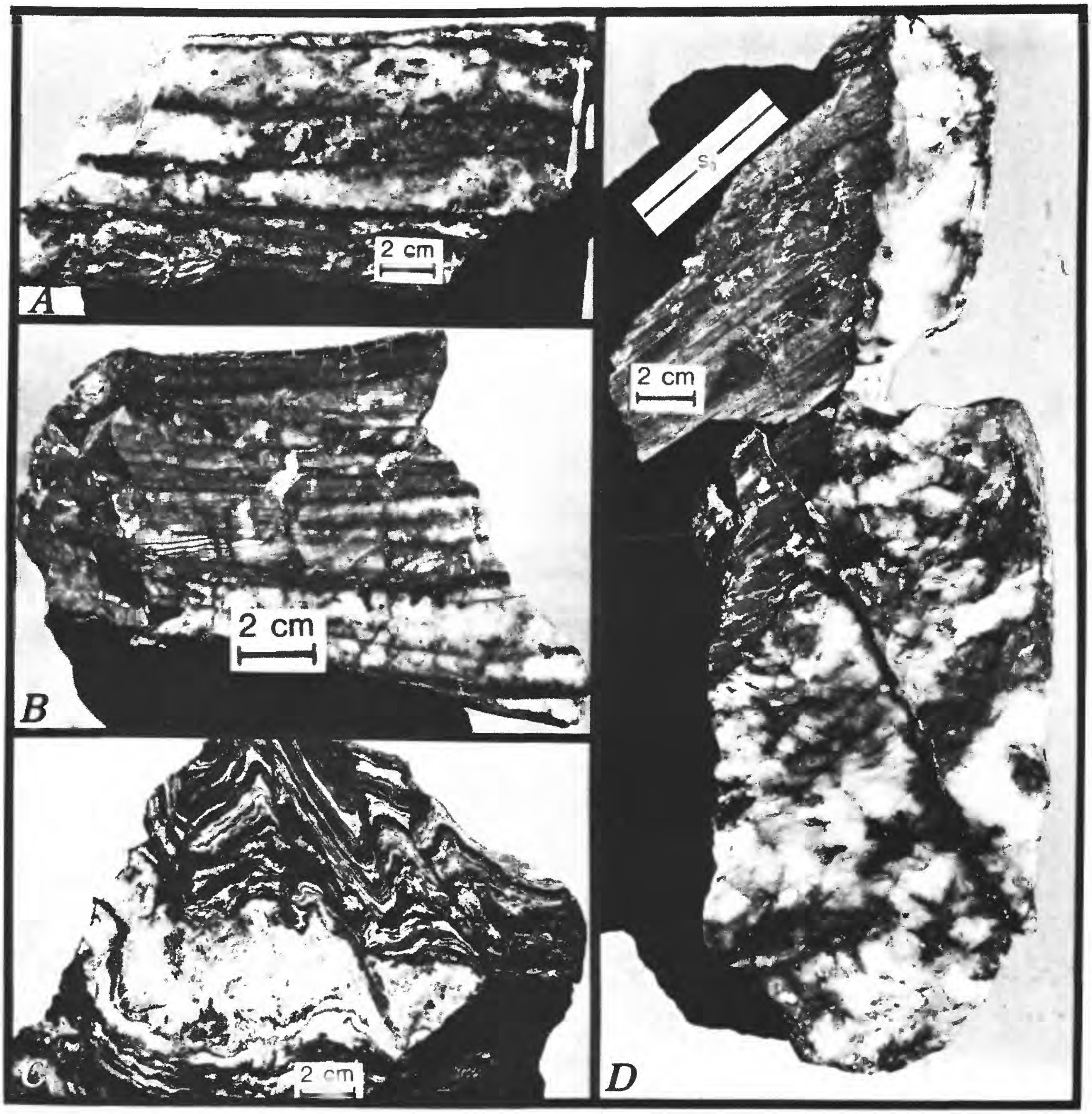


Figure 12. Photographs showing mylonites from Penokean thrust faults in Upper Michigan. *A*, Banded quartz mylonite with graphitic slate layers from the Falls River thrust at the Silver River locality (fig. 2). *B*, Banded and crossbanded quartz mylonite from one of several minor thrusts along the West Branch Huron River (fig. 2). *C*, Contorted quartz mylonite with slate bands from a minor thrust along the Covington section (fig. 1). *D*, Quartz mylonite with cherty iron-formation fragments, cutting bedding in cherty iron-formation. Sample is from the Taylor mine area (fig. 1) and occurs along a large thrust, shown as a saddle reef in Klasner (1978, fig. 3).

(fig. 2). The structural setting is characterized by a series of northward-verging folds that have been transported northward along a set of imbricate thrusts, possibly of "sledrunner" morphology.

The general term "nappe" is widely used to describe tectonic settings such as those observed in the Michigan and Minnesota terranes, and is synonymous with "overthrust" (Rodgers, 1990) or "thrust sheet" (Ramsay and Huber, 1987, p. 521). The term "fold nappe" (Ramsay, 1967) refers specifically to the classic nappe model, consisting of a regional-scale recumbent fold with a downward-facing hinge, an inverted lower limb, and an upward-facing "root" zone (Tobisch and Glover, 1971; Ager, 1980, fig. 14.5). The term "thrust nappe" (Ramsay, 1967) is used to describe a large recumbent fold nappe partially or completely cut by a lower thrust (Ramsay and Huber, 1983, fig. 11.10; Park, 1983, fig. 12.3); a series of imbricate thrusts in nearly undisturbed rocks (imbricate fans or schuppen-structure); or even a series of folds with relatively constant fold vergence (Hobbs and others, 1976, p. 411).

