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Nature of the Taconic Orogeny in the Cupsuptic Quadrangle, West-Central Maine

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Trip H

Nature of the Taconic orogeny in the Cupsuptic quadrangle, west-central Maine[1/]

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Introduction

In recent years considerable evidence has been accumulated in the central and northern Appalachians to support the major hiatus between the pre-Silurian and Silurian rocks, proposed by Rogers (1838, p. 37) more than a century ago. Although the nature of the structural break and the length of the hiatus vary considerably over the Appalachian region, the data summarized by Pavlides and others (1968) indicate that diastrophism took place in much, but not all, of the northern Appalachians from late Middle Ordovician to Early Silurian. Hall (1969) and Zen (1967, 1968) have recently presented evidence for early Middle Ordovician tectonism which Zen (1968, p. 131) suggested was an early phase of the Taconic orogeny.

In the southeastern part of the Cupsuptic quadrangle there is an angular discordance between the upper Middle Ordovician (graptolite zone 12) rocks and the overlying Lower Silurian rocks. It is the purpose of this trip to examine the rocks and structural styles above and below the pre-Silurian--Silurian boundary in the hope of demonstrating some of the effects of the Taconic orogeny.

The trip involves two fairly long but not very strenuous hikes over old logging roads and through relatively open woods. Hiking boots are strongly recommended.

Stratigraphy

Stops will be made to look at the Albee, Dixville, and Rangeley Formations in the southeastern part of the Cupsuptic quadrangle. Brief lithologic descriptions of these formations with references to more detailed descriptions as well as discussions of their age and stratigraphic order are given in trip A-3 (Harwood, Green, and Guidotti, this volume) and trip A-1 (Moench and Boudette, this volume) and will not be repeated here. Lithologic descriptions of specific rocks seen on this trip are given under the appropriate stops in the accompanying road log.

