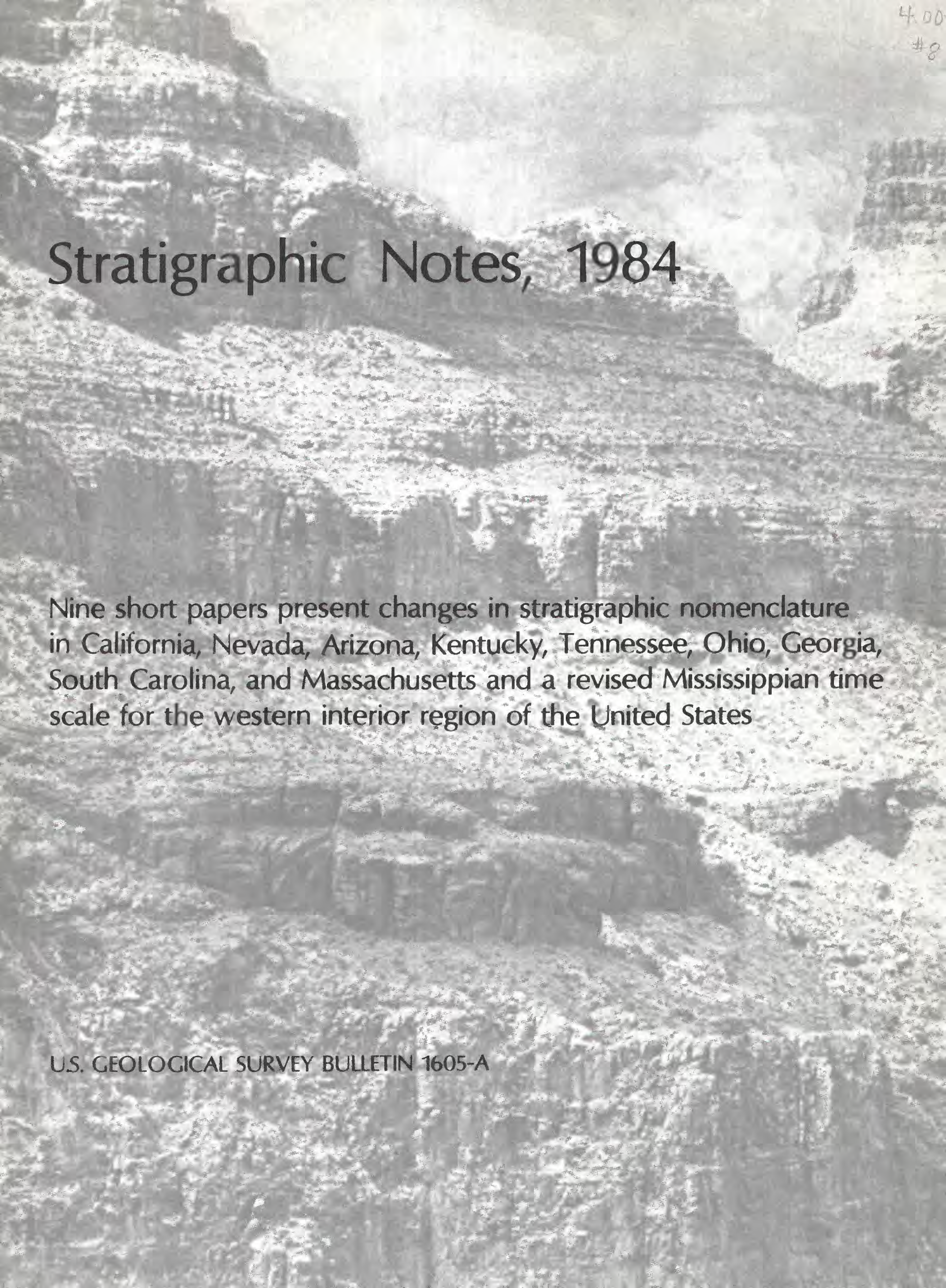
Stratigraphic Notes, 1984

U.S. GEOLOGICAL SURVEY BULLETIN 1605-A

COVER PHOTOGRAPH: *Type section of the Surprise Canyon Formation, Grand Canyon, Arizona, by Stanley S. Beus, 1984*



Stratigraphic Notes, 1984

Nine short papers present changes in stratigraphic nomenclature in California, Nevada, Arizona, Kentucky, Tennessee, Ohio, Georgia, South Carolina, and Massachusetts and a revised Mississippian time scale for the western interior region of the United States

U.S. GEOLOGICAL SURVEY BULLETIN 1605-A

DEPARTMENT OF THE INTERIOR
DONALD PAUL HODEL, Secretary

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Dallas L. Peck, Director

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CONTENTS

REVISIONS IN THE TERTIARY STRATIGRAPHY OF THE EAST FLANK OF THE DIABLO RANGE, CENTRAL CALIFORNIA

By J. Alan Bartow[1]

Abstract

The San Pablo Formation is geographically restricted from the area south of Hospital Creek on the east flank of the Diablo Range. Strata previously assigned to the basal part of the San Pablo Formation and lower previously unnamed strata are herein reassigned to the Poverty Flat Sandstone (new) of late middle and late Eocene age. The Valley Springs Formation (upper Oligocene and lower Miocene) is extended from the east side of the northern San Joaquin Valley to include part of the original San Pablo Formation. The remainder of the strata previously assigned to the San Pablo Formation are reassigned to the Neroly Sandstone (upper Miocene).

INTRODUCTION

Recent geologic mapping and stratigraphic studies of Cenozoic strata bordering the northern San Joaquin Valley, together with study of subsurface data from strata under the valley, have provided new information about the regional stratigraphic relations that require revisions in the Tertiary stratigraphy. Lithologic similarities between upper Eocene to upper Miocene sequences exposed on the east side of the northern San Joaquin Valley and on the west side between Hospital and Garzas Creeks (fig. 1) strongly suggest a one-to-one correlation of units. Lithologic equivalents of the east-side units—the Ione, Valley Springs, and Mehrten Formations—are recognized on the west side of the valley, mostly as part of what had been previously mapped as the San Pablo Formation of Anderson and Pack (1915). The San Pablo Formation is here geographically restricted from the area south of Hospital Creek on the east flank of the Diablo Range, and its rocks are here reassigned to the Poverty Flat Sandstone (new), the Valley Springs Formation, and the Neroly Sandstone.

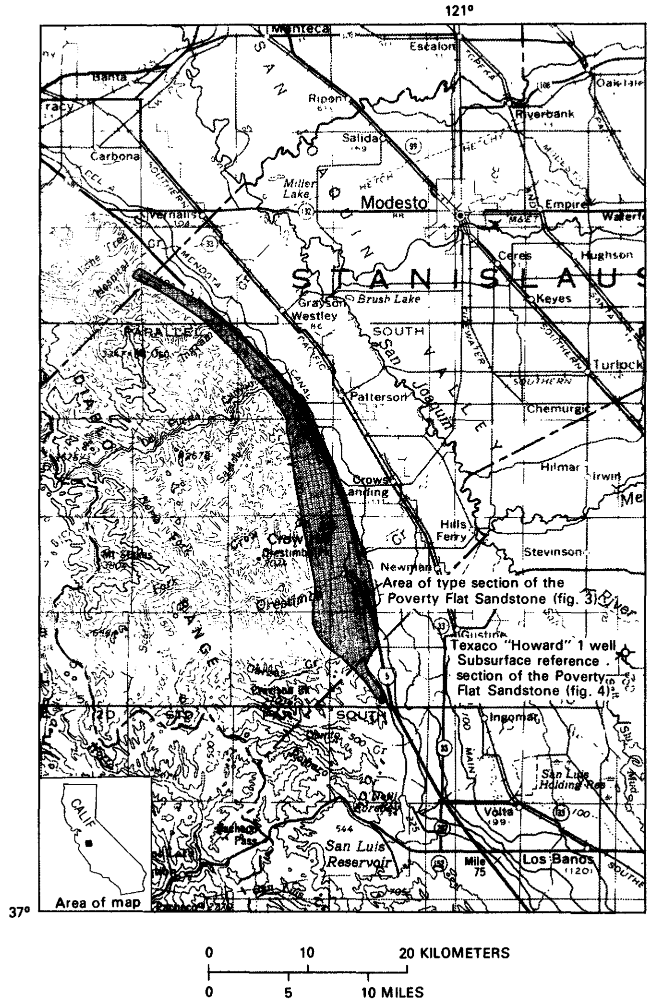

Figure 1. Index map of eastern flank of the Diablo Range and adjacent San Joaquin Valley showing the generalized outcrop area of the Tertiary strata (shaded area) and the locations of the type and subsurface reference sections of the Poverty Flat Sandstone.

POVERTY FLAT SANDSTONE

A predominantly sandstone sequence, above the Kreyenhagen Shale and below the Valley Springs Formation, that is exposed for 4 to 5 km along the California Aqueduct south of Orestimba Creek (fig. 1) is here named the Poverty Flat Sandstone, a name taken from the shallow, broad valley just east of the aqueduct. The Poverty Flat includes the "undifferentiated Miocene between Garzas and Crow Creeks" of Anderson and Pack (1915, p. 89, 90) and the basal conglomerate and conglomeratic sandstone of the overlying San Pablo Formation of Anderson and Pack (1915, p. 99) (fig. 2). The Poverty Flat comprises units 15, 16, and 17 in the stratigraphic section from near Crow Creek described by Anderson and Pack (1915, p. 99, 100). Subsequent workers have followed the Anderson and Pack definitions, although an Oligocene age was proposed for the "undifferentiated Miocene" of Anderson and Pack (1915) in a 1922 University of California student report by H. O. Elftman (cited in Jenkins, 1931). The name "Poverty Flat" was first used for the undifferentiated unit by Collins (1950), who also considered it to be Oligocene (fig. 2).

The Poverty Flat Sandstone, at its type section along the California Aqueduct in secs. 20, 29, and 32, T. 7 S., R. 8 E., Newman 7 1/2-minute quadrangle (fig. 3) and in the subsurface reference section in the Texaco "Howard" 1 well about 25 km to the east (figs. 1, 4), is mostly sandstone with subordinate amounts of siltstone or claystone. The lower part of the formation is composed

[1]U.S. Geological Survey, Menlo Park, CA 94025.

U.S. Geological Survey Bulletin 1605-A, 1985, p. A1-A6.

<table>
<tr><th>Anderson and Pack (1915)</th><th>Collins (1950)</th><th>THIS REPORT</th></tr>
<tr><td>Tulare(?) Formation</td><td>Tulare(?) Formation</td><td>Miocene and Pliocene(?) fanglomerate</td></tr>
<tr><td rowspan="3">San Pablo Formation</td><td rowspan="3">San Pablo Formation</td><td>Neroly Sandstone</td></tr>
<tr><td>Valley Springs Formation</td></tr>
<tr><td rowspan="2">Poverty Flat Sandstone</td></tr>
<tr><td>Undifferentiated Miocene [rocks]</td><td>Poverty Flat Siltstone</td></tr>
<tr><td rowspan="2">Kreyenhagen Shale</td><td rowspan="2">Kreyenhagen Formation</td><td>Kreyenhagen Shale</td></tr>
<tr><td>Domengine Sandstone</td></tr>
<tr><td>Tejon Formation</td><td>Tesla Formation</td><td>Tesla Formation</td></tr>
</table>

Figure 2. Correlation chart showing the development of the stratigraphic nomenclature for Tertiary units of the east-central Diablo Range.

mostly of gray or blue-gray lithic sandstone, and the upper part of interbedded light-gray to white kaolinitic quartzose sandstone and siltstone, carbonaceous shale, and kaolinitic claystone. Relative proportions of these lithologies are variable, especially in the upper part of the unit. A characteristic of the formation is the red or orange mottling in the upper part. A conglomerate at the top of the formation south of Crow Hill and north of Orestimba Creek is composed principally of red radiolarian-chert pebbles derived from the Franciscan Complex together with a few badly weathered pebbles of other Franciscan and ophiolite lithologies (diabase, gabbro, blueschist) (fig. 5). Just north of Crow Creek, pebbly sandstone with a similar clast composition is heavily cemented with hematite.

Previous workers (Anderson and Pack, 1915; Collins, 1950) have stated that the unit, here named the Poverty Flat Sandstone, unconformably overlies the Kreyenhagen Shale. Their evidence for this was the large clasts of white diatomaceous claystone included in the basal part of the sandstone that were inferred to have been eroded from the underlying Kreyenhagen Shale. However, as shown in the type section (fig. 3), claystone of the same lithology as the clasts is actually interbedded with the sandstone at the base of the Poverty Flat as well as occurring at the top of the Kreyenhagen. Both the Kreyenhagen diatomite and the claystone interbeds in the basal part of the Poverty Flat contain late middle Eocene diatoms (J. A. Barron, written commun., 1981). It therefore seems likely that the claystone clasts resulted from local scouring by currents during deposition of the basal Poverty Flat sand and do not indicate a significant unconformity.

The Poverty Flat Sandstone is unconformably overlain by the Valley Springs Formation, which truncates both the Poverty Flat and the Kreyenhagen Shale a short distance north of Crow Creek. No definite outcrops of the Poverty Flat are known south of the type area, but a red- and white-mottled quartzose sandstone that occurs locally between the Kreyenhagen Shale and a Tertiary laterite developed on top of the Kreyenhagen near Laguna Seca Creek, about 50 km to the southeast (Lettis, 1982), may be correlative with the Poverty Flat. The Poverty Flat can be recognized in many well cores and logs in the San Joaquin Valley from the Merced area, east of the type locality, at least as far south as Madera on the east and Los Banos on the west.

Diatoms from claystone at the base of the Poverty Flat Sandstone indicate a late middle Eocene age. A sparse molluscan assemblage from the same beds indicates an age of late Eocene or early Oligocene (E. J. Moore, written commun., 1981). The upper part of the formation is apparently nonfossiliferous; however, the prevailing quartz-kaolinite mineralogy of the sandstone and conglomerate, together with the red or orange mottles and concentrations of hematite cement, indicates subaerial weathering under a warm wet (tropical) climate. An early Paleogene tropical climate was followed by a well-documented climatic deterioration at the end of the Eocene (Wolfe, 1971; Frakes, 1979). Warmer climates returned during the late Oligocene to middle Miocene, but they were drier and not fully tropical (Wolfe, 1971; Frakes, 1979). Thus the deposition and weathering of the sand and gravel must have been completed before the terminal Eocene climatic deterioration. The age of the Poverty Flat, then, is late middle and late Eocene.

The Poverty Flat Sandstone is approximately correlative with the uppermost or late Eocene part of the Kreyenhagen Shale of the central part of the San Joaquin Valley and is probably correlative with the Wagonwheel Formation (Dibblee, 1973) of the southwestern San Joaquin Valley.[2] The Poverty Flat is probably partly correlative with the Ione Formation of the Sierra Nevada foothills to the east, but the age range of the Ione is not known well enough to make a definite correlation.

The Poverty Flat Sandstone represents a shallowing-upward sequence of depositional environments. The sandstone in the lower part of the formation is probably a marine shelf deposit overlying the upper-slope diatomite and diatomaceous claystone of the uppermost part of the Kreyenhagen Shale. Sandstone near the top of the Poverty Flat appears to have been deposited in a fluvial environment. A more thorough sedimentologic study is needed before the depositional environments can be described in any more detail. The formation represents the final regressive phase of the middle and late Eocene transgressive-regressive cycle that began with the deposition of the Domengine Sandstone and lower part of the Kreyenhagen Shale.

VALLEY SPRINGS FORMATION

The Valley Springs Formation was originally named and described in the Sierra Nevada foothills, where it lies above the Ione Formation and below the Mehrten Formation (Piper and others, 1939). The distinctive lithology of the Valley Springs Formation, which consists of crudely bedded yellowish-gray and tan clayey sandstone, sandy tuffaceous claystone, and vitric tuff, can be recognized in well cores throughout the northern San Joaquin Valley

[2]Dibblee originally assigned the Wagonwheel Formation to the Oligocene because he considered the Refugian Stage to be Oligocene. The Refugian, at the time this report was written, was considered to be late Eocene (Poore, 1980).

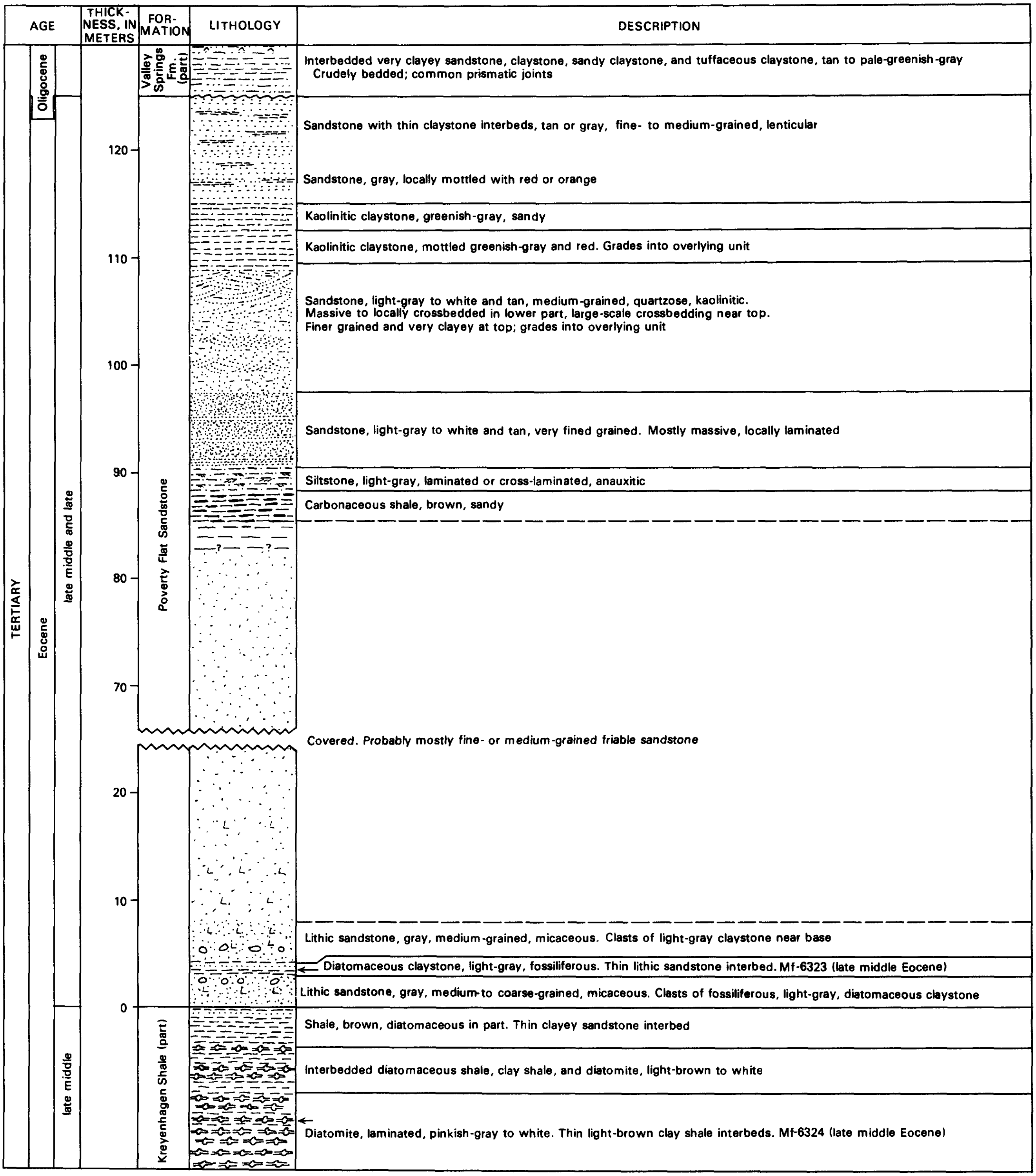

Figure 3. Type section of the Poverty Flat Sandstone along the California Aqueduct south of Orestimba Creek, secs. 20, 29, and 32, T. 7 S., R. 8 E. Arrows indicate location of U.S. Geological Survey diatom localities (Mf-6323, Mf-6324).

and is now recognized in Diablo Range outcrops. Correlation of the Diablo Range outcrops with the Valley Springs in its type area is based on similar lithology, relative stratigraphic position, and trace- and minor-element chemistry of the glass in vitric tuff interbeds (A. M. Sarna-Wojcicki, written commun., 1981).