Thrust nappes appear to be typical of many foreland regions, where fold vergence is uniformly towards the stable craton, and where no evidence for a lower recumbent fold limb exists. Fold nappes, on the other hand, seem to be associated with highly strained (Ramsay and Huber, 1983) rocks within or adjacent to metamorphic core zones or tectonic root zones. The foreland terranes in Michigan and Minnesota are best described as thrust nappes because there appears to be no direct evidence for a large-scale lower limb in either area.

Although the location of the Falls River thrust is relatively well defined, the location of the southern boundary of the Falls River slice is as yet unknown. This boundary may occur against lower parautochthonous or allochthonous rocks, in which case the slice would be a klippe (Zen, 1972, p. 71), against an overlying allochthonous slice, or against the Great Lakes tectonic zone (Sims and others, 1980). The possibility of a parautochthonous terrane to the south of the Falls River slice seems unlikely, but its existence could be demonstrated in the field by the absence of pervasive B_2 structures, as is the case in the Huron Bay parautochthon. Klasner (1978) reported B_2 structures at the Taylor mine area and in the Lake Michigamme area along the western Marquette syncline (fig. 1). Holst (1989a), on the other hand, reported a boundary along the Baraga–Iron County line, 15 km south of Covington (fig. 1), which separates a southern terrane deformed by B_2 folds from a northern terrane lacking second generation structures. He suggested that the boundary, referred to as a "nappe front," correlated with a similar boundary that he delineated in Carlton County, Minn. (Holst, 1984b).

Correlation of Thrust Boundaries

Numerous reports, including this one, have shown the existence of B_2 structures in rocks north of Holst's boundary. Furthermore, recent mapping by me along the Covington transect shows evidence for large-scale post-B_1 rotations of early folds along a fault located approximately 7 km south of Covington (figs. 1, 13), and well within Holst's northern terrane. At this location an abrupt change in the orientation of S_1 cleavage and the L_1 intersection lineation occurs, with B_1 fold axes showing a more northerly plunge on the south side of the fault. This location corresponds closely with the boundary delineated by Barovich and others (1989) on the basis of Nd isotope studies. The relative scarcity of S_2 along this section, as reported by Holst (1989a), may be caused by the dominance of psammitic rocks, which do not readily form crenulation cleavages, rather than the absence of a B_2 deformational event.

Because of the strong similarity in structural setting, I suggest that Holst's boundary in Carlton County, Minn., correlates with the Falls River thrust. The terrane south of the boundary in both areas contains first generation folds that are in places isoclinal and recumbent, and which contain an axial-planar slaty cleavage. These folds are overprinted in both areas by open, upright second generation folds and a steeply dipping crenulation cleavage, neither of which occur in the northern terranes.

In addition to the strong structural similarity between the Baraga and Carlton County areas, evidence exists for a similar sedimentological contrast across the boundaries in both areas. In Minnesota the amount of graphitic sedimentary rocks increases sharply to the south across the fault (Holst, 1984b). In Baraga County, the rocks in the northern area are weakly graphitic, whereas those south of the boundary are very graphitic, and economic graphite deposits are known to exist (Johnson, 1977).

Correlation of Mesoscopic Structures

In Minnesota, Holst (1981, 1982, 1989b) correlated the second generation folds that occur in the southern Minnesota terrane with the first generation folds to the north on the basis of similar cleavage orientation. He theorized that the northern terrane contained no folds or cleavage at the time the isoclinal F_1 folds and slaty cleavage developed in the southern terrane. He then postulated that F_1 folds were rotated and overprinted by open F_2 folds and crenulation cleavage during nappe emplacement in the southern terrane, with synchronous development of tight folds and slaty cleavage in the footwall rocks of the northern terrane. Using this model, Holst assumed that the northern rocks recorded only the strains due to F_2 deformation. He then used strain measurements made in the northern area in a subtractive manner, to calculate supposed finite strain due to F_1 deformation south of the boundary. This calculation led him to propose total strains that are uncommonly high for a typical fold and thrust belt setting (Ramsay and Wood, 1973), and which demand a peculiar subhorizontal orientation of the principal plane of shortening (λ_1/λ_2)