1/Publication authorized by the Director, U. S. Geological Survey

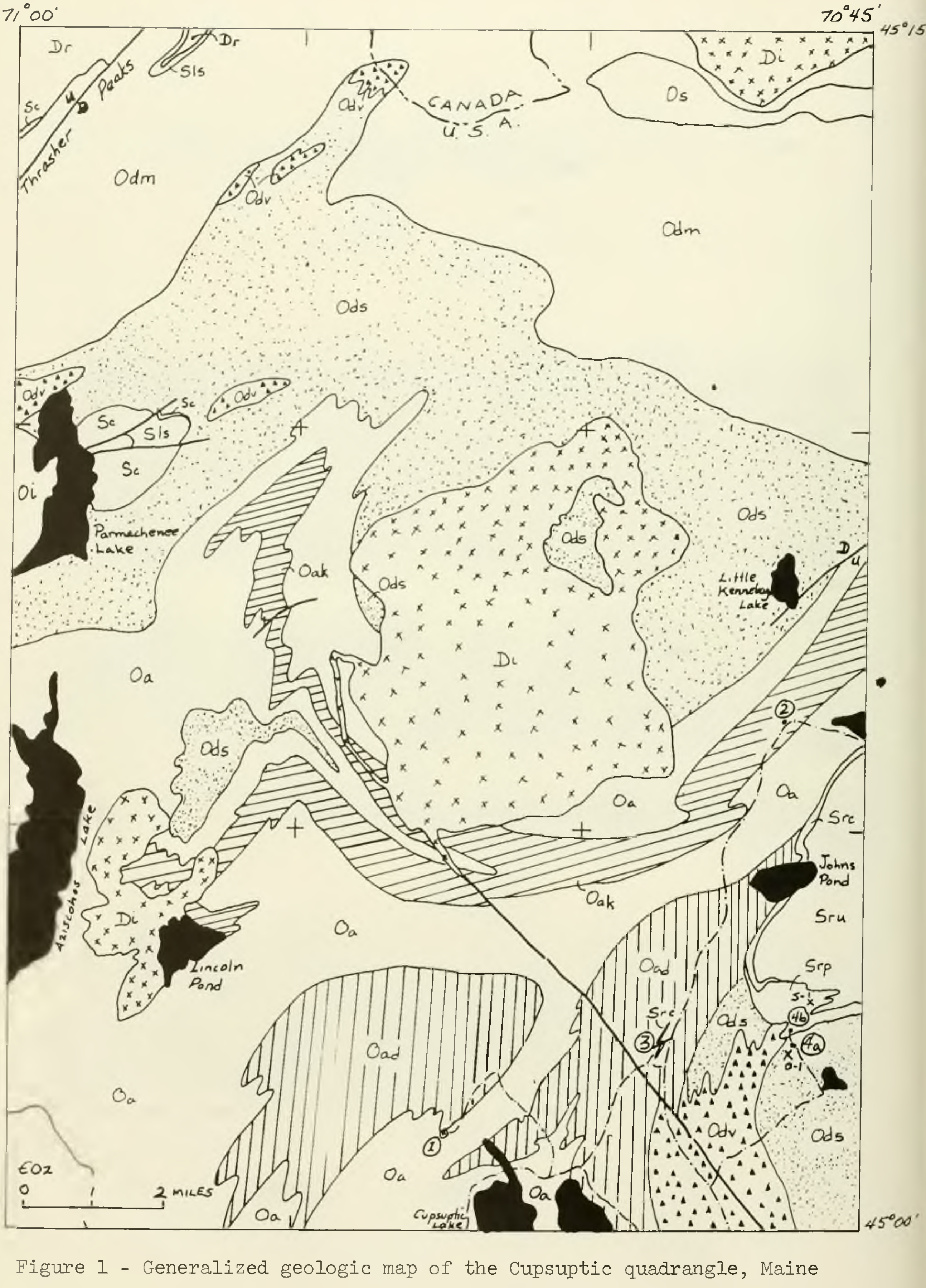

Figure 1 - Generalized geologic map of the Cupsuptic quadrangle, Maine

EXPLANATION

Devonian

Dr

Gray slate and feldspathic quartzite

Silurian

Sls / Sc

Unnamed rocks

Sls- gray clacareous slate, limestone, and limestone conglomerate

Sc- quartz-pebble conglomerate and minor polymict conglomerate

Sru / Src / Srp

Rangeley Formation

Sru- gray slate and feldspathic quartzite with scattered layers of polymict conglomerate

Src- quartz-pebble conglomerate and quartzite

Srp- polymict conglomerate and feldspathic quartzite

Rangeley Formation considered to Lower Silurian (?)

UNCONFORMITY

Ordovician

Odm / Ods / Odv

Dixville Formation

Odm- Magalloway Member; feldspathic graywacke granule conglomerate, minor slate and metavolcanic rocks

Ods-black sulfidic slate and feldspathic quartzite

Odv- mafic metavolcanic rocks

Oa / Oad / Oak

Albee Formation

Oad- green, purplish-gray and black phyllite with less than 10 percent of interbedded "pinstripe" quartzite

Oa- green to greenish-gray slate and phyllite with 10 to 50 percent "pinstripe" quartzite

Oak- red slate with greater than 10 percent of white weathering "pinstripe" arenaceous rocks

Camb. & Ord.

O€z

Aziscohos Formation of Green (1959)

Green phyllite and chlorite schist with abundant stringers and pods of white quartz

Di

Devonian intrusive rocks

Oi

Ordovician intrusive rocks

Os

Ordovician serpentinite and related mafic rocks

④

Field trip route showing location of stops

O-1
X

Fossil locality

D
U

Fault

D,downthrown side; U, upthrown side

Pre-Silurian--Silurian boundary

Near Johns Pond (fig. 1) polymict and quartz-pebble conglomerate of the Rangeley Formation truncate the steeply dipping contact between the Middle Ordovician Dixville Formation and the underlying Albee Formation. The contact is not exposed here, but the bedding attitudes shown on figure 2 clearly indicate an angular discordance between the Silurian and pre-Silurian rocks. About 10 miles to the southeast, however, the Rangeley Formation at the type locality appears to rest conformably above the Greenvale Cove Formation which, in turn, rests conformably above the Quimby and Dixville Formation (see trip A-1, Moench and Boudette, this volume).

The pre-Silurian--Silurian boundary near John Pond has been interpreted by Harwood (1966; Harwood and Berry, 1967) as an angular unconformity because the polymict conglomerate contains abundant clasts of the subjacent pre-Silurian rocks, and there is no evidence of brittle, low-angle thrust movement in the rocks near the contact. The absence of the Greenvale Cove and the Quimby Formations is due, in this interpretation, to either nondeposition or erosion on the unconformity.

Recently Moench (1970, p. 1492) has suggested that the inferred unconformity might be a major premetamorphic fault along which wet sediments of the Rangeley Formation slid southeastward, becoming folded and cleaved in the process. This interpretation explains the absence of the Greenvale Cove and Quimby Formations and the absence of brittle fault deformation in the Silurian rocks, but it is incompatible with the fact that the pre-Silurian--Silurian boundary is folded into a northeast-trending syncline at Johns Pond. The premetamorphic faults mapped by Moench to the southeast are not folded about northeast axes; in fact, Moench's concept of down-to-basin faulting and folding requires that the northeast-trending folds be generated by slumping on unfolded premetamorphic faults.