In the Diablo Range, the Valley Springs Formation was originally included in the San Pablo Formation of

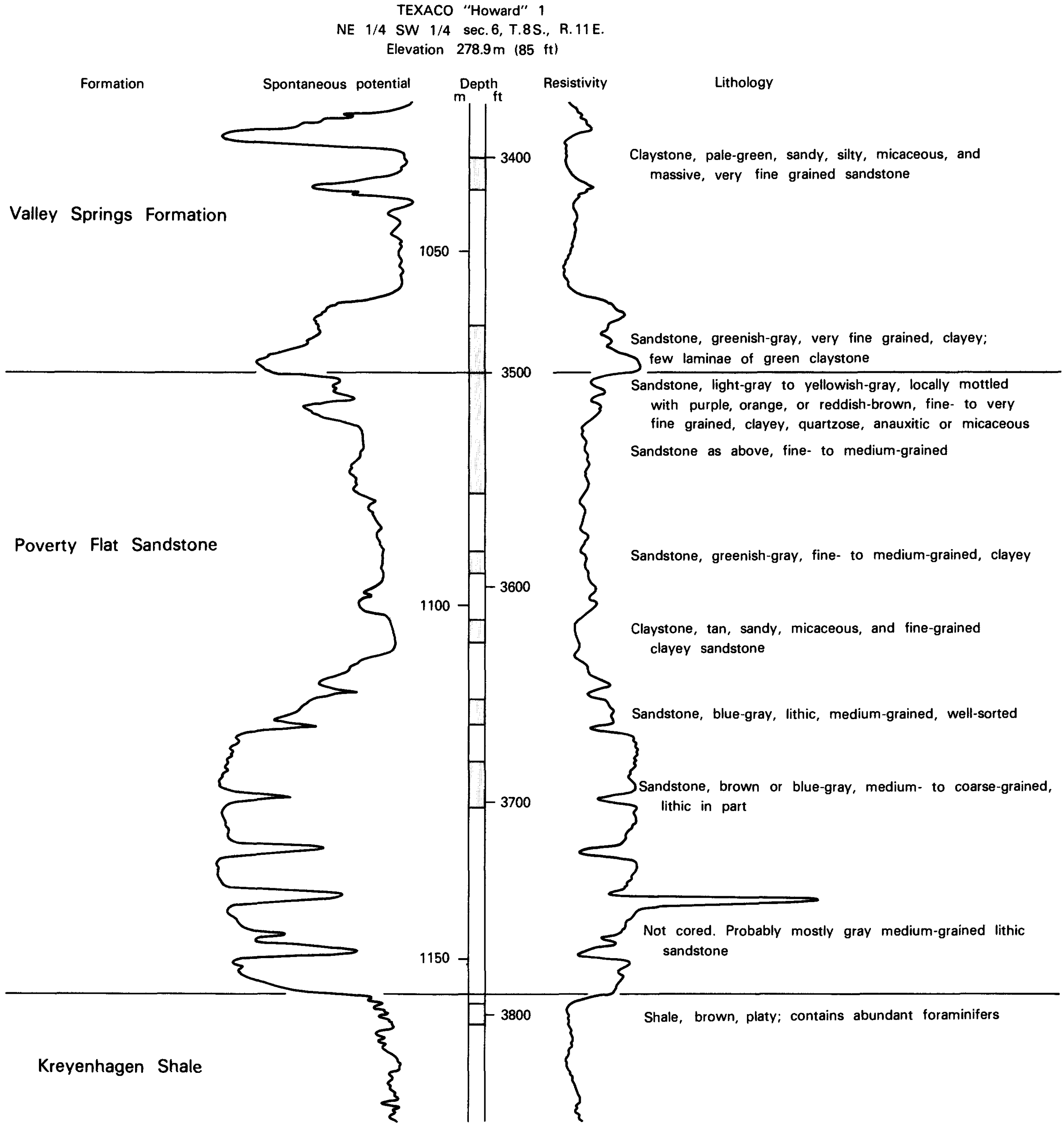

Figure 4. Subsurface reference section of the Poverty Flat Sandstone between the depths of 1,067 m (3,500 ft) and 1,155 m (3,790 ft) in the Texaco "Howard" 1 well. Shaded parts of column indicate cored intervals.

Anderson and Pack (1915) and comprises units 3 through 14 of their measured section near Crow Hill (1915, p. 99, 100). As they pointed out, the lithology of the strata that they assigned to the San Pablo Formation changes markedly near Ingram Creek. The blue-gray sandstone that makes up their San Pablo north of Ingram Creek overlaps strata of Valley Springs lithology for 1 to 2 km south of Ingram Creek but is absent farther south. All of the original San Pablo south of the overlap near Ingram Creek, exclusive of the conglomerate and sandstone that is herein reassigned to the Poverty Flat Sandstone, is here reassigned to the Valley Springs Formation.

The strata herein reassigned to the Valley Springs Formation were originally, as part of the San Pablo Formation, considered to be late Miocene in age. No new information bearing on the age of the Valley Springs was obtained from the Diablo Range outcrops, and, on the basis of K-Ar ages for tuff interbeds in the type area

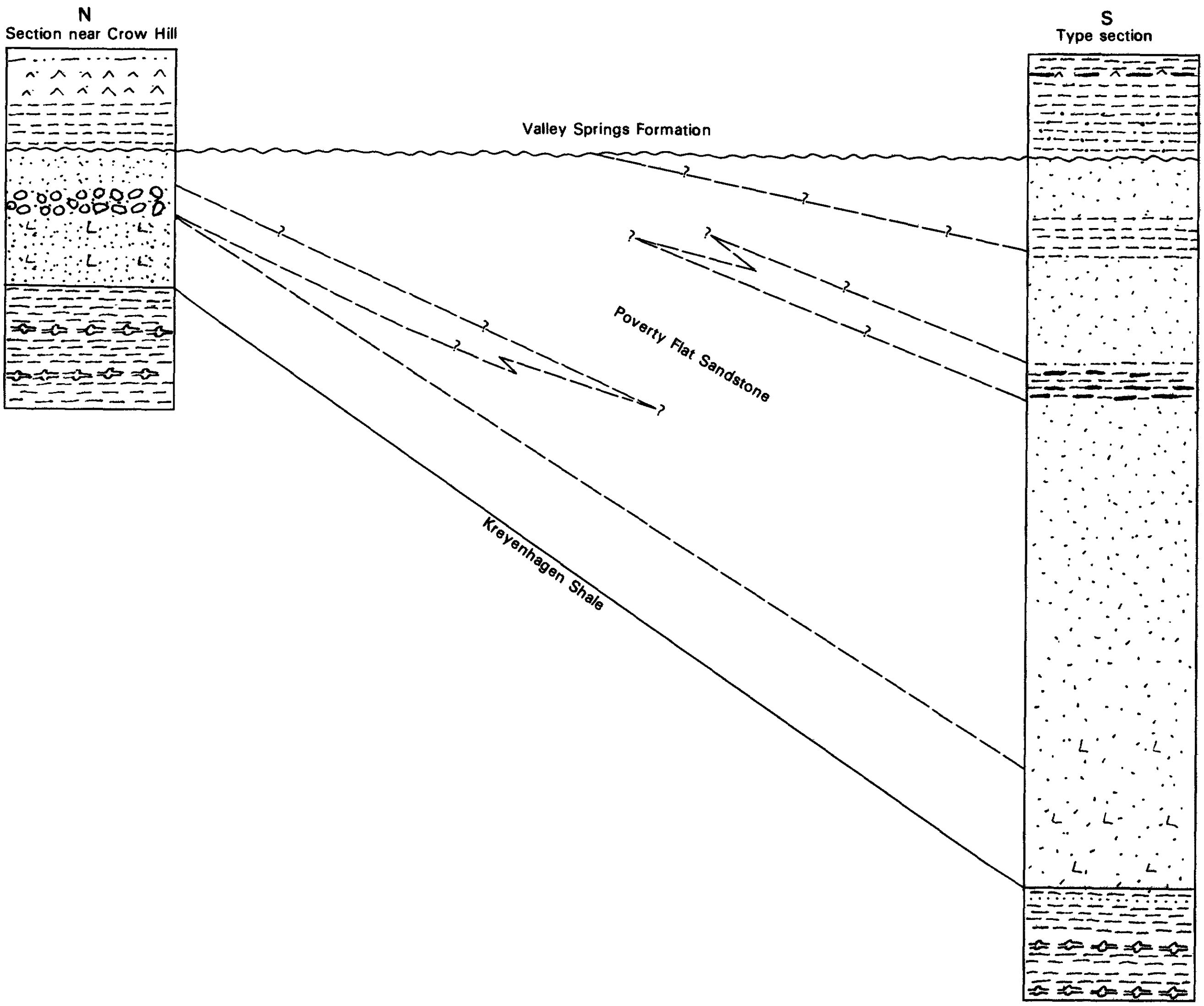

Figure 5. Diagram showing inferred relations of Poverty Flat Sandstone sections north and south of Orestimba Creek. Section near Crow Hill from Anderson and Pack (1915) section of San Pablo Formation (units 15, 16, and 17); type section generalized from figure 3.

(Dalrymple, 1963, 1964), it is considered to be late Oligocene and early Miocene in age.

NEROLY SANDSTONE

Blue-gray andesitic sandstone that unconformably overlies the Valley Springs Formation or older units and is conformably overlain by upper Miocene and lower Pliocene(?) fanglomerate is assigned to the Neroly Sandstone. The Neroly was included in the San Pablo Formation by Anderson and Pack (1915) and was assigned to the Neroly Formation of the San Pablo Group in the Tesla quadrangle a few kilometers to the northwest by Huey (1948). In the area south of Hospital Creek, the Neroly is much thinner than it is in the Tesla area and is composed entirely of sandstone and pebbly sandstone. It pinches out just south of Ingram Creek.

Sandstone and conglomeratic sandstone of the Neroly Sandstone are virtually identical petrologically to sandstone and conglomerate of the Mehrten Formation in the Sierra Nevada foothills; the volcanic detritus in both units was derived from andesitic eruptions farther east, near the crest of the Sierra Nevada. The Neroly and the Mehrten are at least partly correlative, although the Mehrten probably represents a somewhat longer span of time. Clarendonian age vertebrates from the Neroly (Raymond, 1969) indicate a late Miocene age.

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U.S. Geological Survey, August 29, 1984.

THE BIG ISLAND FORMATION, A MIOCENE FORMATION IN NORTHERN ELKO COUNTY, NEVADA, AND ADJACENT IDAHO, INCLUDING A CONSIDERATION OF ITS COMPOSITION AND PETROGRAPHIC CHARACTER

By Robert R. Coats[1]

Abstract

The Big Island Formation is a sequence of nonmarine sedimentary and volcanic rocks exposed on the western part of Big Island, a plateau between the Jarbidge River and its East Fork, and on the west and east sides of the Jarbidge River, in Elko County, Nev., and adjacent Owyhee and Twin Falls Counties, Idaho. The formation overlies the late Miocene Cougar Point Welded Tuff and is overlain discontinuously by Quaternary unconsolidated deposits.

The Big Island Formation consists of three distinct units, broadly lenticular. At the type section, the lowest unit is a boulder gravel about 100 m thick. It is overlain in several places by the middle unit, which consists of not more than 6 m of weathered dacitic tuff. This is overlain in turn by the upper unit, about 60 m of dark-gray tholeiitic olivine basalt. The basalt is the most widespread unit, and its relatively consistent petrography and age at widely scattered localities appear to offer the best criteria for correlation from place to place along the southern borders of Owyhee and Twin Falls Counties, Idaho, and in adjacent northern Elko County, Nev. The petrographic and chemical characters that are peculiar to the basalt flows of the Big Island Formation and which permit their distinction from other basalts of this region are described.

Much of the basalt was originally included in the Banbury Volcanics. Discrepancies in the age of the unit, renamed the Banbury Basalt, led researchers to conclude that the Banbury does not represent a formation of limited stratigraphic position nor of limited age range. The Banbury Basalt of the type locality is not continuous with the great area of basalt that surfaces the plateau of southern Twin Falls and Owyhee Counties, Idaho, and adjacent Elko County, Nev. The range of chemical and petrographic characters in these basalts is small; they are here all reassigned to the Big Island Formation, resulting in the restriction of the name Banbury from these areas.

DESCRIPTION

The Big Island Formation is here named for a sequence of nonmarine sedimentary and volcanic rocks exposed on the western part of Big Island, a plateau between the Jarbidge River and its East Fork, and on the east and west sides of the Jarbidge River, in Elko County, Nev., and adjacent Owyhee County, Idaho. The formation overlies the late Miocene Cougar Point Welded Tuff and is overlain discontinuously by Quaternary unconsolidated deposits. The type section of the formation is on the west wall of the Jarbidge River Canyon (fig. 1), in the Dishpan quadrangle, Idaho, beginning at a point having Idaho (W. zone) State plane coordinates of 591,300 ft east and 129,520 ft north. This point is in C, N 1/2, sec. 28, T. 16 S., R. 9 E., about 122 m west-northwest of the confluence of Buck Creek and the Jarbidge River. The section starts at the bottom of the Jarbidge River Canyon and extends for 310 m westward to the upper edge of the canyon wall. The distribution of the Big Island Formation in Elko County, Nev., is shown in figure 2.

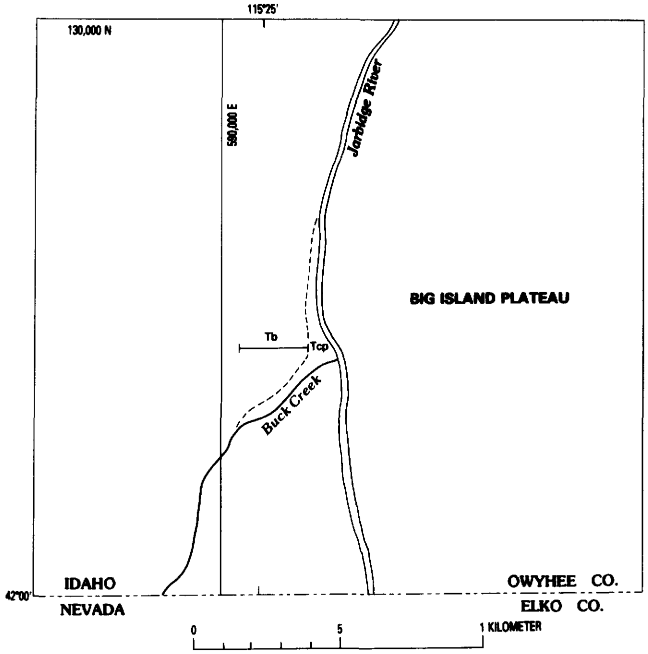

Figure 1. Sketch map showing location of type section Big Island Formation, Owyhee County, Idaho.

The Big Island Formation consists of three distinct units, broadly lenticular. At the type section, the lowest unit is a boulder gravel about 104 m thick. It is overlain in several places by the middle unit consisting of not more than 6 m of weathered, rhyolitic tuff. This, in turn, is overlain by the upper unit, about 60 m of dark-gray, tholeiitic olivine basalt. The thickness and character of the lower unit range widely within short distances; the unit pinches out to the south within a few kilometers. The tuff is also lenticular, and exposures are rare because the tuff is commonly concealed by talus. The basalt is the most widespread unit (fig. 1), and its relatively consistent petrology and age at widely scattered localities appear to offer the best criteria for correlation from place to place along the southern borders of Owyhee and Twin Falls Counties, Idaho, and in adjacent northern Elko County, Nev. Most of this paper is a description of the petrographic and chemical characters that are peculiar to the basalt flows of the Big Island Formation.

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U.S. Geological Survey Bulletin 1605-A, 1985, p. A7-A13.

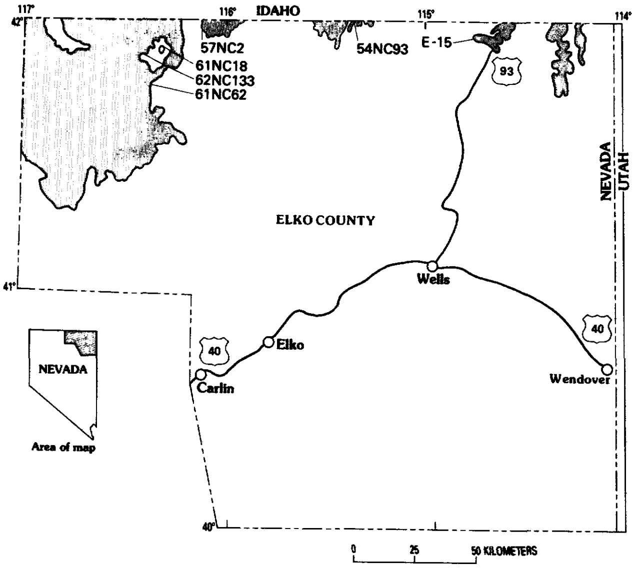

Figure 2. Map showing distribution of Big Island Formation in Elko County, Nev. (Hope and Coats, 1976), and location of collection of analyzed samples.

Lower Unit

The gravel and boulder gravel, which constitute the lower unit, differ in lithology, grain size, and thickness from place to place. In the type section, the clasts range from pebble to boulder size, with a maximum diameter of about 0.5 m. The clasts were derived from both the Jarbidge Rhyolite (middle Miocene) and the Cougar Point Welded Tuff (late Miocene). The unit is also exposed in the Hat Peak quadrangle about 76 km farther west where it had been assigned to the Banbury Formation by Coats (1968). At this locality it includes unconsolidated gravel consisting mostly of chert and other resistant rocks derived almost exclusively from Paleozoic rocks, like those exposed in the Owyhee quadrangle, southeast of the Hat Peak quadrangle, and with a maximum size of about 3 cm.

Middle Unit

The middle unit of the Big Island Formation consists of rhyolitic tuff. It is only sparsely exposed, either because of nondeposition or because of erosion prior to the extrusion of the basalt. In places where it does not crop out, the tuff may be largely concealed by talus from the overlying basalt; this is believed to be true in the type section. Tuff is present in several places beneath the basalt on the west side of Buck Creek Canyon, just south of the type section, and in the Jarbidge River Canyon, near the margin of the basalt. Outcrops would be even scarcer were it not that the upper part of the tuff has been hardened by heat from the overlying basalt flow.

Most of the tuff is exceedingly fine grained, recognizable fragments and crystals averaging less than 0.1 mm. In general, the tuff is dacitic, as quartz is almost omnipresent. The plagioclase has a wide range of composition, from oligoclase to bytownite. Sanidine is scarce; augite is the most common mafic mineral, and biotite and hornblende are less common. Some of the hornblende, which ranges from colorless to blue green, may be from metamorphic rocks in an unknown source area. Reddening of the biotite and hornblende indicate that the baked tuffs were heated by overlying flows.

Upper Unit

The basalt flows that form the upper unit are uniform in character. Just south of the type section (sample locality 54NC93), the three uppermost flows, in ascending order, are 5.6, 10, and 9 m thick. The rock here is nearly holocrystalline, with olivine phenocrysts as much as 3 mm in diameter, in a subophitic groundmass of pyroxene, plagioclase, granular olivine, magnetite, and ilmenite.

The pyroxene is a pale-brown augite with 2V of about 60°. The plagioclase crystals are zoned from An_{77} to about An_{52}, the most calcic. Groundmass olivine is interstitial to the ophitic pyroxene and about 0.1 mm in size. Also interstitial are opaque iron oxides, probably both magnetite and ilmenite, and apatite. Some of the rocks contain clots of early-formed plagioclase and olivine, the prevalence of which may result in slightly larger contents of MgO, CaO, and Al_2O_3, as in sample 54NC93, table 1.

In several places in the Owyhee desert, dikes of the basalt were recognized, and, in places where the extrusion of basalt continued for a longer than ordinary time, small hills were built up, none of which rise much more than 300 ft above the general land surface. These hills are either small shield volcanoes or small cinder cones that have been preserved for a long time because of the permeability of the material; they are the youngest volcanic features on the basaltic plateau.

PREVIOUS USAGE

The name Banbury Volcanics was applied by Stearns (1936, p. 435) to "* * * massive dark-brown weathered basalt flows and coarse and fine tuff beds * * *" exposed "* * * near Banbury Hot Springs, Twin Falls County, Idaho." Malde and Powers (1962, p. 1204) applied the name Banbury Basalt to its type area where they divided it into three parts. "* * * The lowest part consists of at least 400 feet of greatly decomposed olivine basalt that usually occurs as amygdular lava flows about 15 feet thick, although some makes columnar flows about 30 feet thick. The middle part, about 100 feet thick, consists mainly of brownish sand and pebble gravel in lenticular stream deposits but includes some light-colored clay, silt, and diatomite in lake deposits, as well as beds of siliceous volcanic ash. The upper part, in places 500 feet thick, is composed mainly of olivine basalt and some porphyritic plagioclase-olivine basalt in columnar lava flows as much as 50 feet thick. Thin lenticular beds of silt and sand locally separate the flows. The upper part, although considerably weathered, is less decomposed than the lower part." Malde and Powers (1962) point out that an area of basalt, also referred by them to the Banbury Basalt, crops out "* * * about 15 miles south of the Snake River, between Bruneau and Hagerman * * * to form a basalt plateau that extends southward about 40 miles to the foothills of the Jarbidge Mountains."

The name Banbury Formation was used by Coats (1964, p. M16) in the Jarbidge quadrangle, Nevada, in preference to Banbury Volcanics or Banbury Basalt, because of the large amount of sedimentary material that he included in the unit at the distal edge of the plateau referred to above by Malde and Powers. The basalt of the type Banbury at Banbury Hot Springs was not mapped by Malde and Powers (1972), in their definitive map of

Table 1. Major-element chemistry of basalt Big Island Formation, Nevada
[Analyst: Edythe Engelman, (standard chemical methods)]

	61NC62	E-15	57NC2	54NC93	61NC18	62NC133
SiO_2	48.19	48.00	47.99	48.11	47.88	48.41
Al_2O_3	15.59	16.96	16.45	16.81	15.68	15.25
Fe_2O_3	2.99	2.41	1.68	1.36	2.68	1.92
FeO	9.91	7.48	9.42	8.32	8.93	8.37
MgO	6.64	8.80	7.83	9.65	8.34	8.77
CaO	9.90	12.02	10.66	11.98	10.70	11.50
Na_2O	2.44	2.09	2.44	2.13	2.44	2.10
K_2O	.59	.16	.43	.15	.38	.38
H_2O^+	.48	.24	.35	.14	.39	.45
TiO_2	2.42	1.21	1.45	.96	1.74	1.43
P_2O_5	.47	.14	.25	.09	.26	.37
MnO	.21	.18	.20	.18	.19	.17
CO_2	.01	.01	.03	—	.01	.15
Cl	.01	.01	.03	—	—	.01
F	.05	.02	.03	.02	.03	.04
Subtotal	99.90	99.73	99.21	99.90	99.65	99.32
Less 0 = F	.02	.01	.01	.01	.01	.02
Total	99.88	99.72	99.20	99.89	99.64	99.62
CIPW norms						
Q	.001	.000	.000	.000	.000	.000
or	3.49	.948	2.56	0.88	2.253	2.26
ab	20.593	17.66	20.81	18.04	20.72	17.82
an	29.91	36.56	32.92	35.90	30.817	31.31
hl	.017	.017	—	—	—	.02
di	13.15	17.930	15.13	18.59	16.76	18.41
di-wo	6.70	9.27	7.72	9.57	8.62	9.47
di-en	3.72	5.93	4.33	5.87	5.22	5.81
di-fs	2.73	2.73	3.08	3.15	2.93	3.12
hy	22.24	12.71	11.61	7.70	12.61	17.80
hy-en	12.84	8.71	6.78	5.00	8.07	11.58
hy-fs	9.40	4.00	4.83	2.69	4.53	6.22
ol	.000	7.75	10.69	14.70	8.57	5.13
ol-fo	—	5.14	5.99	9.23	5.29	3.22
ol-fa	—	2.60	4.70	5.46	3.28	1.91
wo	.00	.000	.000	.000	.000	.000
mt	4.34	3.50	2.46	1.97	3.90	2.80
il	4.60	2.30	2.78	1.83	3.32	2.73
ap	1.11	.33	.60	.21	.61	.88
cc	.02	.02	.07	—	.02	.34
fr	.02	.02	.02	.03	.01	.01
Total	99.50	99.75	99.64	99.85	99.60	99.53
salic	54.01	55.18	56.30	54.83	53.79	51.41
femic	45.49	44.57	43.34	45.03	45.81	48.12

Table 1. Continued

	61NC62	E-15	57NC2	54NC93	61NC18	62NC133
			Barth cations			
Si	45.68	44.65	45.10	44.38	44.89	45.45
Al	17.42	18.59	18.22	18.28	17.33	16.87
Fe^{+3}	2.13	1.69	1.19	0.94	1.89	1.36
Fe^{+2}	7.86	5.82	7.40	6.42	7.00	6.57
Mg	9.38	12.20	10.97	13.27	11.65	12.27
Ca	10.05	11.98	10.73	11.84	10.75	11.57
Na	4.48	3.77	4.45	3.81	4.44	3.82
K	.71	.19	.52	.18	.45	.46
H	3.03	1.49	2.19	0.86	2.44	2.82
Ti	1.73	.85	1.02	.67	1.23	1.01
P	.38	.11	.20	.07	.21	.29
Mn	.17	.14	.16	.14	.15	.14
C	.01	.01	—	—	.01	.19
Cl	.02	.02	—	—	.00	.02
F	.15	.06	.09	.06	.09	.12
			Niggli values			
Al*	21.29	21.57	21.67	20.89	20.36	19.86
FM*	47.77	46.05	46.90	47.49	48.64	47.87
C*	24.58	27.79	25.53	27.07	25.26	27.23
Alk*	6.35	4.59	5.90	4.56	5.75	5.04
Si	111.68	103.58	107.27	101.46	105.49	106.99
ti	4.22	1.96	2.44	1.52	2.88	2.38
P	.46	.13	.24	.08	.24	.35
h	3.71	1.73	2.61	.98	2.87	3.32
k	.14	.05	.10	.04	.09	.11
mg	.48	.61	.56	.64	.56	.60
Si'	125.42	118.37	123.60	118.23	122.98	120.14
Qz	-13.73	-14.79	-16.33	-16.77	-17.49	-13.15
Differentiation Index	24.085	18.607	23.372	18.929	22.972	20.078

Location of analyzed samples (Nevada E. zone State plane coordinates)

	Coordinates		
Sample number	East	North	Quadrangle
E-15 (collected by R. A. Hope)	753,400	2,625,300	Delaplain
54NC93	545,700	2,638,700	Jarbidge
57NC2	366,800	2,639,600	Owyhee
61NC18	281,000	2,604,250	Hat Peak
61NC62	290,900	2,576,700	Hat Peak
62NC133	271,700	2,605,800	Hat Peak

this area, as being continuous with the plateau beginning "* * * about 15 miles south of the Snake River, between Bruneau and Hagerman."