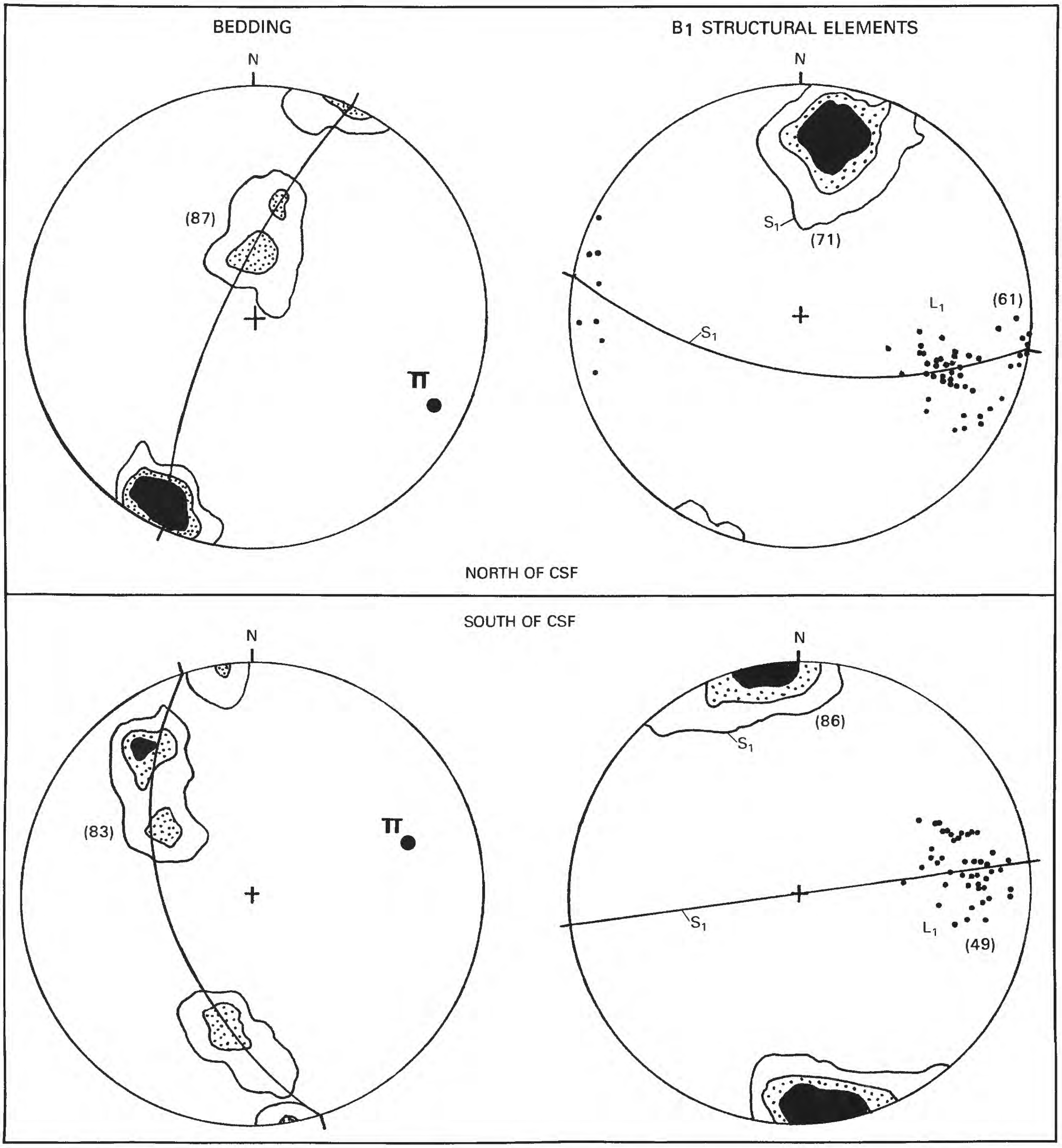


Figure 13. Equal-area hemispherical plots showing rotation of B_1 structural elements across the fault within the Covington section, Michigan. S_1, slaty cleavage; L_1, bedding-cleavage intersection lineation; CSF, Covington section fault (fig. 1). Contours are 1 point, 5 percent and 10 percent per 1 percent area for all plots. π, pole axis S_0. Numbers, number of measurements.