The distribution of sedimentary facies in the Silurian rocks supports the concept of a hinge line between Johns Pond and Rangeley Lake proposed by Moench (1970) to separate a basin of apparently continuous sedimentation to the southeast from an eroding positive area to the northwest. At Johns Pond the quartz pebble conglomerate overlaps the northwestern edge of a wedge of polymict conglomerate that thickens from about 1,000 feet south of Johns Pond to about 9,000 feet near Rangeley Lake (Moench, 1970). The quartz-pebble conglomerate apparently marks a local southeastern limit of the beach environment in Llandovery time. The northwestern extent of the Llandovery beach, or, conversely the southeastern limit of the positive area in Llandovery time is not precisely known, but it must have been between Johns Pond and Thrasher Peaks (fig. 1) because the basal Silurian rocks near Thrasher Peaks are most probably Ludlow (Green, 1968, p. 1516).

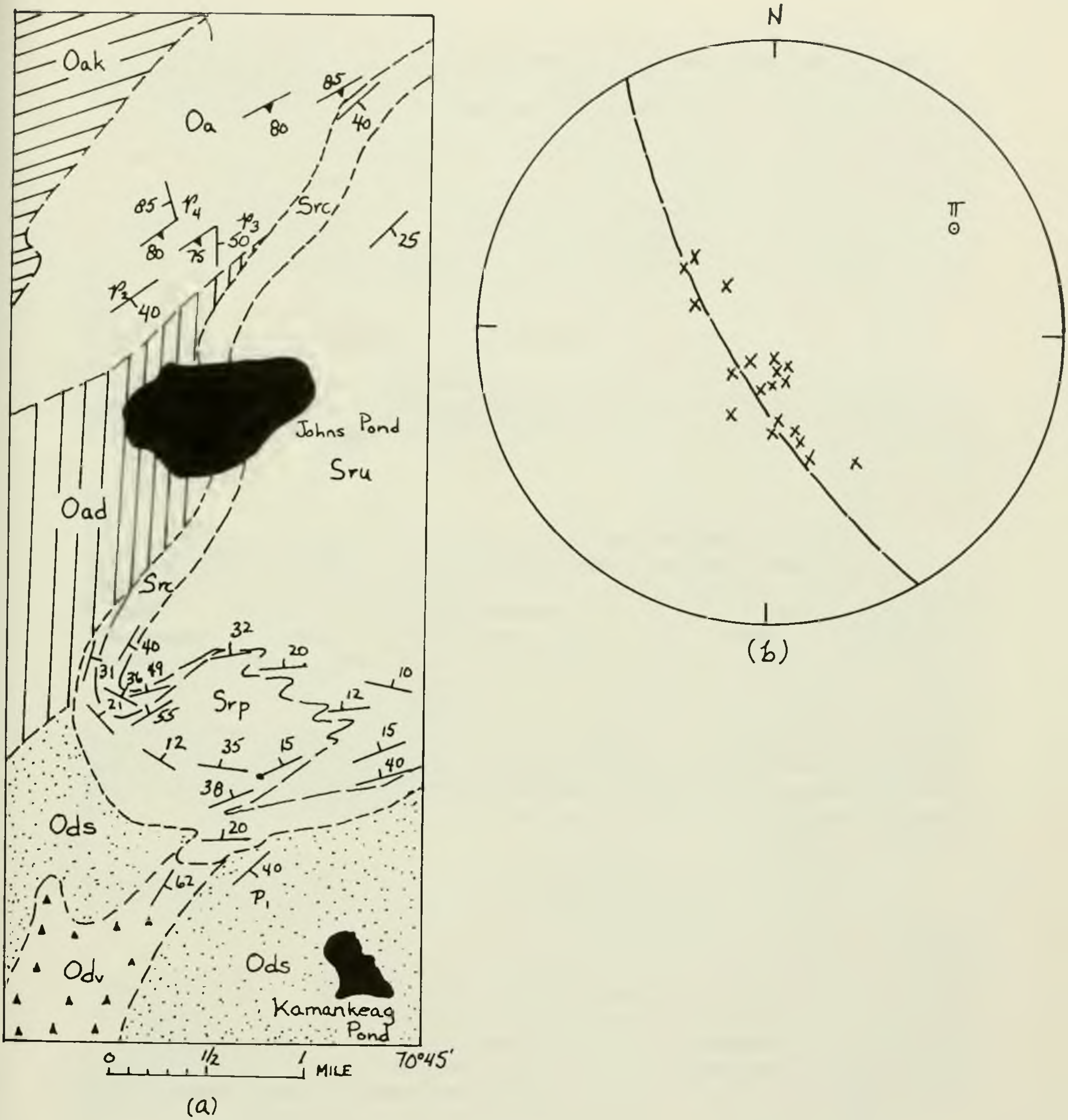

Figure 2. a) Geologic map of Silurian and pre-Silurian rocks near Johns Pond showing bedding (40) and cleavage (85).

b) Point diagram showing girdle (dash great circle) and girdle axis (π) defined by poles to bedding (x) for Silurian rocks near Johns Pond, southeastern part of the Cupsuptic quadrangle.