K-Ar dates reported by Armstrong, Leeman, and Malde (1975, p. 238) for "* * * fresh-looking Banbury from the type locality are 4.4±0.6 and 4.9±0.6 m.y. * * *. The rocks dated are bracketed by 4.4 m.y. for the overlying Lucerne School lava flow and 6.25±0.13 m.y. for feldspar from the Idavada Volcanics at Shoshone Falls."

In 1980, Armstrong, Harakal, and Neill (1980, p. 7) published a redetermined age on supposed Banbury Basalt from the Mount Bennett Hills, originally dated at 13.5±1.5 m.y. (Armstrong and others, 1975, p. 230); the redetermination gave an age of 8.1±0.7 m.y. On the basis of these dates (and others?), Armstrong, Harakal, and Neill (1980) estimate the age of the Banbury as approximately 9.4 m.y. because of ages as young as 9.6 m.y. for the underlying Idavada Volcanics. They did not resolve the problem of the 6.25±0.13 m.y. date (Armstrong and others, 1975, p. 231) for the Idavada Volcanics at Shoshone Falls, which underlies basalt traceable into the type Banbury, nor with the Banbury Basalt overlying the rhyolite of Magic Reservoir; according to Armstrong, Leeman, and Malde (1975, p. 238), an intrusion "* * * considered to be related to the rhyolite * * *" was dated 3.06±0.04 m.y. on feldspar. The discrepancies that still remain on the age of the so-called Banbury lead one to agree with Armstrong, Leeman, and Malde (1975, p. 238) that "* * * as surmised by Malde, Powers, and Marshall (1963), the name Banbury has been used to designate basalts of several ages sandwiched between older silicic volcanic rocks and younger units * * *. The Banbury Basalt does not, therefore, represent a formation of limited stratigraphic position nor a formation of limited range in age."

This paper names and defines a formation, largely basalt, that includes part of what has been previously called Banbury. The Banbury Basalt of the type locality was not mapped by Malde, Powers, and Marshall (1963) as being continuous with the great area of basalt that surfaces the plateau of southern Twin Falls and Owyhee Counties, in some places reaching to, or even south of, the Nevada-Idaho State line. All of the samples described in this paper were collected from the Nevada part of this basalt field. The chemical and petrographic characteristics of the basalt are quite coherent, and these basalts, including those previously assigned to the Banbury Formation in the Jarbidge quadrangle by Coats (1964), are herein reassigned to the Big Island Formation. The age of the Big Island Formation is considered late Miocene on the basis of its stratigraphic position above the late Miocene Cougar Point Welded Tuff and radiometric age determinations of samples of its basalt (table 2), as those dates compare with those employed in the 1983 Geologic Time Scale (Palmer, 1983, p. 504). The age span represented by these samples is small compared with the total age span of all rocks previously assigned to the Banbury. Information is not available to permit adequate consideration of the whole range of ages and compositions of rocks that have been mapped as Banbury Basalt by various authors in various parts of Idaho. However, adoption of the name Big Island Formation for rocks in parts of Owyhee County, southern Idaho, and adjacent Elko County, northeastern Nevada, results in restriction of the name Banbury from these areas.

Table 2. Age of samples of basalt from the Big Island Formation K-Ar determinations by E. H. McKee, in Mark and others, 1975

Sample number	Age (m.y.) ±
E-15	8.2±0.6
54NC93	7.9±0.5
61NC18	10.6±1.0

In table 3 below (quoted from Mark and others, 1975, table 3), the potassium, rubidium, and strontium contents and the strontium isotopic ratios, as measured by isotope dilution, are listed for four samples of basalt collected from the Big Island Formation. As Mark and others (1975, p. 1673) point out, the basalt generally contains less potassium, rubidium, and strontium than do the tholeiites from the Snake River Plain (Leeman and Manton, 1971), including the basalts of the type locality of the Banbury Basalt.

Table 3. Potassium, rubidium, and strontium concentrations (ppm) and strontium isotopic ratios of four samples of basalt from the Big Island Formation measured by isotope dilution mass spectroscopy

[Concentrations in parts per million. Sr isotopic ratios are normalized to $^{86}Sr/^{88}Sr$=0.1194 and adjusted to a value of 0.71014 for NBS SRM 987 (0.7080 for Eimer and Amend $SrCO_3$ standard). As a result of low Rb/Sr ratios and ages, the isotopic ratios have not been corrected for growth of ^{87}Sr since eruption]

	E-15	54NC93	61NC18	62NC133
K	1,462	1,246	3,119	3,366
Rb	2.63	1.92	3.88	6.20
Sr	183	139	239	239
K/Rb	556	649	530	543
Rb/Sr	.0144	.0138	.0246	.0259
$^{87}Sr/^{86}Sr$	.7076±1	.7064±1	.7056±3	.7069±1 (±2σ)

PETROCHEMISTRY OF BASALT OF BIG ISLAND FORMATION

In table 1 are summarized the chemical analyses of six samples of basalt from the Big Island Formation, collected in northern Elko County. For comparative purposes, the CIPW norms, Barth cation numbers, Niggli numbers, and the differentiation indices have been calculated for each of these. An ALK-F-M triangular diagram (fig. 3) shows the chemical coherence of the six basalts, which cover an age span of about 1 to 4 m.y. (table 2) and extend over nearly 130 km. The circle on the F-M boundary represents the molar ratio of the olivine that is present in these rocks, as determined petrographically. All of these rocks could have been derived from a common ancestral magma by the separation of olivine slightly more magnesian than that present in the rocks now. The chemical analyses, however, indicate that such crystallization differentiation must have been slight. A triangular diagram of normative Or-Ab-An values (fig. 4), recalculated to total 100 percent, indicates how little change in the normative feldspar composition spans the range of variation in these rocks.

The sum of the alkalies has been plotted against silica content (fig. 5) in weight percent for the same six rocks. The diagonal line trending upward to the right is the line used by Macdonald and Katsura (1964) to separate analyses of the tholeiitic rocks below the line from those of the alkali basalts above. Each sample is represented by the same symbol in each of the diagrams.

Leeman and Manton (1971, p. 428) give rubidium and strontium values for samples of so-called Banbury Basalt from five localities in the Snake River Plain (table 4, this report). The first three of these were collected from the type area of the Banbury Basalt, the last two from basaltic units that may correlate with the Big Island Formation.

Potassium, rubidium, and strontium contents of olivine tholeiites in this region may afford useful criteria for distinguishing the basalts here assigned to the Big Island Formation from the younger basalts of the Banbury Basalt.

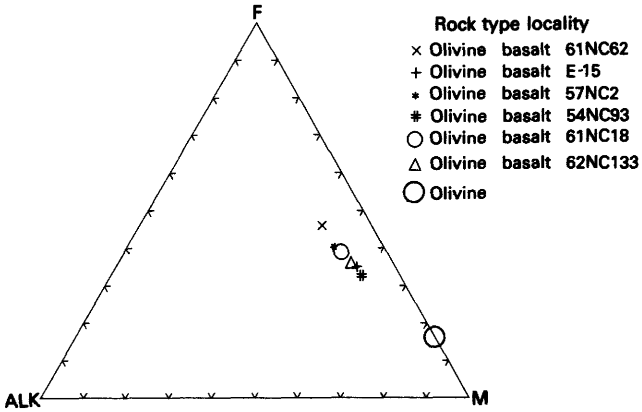

Figure 3. ALK-F-M diagram of six samples of basalt from the Big Island Formation.

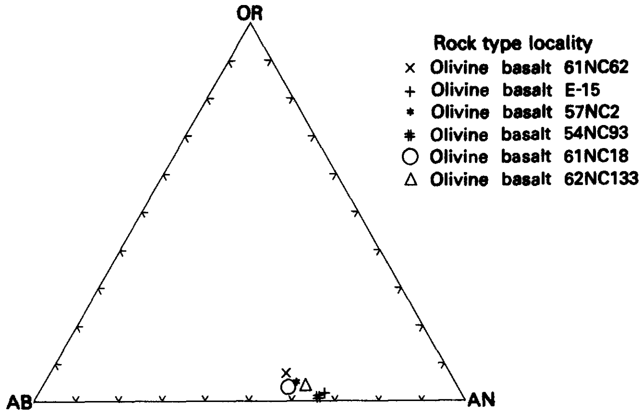

Figure 4. Normative Or-Ab-An values in six samples of basalt from the Big Island Formation, recalculated to a 100-percent basis.

Table 4. Rubidium and strontium contents for samples of basalt from five localities of the so-called Banbury Basalt in the Snake River Plain, Idaho

[Contents in parts per million. From Leeman and Manton (1971), table 3]

Sample	Description	Location: Latitude	Location: Longitude	Rb	Sr
D-1654	Upper part of Banbury Basalt.	42°43'	114°51'	8.2	325
D-1653	Lower part of Banbury Basalt.	42°44'	114°51'	6.7	285
A-196	Lower part of Banbury Basalt.	42°54'	115°00'	6.4	310
WPL-80	Banbury Basalt, near Murphy, Idaho.	43°10'	116°30'	7.7	210
P-37	Banbury Basalt, near Roseworth, Idaho.	42°25'	114°58'	5.8	205

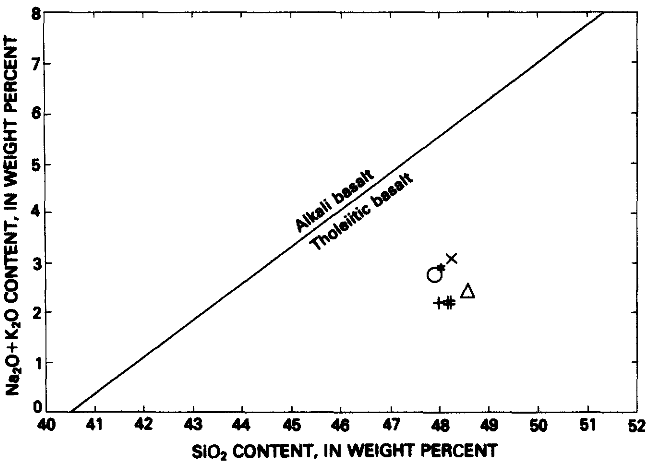

Figure 5. Graph showing sum of alkali oxides plotted against silica content in weight percent for samples of basalt from the Big Island Formation.

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REVISED MISSISSIPPIAN TIME SCALE, WESTERN INTERIOR REGION, CONTERMINOUS UNITED STATES

By William J. Sando[1]

Abstract

Integration of biozones based on foraminifers, conodonts, and corals in the Mississippian of the western interior region provides an excellent biometric base for this segment of the Phanerozoic time scale. The biometric scale now can be calibrated radiometrically with greater apparent precision because of recent revision of the radiometric scale. A composite biozonation consisting of 27 zones that have resolutions ranging from 0.2 to 4.9 m.y. provides a better biostratigraphic capability than do any of the biozonations based on one fossil group. Problems of analytical and fitting errors, paucity of North American radiometric ages, and radiometric weighting of foraminifer zones used as a standard for calibration of the other biozones should be the focal points of future work.

INTRODUCTION

The richly fossiliferous, predominantly carbonate Mississippian rock sequence of the western interior region affords excellent opportunities for biozonation and biochronometry. During Mississippian time, sedimentation took place over a large area between the Antler Orogenic highlands on the west and the Transcontinental arch on the east (fig. 1). The area of sedimentation was differentiated into a foreland basin or miogeosyncline on the west and a cratonic platform on the east (Sando, 1976; Gutschick and others, 1980; Sandberg and others, 1982). The line of demarcation between these areas marked the approximate location of a shelf margin during most of the Mississippian.

Intensive studies, begun in 1957, of the distribution of corals by the writer, of brachiopods by J. T. Dutro, Jr., and of foraminifers by B. L. Mamet led to a twofold biozonation of the western interior Mississippian (Sando and others, 1969). This biochronometric scale was later calibrated radiometrically (Sando, 1975, 1980) by using data from the British Phanerozoic Time Scale Symposium (Francis and Woodland, 1964) and its supplement (Lambert, 1971). At that time, only three radiometric checkpoints were available for calibration of the Mississippian (fig. 2).

Recent revisions of the geologic time scale (Harland and others, 1982; Odin, 1982) are based on many more radiometric checkpoints for the Phanerozoic and provide a total of 30 ages useful for calibrating Mississippian rocks. The revision by Harland and others (1982) is more useful for calibrating the western interior Mississippian scale because their revision presents more dated biostratigraphic boundaries that can be correlated to boundaries in North America.

Since 1975, western interior biozonation (Sando and others, 1969) also has been improved significantly. A system of conodont zones has been added to the previously recognized foraminifer and coral-brachiopod zonations, largely through the work of C. A. Sandberg (in Sando and others, 1981, and in Sandberg and others, 1982) and, more recently, by B. R. Wardlaw (unpublished data, 1984). The coral zonation has been revised by Sando and Bamber (1984, 1985). This threefold zonation system (foraminifers, conodonts, and corals) has been strengthened by analysis of many new samples from carefully studied stratigraphic sections.

This report revises the western interior Mississippian time scale, estimates the radiometric durations of Mississippian biozones more accurately and assesses their relative time values, develops and evaluates a composite biozonation, and points out problems with the time scale and possible solutions to these. Validity of the dating of biostratigraphic boundaries in the British radiometric scale (Harland and others, 1982) is a basic first assumption in the following radiometric calibration of the Mississippian biometric scale for the western interior. The precision of durations of biozones derived from this calibration is no greater than that of the primary data incorporated in the British scale. Consequently, all radiometric values derived in this report are only estimates.

ACKNOWLEDGMENTS

I am grateful to T. W. Henry and J. T. Dutro for their helpful comments on an early version of this report. Critical reviews by B. L. Mamet, J. D. Obradovich, and J. I. Tracey, Jr., were very helpful in bringing to my attention the need to qualify the precision of conclusions derived from original data.

REVISION OF THE TIME SCALE

The first step in revision of the time scale (fig. 3) was correlation of European stages of the Lower Carboniferous with North American Mississippian provincial series. The correlation was necessary because only two North American radiometric dates were used for calibration of the North American Mississippian, and only European boundaries were directly calibrated radiometrically by Harland and others (1982, chart 2.6). My revision of the western interior time scale incorporates correlations based on foraminifers by Mamet (in Mamet and Skipp, 1970; written commun., 1983), which differ slightly from those of Harland and others (1982, chart 2.6). In the Mamet scheme, the top of the Kinderhookian equates approximately with the top of Tn2a of the Tournaisian, and the top of the Meramecian correlates with a level just below the top of V3b of the Visean. Mamet's correlations of the other major biostratigraphic boundaries are essentially the same as those of Harland and others (1982, chart 2.6).

The next step in the revision was the integration of the western interior biozones into the time scale. In figure 3, the Upper Devonian part of the chart is based on Sandberg's (in Sandberg and others, 1982, p. 694-695) determination of conodont zone durations in the Famennian, independent of recent radiometric constraints. The base of the Mississippian was placed at the base of Tn1b,

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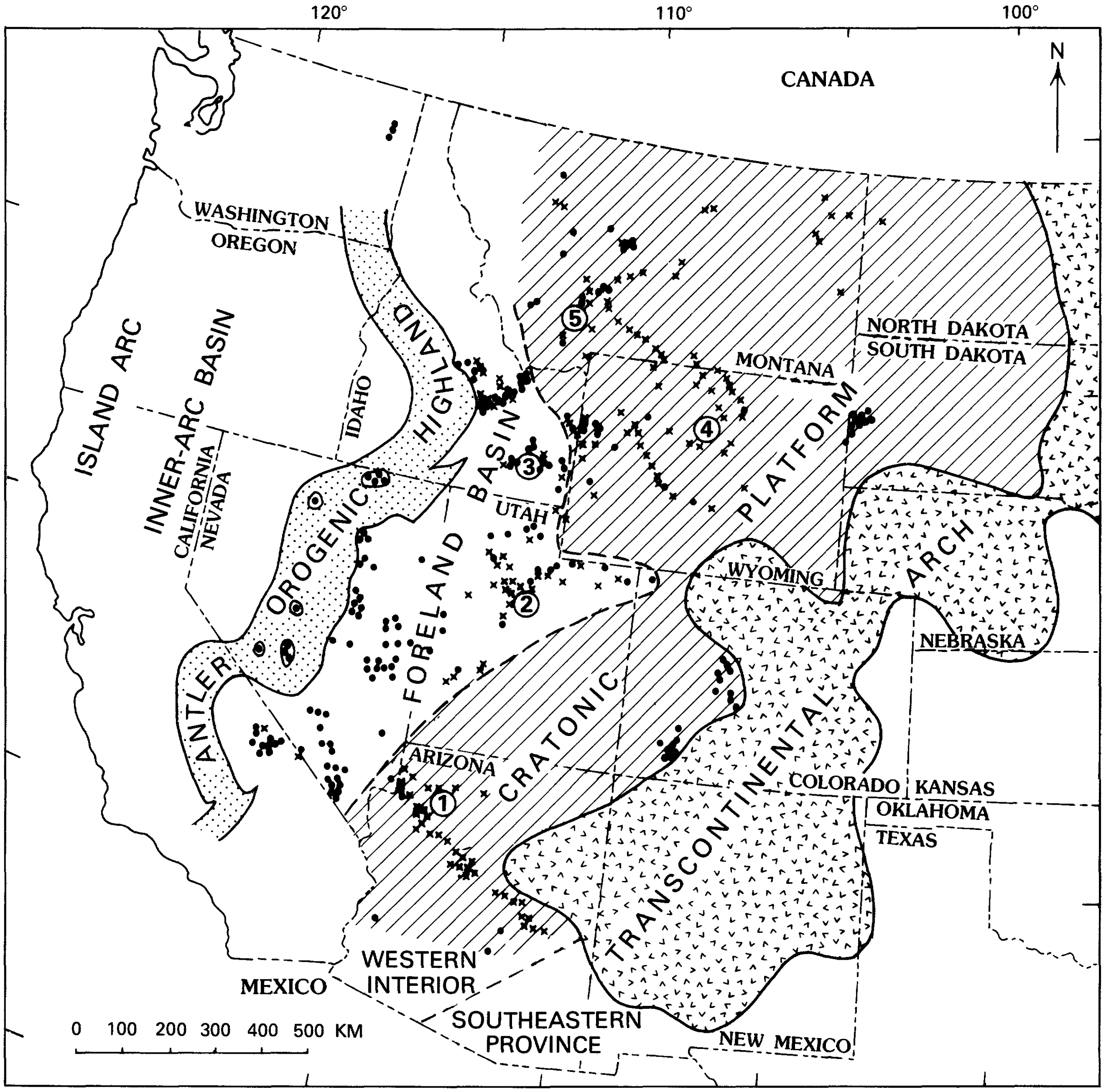

Figure 1. Major paleotectonic elements of the western interior region during Mississippian time (from Sando and Bamber, 1984). Dots and crosses indicate important control points for biozonation and biochronometry. Circled numbers indicate locations of selected stratigraphic sequences discussed in Sando and Bamber (1984).

which equates approximately with the base of the *Siphonodella sulcata* conodont zone at 360 m.y. B.P., the calibrated base of the Carboniferous of Harland and others (1982, chart 2.6). The top of the Kinderhookian was first calibrated at 355 m.y. B.P. by reference to Harland and others (1982, chart 2.6), but this calibration had to be adjusted to 355.5 m.y. B.P. to accommodate zonal correlations and zonal rock-thickness ratios. Kinderhookian conodont zones were fitted into the radiometric scale by assigning equal time durations to the lower four zones and twice that value to the upper zone, following Sandberg's (in Sandberg and others, 1982, fig. 4) rationale, but using a Kinderhookian duration of 4.5 m.y. instead of the 9-m.y. figure used by Sandberg.