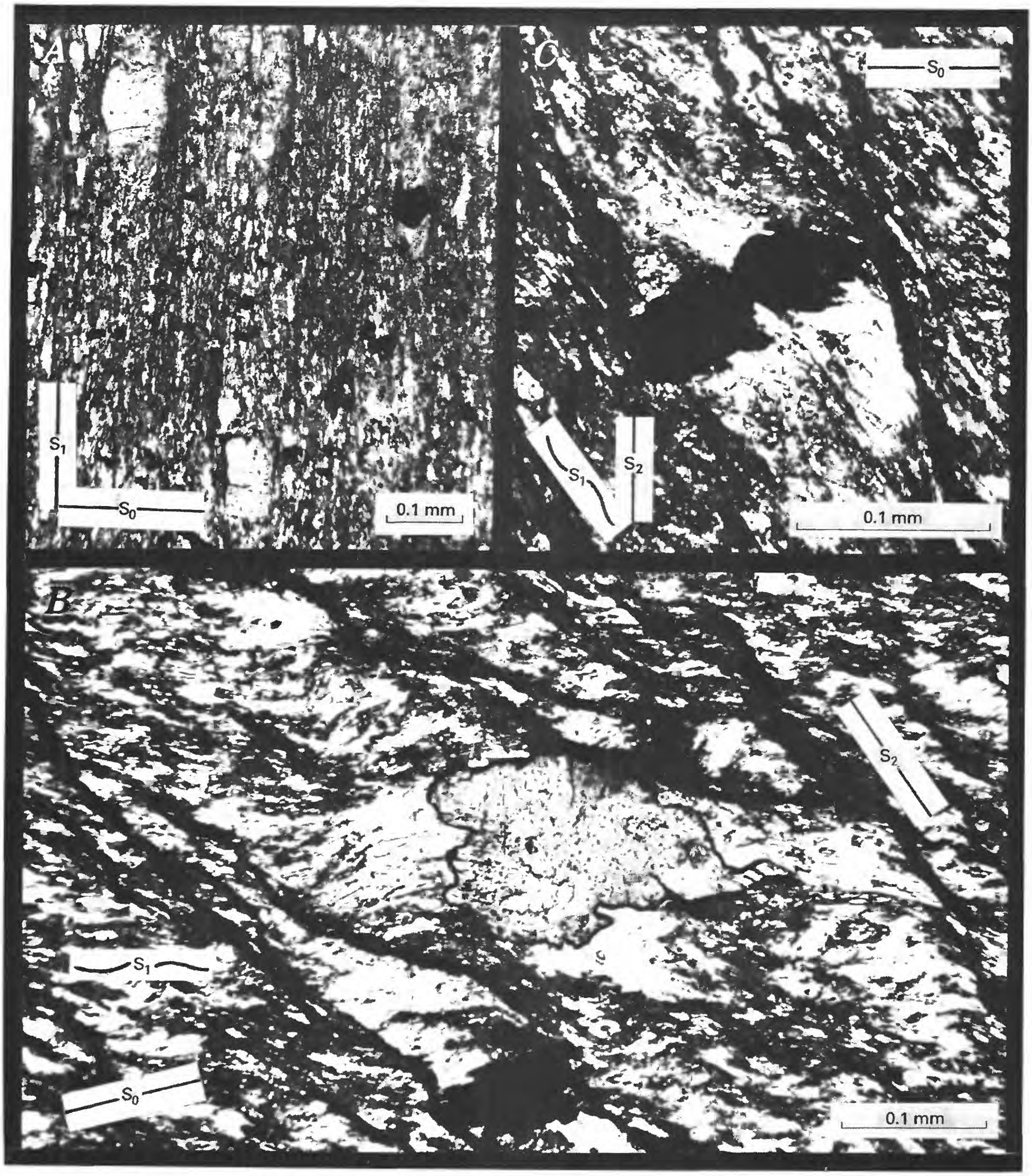


Figure 14. Photomicrographs showing microscopic structures in slate, Carlton County, Minn. *A*, Large chlorite–white mica aggregates from the Thompson dam, showing original bedding-parallel orientation of (001), and development of strain shadows parallel to S_1 slaty cleavage. *B*, Chlorite–white mica aggregate deformed by S_2 crenulation cleavage (from Gillogly Road location in Holst's southern terrane). Strain shadows parallel to S_1 remain visible despite S_2 overprint. *C*, Large opaque porphyroblasts with strain shadows parallel to S_1, subsequently deformed by S_2 (from same location as *B*). Compare with identical textures in figure 10.

during the first deformation (Holst, 1985; Holm and others, 1988).

My recent studies in Carlton County, Minn., suggest an alternative interpretation of the relative timing of the foliations and associated folds in the northern and southern terranes. The slaty cleavage associated with first generation folds in the southern area immediately adjacent to the boundary appears to be identical in structural petrography to the cleavage associated with first generation folds to the north. Like the rocks of the Baraga, Mich., area, the slaty cleavage can be readily correlated on both sides of Holst's boundary by its relationship with diagenetic chlorite porphyroblasts (fig. 14*A*, *B*). Although these grains are deformed by first and second generation folding south of the boundary, they display strain shadows parallel only to S_1 slaty cleavage. These strain shadows are later crenulated by S_2 (fig. 14*B*), thus allowing an easy distinction between the foliations when the chlorite porphyroblasts are present.

The structural correlation presented above is based on style group criteria, that is, morphological similarity of foliations, lineations, and folds, and overprinting relationships. Holst's correlation is based on similarity in orientation between structural features, an approach which has been shown to be problematic (Weiss and McIntyre, 1957; Turner and Weiss, 1963; Hobbs, 1965; Hobbs, Means, and Williams, 1976). Furthermore, no compelling reason presents itself to make the unusual correlation (Turner and Weiss, 1963, p. 464) of a regional slaty cleavage with a late crenulation cleavage.

The field evidence in Carlton County strongly supports the progressive deformation model proposed by Connolly (1981), whereby slaty cleavage developed synchronously in first generation folds in rocks from both the north and south terranes, with deformation increasing to the south. The evidence further suggests that the southern area was subsequently overthrust to the north, with the development of B_2 folds in the southern transported hanging-wall terrane, rather than in the northern footwall terrane (Holst, 1985).

COMPARISONS WITH NORTHERN APPALACHIAN TECTONICS

Similarity in Terrane Assemblages

Numerous authors have suggested a collisional plate tectonic model for the Early Proterozoic Penokean orogeny in Upper Michigan and Minnesota (Cambray, 1977, 1978a, 1978b; Larue, 1983; Schulz, 1984; Holst, 1984a; Sims and others, 1987). The work presented here extends this concept, providing field documentation for the existence of a typical foreland fold and thrust belt in Upper Michigan (fig. 1). The *terrane assemblage* in this area, from the parautochthonous rocks at the northern tectonic front, to the metamorphic core zone, and finally to the well-documented Niagara fault zone in the south (Larue, 1983; Sims and others, 1984; LaBerge and others, 1984) is remarkably similar to that of the Northern Appalachians.

Figure 15 is a structural section across the Northern Appalachians, also showing the size and sequence of similar terranes in northern Michigan. The section runs from Glens Falls, N.Y., to Claremont, N.H., and is situated close to lines 1, 3, 4, and 5 of the COCORP New England seismic survey. Seismic "picks" from Brown and others (1983, fig. 2) and Ando and others (1984, fig. 3) were incorporated in the structure section. The geology of the parautochthonous and allochthonous foreland terranes is based largely upon the work of Rowley and Kidd (1981), Bosworth and Rowley (1984), and Stanley and Ratcliffe (1985). Interpretations in the Green Mountains, Ludlow Valley, and Chester dome areas are based on sections by Ando and others (1984), and on field work by Thompson (1950), Nisbet (1976), and Gregg (1975, 1989b).

The foreland in the Northern Appalachians consists of a sequence of parautochthonous Cambrian and Ordovician sediments, overlain by the highly deformed allochthonous Taconic slices (Zen, 1967; 1972). To the east of the foreland lie three distinct terranes that make up the metamorphic core zone. The westernmost of these, the Green Mountains, is a large Precambrian massif that is detached from the basement along a shallow decollement (fig. 15 and Brown and others, 1983). On the east side of this massif is the complexly deformed Ludlow Valley sequence of the Vermont Valley slices. This zone, formerly referred to as the Townshend-Brownington syncline, consists of a tectonic stacking sequence of fault-bounded, polyphase-deformed metasedimentary rocks (Gregg, 1975). Finite strains within this pseudo-stratigraphic sequence are extremely high; as many as three secondary foliations are present, and transposition is widespread, commonly resulting in tectonic pseudo-conglomerates (Gregg, 1990). Tectonic slices of ultramafic rocks occur along many of the faulted lithologic contacts in this terrane. These ultramafics, marking an earlier overthrust, have been redistributed by a major doming event that marks the next terrane.