From these relationships it is concluded that the pre-Silurian--Silurian boundary in this area is a time-transgressive unconformity that developed on an eroding positive area underlain in part at least by pre-Silurian rocks occupying the core of the present Boundary Mountain anticlinorium. Cady (1968, p. 156) shows this positive area as the Somerset geanticline, and Boucot (1968, p. 88) shows it as Somerset Island. The onlap of the Silurian sea represents only the final curtain call to a period of diastrophism, and we logically ask, when did the show open and what were the main events?

Affects of the Acadian orogeny

All the stratified rocks in the Cupsuptic area were deformed during the Acadian orogeny in early Middle Devonian time. This deformation must be identified and effectively removed from the total deformation of the pre-Silurian rocks in order to examine the effects of the Taconic orogeny.

Bedding attitudes and graded beds indicate that the Silurian rocks near Johns Pond have been folded into a northeast-trending syncline. Poles to bedding, shown in figure 2, define a girdle whose axis plunges 18° N. 58° E. and represents the local axis of Acadian folding. The trend of this axis agrees fairly well with the trend of fold axes in the Deer Brook syncline (Harwood, 1969) to the northwest and with the trace of most of the folds outlined by Silurian and Devonian rocks in the Merrimack synclinorium to the southeast (see Osberg, and others, 1968, fig. 18-2, p. 242). Northwest-trending folds are found locally in the Silurian and Devonian rocks, but these folds appear to be local anomalies in the regional northeast trend related to deformation produced by large Devonian plutons (Guidotti, 1965).

The trend of the axial surfaces in many folds in the pre-Silurian rocks in the Cupsuptic quadrangle is between N. 40° E. and N. 50° E. and thus closely parallels the trend of the folds in the Silurian rocks. Generally, however, this is the only geometric feature that the folds above and below the unconformity have in common. Typically the folds in the Silurian rock are relatively open with gently dipping limbs, gently plunging linear axes, and simple forms. Pre-Silurian folds with a northeast trend, on the other hand, are typically isoclinal with steeply plunging commonly curvilinear axes and have complex forms.

At Stop 1 the complex fold pattern in the Albee Formation can be related to the superposition of northeast-trending folds on west-northwest-trending folds which has produced a series of small domes and basins separated by saddle-shaped areas. At Stop 2 isoclinal northeast-trending folds in the Albee have strongly curved axes resulting in "hog-back" shaped minor folds, but there is no evidence of a preexisting fold pattern or of a folded cleavage. In fact, most of the axes of the minor folds in the pre-Silurian rocks are steeply plunging or vertical, and these folds show no refolded linear or planar features.

The different fold patterns above and below the unconformity suggest that the Acadian folds were superimposed on previously folded beds, but the absence of an early pervasive flow cleavage argues against intense deformation in pre-Silurian time.

Pre-Acadian attitude of the pre-Silurian rocks

The pre-Acadian attitude of the pre-Silurian rocks near Johns Pond can be determined graphically by unfolding the Silurian rocks about the axis of Acadian folding shown in figure 2. Only an approximate orientation can be obtained because there is no way of accurately determining the slope of the erosion surface that developed on the pre-Silurian rocks. For simplicity, it will be assumed that the erosion surface was flat and thus comparable to the one beneath the Atlantic Coastal Plain sediments in New Jersey which dips slightly less than 1° (J. P. Minard, 1970, pers. comm.).

The attitude of bedding in the Dixville Formation near fossil locality 0-1 (fig. 1) rotates from N. 43°E./40°SE. (p_1, fig. 3a) to N.62°E./62°SE. (p_1^1, fig. 3a) when the Acadian axis of folding is restored to the horizontal and the overlying Silurian rocks are rotated to the horizontal. Greater variation is shown in the bedding attitudes in the Albee near the Silurian rocks north of Johns Pond; the present and pre-Acadian attitudes of bedding for three localities are shown in figures 3b, c, and d.

Although this specific geometric restoration cannot be applied throughout the Cupsuptic area because the sense of rotation must be obtained from immediately adjacent Silurian rocks, the examples in figure 3 clearly show that some of the pre-Silurian rocks were locally steeply inclined prior to Silurian time. The assumption that the erosion surface on the pre-Silurian rocks dipped a significant amount to the southeast toward the maximum thickness of the Silurian polymict conglomerate only steepens the pre-Acadian dips in the pre-Silurian rocks near Johns Pond.