The bases of Mamet Foraminifer Zones pre-7 and 7 were then fitted into the time scale by correlation with the upper two Kinderhookian conodont zones. Durations of the remaining foraminifer zones to the top of the Mississippian were calculated by determining the proportions of corresponding radiometrically calibrated total rock intervals occupied by these zones. This procedure provides a radiometrically calibrated foraminifer-zone scale to use as a standard for determining conodont and coral zone durations above the Kinderhookian. The radiometric top of the Meramecian (336.6 m.y. B.P.) also was determined by this construction.

The final step in the revision was the construction and radiometric calibration of a composite zonation (fig.

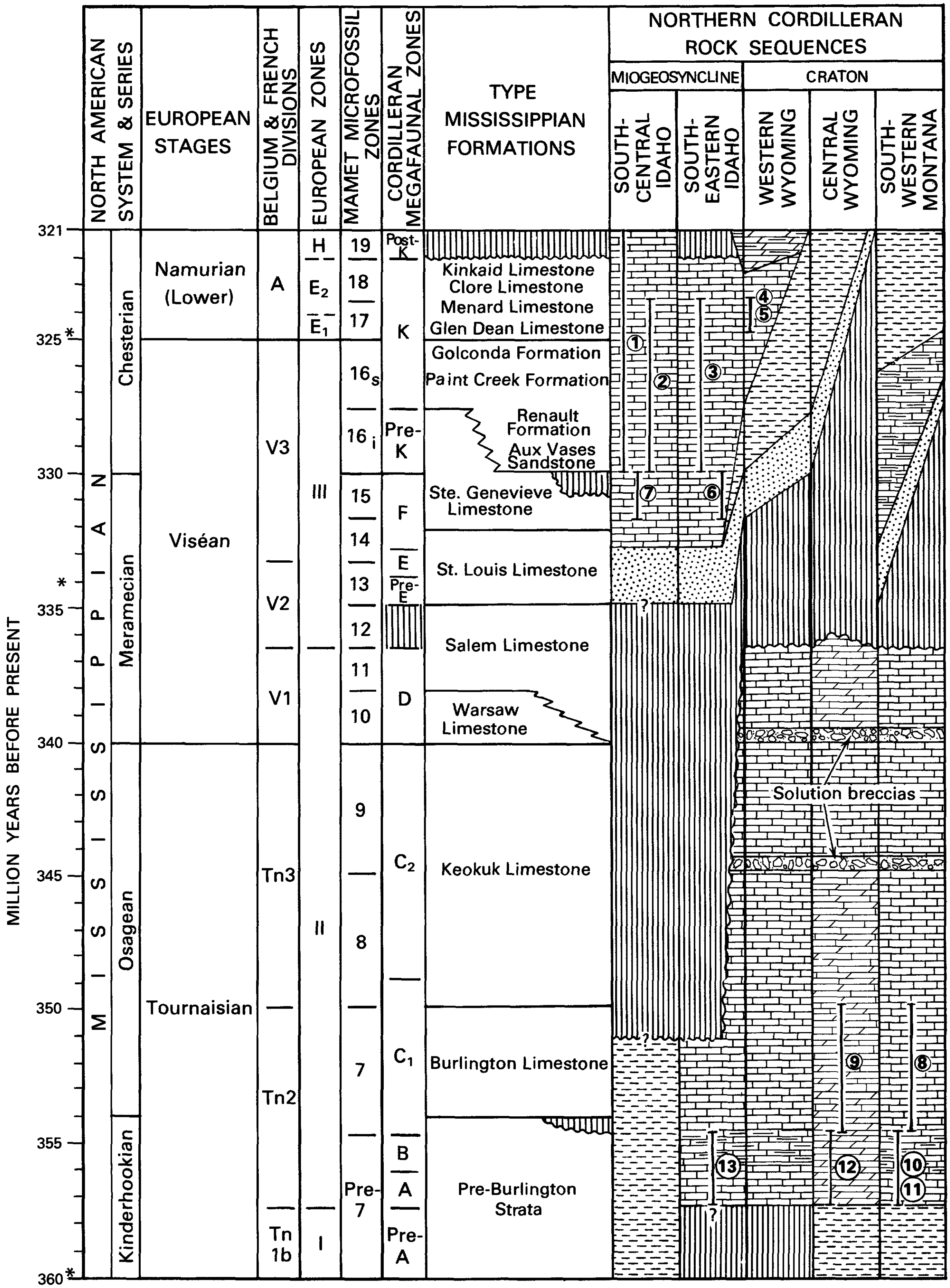

Figure 2. Previous radiometric calibration of twofold biozonation in the western interior region (Sando, 1975, 1980). Asterisks in radiometric scale mark checkpoints derived from radiometric data (Francis and Woodland, 1964; Lambert, 1971). Circled numbers refer to sequences discussed in Sando (1975, 1980).

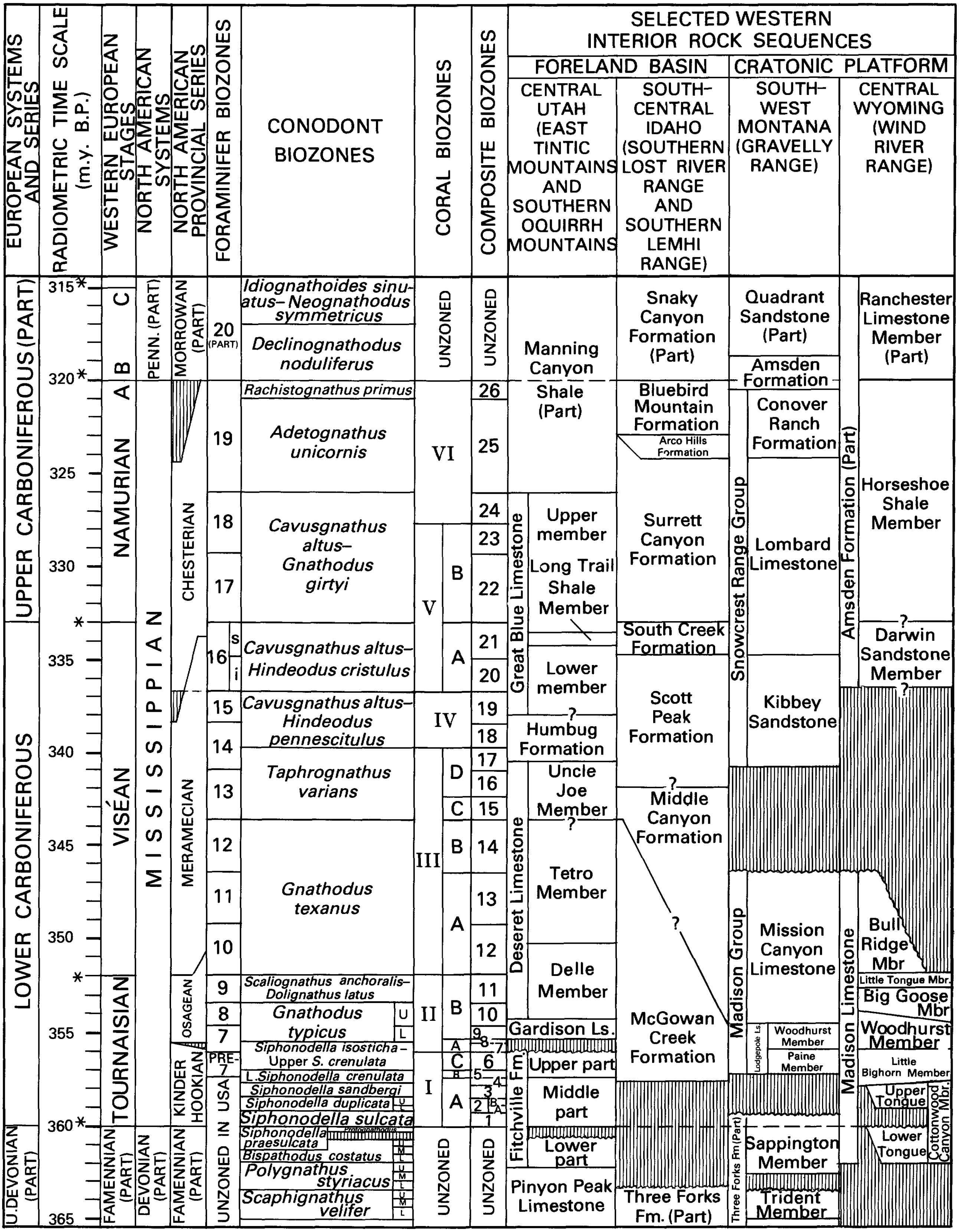

Figure 3. Revised western interior Mississippian time scale showing biozonations based on foraminifers, conodonts, and corals and showing correlation of stratigraphic units in selected western interior rock sequences. Asterisks in the radiometric scale mark radiometric checkpoints from Harland and others (1982). Conodont zones to top of Gnathodus texanus Zone are worldwide-standard first-occurrence zones used currently by C. A. Sandberg. Succeeding conodont zones are unpublished assemblage zones used currently by B. R. Wardlaw.

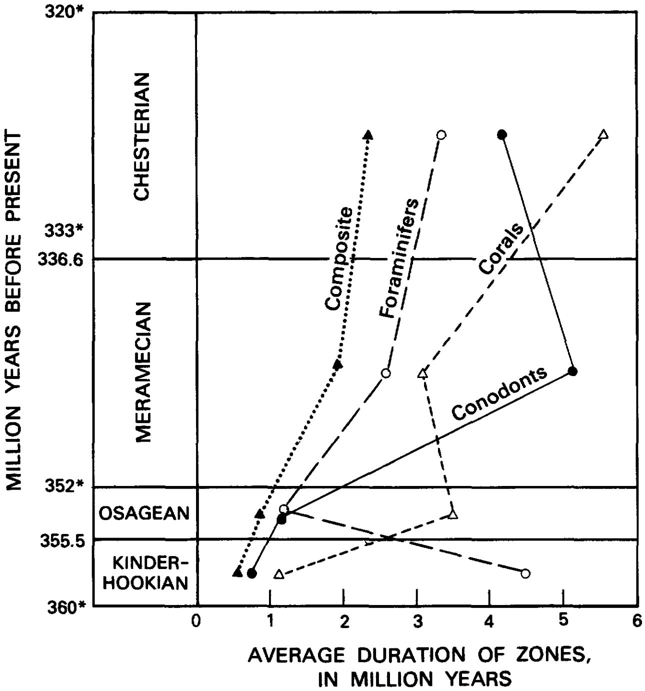

Figure 4. Estimated average durations of biozones in radiometrically calibrated North American provincial series in the western interior region. Asterisks mark radiometric checkpoints from Harland and others (1982).

3). This was accomplished by using all boundary positions of the three biozonations to define composite zones. The composite zonation consists of 27 biozones that can be recognized in places where some or all of the biozones can be determined.

Correlation of stratigraphic units in selected western interior rock sequences is shown on figure 3 to provide a lithostratigraphic perspective to the time scale.

BIOZONAL RESOLUTION

One of the interesting exercises made possible by radiometric calibration of the biozonations is the evaluation of the radiometric time resolution of the various biozones. Figure 4 shows average durations of the biozones ranging from 0.6 m.y. to 5.5 m.y. in radiometrically calibrated provincial series. Note that the conodonts reign supreme only in the Kinderhookian, which represents only a small fraction of the total Mississippian. The foraminifers are the best single group for resolution in the post-Kinderhookian time span. However, the composite zonation is by far the best tool; it is better than any of the single resolutions in all of the provincial series, having an average resolution of 0.6 to 2.4 m.y. All the resolution factors tend to diminish from older to younger levels of the Mississippian.

Analysis of durations of all the individual biozones (fig. 5) confirms the more general pattern of relative resolution capabilities of the different fossil groups. This analysis shows the dominance of conodonts as chronometric tools in the Kinderhookian, where their zonal durations range from 0.4 to 1.5 m.y. In the post-Kinderhookian interval, however, conodont zone resolution ranges from 0.9 m.y. in the Osagean to as much as 8.4 m.y. in the Meramecian. The foraminifer zones are better dating tools overall, having durations of an estimated 1.2 to 5.9 m.y. Coral resolutions are surprisingly good for fossils that are commonly regarded as poor chronometers, ranging from about 0.4 m.y. in the Kinderhookian to about 7.7 m.y. in the Chesterian.

Actually, comparison of the three groups to determine which is the best for chronometric resolution is not particularly constructive. The best chronometric capability is attained by using the composite biozonation, which affords a minimum resolution of about 4.9 m.y. in the Chesterian and ranges to a possible resolution of 0.2 m.y. in the Osagean.

PROBLEMS AND SOLUTIONS

Construction of the Mississippian time scale presented in this report revealed a number of problems that affect the actual precision of the scale. These problems are discussed below, with some possible solutions.

Analytical error and fitting errors

Estimated analytical errors in individual dates used to construct the Mississippian radiometric scale range from 3 m.y. to 20 m.y. (tables 1 and 2). Calculated durations of Mississippian provincial series range from 3.5 m.y. to 16.6 m.y. (fig. 3). Comparison indicates that estimated errors in ages used to construct the radiometric scale are of provincial series magnitude. Moreover, only 7 of the 30 available ages used to construct the radiometric scale are reasonably consonant with their biochronometric ages (tables 1 and 2). Although methods used to construct the radiometric scale may be statistically valid, more precision in the determination of these ages is needed, as well as more determinations that are well dated biometrically.

Critics of this calibration exercise (B. L. Mamet, written commun., 1984; J. D. Obradovich and J. I. Tracey, Jr., 1985) point out that derived radiometric durations of the biozones are much smaller than the possible analytical errors of the primary ages upon which the radiometric scale was constructed. Proration of these analytical errors helps to make the zonal duration estimates slightly more reliable but does not remove this valid criticism of the analysis. However, the alternative to using the best data available for such constructions is not even to attempt to calibrate the biometric scale, which might suggest that the radiometric scale is worthless.

Paucity of North American radiometric dates

The precision of the western interior scale is weakened by its reliance on controversial biostratigraphic correlations of radiometric checkpoints determined outside of North America. Only 4 of the 30 radiometric dates for the Mississippian are from North America, and none of these is consonant with its biometrically determined stratigraphic age (tables 1 and 2). As stated earlier (see p. A15), the correlations of radiometrically dated European biostratigraphic boundaries used in this paper differ from those of Harland and others (1982, chart 2.6).

Obviously, more reliable radiochronology for North American dates is needed, particularly from the western

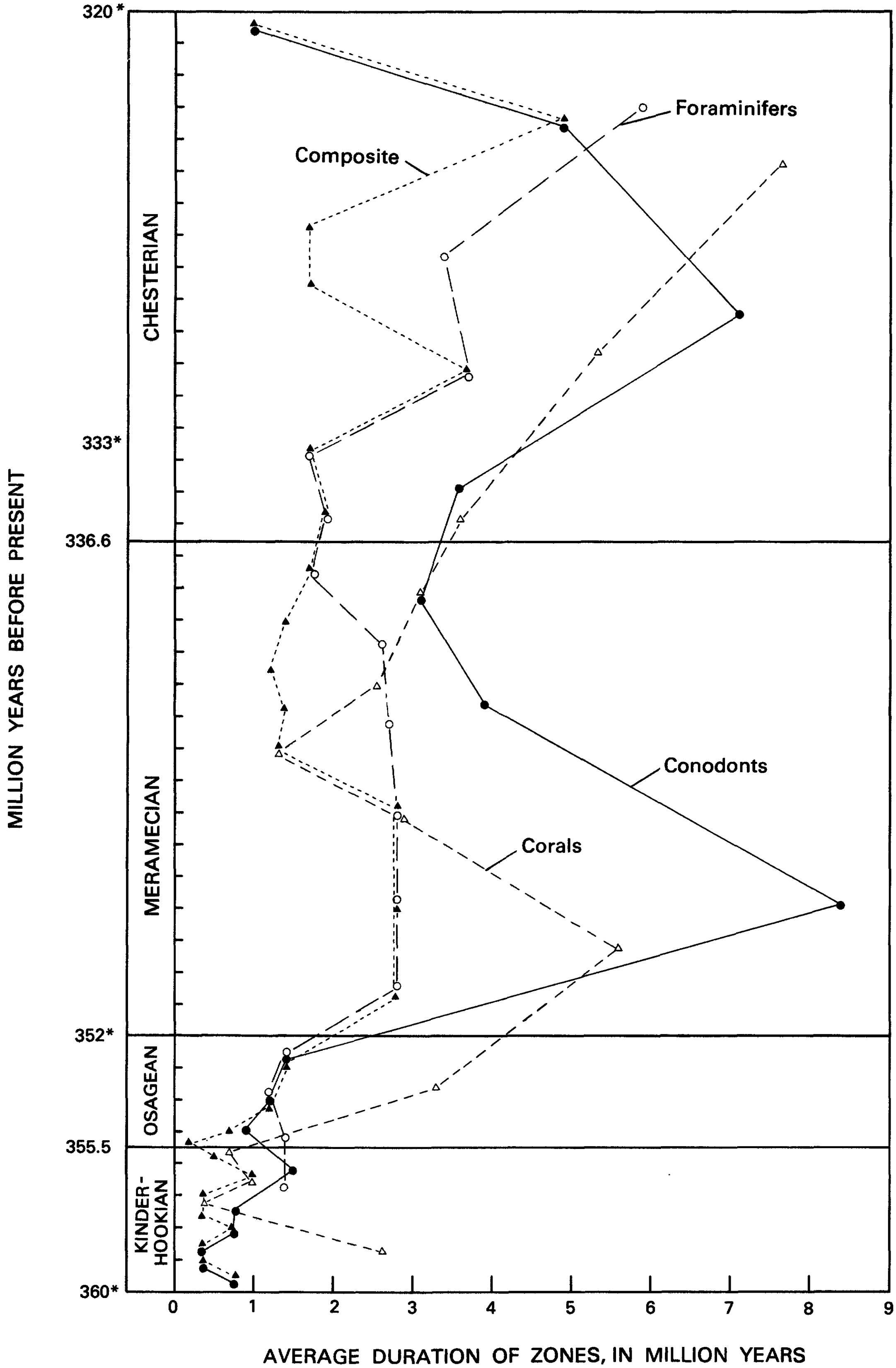

Figure 5. Estimated durations of individual biozones in the Mississippian of the western interior region. Asterisks mark radiometric checkpoints from Harland and others (1982).

interior region, so that the western interior time scale can be calibrated directly, without reliance on biostratigraphic correlations from localities elsewhere in the world. If we look at the ways in which radiometric ages were fitted into the Mississippian part of the time scale of Harland and others (1982), we find the following cases: (1) Seventeen (minimum) ages are from extrusive igneous rocks, predominantly basalts, interleaved with biometrically dated sedimentary rocks. (2) Nine (minimum) ages are from intrusive igneous rocks, predominantly granites, intruded into biometrically dated sedimentary rocks. (3) Four ages are from sedimentary rocks dated directly from contained glauconite (minimum, controversial) or uranium-lead minerals.

Prospects for radiometric dating of western interior sedimentary rocks by using minerals in intrusive and extrusive igneous rocks seem virtually nil. I know of no unmetamorphosed Mississippian volcanic rocks in the western interior, and Laramide plutonic events have overprinted any Mississippian plutonism. However, some western interior Mississippian rocks do afford opportunities for radiometric dating by means of contained glauconite or uranium minerals. These rocks are (1) the basal glauconitic limestone of the Kinderhookian Paine Member of the Lodgepole Limestone in Montana, Idaho, Wyoming, and Utah, (2) the glauconitic and sparsely uraniferous Osagean and Meramecian phosphatic shale member of the Deseret Limestone in Utah and equivalent rocks in Utah and Idaho, and (3) the very glauconitic, uraniferous upper tongue (Kinderhookian) of the Cottonwood Canyon Member of the Lodgepole Limestone and Madison Limestone in Wyoming and Montana. These units are all well dated biometrically (Gutschick and others, 1980) but have not been investigated radiometrically. The controversial nature of ages derived from glauconites and phosphates limits the precision attainable from these occurrences.

Elsewhere in North America, a cursory survey reveals several other opportunities for radiometric dating:

(1) *Volcanic ash and pyroclastic flows in the Chesterian part of the Stanley Shale of Oklahoma.*—Whole rock Rb-Sr analysis of samples from the Hatton Tuff Lentil of the Stanley gave an age of 310 m.y.±15 m.y. (Mose, 1969, p. 2374; Johnson, 1979, p. R23). Recalculation of this date, using the new Rb-Sr decay constant (Steiger and Jager, 1977), results in an age of 317 m.y.±15 m.y., which would place the samples in the Early Pennsylvanian (early Morrowan) on the revised time scale. Inasmuch as the Hatton is closely dated biometrically as early Chesterian (Gordon and Stone, 1977, p. 74), the Hatton age needs to be reinvestigated.

(2) *Volcaniclastic sediments in the Baird and Bragdon Formations in the eastern Klamath Mountains of northern California.*—Here, Watkins (1973, 1979) found *Striatifera* and other brachiopods of Late Mississippian age in a sequence of tuffs and tuffaceous terrigenous sedimentary rocks. No radiometric dates from these rocks have been published.

(3) *Volcaniclastics, intrusives, and flows in the Peale Formation in the northern Sierra Nevada of California.*—Brachiopods of Early Mississippian age occur at the base of the upper member of the Peale directly above silicic pyroclastic rocks at the top of the lower member in the Taylorsville area (McMath, 1966; D'Allura and Moores, 1979). No radiometric ages from these pyroclastics have been published, but a Pb-U age of 300 m.y. was obtained from an albitite in the Feather River peridotite belt immediately west of the Peale Formation outcrop belt (Weisenberg and Ave Lallemant, 1977).

(4) *Volcaniclastic sediments and flows interbedded with sedimentary rocks in the Upper and Lower Mississippian part of the Lisburne Group and in the Lower Mississippian Kayak Shale in the Brooks Range, northern Alaska.*—On the Ivishak River, breccia, tuff, volcanic conglomerate, tuffaceous limestone, and flows occur just above limestone containing probable Chesterian (Mamet Zone 18 or younger) foraminifers; furthermore, Mississippian brachiopods, bryozoans, and echinoderms occur in tuffaceous limestone within the volcanogenic sequence (Reiser and others, 1979). In the headwaters of the John River, tuff conglomerate, tuffaceous limestone, and possible basaltic andesite flows include limestone lentils that contain probable late Kinderhookian brachiopods (Reiser and others, 1979). Samples submitted for radiometric analysis were rejected because of inappropriate mineralogy, and no whole rock analyses were attempted (J. T. Dutro, Jr., oral commun., 1984).