To the east of the Ludlow Valley sequence lies a series of mantled gneiss domes (Thompson, 1950), typified by the Chester dome in figure 15. Strong seismic reflectors occur beneath the domes (Brown and others, 1983; Ando and others, 1984) and appear to truncate the root zones of the early thrusts present in the eastern fold and thrust terrane (Stanley and Ratcliffe, 1985).

Origin of Gneiss Domes

Since their initial description by Eskola (1949), mantled gneiss domes have been considered to be either wholly or partly diapiric in origin by many workers (Thompson, 1950; Thompson and others, 1968; Ramberg, 1963; Stanley, 1975; Miller, 1983; Sims and others, 1985; Dietsch, 1989). A few workers have pointed out the strong possibility that the domes are compressional structures displaying large-scale fold interference patterns (Ramsay, 1967). In the Northern Appalachians, Nisbet (1976) documented the compressional nature of the Chester dome, showing its large-scale box-fold form and its close association with a macroscopically penetrative synkinematic axial-planar crenulation cleavage. He showed that the dome event overprinted two earlier style groups of folds and associated thrusts, forming the Butternut Hill fold nappe on the west flank of the Chester dome (Nisbet, 1976, p. 152), and folding the main ultramafic suture zone in central Vermont.

The broad antiformal gneiss domes in the Northern Appalachians, and in other areas such as Upper Michigan, are typically paired with tightly appressed synforms developed in the overlying metasedimentary rocks. The characteristic form of the gneiss dome–synform pair fits Ramsay's (1967, p. 383) mechanical model, in which a high viscosity contrast along a boundary between basement and cover causes the crystalline basement rocks to develop broad, rounded folds of large amplitude, while the overlying low-viscosity sedimentary rock is deformed into small-wavelength, "pinched" synforms. I suggest that gneiss dome-and-pinched synform structure originated during the buckling of crystalline thrust sheets (Hatcher and Williams, 1986) after detachment from the basement. This buckling may explain the box fold or angular fold profile observed in gneiss domes, because folds of this style commonly form above decollements in materials characterized by closely packed competent layering, or around single buckling layers (Ramsay, 1967, fig. 7–81). The varieties of crystalline thrust sheets described by Hatcher and Williams (1986) are all capable of producing thin basement slices that could approximate the single-layer case. In the Chester dome, for example, the buckling is superimposed upon a composite-type crystalline thrust sheet, whereas domes in other areas, such as northern Michigan, may be developed in thin-skinned sheets.

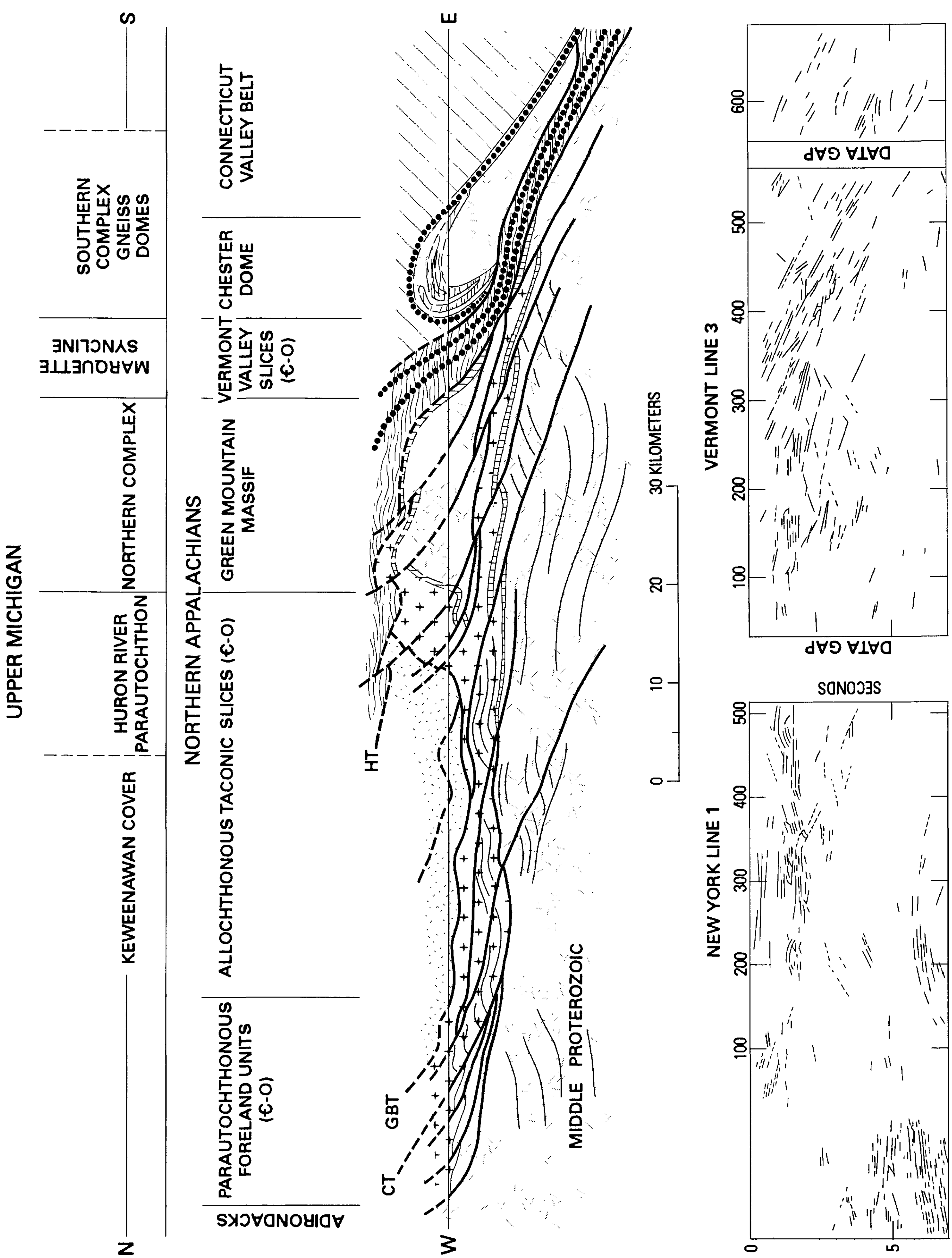
Basement Detachments in the Penokean Orogen