Beginning of the Taconic orogeny

Zen (1967, 1968) recently summarized the evidence for Ordovician diastrophism in the Taconic area, from which he concluded that tectonic events began by block faulting, minor volcanism, and folding in earliest Middle Ordovician time. These events were followed by the deposition of Middle Ordovician carbonate rocks and black mud (graptolite zone 11 and 12 time) and eventually by emplacement of at least the early slices of the Taconic allochthon (graptolite zone 13 time). Both the allochthonous and autochthonous rocks are unconformably overlain by Silurian rocks in eastern New York. Hall (1969) also has recognized an unconformity at the base of Middle Ordovician clastic

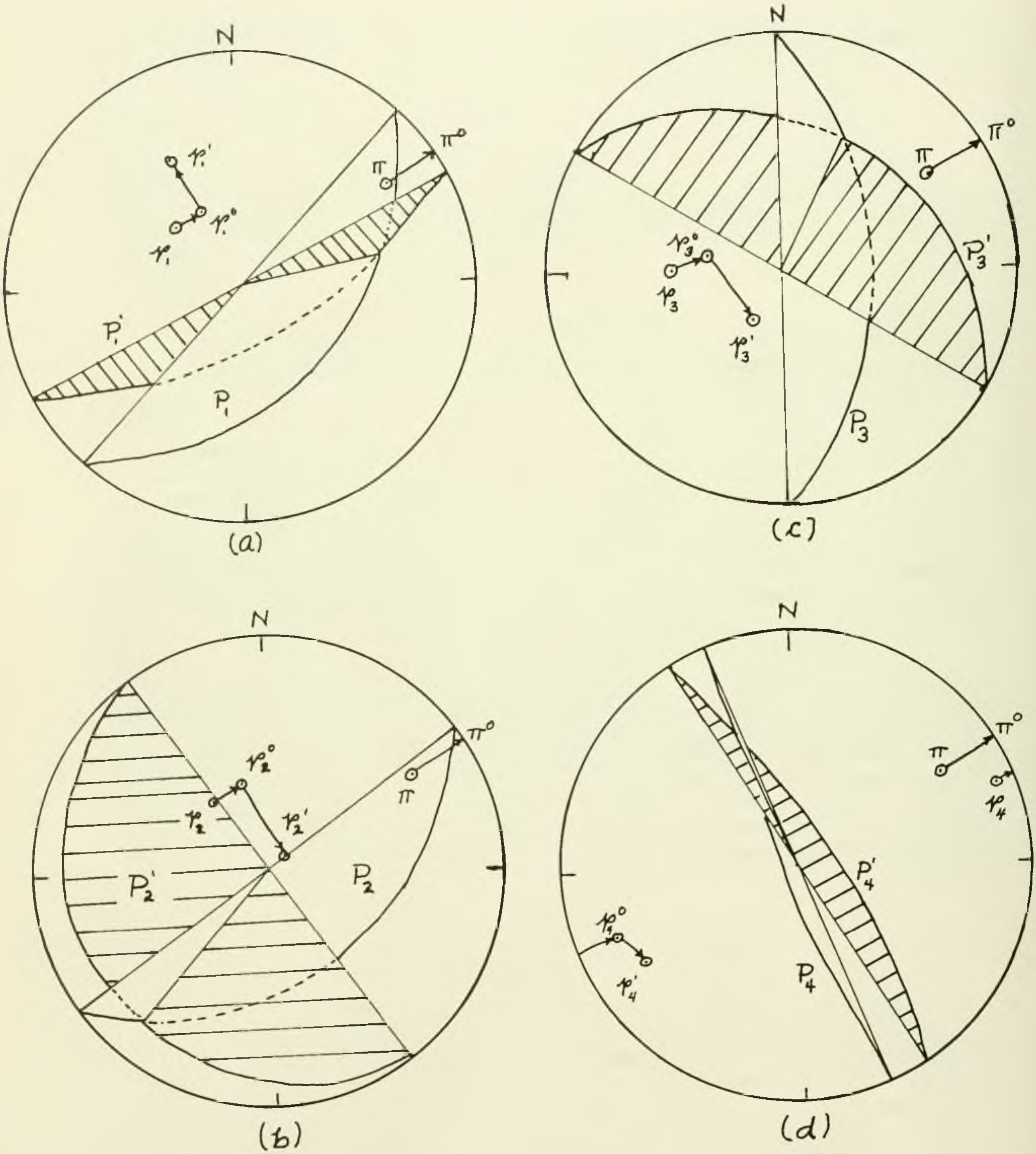
N
π
π°
P1'
P1
(a)
N
π
π°
P3'
P3
(c)
N
π
π°
P2'
P2
(b)
N
π
π°
P4'
P4
(d)

Figure 3(a). Pre-Silurian orientation of bedding plane in Dixville Formation at fossil locality 0-1.

P_1 = bedding plane in present, post-Acadian, orientation

p_1 = pole to plane P_1

π = axis of Acadian folding in Silurian rocks near Johns Pond determined in figure 2.