(5) *Intrusive igneous rocks overlain unconformably by Upper Mississippian limestone belonging to the Lisburne Group in the Brooks Range, northern Alaska.*—In the Navy East Teshekpuk well, sandy limestone dated as late Chesterian (Mamet Zones 18-19) on the basis of foraminifers rests unconformably on granite dated as 332 m.y.±10 m.y. by radiometric analysis of potassium feldspar (Bird and others, 1978).

(6) *Extrusive rocks interbedded with Mississippian limestone in southeastern Alaska.*—In the Keku Islets, the Saginaw Bay Formation in its lower part includes limestone containing Mississippian fossils interbedded with tuffaceous sandstone and limestone, aquagene tuff, pillow flows, and basaltic flow breccia (Muffler, 1967, p. C20). In the Soda Bay area, the Peratrovich Formation, which contains Kinderhookian or early Osagean fossils in its lower part, conformably overlies a calcareous basaltic tuff at the top of the Port Refugio Formation, dated as Late Devonian on the basis of brachiopods (Eberlein and Churkin, 1970, p. 49-50). Neither of these localities has been investigated radiometrically (J. T. Dutro, Jr., oral commun., 1984).

(7) *A possible constraint on the top of the Mississippian obtainable from the Wise Formation (upper Morrowan or basal Middle Pennsylvanian) of Virginia, where Nelson (1959) reported a sanidine-bearing bentonite (J. D. Obradovich, 1985, written commun.).*—Possibly correlative beds in the Fire Clay coal bed in the Breathitt Formation of eastern Kentucky, where Seiders (1965) reported a sanidine-bearing flint clay of possible volcanic origin, may offer another opportunity for narrowing the dating of the Mississippian-Pennsylvanian boundary (J. D. Obradovich, written commun., 1985).

(8) *A bentonite bed in the Exshaw Shale near Nordegg, Alberta, that bears both sanidine and zircon (J. D. Obradovich, written commun., 1985).*—This bed, which may be of Late Devonian or Early Mississippian age, might be useful in dating the base of the Mississippian.

(9) *Basal Mississippian volcanic rocks on Cape Breton Island, Nova Scotia.*—These rocks also may provide an opportunity for dating the base of the Mississippian (J. D. Obradovich, written commun., 1985).

Radiometric weighting of Mamet foraminifer zones

A third problem in construction of the western interior Mississippian time scale is the determination of the radiometric durations of Mamet foraminifer zones used as a standard for calibration of the other biozones above the Kinderhookian. Available data did not permit

Table 1. Mississippian radiometric dates (320-360 m.y. B.P.) used by Harland and others (1982) for construction of the time scale

Reference number	Standardized age (m.y. B.P.)	Estimated error (m.y.)	Country	Geologic unit	Radiometrically determined stratigraphic age	Biometrically determined stratigraphic age
PTSS 339	321.89	10.00	U.S.S.R.	Keregetass Volcanics.	Namurian A	Namurian B-C to Stephanian.
PTS 191	328.16	12.00	Scotland	Hillhouse Basalt.	Early Namurian	Earliest Namurian.
PTS 171	331.27	8.50	France	Vosges Granite.	Early Namurian	Post-early Viséan, Pre-Stephanian.
PTS 31	333.23	8.50	Germany	Harzburger Gabbro.	Latest Viséan	Post-Viséan, Pre-Stephanian.
PTS 66	333.23	8.80	Australia	Lower Kuttung Lavas.	Latest Viséan	Namurian.
A 440	336.49	10.00	U.S.S.R.	Aksahut River Basalt.	Late Viséan	Famennian.
PTS 172	339.55	9.00	France	Tuff	Late Viséan	Late Viséan.
A 2360	340.32	17.00	Great Britain.	Little Wenlock Basalt.	Late Viséan	Westphalian.
PTS 174	341.36	8.20	East Germany.	Granitoid	Middle Viséan	Early Carboniferous.
B 1	341.62	6.00	France	Huisserie Formation.	Middle Viséan	Early Tournaisian.
A 3360	344.37	4.00	Great Britain.	Burntisland Basalt.	Middle Viséan	Westphalian.

PTS 173	345.76	7.00	France	Gren-sur-Cur Granite.	Early middle Viséan.	Latest early Viséan.
A 424	346.22	10.00	U.S.S.R.	Central Kazakhstan granite.	Early middle Viséan.	Tournaisian.
A 442	346.22	10.00	U.S.S.R.	Central Kazakhstan granite.	Early Middle Viséan.	Frasnian.
A 443	346.22	10.00	U.S.S.R.	Ugum Range granite.	Early middle Viséan.	Late Devonian.
PTS 2	346.33	10.00	U.S.A.	Chattanooga Shale.	Early middle Viséan.	Late Devonian.
PTS 94	346.40	6.00	U.S.A.	Chattanooga Shale.	Early middle Viséan.	Late Devonian.
A 4360	353.49	7.00	Great Britain.	Arthurs Seat Basalt.	Late Tournaisian	Viséan.
A 441	355.95	20.00	U.S.S.R.	Cahut River granite.	Middle Tournaisian	Late Devonian.
PTS 95	356.53	9.30	Australia	Snobs Creek Rhyodacite.	Middle Tournaisian	Early Carboniferous(?).
A 425	356.93	10.00	U.S.S.R.	Kursky district sedimentary rock.	Middle Tournaisian	Frasnian.
A 413	359.22	9.00	France	Massif Central granite.	Earliest Tournaisian.	Viséan.

Table 2. Mississippian radiometric dates (320-360 m.y. B.P.) used by Odin (1982) for construction of the time scale

Reference number	Standardized age (m.y. B.P.)	Estimated error (m.y.)	Country	Geologic unit	Radiometrically determined stratigraphic age	Biometrically determined stratigraphic age
NDS 230	320	8	Scotland	Machrihanish lavas.	Namurian A/B boundary.	Viséan.
NDS 167	326	7	Scotland	Clyde Plateau lavas.	Namurian A	Viséan.
NDS 133	331	6	France	Chateaulin Basin volcanics.	Early Namurian A	Early Viséan.
NDS 133	342	13	France	Laval Basin volcanics.	Middle Viséan	Early Tournaisian.
NDS 152	348	3	U.S.A.	Houy Formation.	Early Meramecian	Middle Kinderhookian.
NDS 166	353	7	Scotland	Garleton Hills lavas.	Late Tournaisian	Early Viséan.
NDS 153	354	3	U.S.A.	Barnett Formation.	Early Osagean	Middle Osagean.
NDS 165	361	12	Scotland	Birrenswark and Kelso lavas.	Latest Devonian	Late Devonian to early Tournaisian.

precise measurement of all Mamet zonal rock thicknesses because of incomplete sample control and because of the presence of unconformities in the rock sequence. Adjustments had to be made so that all the zones fit the radiometrically calibrated rock intervals. A possible solution to this problem lies in the Mississippian of western Canada, which provides an excellent, uninterrupted foraminiferal sequence (Mamet, 1976) where the rock thicknesses of all the Mamet zones can be determined more precisely.

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THE SURPRISE CANYON FORMATION—AN UPPER MISSISSIPPIAN AND LOWER PENNSYLVANIAN(?) ROCK UNIT IN THE GRAND CANYON, ARIZONA

By George H. Billingsley[1] and Stanley S. Beus[2]

Abstract

The Surprise Canyon Formation is a new name applied in this report to rocks previously considered to be part of the Redwall Limestone and the Watahomigi Formation, or part of a valley-fill sequence that post-dated the Redwall and predated the Watahomigi. The Surprise Canyon consists of fossiliferous nonmarine and marine sediments of Late Mississippian (Chesterian) and perhaps locally very early Pennsylvanian (Morrowan) age that fill large- and small-scale erosional valleys cut through the Horseshoe Mesa Member and into the upper part of the Mooney Falls Member of the Redwall Limestone in the Grand Canyon, Ariz. Strata of the Surprise Canyon Formation are lithologically distinct from and unconformable with the underlying Mississippian Redwall Limestone and the overlying Watahomigi Formation of the Pennsylvanian-Permian Supai Group. The formation is named for Surprise Canyon, a northern tributary canyon to the Colorado River in the western Grand Canyon. The type section is located near the Bat Tower viewpoint south of the Colorado River in the western Grand Canyon.

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INTRODUCTION

Observations in the Grand Canyon, 1975 through 1983, revealed the presence of scattered large- and small-scale erosional valleys cut through the Horseshoe Mesa Member, into the Mooney Falls Member of the Redwall Limestone, and filled with nonmarine and marine deposits (Billingsley, 1979; Billingsley and McKee, 1982; Beus and Billingsley, 1982). These strata record local deposition during part of the Late Mississippian and possibly very early Pennsylvanian, an interval previously considered a depositional gap in the Grand Canyon rock record.

The valley-fill strata, herein designated the Surprise Canyon Formation, form a mappable rock unit that is confined mainly to erosional valleys and shallow depressions but also occurs locally in caves within the Mooney Falls and Horseshoe Mesa Members of the Redwall Limestone. The formation is separated from the Watahomigi Formation by an erosional unconformity. The red-brown rocks are clearly distinguishable in lithology and color from the underlying gray Redwall Limestone and overlying purplish-gray Watahomigi Formation. The formation is named after Surprise Canyon (fig. 1), a large northern tributary canyon to the Colorado River in the western Grand Canyon (U.S. Geological Survey, 1967). Cross sections of the formation are exposed at five different localities in the walls of Surprise Canyon. An outcrop at long 113°48', lat 36°00', 1.6 km west of the Colorado River and 2.4 km southwest of the mouth of Tincanebits Canyon, in a small tributary canyon, was chosen as the type section because this locality contains one of the thickest and most continuously exposed sections of the formation (fig. 2) and is more accessible than other localities. Surprise Canyon is the closest unused geographic name to the type section (18 km to the east).

DISTRIBUTION

The Surprise Canyon Formation is not a continuous rock unit but rather crops out as isolated patches throughout much of the Grand Canyon and part of Marble Canyon to the northeast. The formation may occur as shoestring deposits beneath the Colorado Plateau adjacent to the Grand Canyon. It has not, however, been recognized in the subsurface. The valleys in which the formation occurs average about 300 m in width and 75 m in depth in the western Grand Canyon and become progressively shallower and wider toward the eastern Grand Canyon. Thickness of the formation essentially corresponds to the depth of the valleys in which it was deposited. In the western part of the Grand Canyon, the formation is generally between 60 and 100 m thick. The thickest section measured is 122 m at Quartermaster Canyon. In the central part of the Grand Canyon, the thickest exposures are about 45 m, and in the eastern Grand Canyon and Marble Canyon the maximum thickness does not exceed 25 m.

Most exposures of the Surprise Canyon are gently U-shaped or V-shaped lenses notched into the top of the Redwall Limestone. The lower layers of the valley-fill strata, along the axis of the major valleys, are everywhere deposited on the Mooney Falls Member of the Redwall Limestone because the host valleys cut entirely through the Horseshoe Mesa Member of the Redwall and into the Mooney Falls. The upper layers of the Surprise Canyon deposits are more widely distributed, in places spreading out as much as 0.5 km on either side of the main valley axis, and the strata rest on the Horseshoe Mesa Member of the Redwall (fig. 3).

Some outcrops of the formation in the western Grand Canyon are shown as Mississippian channels on geologic maps by Huntoon and others (1981, 1982) and Billingsley and Huntoon (1983). A single isolated outcrop of Chesterian age rock was recorded as an unnamed unit at the top of the Redwall Limestone near the Bright Angel Trail in the Grand Canyon (McKee and Gutschick, 1969, p. 74) and is shown as middle Chesterian age by Betty Skipp (1979, p. 298). This isolated outcrop is herein reassigned to the Surprise Canyon Formation because the outcrop is correlative to the upper limestone beds of the

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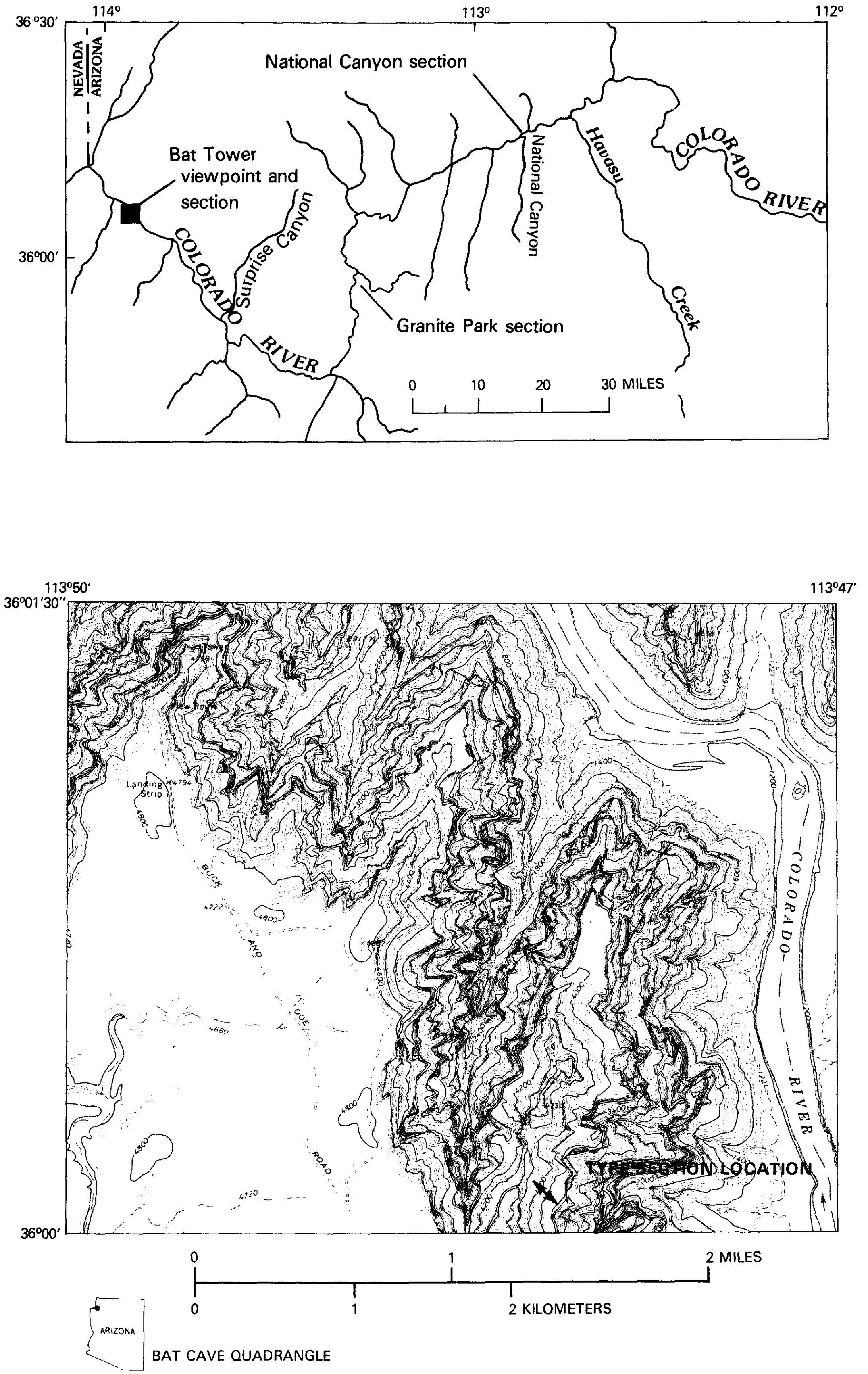

Figure 1. Location of the Surprise Canyon Formation type locality, near the Bat Tower viewpoint in the western Grand Canyon, Ariz.

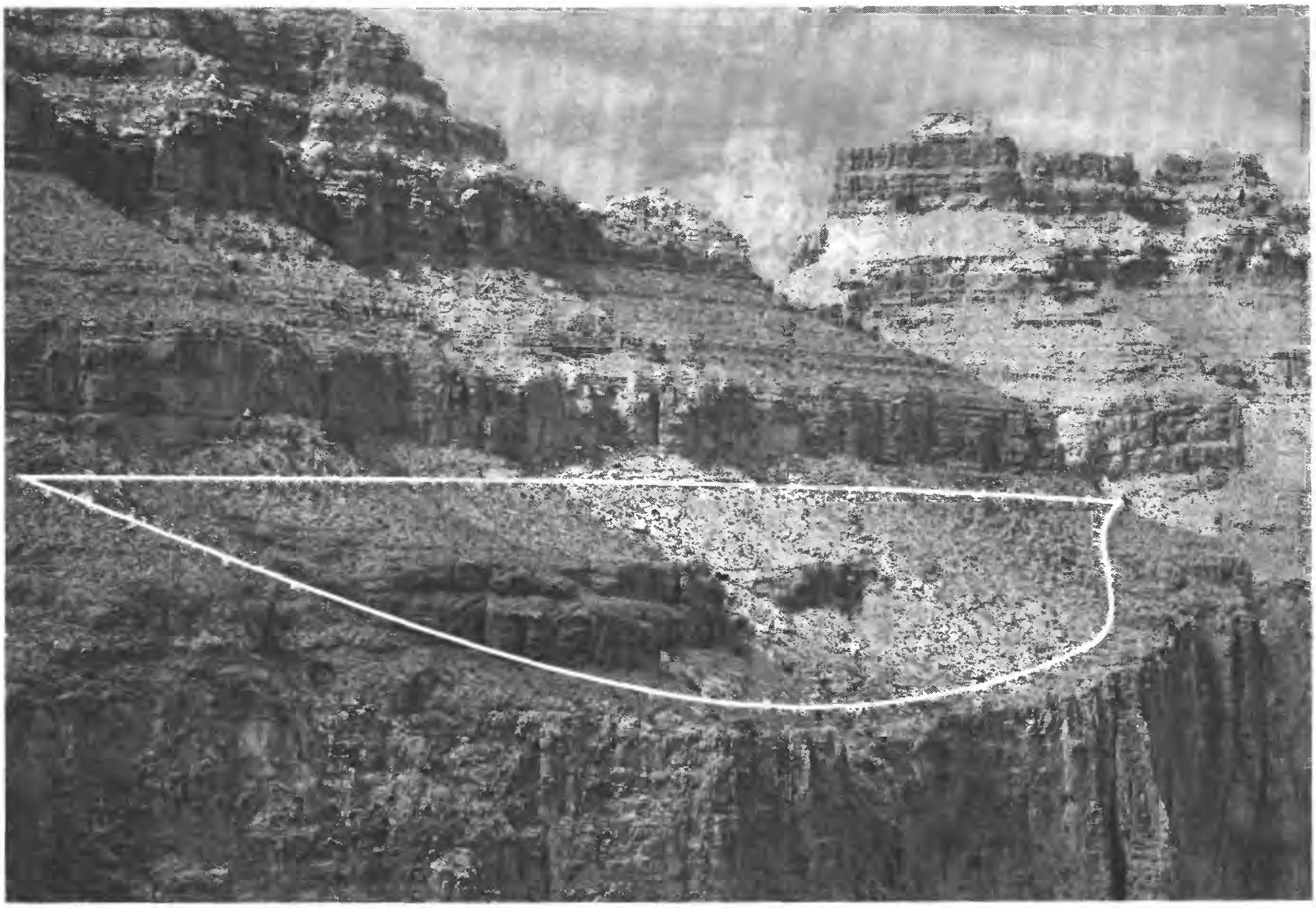

Figure 2. Type section of Surprise Canyon Formation, long 113°48', lat 36°00', 1.6 km west of the Colorado River (mi 263) and 2.4 km southwest of the mouth of Tincanebits Canyon, western Grand Canyon, Ariz.

Surprise Canyon Formation of the western Grand Canyon, as suggested by Skipp (1979, p. 298). The uppermost Redwall Limestone (Horseshoe Mesa Member), as herein recognized in the Grand Canyon, includes rocks no younger than Meramecian age.

The Watahomigi Formation (McKee, 1975) of middle Morrowan to Atokan age overlies the Redwall Limestone throughout the Grand Canyon. Locally, the basal Watahomigi beds on the Redwall surface, as described by McKee and Gutschick (1969, p. 76-80), are conglomerate beds or lenses of conglomerate within red slope-forming, gnarly bedded mudstones, from 0.9 to 12.2 m in thickness. These sediments are found also in solution caves within the Mooney Falls and Horseshoe Mesa Members of the Redwall. The conglomerate clasts are of resistant rock types such as chert and commonly include indigenous marine fossils (McKee, 1982, p. 39-46). The rocks described by McKee fit the lithologic description and stratigraphic position of the Surprise Canyon Formation rocks. Therefore, on the basis of lithology, color, and stratigraphic position, the lowermost conglomerate and gnarly mudstone deposited in shallow depressions and caverns within the Redwall Limestone are all reassigned to the Surprise Canyon Formation. The base of the Watahomigi, as herein redefined, is marked by a thin, widespread, locally discontinuous limestone pebble conglomerate that contains minor chert clasts. This conglomerate generally is overlain by a purplish-red calcareous siltstone and mudstone of the lower slope unit of the Watahomigi (fig. 3). This boundary is several meters below the Atokan boundary, as described by McKee (1975, p. J4).

LITHOLOGY

The Surprise Canyon Formation is subdivided into two units—a basal unit of terrigenous clastic sedimentary rocks and an upper unit of mostly marine carbonate and fine-grained clastic rocks (fig. 4). The basal strata in most localities consist of detrital sediments of cobble and pebble conglomerate and calcareous sandstone. This conglomerate varies in thickness, is massively bedded, and contains some interbedded sandstone layers and lenses. The conglomerate consists chiefly of white, yellow, and red chert pebbles and cobbles that range in diameter from 0.6 to 12.7 cm; locally the conglomerate includes limestone clasts. A few subrounded chert and limestone boulders as much as 1 m in diameter occur at some localities. The fossiliferous chert and limestone cobbles clearly are derived from the chert lenses and beds of the Redwall Limestone. The gravel clasts are in a matrix of poorly sorted, subrounded, coarse-grained quartz sand with siliceous, calcareous, and ferruginous cements.