The terrane assemblage outlined above is quite similar to the sequence present along a north-south line through the Precambrian of Upper Michigan (figs. 1, 15). The foreland in Upper Michigan is represented by the Huron River parautochthon. The Falls River thrust marks the southern limit of this terrane, separating it from allochthonous rocks to the south. This fault also appears to define the boundary between the Huron River parautochthon and the northern complex. I suggest, therefore, that the northern complex may be a detached massif similar to the Green Mountains in Vermont, transported northward with the polyphase-deformed metasediments of the Falls River slice along a shallow decollement. The very low finite strains present in the Huron River parautochthon, and the absence of allochthonous terranes in front of the northern complex, suggest that the amount of transport was significantly less than that which occurred in the Northern Appalachians.

Several small bodies of Archean rocks, completely surrounded by Baraga Group (Early Proterozoic) sedimentary rocks, occur along the northern boundary of the northern complex. The largest of these, at Arvon Hill (fig. 2), is a thin slice of Archean gneisses approximately 550 m wide and 3,500 m long. Cannon (1986) has suggested that this body is bounded by steep, east-west-trending faults. An alternative explanation might be that the body, along with an intact "cap" of Goodrich Quartzite, is an exposed ramp formed by a south-dipping thrust sheet. Field mapping has been inconclusive in regard to this problem, which might be better investigated with geophysical methods.

Because of the lack of modern structural mapping in the region northwest of Marquette (fig. 1), complete integration of the eastern Clark Creek and Dead River slate belts into the tectonic picture is not yet possible. Both belts lie south of the proposed extension of the Falls River thrust, so their current status as undisturbed sedimentary basins is questionable. If they are allochthonous terranes, they represent either slices that dip beneath part of the northern complex, or klippen resting on shallow decollements above the Archean rocks.

In northern Michigan, the second terrane of the metamorphic core is represented by the Marquette syncline, which lies between the basement massif and the southern complex gneiss domes. Both the Marquette syncline and the analogous Ludlow Valley sequence fit the gneiss dome–synform model presented in this report; however, the rocks in the Marquette syncline show far less deformation (Cambray, 1984) than the Ludlow Valley rocks (Gregg, 1975) and contain no ultramafic zones of Penokean age. It


UPPER MICHIGAN
N
S
KEWEENAWAN COVER
HURON RIVER PARAUTOCHTHON
NORTHERN COMPLEX
MARQUETTE SYNCLINE
SOUTHERN COMPLEX GNEISS DOMES
NORTHERN APPALACHIANS
ADIRONDACKS
PARAUTOCHTHONOUS FORELAND UNITS (Є-O)
ALLOCHTHONOUS TACONIC SLICES (Є-O)
GREEN MOUNTAIN MASSIF
VERMONT VALLEY SLICES (Є-O)
CHESTER DOME
CONNECTICUT VALLEY BELT
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E
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MIDDLE PROTEROZOIC
0
10
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30 KILOMETERS
NEW YORK LINE 1
VERMONT LINE 3
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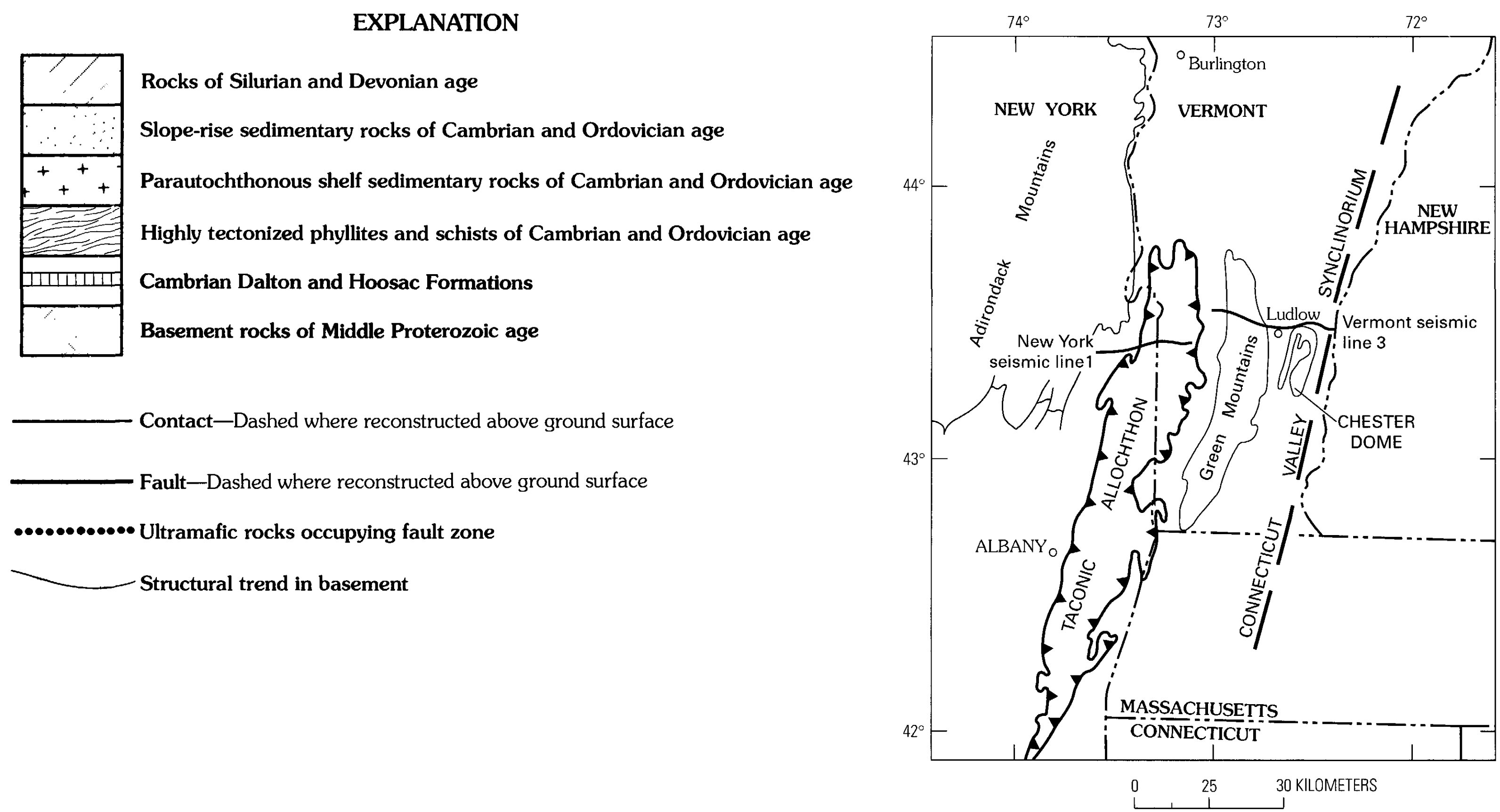