π° = axis of Acadian folding restored to horizontal

P_1^1 = bedding plane P_1 in pre-Silurian orientation

p_1^1 = pole to bedding plane P_1^1.

Pole p_1 rotates to p_1° as axis of Acadian folding (π) is restored to horizontal (π°) along east-west axis of the net. With π° at the north pole of the net, p_1° rotates along a small circle of the net to p_1^1 as Silurian rocks on south limb of syncline are rotated clockwise 24° to horizontal position.

3(b), (c), (d). Pre-Silurian orientations of bedding planes in the Albee Formation adjacent to Silurian rocks north of Johns Pond. Symbology the same as in figure 3(a) except that p_2°, p_3°, p_4° rotate to p_2^1, p_3^1, and p_4^1, respectively, as overlying Silurian rocks are rotated 40° counterclockwise to horizontal.

rocks and has recognized different fold styles above and below this unconformity in north-central Maine. He concluded that folding, uplift, and erosion occurred prior to graptolite zone 11 time and again between graptolite zone 13 time and Early Silurian time.

Folding of the pre-Silurian rocks in the Cupsuptic area must have occurred after graptolite zone 12 time and before C-3-C-5 time of the late Llandovery. This corresponds very well with the second period of folding recognized by Hall in north-central Maine and with the time of emplacement of the early slices of the Taconic allochthon determined by Zen. To date, however, there is no unequivocal evidence in the Cupsuptic area for an unconformity or different fold styles within the pre-Silurian rocks which might indicate early Middle Ordovician tectonism comparable to that found elsewhere. To further complicate matters, no fossils have been found in rocks below the zone 12 graptolite-bearing slate, so there is no way of knowing whether a significant hiatus exists within these older rocks. Therefore, the stratigraphic position of the pre-zone 11 break reported elsewhere and the geologic history in this area during that time are a matter of speculation.

Hall (1969) proposed that the contact between the Aziscohos Formation of Green (1959) and Albee Formations might correspond to the unconformity he mapped beneath graptolite zone 11-bearing rocks. This possibility cannot be refuted, but certainly the most profound change in the sedimentary record below the zone 12 graptolite-bearing slate is found at the Albee-Dixville boundary. Here hematite- and magnetite-bearing pelitic rocks in the Albee are succeeded abruptly and conformably with minor interbedding by sulfide-bearing carbonaceous pelitic rocks and thick accumulations of volcanic rocks and feldspathic graywacke. In fact, the Magalloway Member of the Dixville is lithologically similar to the graywacke, granule conglomerate, slate and volcanic rocks assigned by Hall (1969, p. 468) to his Group B which contains graptolites indicative of zone 11 and 12 and which rests unconformably above pre-zone 11 rocks.

The distribution of rock types in the Dixville is compatible with uplift and erosion north and northeast of the Cupsuptic quadrangle. The thick sequence of feldspathic graywacke and granule conglomerate mapped as the Magalloway Member of the Dixville (fig. 1) may have been derived from a positive area to the north and east. The Magalloway Member is succeeded to the south by black slate containing scattered patches of mafic volcanic rocks. Along the southeastern flank of the Boundary Mountain anticlinorium a thick accumulation of mafic volcanic rocks is in contact with the Albee and is succeeded upward by black slate which, in turn, is overlain by graywacke and felsic metavolcanic rocks of the Quimby Formation.

This distribution of rocks suggests rapid deposition of the Magalloway Member from the north into a euxinic basin that was concurrently receiving large amounts of mafic volcanic rocks further to the south. It implies, furthermore, that the map units in the Dixville Formation are lithofacies, an implication supported by some fossil evidence. Neuman (1968, p. 40) equates brachiopods found by E. V. Post in the Brassua Lake quadrangle in feldspathic graywacke (see U. S. Geol. Survey, 1961, p. A12) similar to the Magalloway Member with the zone 12 graptolites found in black slate at locality O-1 (fig. 1).

It is tentatively concluded, therefore, that the early Middle Ordovician tectonism marked by an unconformity at the base of zone 11 graptolite-bearing rocks in the Taconic area and in north-central Maine is marked by the conformable change from Albee-type sedimentation to Dixville-type sedimentation in the Cupsuptic area.

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Road log.

The route of this trip is in the Rangeley, Oquossoc, and Cupsuptic 15-min. quadrangles, but stops will be made only in the Cupsuptic quadrangle. Assemble at 8:00 a.m. at the junction of Routes 16 and 4 in the center of the village of Rangeley.

Mileage

0.0 Drive west on Routes 16 and 4 toward Oquossoc. Stratigraphic units crossed along this part of the route are discussed in the road logs of trips A-1 and A-3 in this volume.