The sandstone of the basal unit is dark gray to red brown, medium to coarse grained, and locally cross stratified. In many localities, the sandstone is inter-

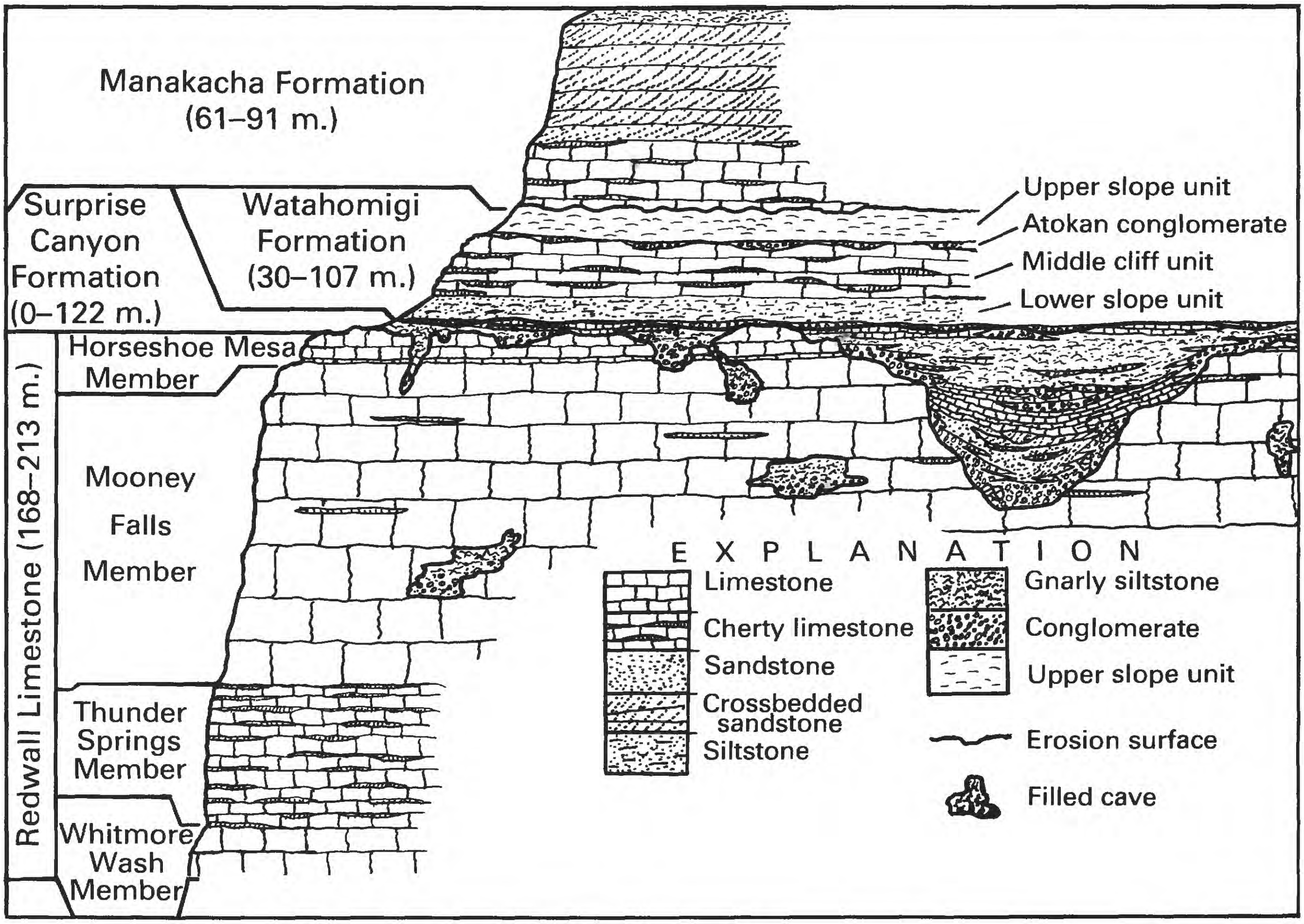

Figure 3. Cross-sectional view showing stratigraphic position and contacts of the Surprise Canyon Formation in relation to the Redwall Limestone (McKee and Gutschick, 1969, p. 78-80) and the Watahomigi Formation (McKee, 1982, p. 32).

bedded with medium- to fine-grained siltstone that has abundant carbonaceous plants, plant and wood fragments, and tree impressions, especially at Granite Park (figs. 1 and 4, subunits 2, 4, and 5). These terrestrial fossiliferous sandstones are exposed in most outcrops of the Surprise Canyon Formation throughout the Grand Canyon; they vary in thickness and grade upward into marine sediments of the upper unit. At several localities, an erosional unconformity marks the boundary between the lower units and the upper marine sediments.

The upper unit is thickest in the western Grand Canyon and gradually thins and pinches out to the east. The lowest marine strata in the upper unit consist of a brown to yellow-gray, coarse-grained, fossiliferous limestone that forms a cliff or series of steep ledges. The limestone is overlain by red-brown, thin-bedded, ripple-

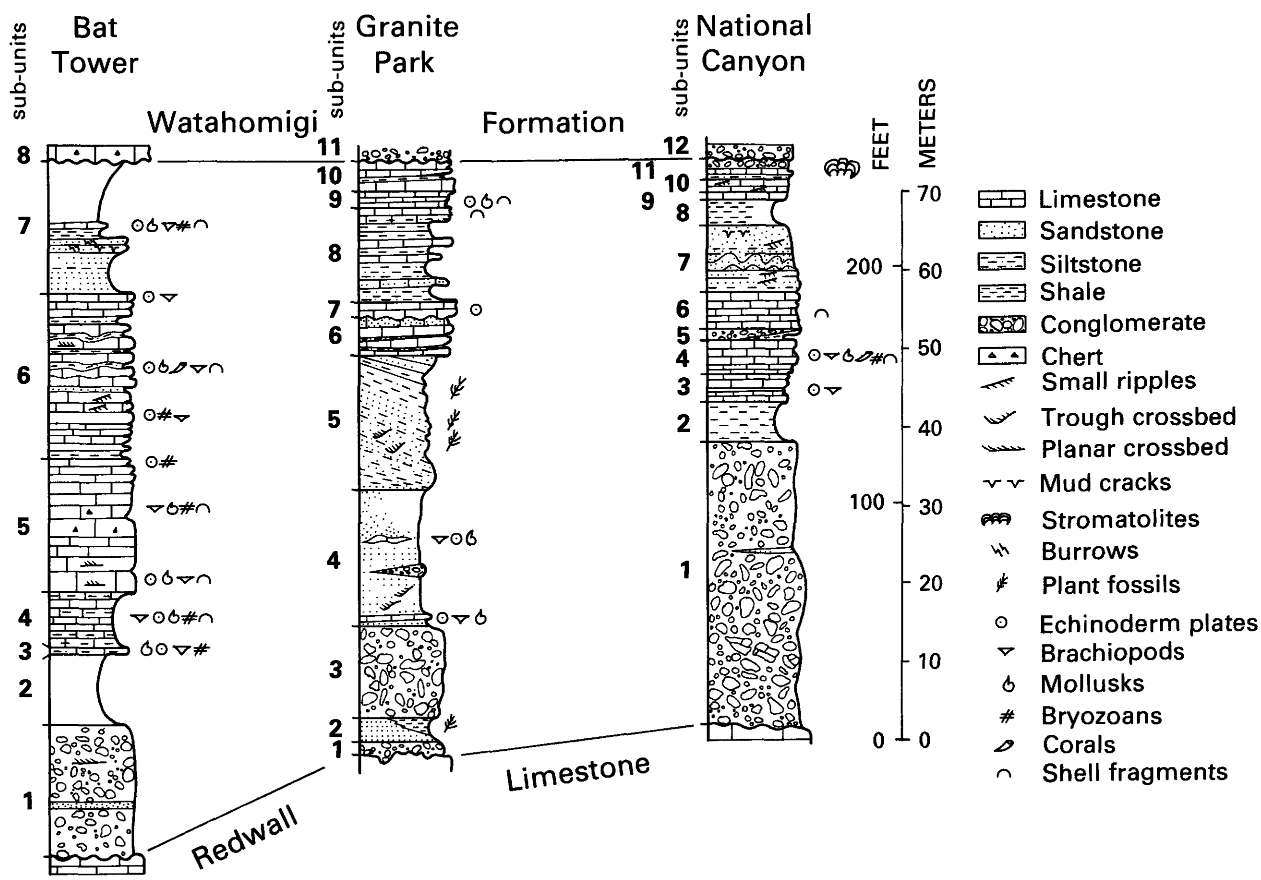

Figure 4. Columnar sections of the Surprise Canyon Formation at Bat Tower viewpoint, Granite Park, and National Canyon localities, the western Grand Canyon, Ariz. See section "Description of type locality" for detailed description of subunits 1-7.

laminated sandstone and siltstone that generally form slopes, except in the National Canyon area where they form a cliff, (National Canyon, subunit 5, fig. 4). In most areas of the western and central Grand Canyon, but not the eastern Grand Canyon, an upper ledge of gray, coarse-grained limestone occurs above the sandstone. Stromatolites occur locally in the upper limestone. In the western Grand Canyon area, this sequence generally weathers into two weak cliffs separated by a slope. In the central Grand Canyon, the upper strata generally weather into a single weak cliff with several recesses (National Canyon, subunits 5-8, fig. 4). The upper marine unit thins and is locally absent in the eastern Grand Canyon, where the Surprise Canyon Formation consists mainly of widespread slope-forming, red-brown terrestrial deposits.

FOSSILS AND CORRELATION

Marine invertebrate fossils are abundant in limestones of the upper unit in the Surprise Canyon Formation. Preliminary studies of the fauna from sections in the western Grand Canyon by MacKenzie Gordon, Jr. (brachiopods and mollusks), Betty Skipp (foraminifers), and W. J. Sando (corals) recognize the following (Billingsley and McKee, 1982):

Lower limestone beds

11 brachiopods, including *Inflatia* aff. *I. clydensis* (Girty), *Torynifier setiger* (Hall)?, and *Composita* spp.
10 mollusks, including *Glaborocingulum* cf. *G. quadrigatum* Sadlich and Nielson
1 trilobite—*Paladin* sp.
The corals *Mechelinia* sp. and *Barytichisma* sp.

Upper limestone beds

Foraminifera:
Endothyra excellens (D. E. N. Zeller)
Eosigmoilina explicata Ganelina
"*Eosigmoilina rugosa*" Brazhnikova
Neoarchaediscus sp.
Pseudoglomospira sp.
Volvotextularina sp.
Zellerina sp.
Brachiopods:
Anthracospirifer aff. *A. curvilateralis* (Easton)
Composita ovata Mather
Inflatia sp.

In the lower sandstone and siltstone beds of the formation, *Lepidodendron* logs and palynomorphs (spores) indicative of Upper Mississippian strata are locally abundant. From the Granite Park section, 22 palynomorph species were identified by R. M. Kosanke (Billingsley and McKee, 1982, p. 144). Plant and invertebrate remains and vertebrate teeth from other sections in the western and central Grand Canyon are currently under study.

Fossil evidence from the spores, foraminifers, and corals obtained thus far documents a Chesterian and locally an early Morrowan age for the Surprise Canyon Formation. In particular, the eosigmoilinid foraminifers are believed by Betty Skipp to indicate zone 19 (latest Chesterian age) (Billingsley and McKee, 1982, p. 144). The formation is thus considered to be slightly younger than the Paradise Formation of Meramecian and Chesterian age (Armstrong and Repetski, 1980) in southern Arizona and similar in age (Chesterian) to part of the Indian Springs Formation (Brenckle, 1973) of southern Nevada. The Surprise Canyon Formation may, in part, be equivalent to the late Chesterian Log Springs Formation (Armstrong and Mamet, 1974) of New Mexico.

INTERPRETATION

The Surprise Canyon Formation is made up of continental and marine rocks that filled caves and deep erosional valleys cut into the Redwall Limestone in the Grand Canyon in Late Mississippian time. Regional uplift allowed stream erosion to cut valleys as deep as 122 m, primarily in the western Grand Canyon. The oldest deposits of the formation are gravel, sand, and silt of mainly fluvial origin. Fossil plant materials in the basal units indicate that vegetation grew along the stream valleys and perhaps elsewhere on the limestone surface adjacent to the valleys.

As the valleys began to subside, or as sea level rose, marine waters slowly inundated the stream valleys from the west and deposited marine sediments and fossils in the upper unit of the Surprise Canyon Formation. Two marine transgressions, separated by a short regression, may have extended to the eastern Grand Canyon area. A short period of erosion occurred after deposition of the Surprise Canyon Formation and before deposition of the Watahomigi Formation.

DESCRIPTION OF TYPE LOCALITY

Type section of the Surprise Canyon Formation measured at top of the Mooney Falls Member of the Redwall Limestone about 2.4 km southeast of the Bat Tower viewpoint and on the east-facing cliffs of the main canyon wall above Lake Mead, 2 km southwest of the mouth of Tinecannebitts Canyon (Colorado River, mile 263). Width of channel, 305 m; depth of channel, 91 m.

Subunits (numbers refer to numbered sequences in sections, fig. 3):

(7) Covered slope: Some red shaly siltstone with horizontal burrows and mud cracks. Middle part includes thin gray skeletal-lime grainstone beds with abundant echinoderm plates and a few mollusks, bryozoans, and trilobite fragments. Thin beds, 3-4 cm thick, in red-shale slope about 5.5-7.3 m above base of covered slope; upper contact with Watahomigi not exposed. 18.3 m thick.

(6) Sandy limestone: Dark-purple-gray, medium-crystalline grainstone, thin beds 5-20 cm thick separated by silty limestones that form recesses 3-14 cm thick; begins as a recess below irregular slabby ledges and an upper thin recess. Some low-angle tabular and wedge crossbedding; gently rippled on top of limestone beds. Some small-scale trough crossbeds, asymetrical ripples, axis 220°, current direction 310°; abundant echinoderm plates throughout most of the subunit, locally abundant productid brachiopods occur at 11 and 18 m above the base; forms receding ledges. 22 m thick.

(5) Limestone: Dark-red-brown to yellowish-gray skeletal grainstone, small-scale crossbedding in some beds; coarsely crystalline, thin-bedded, fossiliferous throughout; mostly shell hash composed of shells of Composita, productids, small spirifers, gastropods, and crinoids. Forms cliff that has a small recess of darker red brown limestone containing small angular red chert pebbles. 17.5 m thick.

(4) Silty limestone: Yellow-gray wackestone, contains some light-gray bands of thin (5-10 cm) sandstone; forms slope. 7.3 m thick.

(3) Silty limestone: Red-brown to yellow-brown, hematitic grainstone, irregular bedding, fossiliferous. Contains small productids, brachiopods, bryozoans, echinoderms, and gastropods; forms ledge. 0.4 m thick.

(2) Covered slope. 8.4 m thick.

(1) Conglomerate: Dark-red-brown, chert pebbles, cobbles, and boulders in sandy matrix. Rounded pebbles and boulders, matrix supported; rests on irregular surface of Redwall Limestone. Includes some dark-red-brown, iron-rich sandstone beds 7.3 m above base; also includes a few huge blocks of Redwall as much as 3 m in diameter. Some crude bedding and planar crossbedding in pebbly sandstone beds; forms irregular cliff. 17.5 m thick.

Total thickness: 91 m

Bottom: Mooney Falls Member of the Redwall Limestone; gray limestone that has beds of brown and gray chert.

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U.S. Geological Survey, March 13, 1985.

LOWER AND MIDDLE PENNSYLVANIAN NOMENCLATURE IN THE CUMBERLAND OVERTHRUST SHEET OF SOUTHEASTERN KENTUCKY AND EASTERN TENNESSEE: A DISCUSSION

By Robert C. McDowell[1], Charles L. Rice[1], and Wayne L. Newell[1]

Abstract

The sequence of largely Middle Pennsylvanian, coal-bearing sedimentary rocks overlying the Lee Formation in the Cumberland overthrust sheet of southeastern Kentucky and adjacent Tennessee has been divided into the Hance, Mingo, Catron, Hignite, and Bryson Formations of the Breathitt Group. A systematic program of quadrangle mapping in recent years has demonstrated that these formations, as originally defined and subsequently modified, do not accurately portray lateral and vertical lithologic variations within the sequence. Additional redefinition of these units is not practical and would probably be misleading. Rather than use these formations, we have treated and will continue to treat the Breathitt interval as a formation, in conformity with U.S. Geological Survey usage in the rest of eastern Kentucky and adjacent Tennessee.

INTRODUCTION

Lowermost rocks of Pennsylvanian age in much of the Appalachian basin of eastern Kentucky, adjacent Tennessee, and Virginia are massive, pebbly orthoquartzites of the Lee Formation. These distinctive, resistant rocks are overlain by a sequence of siltstones, claystones, subgraywacke sandstones, and thin interbeds of coal, underclay, ironstone, and limestone. This Lower and Middle Pennsylvanian sequence has a maximum thickness of 3,100 ft (950 m) and consists of largely deltaic sedimentary rocks that contain many commercially exploited coal beds. These strata have been described recently by Rice and others (1979), and the distribution of the strata is shown on the new Geologic Map of Kentucky (McDowell and others, 1981). Rice and others (1979) and McDowell and others (1981) refer to this stratigraphic unit as the Breathitt Formation. However, on the Cumberland overthrust sheet (south of the Pine Mountain fault), southeastern Kentucky (fig. 1), the sequence overlying the Lower Pennsylvanian Lee Formation traditionally has been subdivided into five formations: in ascending order, the Hance, Mingo, Catron, Hignite, and Bryson Formations. These Middle Pennsylvanian formations (the Hance Formation may locally contain strata of Early Pennsylvanian age) are shown on 24 geologic quadrangle maps published during the recently completed cooperative mapping program of the U.S. Geological Survey and the Kentucky Geological Survey (fig. 1). Thus, two different nomenclatures have been applied recently to the same sedimentary sequence south of the Pine Mountain fault. This report reviews the origins of the current incompatible usages and analyzes alternative solutions.

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STRATIGRAPHY

The largely Middle Pennsylvanian rocks overlying the orthoquartzites of the Lee Formation in the Cumberland overthrust sheet were assigned to the Hance, Mingo, Catron, Hignite, and Bryson Formations by Ashley and Glenn (1906, p. 32-33) mainly on the basis of locally identified and mapped, informally named coal beds. These formations were later combined as the Breathitt Group (fig. 2) by Englund and others (1963, p. B15). These names, used principally in Kentucky, have been used only locally in adjoining quadrangles of Tennessee and Virginia (fig. 1) south of the Pine Mountain fault.[2]

Ashley and Glenn (1906) defined most of the boundaries of the five formations of the Breathitt Group at the top or bottom of arbitrarily selected coal beds, even though they recognized the uncertainty of some of their coal bed correlations. Moreover, type sections or type areas of the formations were widely spaced; three were located in the Log Mountains area west of the Rocky Face fault, a major structural discontinuity, while the remaining two were in the Black Mountains area, east of the fault (fig. 3).

U.S. Geological Survey quadrangle mapping has shown that these "Breathitt Group" formations are not distinguishable on the basis of lithic character and that the coal beds, which define the tops and bottoms of the formations, are not laterally persistent throughout the Cumberland overthrust sheet but are missing over tens to hundreds of square miles. Thus artificial gaps or overlaps in the stratigraphic section have resulted from the arbitrary projection of formations whose type sections are widely spaced and whose delineating coal beds are absent locally or concealed. The published quadrangle maps shown in figure 1 reflect the diversity of interpretations that has resulted from the existing nomenclature (see also Rice, 1984). Details of this diversity of usage are presented below.

Hance Formation

The Hance Formation was defined by Ashley and Glenn (1906, p. 37) as extending from the top of the Naese Sandstone Member of the Lee Formation "at the locality of the Naese upward to the bottom of the Lower Hance coal." Where the Naese is locally absent, the base of the Hance is generally placed at the top of the highest orthoquartzite. In the western part of the area, the top of the Naese Sandstone Member of the Lee Formation forms this boundary, and in the eastern part, the top of the Bee Rock Sandstone Member of the Lee forms the boundary (fig. 4). The contact is commonly within a sequence in which the quartzose sandstone

[2]The names Hance, Mingo, Catron, and Hignite also have been used locally as formations or as members of the Breathitt Formation north of the Pine Mountain fault (Englund, 1966, 1969).

U.S. Geological Survey Bulletin 1605-A, 1985, p. A35-A43.