Figure 15. Comparison of tectonic terranes in the Northern Appalachians and Upper Michigan. The terranes occur in a similar order and are approximately the same size. The seismic section for the Northern Appalachians is based on combined interpretations of COCORP data by Ando and others (1984) and Brown and others (1983); numbers, stations along seismic lines. CT, Champlain thrust; GBT, Giddings Brook thrust; HT, Hinesburg thrust. No vertical exaggeration. Index map modified from Ando and others (1984).

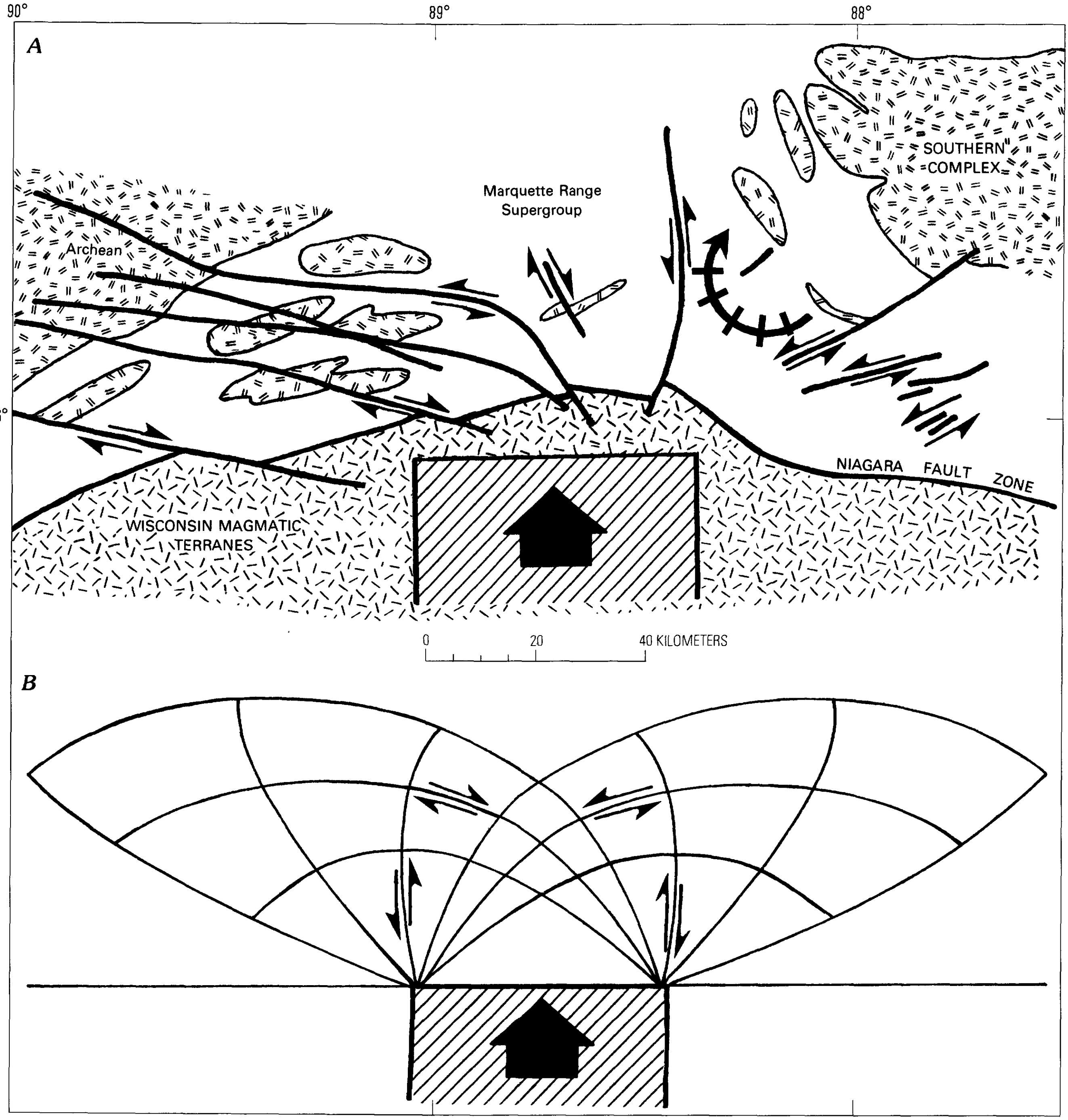

Figure 16. Structural model for development of late Penokean faults in northern Michigan. *A*, Structural features in Wisconsin-Michigan border region (modified from Cannon, 1986). Heavy lines, faults; barbs show direction of relative movement where known. Large curved hachured arrow shows clockwise sense of rotation near southern complex, possibly causing reorientation of gneiss domes. Large, wide arrow, direction of tectonic transport. *B*, Slip-line model showing fault pattern similar to *A* developed at the front of an irregular indentor (large, wide arrow) during continental collision (modified from Tapponier and Molnar, 1976).

seems likely, therefore, that major pre-doming detachments did not extend to this zone from the deeper Niagara suture (fault zone) to the south, and that faulting such as that along the Covington section (fig. 1) was limited to minor detachments associated with later stage crustal shortening. Larger movements may have occurred along the proposed detachment beneath the northern complex, or along higher detachments that may have ramped over the gneiss dome–synform structures in the southern complex.

The gneiss domes of Upper Michigan, like the synforms, appear to be less deformed counterparts of the Grenville domes in Vermont, having developed on the surface of a large, relatively undisturbed basement block, rather than on a complexly deformed basement slice. They nevertheless show the usual box-fold or angular fold styles when viewed in the profile plane, and are bounded by a number of tightly appressed synforms. The gneiss dome–synform model suggests a compressional origin for the domes, as well as the possibility of their northward transport along a shallow decollement during the Penokean orogeny. The model supports the suggestion of Ueng, Larue, and Sedlock (1984) that the Amasa dome represents "crossfolding" rather than a "gneiss dome," if one assumes that their reference to gneiss domes implies the classic diapiric model. It also supports the work of Klasner (1984) and Attoh and Klasner (1989) who showed intense involvement of basement during Penokean thrusting in the Wisconsin magmatic arc terranes (Schulz, 1984), with northward transport of the Dunbar gneiss dome toward the continental foreland along southward-dipping overthrusts.

Origin of Metamorphic Nodes