5.6 Enter Oquossoc quadrangle.

6.5 Turn right (northwest) on Route 16 at the northwest end of Rangeley Lake.

8.1 Cross Kennebago River.

9.7 Enter Cupsuptic quadrangle.

11.7 Turn right (north) from Route 16 at Cupsuptic Nursery.

13.9 Cross Cupsuptic River at the Cupsuptic Sporting Camps. Turn left on the west side of Cupsuptic River and proceed southwest toward the Moochers Home.

14.0 Park cars in the fields west of Moochers Home and continue on foot along an old road along the west side of Cupsuptic River.

14.8 STOP 1. NO HAMMERING PLEASE. We are on the 1740-foot knob on the northwest-trending ridge between the Cupsuptic River and Cold Brook. Outcrops on the top of the knob are rusty-weathering black phyllite, the lowest lithologic unit in the upper member of the Albee Formation. Overlying the black phyllite is green, gray, and purplish-gray phyllite containing less than 10 percent of interbedded "pinstripe" arenaceous rocks that make up most of the upper member of the Albee. Below the black phyllite is greenish-gray and green phyllite containing more than 10 percent of "pinstripe" feldspathic quartzite typical of the bulk of the Albee Formation.

The minor folds in the "pinstripe" arenaceous beds below the black phyllite are of particular interest because they clearly show two widely divergent directions of folding. Axial surfaces of the early folds are variable in trend but generally strike west northwest and dip about 50° S. Axes of these early folds plunge from 10° to 20°E.

to as much as 30°W. and describe a helical pattern across the axial trace of the late folds. Axial surfaces of the late folds are more consistent in strike, being about N. 20° E., but the axes vary in plunge from 10° to 25° NE. to as much as 50° SE. Individual beds traced around this exposure give a three-dimensional picture of dome, basin, and saddle-shaped fold segments typical of superimposed, nearly orthogonal, steeply dipping fold systems.

The regional slaty cleavage is parallel to the axial surface of the late folds. There is no recognizable regional cleavage parallel to the axial surface of the early folds, but a northwest-trending cleavage is faintly visible in the cores of some early folds in this outcrop.

Retrace marked trail back to the cars and retrace the route along the Cupsuptic Tote Road to Route 16.

17.9 Turn left (east) on Route 16.

18.2 Turn left (north) from Route 16 onto private road to north end of Kennebago Lake.

20.1 Cross Kennebago River.

20.3 Turn left (north) onto abandoned Maine Central Railroad grade and proceed north toward Kennebago Lake.

25.5 STOP 2. The gorge of the Kennebago River below the dam at Kennebago Lake is cut in red slate and interbedded "pin-stripe" feldspathic quartzite of the lower member of the Albee Formation. Note that, except for the color caused by microscopic plates of hematite, the red slate and feldspathic quartzite are identical to the green slate and arenaceous rocks of the typical Albee seen at STOP 1. Downstream near the lower power dam are thin discontinuous layers of greenstone in the red slate and a few highly folded beds of relatively clean quartzite about 2 feet thick.

The minor folds on the east side of the river about 100 feet below the upper dam are of particular interest to this trip. Nearly all the minor folds in this large exposure are isoclinal and have near vertical axial surfaces that strike N. 50° E. and axes that plunge 50° to 60° SE. Several hogback minor folds have northeast-trending axial surfaces and axes that plunge from 10° NE. to as much as 50°SW. over a distance of 1 to 5 feet. There is a pronounced cleavage parallel to the northeast-trending axial surfaces of the minor folds but no other cleavage or deformed lineations in this area.

The hogback folds could be the result of differential flowage in the "a" tectonic direction or the result of superimposed northeast-trending late folds on early north-west-trending folds. The fracture pattern near the hinge lines of some of the smaller hogback folds and their diapiric appearance favors differential flowage, but in view of the superimposed folds seen at STOP 1 and the steep plunges of surrounding folds, the alternative interpretation cannot be entirely ruled out.

25.5 Turn vehicles around and retrace the route south along the railroad grade.

29.9 Cross Kennebago River and park vehicles in the abandoned gravel pit about 100 yards west of the bridge. Walk north along the west bank of the Kennebago River to the first outcrops.

30.0 STOP 3. The first outcrop on the west side of the river is white to light-tannish-gray quartz-pebble conglomerate with a light-yellowish-green chloritic matrix. This rock is identical to the quartz-pebble conglomerate member of the Rangeley Formation found near Johns Pond, and thus it is considered to be late Llandovery in age. No fossils have been found at this locality, however, and this isolated sliver of conglomerate cannot be traced into the dated rocks.

A few tens of feet upstream from the first outcrops, the contact between the quartz-pebble conglomerate and green phyllite of the upper part of the Albee Formation is exposed near the river bank. The contact strikes N. 75° E. into the river and is vertical. Bedding is not visible in the Albee, so the existence of an unconformity cannot be unequivocally established.