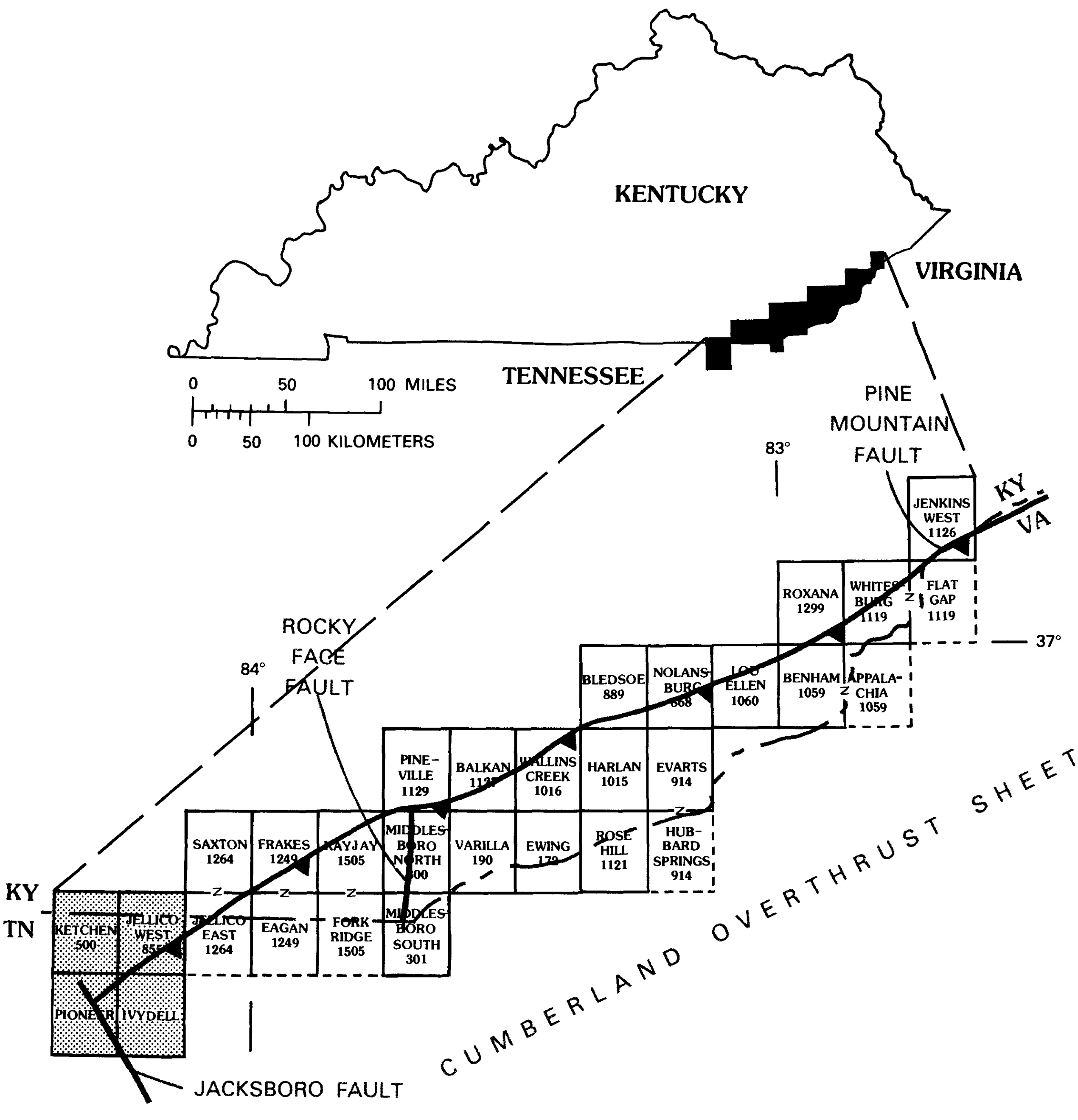


Figure 1. Names and numbers of U.S. Geological Survey geologic quadrangle maps (scale 1:24,000) in which the nomenclature of Ashley and Glenn (1906) was used for stratigraphic units of largely Middle Pennsylvanian age, south of the Pine Mountain fault. The four shaded quadrangles at the western end of the study area are also included in a geologic report of the Elk Valley area (Englund, 1968). Areas in Tennessee and Virginia bounded by dashed lines were not mapped.

grades upward through a few feet to siltstone, shale, or subgraywacke. Locally, the contact is not mapped where the Yellow Creek Sandstone Member of the Hance Formation directly overlies the Naese; there, the contact is within the Yellow Creek and Naese Sandstone Members (undivided) (fig. 4). In places, the Lee and Hance Formations intertongue; where the Bee Rock Sandstone Member forms the top of the Lee Formation, sandstone lenses equivalent in part to the Naese Sandstone Member occur locally in the lower part of the Hance Formation (Froelich and Tazelaar, 1973). In other areas, where the Naese Sandstone Member includes a few feet of silty sandstone and shale at the top, the contact is mapped at the base of the overlying Naese coal bed. In general, the top of the Lee rises stratigraphically westward (Englund and others, 1963, p. B15). The vertical difference in position of the formation boundary, from the top of the Bee Rock Sandstone Member to the top of the Naese Sandstone Member, represents an interval of as much as 250 ft (76 m).

The type locality of the Naese Sandstone Member is at Naese Cliff, about 2,500 ft (760 m) above the mouth of

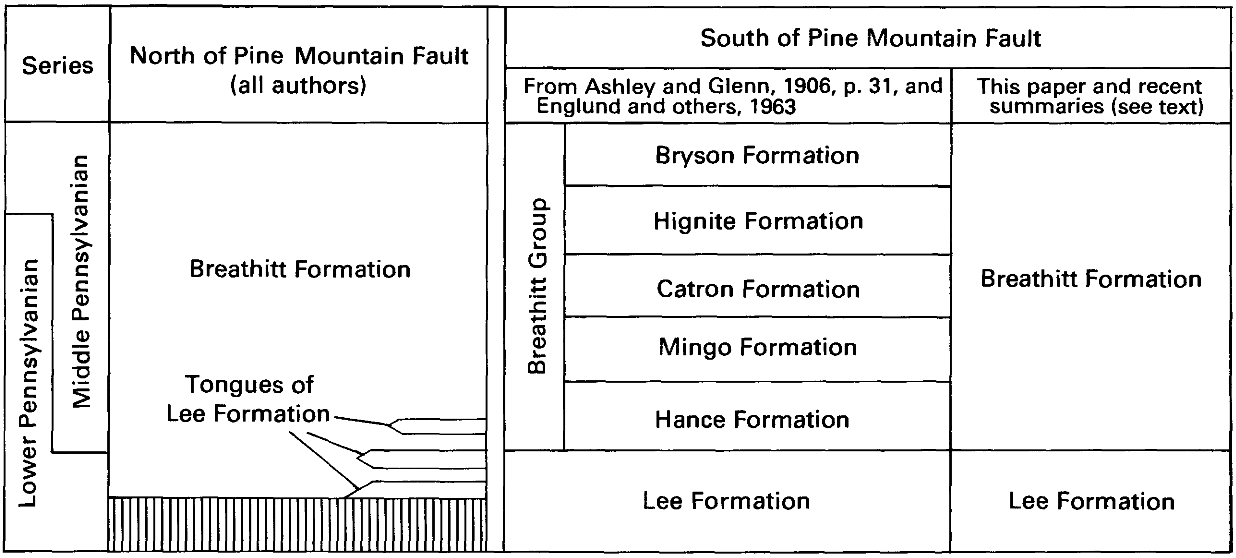


Figure 2. Generalized diagram of Lower and Middle Pennsylvanian nomenclature in eastern Kentucky. The three tongues of the Lee Formation are, from oldest to youngest, the Livingston Conglomerate, the Rockcastle Sandstone, and the Corbin Sandstone Members. Vertical lines indicate missing strata.

Hance Creek on the Cumberland River in the Varilla quadrangle, Black Mountains area (figs. 1 and 3). Ashley and Glenn (1906) apparently intended the Hance Formation, with a "thickness of about 600 ft" (180 m), to extend upward to the Hance coal bed near the northern end of nearby Hance Ridge. However, in their introductory columnar section (Ashley and Glenn, 1906, p. 31), they equated the Harlan coal bed with the Hance coal bed and showed the Hance Formation as extending from the top of the Lee Formation to the base of the "Harlan/Hance" coal bed. Recognizing that the Harlan coal bed was stratigraphically about 600 ft (180 m) above the Hance coal bed, Englund and others (1963, p. B16) placed the top of the Hance Formation at the base of the widely mined Harlan coal bed; following that usage, the Harlan coal bed (and its correlatives, including the Jellico coal bed) has generally been accepted as the upper boundary of the Hance Formation (fig. 5). Raising the top of the Hance to the Harlan coal bed and its correlatives approximately doubles the thickness of the formation as it was originally defined by Ashley and Glenn (1906).

In the easternmost part of the study area, the Harlan coal bed splits into several thin coal beds. In that area, the upper contact is placed at the base of the Harlan coal zone or, where that zone can no longer be traced, at the base of the Collier coal bed, which is about 90 ft (27 m) higher in the section (fig. 5). In the Log Mountains area, Ashley and Glenn (1906, p. 31) placed the top of the Hance Formation at the base of the Bennetts Fork coal bed, which they correctly correlated with the Hance coal bed. Two quadrangle maps followed that usage (fig. 5). In those reports, the Bennetts Fork coal bed is as much as 510 ft (155 m) below the Mingo (Harlan) coal bed.

Mingo Formation

Ashley and Glenn (1906, p. 38) defined the Mingo Formation, named for Mingo Mountain in the Log Mountains area (fig. 3), as extending from the bottom of the Bennetts Fork coal bed upward to the base of the Poplar Lick coal bed, a thickness described as being about 950 ft (290 m). In the Black Mountains area, Ashley and Glenn (1906, p. 31) incorrectly correlated the Poplar Lick coal bed with the stratigraphically higher Wallins Creek coal bed and placed the upper boundary of the Mingo Formation at the base of the Wallins Creek coal bed. The Poplar Lick coal bed correlates with a coal bed in the Creech coal zone (Rice and Smith, 1980), which is as much as 350 ft (110 m) below the Wallins Creek coal bed in the Black Mountains area. The Wallins Creek contains a distinctive flint-clay parting that has long been recognized as one of the most important marker beds in the Middle Pennsylvanian of the central Appalachian basin (Wanless, 1946, p. 56). Unfortunately, the flint-clay parting has not been found in the area of Mingo Mountain or anywhere in the area of the Log Mountains; analysis of

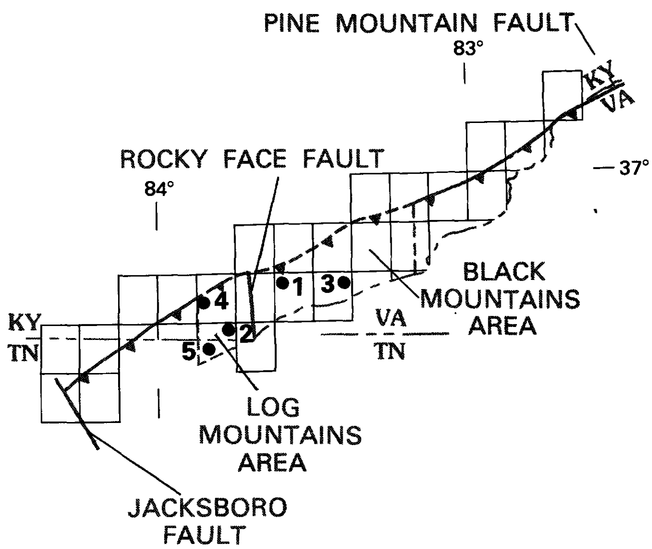
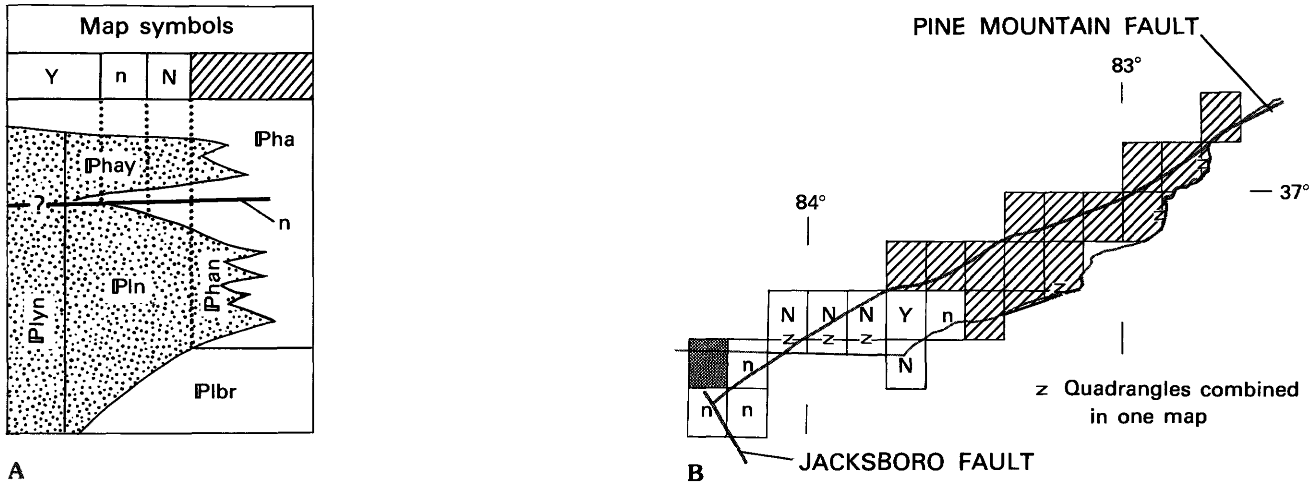


Figure 3. Location of type sections or type areas of formations of Ashley and Glenn (1906). 1, Hance; 2, Mingo; 3, Catron; 4, Hignite; and 5, Bryson. Area mapped by Ashley and Glenn shown by dashed line.



Hance and Lee Formations

IPha, Hance Formation

IPhay, Yellow Creek Sandstone Member of Hance Formation

IPlyn, Yellow Creek and Naese Sandstone Members (undivided) of Lee Formation

n, Naese coal bed

IPln, Naese Sandstone Member of Lee Formation

IPhan, Naese Sandstone Member of Hance Formation

IPlbr, Bee Rock Sandstone Member of Lee Formation

Contact placed at top of Bee Rock Sandstone Member

N Contact placed at top of Naese Sandstone Member

n Contact placed at base of Naese coal bed

Y Contact is within Yellow Creek and Naese Sandstone Members (undivided) or at base of Naese coal bed

Contact below land surface

Figure 4. A, Distribution of various horizons used as the boundary between the Lee and Hance Formations. B, Quadrangles in which various horizons are used to define formation boundary.

the stratigraphic section suggests that the Wallins Creek coal bed may have no correlative there. Thus, in the Log Mountains, which contain the type area of the Mingo Formation, the top of the Mingo has been placed at the base of the Poplar Lick coal bed. Most reports in other areas have placed the contact at the base of the Wallins Creek (fig. 6); locally in these areas, the Wallins Creek coal bed cannot be identified, and the top of the Mingo Formation is shown at the base of the Wallins Creek coal zone (Rice and Wolcott, 1973) or, where the coal bed is absent, is projected as a color break, without either contact or coal bed being shown (Froelich and Stone, 1973).

The flint-clay parting that characterizes the Wallins Creek coal bed also occurs in the southwestern part of the area in the Windrock coal bed of Tennessee; the base of the Windrock coal bed is locally used as the top of the Mingo Formation (Englund, 1968). But coal beds in other parts of the stratigraphic section also locally contain flint-clay partings. Thus, Glenn (1925) identified two coal beds at the southwestern end of the overthrust sheet that contain flint clay in the underclay or as parting in the coal bed. These coal beds are about 260 ft (80 m) apart stratigraphically. Glenn incorrectly projected the horizon of the Windrock coal bed across the Jacksboro fault to the lower of the coal beds in the overthrust sheet. Because that coal bed apparently correlates with the Poplar Lick coal bed, Glenn thereby reinforced the correlation of the Wallins Creek, Poplar Lick, and Windrock coal beds, all of which are used locally to define the top of the Mingo Formation. Glenn (1925) identified the upper coal bed as the Walnut Mountain; although the Walnut Mountain coal bed does correlate with the Windrock and Wallins Creek coal beds (Rice, 1984), it has not been used as a boundary for the Mingo Formation.

Catron Formation

Ashley and Glenn (1906, p. 41) defined the Catron Formation as extending from the bottom of the Wallins Creek coal bed to the top of the Jesse Sandstone Member, as typically exposed in Coon Branch of Catron Creek.[3] In its type area in the Black Mountains, the top of the Catron Formation was most commonly placed on

[3] Because the Poplar Lick and Wallins Creek coal beds are not correlative, the description of the type sections of the Mingo and Catron Formation by Ashley and Glenn (1906) leaves approximately 350 ft (110 m) of strata between the coal beds unassigned to either formation.

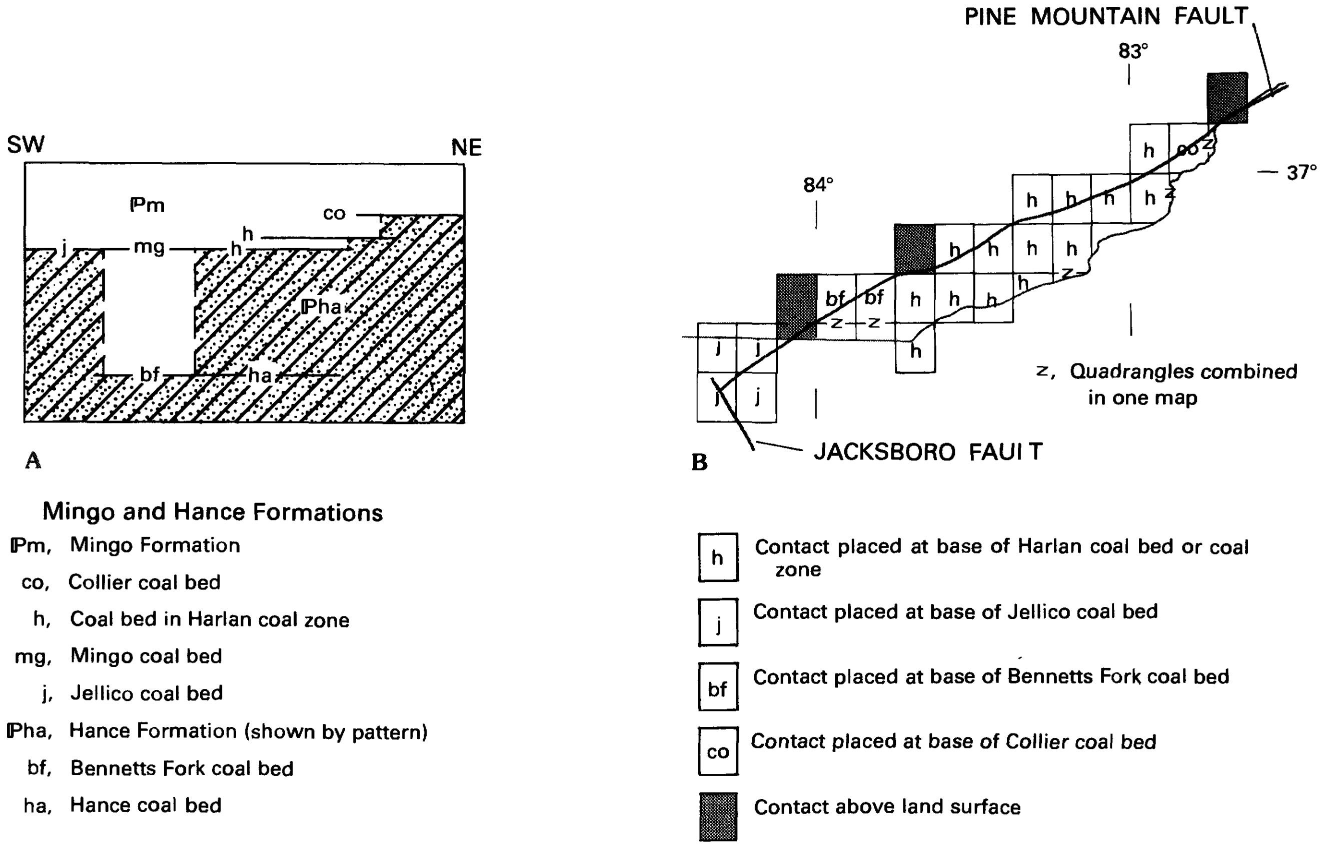


Figure 5. A, Stratigraphic relations between the Hance and Mingo Formations. B, Quadrangles in which various coal beds are used to define formation boundary.

the geologic quadrangle maps at the base of the Limestone coal bed, which in many places overlies the Jesse Sandstone Member (fig. 7). The contact was also locally placed at the base of the Pardee coal bed, which underlies the Jesse Sandstone Member but which is reported to join the Limestone coal bed where the Jesse Sandstone Member is absent.

West of Rocky Face fault (fig. 1), the Jesse Sandstone Member has not been identified, and Ashley and Glenn (1906, p. 41) placed the top of the Catron Formation at the base of the Hignite coal bed in the area of the Log Mountains. This definition of the Catron Formation has been used for quadrangle maps in that area. In other areas at the western end of the overthrust sheet, because none of the above mentioned horizons (Jesse Sandstone Member or Limestone, Pardee, or Hignite coal beds) could be identified, the top of the Catron Formation has been arbitrarily placed at the base of the Magoffin beds of Morse (1931). Drill hole logs and measured sections from outcrops show that the Magoffin is locally as much as 125 ft (38 m) above the Limestone coal bed in the Black Mountains area (see Rice, 1984; Hodge, 1912, p. 57).

The interval representing the difference in stratigraphic position of the contact as shown on geologic quadrangle maps (that is, from the base of the Hignite coal bed to the base of the Magoffin beds) is as much as 325 ft (100 m) thick.

Hignite Formation

The Hignite Formation was named for Hignite Creek in the Log Mountains (fig. 3) and was defined as extending from the bottom of the Hignite coal bed to the top of the Red Springs coal bed (Ashley and Glenn, 1906, p. 43). It is as much as 560 ft (170 m) thick. In geologic quadrangle maps of this area, the upper contact of the Hignite Formation was moved to the base of the Red Springs coal bed, to follow the general practice of placing unit contacts at the base rather than at the top of coal beds. Although this adjustment makes only a few feet difference in the thickness of the formation, the adjustment puts the Red Springs coal bed in the overlying Bryson Formation. In areas beyond the type area, the upper contact of the Hignite Formation was not mapped.