James (1955) reported a series of elliptical metamorphic nodes (fig. 1) associated with the gneiss domes in Upper Michigan and proposed that they developed around a series of separate heat sources. Attoh and Vander Meulen (1984), however, have shown that a magmatic heat source alone is insufficient, and suggested that crustal thickening or doming might have contributed the additional heat required for regional metamorphism. I suggest that the isograd patterns may simply correlate with relative structural position in the gneiss dome–synform model, because mantling sediments in the deep synformal hinges would be expected to show the highest metamorphic grade. Subsequent isostatic uplift, tilting, and erosion of the gneiss dome–synform complex would eventually expose these hinges. The Twin Lake and Grant Lake domes may therefore represent a tilted cross section of the basement-cover relationship, with metamorphic grade decreasing "upwards" out of the Republic syncline to the northwest.

Rotation of Gneiss Domes

Cannon (1973) pointed out that the axial traces of both the southern complex gneiss domes and the major folds in the Proterozoic rocks have very divergent trends, and presented the problems that this situation posed for a horizontal compression model. This redistribution of the gneiss dome–synform structure appears to have been synchronous with the development of a broad pattern of faults that radiate northward from the Wisconsin magmatic arc terranes (Cannon, 1978, 1986; Sims, 1990). I suggest that the dome rotation and the faults were the result of a local, north-northeast-directed tectonic indentation (Dewey and Burke, 1973; Sengor, 1976) by an irregular projection on the north edge of the Wisconsin magmatic terranes. This late transport direction, which differs from the northwest compression direction of the main Penokean event (Sims and Peterman, 1983, fig. 5), may be entirely local.

Figure 16 shows a comparison of the Penokean fault system with the indentor slip-line model of Tapponnier and Molnar (1976, fig. 6). I have modified the model by a 20 percent pure-shear deformation in the thrusting direction. This amount seems reasonable, as it is lower than the strains measured even in remote parautochthonous rocks, and it results in a pattern which closely resembles that shown by the Penokean faults. Agreement is good between the sense of offset shown by the faults and the predicted shear directions in the slip-line model (fig. 16). Near the Amasa dome, the shear sense on the faults shows an overall clockwise rotation of the area, which could account for the rotation of the dome axes out of an original east-west orientation (arrows in fig. 16).

The tectonic interpretations presented in this report suggest that at the beginning of the Proterozoic the southern complex and possibly part of the northern complex existed as a surface of low relief, forming a single large sedimentary basin extending over all the structural terranes, rather than a series of small basins in tectonic troughs (Larue and Sloss, 1980). This basin received uniform sedimentation in a subsiding continental shelf–shallow deltaic environment facing open waters to the south. The Penokean collisional event outlined in Attoh and Klasner (1989) resulted in an assemblage of structural terranes very similar in style to, but less intensely developed than, those found in well-studied Phanerozoic collisional mountain belts. The relatively weak imprint of the collisional event in Upper Michigan might be due to a smaller size of the colliding volcanic-magmatic arc terranes (Schulz, 1984) compared to the Appalachians.

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