The quartz-pebble conglomerate is lithologically similar to much of the Clough Quartzite (Billings, 1937, 1956) in New Hampshire. To the east the conglomerate rests on the Albee Formation north of Johns Pond and above polymict conglomerate of the Rangeley Formation southwest of Johns Pond.

Return to the vehicles and retrace the route south along the private road to Route 16.

31.9 Turn left (east) on Route 16 at the Forest Service station.

33.5 Enter the Oquossoc quadrangle.

35.0 Cross the Kennebago River.

35.2 Turn left (north) from Route 16 onto dirt road that follows the abandoned Maine Central Railroad grade.

36.5 Enter the Cupsuptic quadrangle. Low hogback ridge east of the railroad grade is greenstone assigned to the mafic metavolcanic member of the Dixville Formation.

37.3 Turn right (east) from railroad grade onto logging road to Kamankeag Pond.

38.0 Park vehicles on the side of the logging road and prepare for a hike of about 2 miles to the low hills northwest of Kamankeag Pond.

40.2 STOP 4a. Soon after leaving the cars we crossed from the mafic metavolcanic member of the Dixville into the black slate member. We are presently at fossil locality 0-1 (fig. 1). The slate here contains a collection of graptolites indicative of Berry's zone 12 of the late Middle Ordovician (fossil list given in Harwood and Berry, 1967). The rock is predominantly black slate that contains scattered beds of feldspathic quartzite as much as 2 inches thick and a few scattered lenses of metavolcanic rocks and some stringers of black chert. The beds are folded but in general strike about N. 40° E. and dip 40° to 80° SE.

Along the slopes to the northwest are scattered exposures of calcareous lithic graywacke that contains beds of granule to cobble conglomerate and thin beds and wispy lenticles of black slate. Clasts in the conglomerate bed are gray quartzite, vein quartz, and chert that are contained in a matrix of graywacke or mixed graywacke and black slate. The calcareous lithic graywacke is a lenticular unit within the black slate member. The upper contact is a gradational zone of alternating graywacke and black slate beds a few inches thick. In the one outcrop where the lower contact is exposed, there is a sharp but conformable change upward from black slate to pebble and cobble conglomerate.

There are abundant scattered fragments of various fossils in the calcareous lithic graywacke. None of the fragments could be identified specifically, but R. B. Neuman of the U. S. Geological Survey tentatively identified one brachiopod fragment as *Bilobia* or *Diambonia*, either of which is compatible with the Middle Ordovician age established by the graptolites.

The layer of calcareous lithic graywacke and included conglomerate is interpreted to be an interformational flysch deposit because of the shallow-water nature of the fossil fragments and the chaotic nature of the included wispy lenticles of black slate.

STOP 4b. We have walked northwestward from the graptolite locality and are now at the first outcrop of cobble-to-boulder polymict conglomerate of the Rangeley Formation. The contact between the basal Silurian rocks and the underlying Middle Ordovician rocks is not exposed but probably is in the low swale southeast of this outcrop.

Bedding is not well developed in the basal part of the polymict conglomerate, which is a jumble of cobbles and boulders composed of vein quartz, gray quartzite, chert, black slate, green slate, minor greenstone, and white-weathering fine-grained felsic metavolcanic rocks and some pebbles of medium-grained gray granite. Several boulders of coarsely crystalline gray limestone that contain recrystallized Halysites and horn corals were collected at this locality. The matrix is feldspathic quartzite, and thus it is quite different from the graywacke-black slate matrix of the Middle Ordovician conglomerate seen at STOP 4a.

Above the basal 50 feet the polymict conglomerate becomes distinctly bedded; beds of cobbles alternate with beds of coarse-grained feldspathic quartzite, both ranging in thickness from 1 to 5 feet. The clasts tend to decrease in size, and the amount of stable quartz, quartzite, and chert relative to unstable slate fragments increase. Granitic cobbles tend to be somewhat more abundant in the middle and upper parts of the polymict conglomerate in this region.

Black slate fragments are distinctly cleaved, but the orientation of their cleavage is essentially parallel to the cleavage in the Silurian rocks, indicating that the cleavage was imposed on the slate chips after they were incorporated in the conglomerate. This finding is consistent with the absence of a pervasive cleavage in the pre-Silurian rocks that can be related to Taconic deformation. The black slate fragments are not deformed around adjacent clasts, however, this implies that their protolith was indurated probably to mudstone or argillite before Silurian time.

Several brachiopod fragments identified as Eoplectodonta or Plectodonta by A. J. Boucot and a few corals were found at locality S-2 (fig. 1). Neither these fossils nor the recrystallized corals from the limestone boulders positively date the conglomerate but all are compatible with late Llandovery (C_4 - C_5) age fossils found by E. L. Boudette. This is the last STOP of the trip. Follow the marked trail back to Kamankeag Pond and retrace the logging road to the cars.