Bryson Formation

The Bryson Formation was named for Bryson Peak, Tenn. (fig. 3), where the formation "shows a thickness of about 200 ft" (60 m) (Ashley and Glenn, 1906, p. 44). Ashley and Glenn did not assign a top to the Bryson Formation but described the formation as occupying the "higher tops of the Log Mountains" and the "highest points in the Black Mountains." The formation is shown on the geologic quadrangle maps of the area only in the

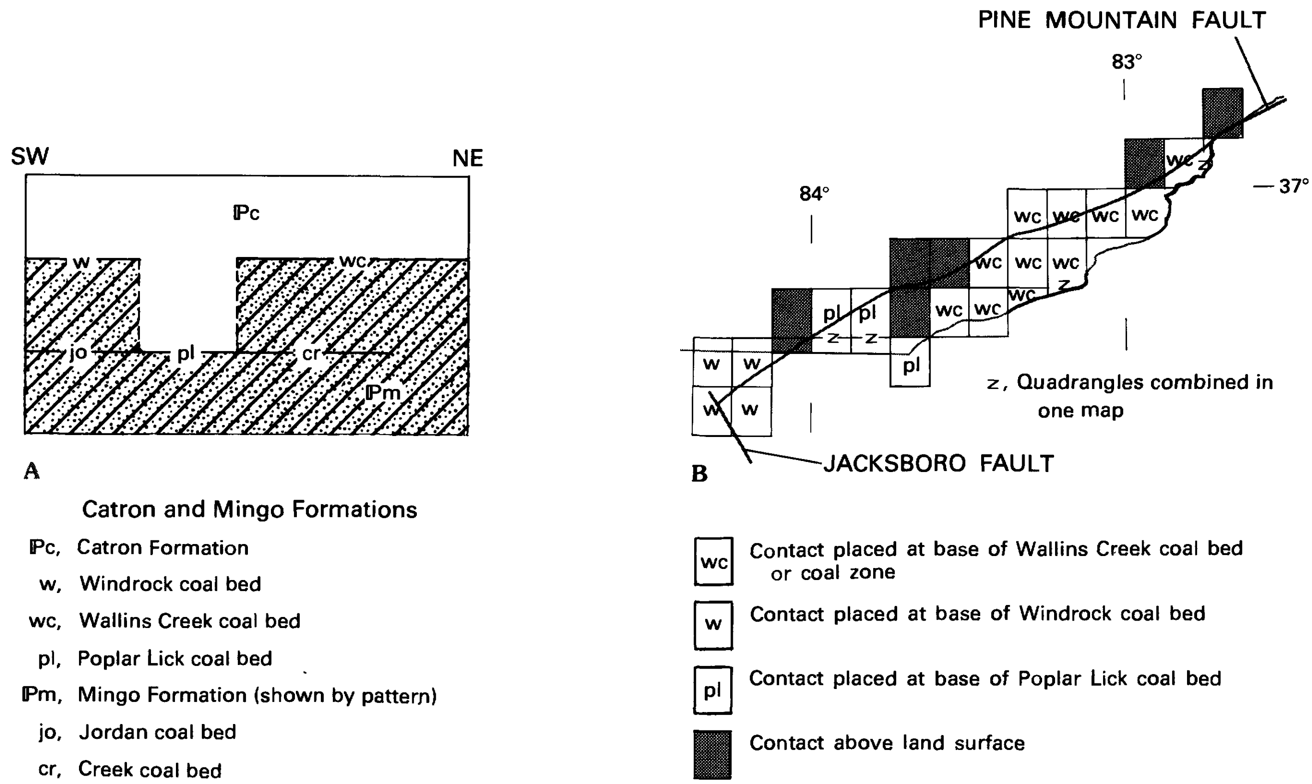


Figure 6. A, Stratigraphic relations between the Mingo and Catron Formations. B, Quadrangles in which various coal beds are used to define formation boundary.

vicinity of the type area. In other areas of the Cumberland overthrust sheet, where rocks of the Bryson Formation could not be identified or where their area of outcrop was considered to be too small to show, these rocks were combined with those of the underlying Hignite Formation. The basal contact for the Bryson in the type area is placed at the base of the Red Springs coal bed. Regional stratigraphic studies suggest that the Red Springs coal bed correlates with the Braden Mountain coal bed in the Elk Valley area at the west end of the study area and with the Morris coal bed in the eastern part of the study area (Rice and Smith, 1980). Figure 8 shows the approximate thickness of mapped Bryson Formation strata and of Bryson Formation strata that have been included in the Hignite Formation in geologic quadrangle maps of areas other than the Log Mountains.

DISCUSSION AND CONCLUSIONS

The foregoing analysis of the nomenclature applied by Ashley and Glenn (1906) to the Pennsylvanian strata of the Cumberland overthrust sheet illustrates the difficulties in the use of these names except in very limited, specific areas. These problems made using these formations on the Geologic Map of Kentucky (McDowell and others, 1981) impractical and led to their being discarded in other detailed regional syntheses in favor of the more convenient and useful grouping of the strata into the Breathitt Formation (Huddle and others, 1963; Rice and others, 1979; Rice and Smith, 1980). We believe these difficulties are inherent in attempts to establish stratigraphic nomenclature in deltaic sequences, which generally involve a succession of fine-grained clastic rocks and coal beds with limited lateral persistence of lithic characteristics. Such sequences typically lack clear lithic contrasts that could serve as a basis for subdivision except in a local sense. The use of arbitrarily selected coal beds as formational boundaries, as in the present case, invites inconsistent application of the nomenclature, as we have shown. To illustrate some of the inconsistencies of nomenclature, figure 9 summarizes the common usages of the Hance, Mingo, Catron, Hignite, and Bryson Formations in three major mapping areas of the Cumberland overthrust sheet (see also Rice, 1984). The lack of success in establishing a consistent series of formations in this largely Middle Pennsylvanian sequence appears to stem mostly from the lack of distinctive, persistent lithic or biostratigraphic breaks that might be used as a basis for subdivision, as indeed was recognized by Ashley and Glenn (1906, p. 32). Apart from a few thin, widely recognizable units, such as the flint clay in the Wallins Creek coal bed of the Catron Formation, marine zones such as the Magoffin beds of Morse (1931) in the Hignite Formation, and the Kendrick Shale of Jillson (1919) in the Catron or Mingo Formations, the entire section is a repetition of several similar lithologies that occur as rock units of limited vertical and lateral extent.

A regional perspective of the Middle Pennsylvanian stratigraphic section in Kentucky suggests that, in fact, the Cumberland overthrust sheet is not a favorable area to establish reliable formational boundaries. Wanless (1946), Englund (1968), and others have long recognized that the Breathitt sequence thickens southeastward toward a depocenter. Englund (1968) used the rate of thickening as an indicator of the magnitude of lateral

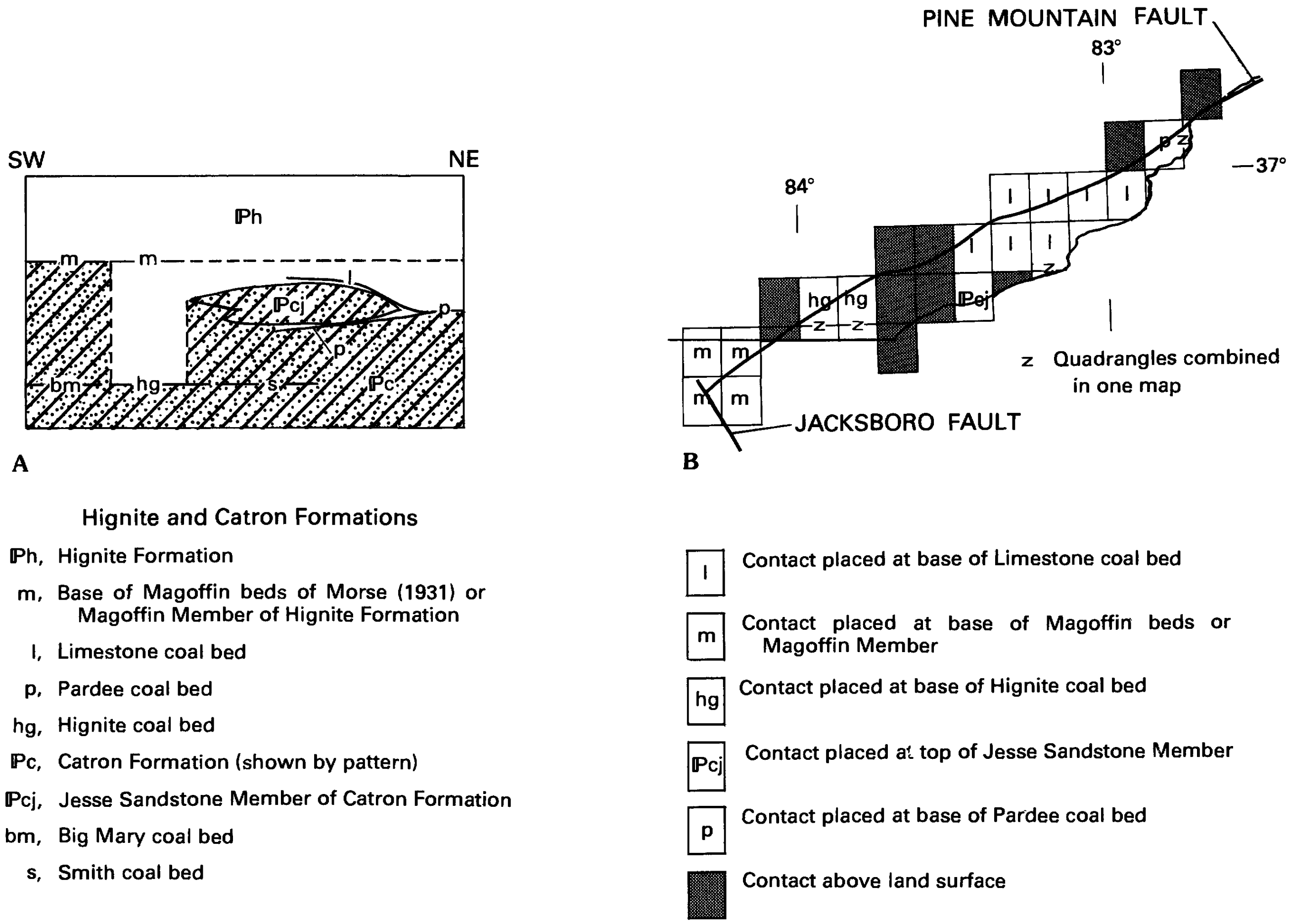


Figure 7. A, Stratigraphic relations between the Catron and Hignite Formations. B, Quadrangles in which various stratigraphic units are used to define formation boundary.

displacement on the Pine Mountain overthrust fault. Wanless (1946) speculated that an abrupt change in thickening rate reflected a hinge line that controlled the location of ramping of the Pine Mountain overthrust fault. Generally, traverses from basins across hinge lines and onto basin margins show nearly continuous deposition in the depocenter, rapid thinning across the hinge line, and significant unconformities on the margins. In none of these three areas has the Breathitt Formation been successfully subdivided. The thicker, basinward sequences south of Pine Mountain record more rapid, continuous depositional sequences. However, even widespread, internally consistent facies that could be treated as formations are lacking there. Much of the section reflects upper and lower delta plain environments, which are known for great internal variation.

Thus field observations, mapping, and regional synthesis all indicate that the Hance, Mingo, Catron, Hignite, and Bryson Formations as presently used are not consistent map units that have uniform or predictable rock stratigraphic boundaries throughout the Cumberland overthrust sheet. We can envision three possible alternative methods of treatment of these strata:

1. Retention of the present formations, but with careful redefinition (beyond any currently in the literature) that would provide a reasonable possibility of regional consistency or with local definitions presented henceforth with each reference to any of these units.

2. Abandonment of the formations and erection of a new nomenclatural system based on division of the sequence into two or three subunits on the basis of marine zones or the most reliable coal beds.

3. Disregard of the present formations and reduction of the Breathitt Group in rank to Breathitt Formation for the purposes of regional consistency in eastern Kentucky. All members of the formations of the Breathitt Group would be retained as members of the Breathitt Formation.

We foresee grave difficulties with the first alternative, redefinition of the present formations. We doubt that a regionally consistent redefinition is possible, and in any case such a redefinition would preserve numerous incompatible usages now in the literature, as would any system of purely local definition. Moreover, we see this solution as failing to conform to the stratigraphic code because the formations are not lithologically distinguishable one from the other (North American Commission on Stratigraphic Nomenclature, 1983, pt. 2, arts. 22 and 24, p. 855-858).

The second alternative, abandonment of the formations and erection of two or three new formations by using key beds such as marine zones and the most

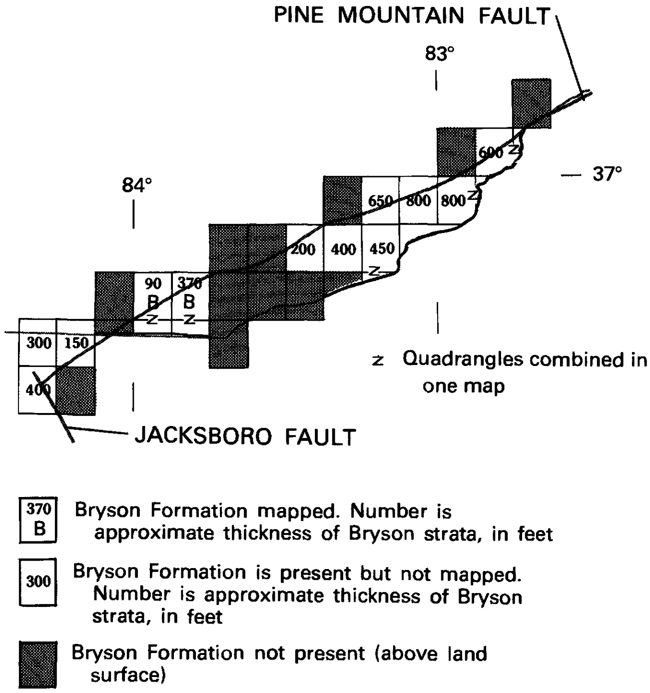


Figure 8. Quadrangles where the Bryson Formation was mapped and where Bryson strata were included in the Hignite Formation. Thickness taken as strata above correlatives of the Red Springs coal bed; that is, the Braden Mountain coal bed in the Elk Valley area and the Morris coal bed in the eastern part of the study area.

reliable coal beds as boundaries, is also at odds with the stratigraphic code and provides no readily apparent benefit. Key beds are the most important stratigraphic units in the Middle Pennsylvanian rock sequence and are commonly mapped; designating them as formational boundaries affords neither new information nor greater ease in conceptualizing the sequence. The North American Commission on Stratigraphic Nomenclature (1983, p. 851), while noting the utility of "marker-defined units" in stratigraphic studies such as the "formats" of Forgotson (1957) or "chronosomes" of Schultz (1982), nevertheless recommended that those units are "best left as informal." Therefore, we suggest that any such subdivision in the Cumberland overthrust sheet be done informally, as on the Geologic Map of Kentucky (McDowell and others, 1981) and on the U.S. Geological Survey geologic quadrangle maps of the areas of eastern Kentucky northwest of the Cumberland overthrust sheet.

We intend, therefore, to treat the Breathitt interval of the Cumberland overthrust sheet as a formation rather than as a group (fig. 2), as it has been treated in the detailed regional syntheses of Huddle and others (1963), Rice and others (1979), Rice and Smith (1980), and McDowell and others (1981). This usage results in a consistent nomenclature throughout the Appalachian basin in Kentucky, in which Lower and Middle Pennsylvanian rocks are assigned to the Lee and Breathitt Formations. Because of the widespread earlier use of the terminology of Ashley and Glenn (1906), especially on geologic quadrangle maps, we do not formally abandon these names, but we believe their continued use, particularly at a regional scale, would be a source of misunderstanding and confusion with virtually no compensating advantages. We therefore caution that the use of these names, even locally, should be accompanied by a clear definition of the usage, including the geographic extent.

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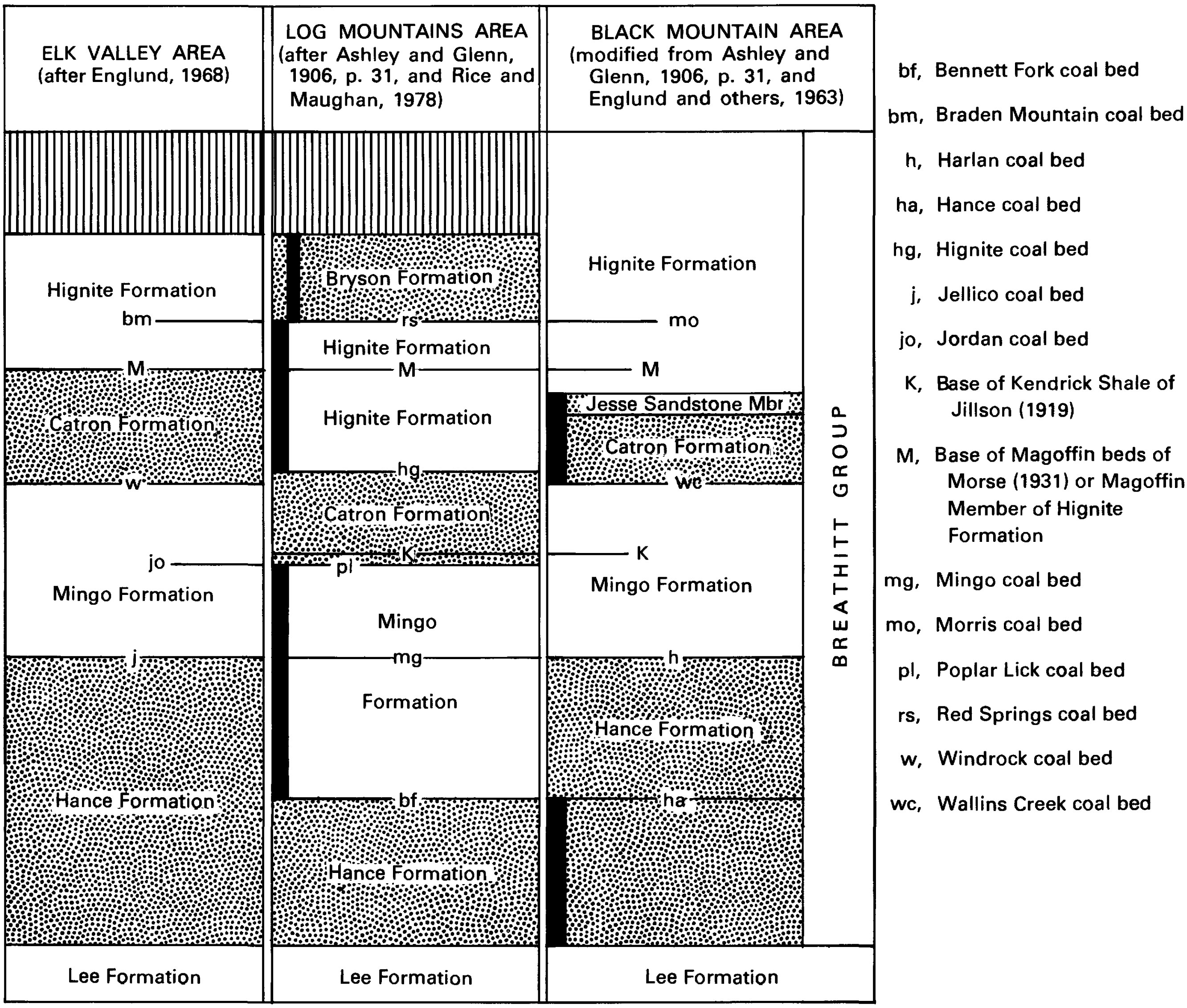

Figure 9. Correlation chart of formations, members, and beds among three mapped areas of the Cumberland overthrust sheet. Vertical bars indicate the area and limits of the type sections of formations as described by Ashley and Glenn (1906). Coal bed correlations from Rice and Smith (1980).

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CORRELATION AND GEOGRAPHIC EXTENT OF SOME MIDDLE AND UPPER DEVONIAN AND LOWER MISSISSIPPIAN BLACK SHALES IN THE APPALACHIAN BASIN

By Wallace de Witt, Jr.,[1] and John B. Roen[1]

Abstract

Nomenclature and correlations of exposed Devonian black shales in the Appalachian basin were well established by 1960. The subsurface extent of the shales, however, was poorly known because techniques for positive identification and tracing of individual units were not available. As a result, subsurface data were scant. The introduction of gamma-ray well logging in the 1950's and its widespread use in the 1960's provided the primary tool for subsurface tracing of black shales by their positive and characteristic signatures on the gamma-ray log.

By correlating gamma-ray and compensated density logs across the basin, we defined the regionally extensive black shales in terms of their named outcrop units, and we synthesized a regional stratigraphic framework for the black gas-producing Devonian shales of the basin.

Under the eastern and central parts of the Appalachian Plateaus, the following extensive units are recognized, in ascending stratigraphic order: the Middle Devonian Marcellus Shale, the Middle and Upper Devonian Geneseo Shale Member of the Genesee Formation, the Upper Devonian Middlesex Shale Member of the Sonyea Formation, the Upper Devonian Rhinestreet Shale Member of the West Falls Formation, the Upper Devonian Pipe Creek Shale Member of the Java Formation, and the Upper Devonian Dunkirk Shale Member of the Perrysburg Formation. All of these shales underlie parts of New York, Pennsylvania, and West Virginia. Several are present also under parts of eastern Ohio, eastern Kentucky, and southwestern Virginia. In central Tennessee, the Rhinestreet Bed is the lower black shale bed of the Upper Devonian Dowelltown Member of the Chattanooga Shale. The Dunkirk Member becomes the Upper Devonian Huron Member of the Ohio Shale in the general area of the Pennsylvania-Ohio State line.

The Huron Member of the Ohio Shale underlies much of eastern Ohio, western West Virginia, western Virginia, and eastern Kentucky. The Huron Member also is present as a bed in the lower part of the Gassaway Member of the Chattanooga Shale in southern Kentucky and adjacent Tennessee and is here formally extended as the Huron Bed. The Gassaway Member of the Chattanooga Shale is more extensive than the older Dowelltown Member and underlies much of eastern Tennessee and adjacent parts of northwest Georgia and northeastern Alabama.

The Cleveland Member of the Ohio Shale is restricted to the western part of the Appalachian basin where the member underlies a broad strip of eastern Ohio and eastern Kentucky. The member is present locally in extreme southwest Virginia. In southeastern Kentucky, the Cleveland is present in the upper part of the Gassaway Member as the Cleveland Bed. It is repre sented by a biostratigraphic zone in the upper part of the Gassaway Member throughout southern Kentucky, much of eastern Tennessee, and adjacent Alabama.

The Lower Mississippian Sunbury Shale, which is the youngest in this sequence of black gas-productive shales, also is present largely in the western part of the basin. It underlies much of eastern Ohio, western West Virginia, southwest Virginia, and eastern Kentucky. In south-central Kentucky, the Sunbury grades laterally into the uppermost part of the Gassaway Member; in eastern Tennessee, the Sunbury is represented at many places by a correlative biostratigraphic zone in the upper few feet or few inches of the Chattanooga Shale.

INTRODUCTION

This paper defines and documents formally the lateral extent of several widespread black shales of Middle and Late Devonian and Early Mississippian age in the subsurface of the Appalachian basin.

Black shales of the Devonian sequence crop out widely in an arcuate belt from eastern New York to west-central Alabama along the northern and western periphery of the Appalachian basin (fig. 1). Some of these shales also are exposed in thrust sheets within the Valley and Ridge fold belt from Pennsylvania to central Alabama. Because of many factors, including the lithologic similarity of individual black-shale units, the absence of diagnostic macrofaunas and floras, the paucity of outcrops at many places, complex facies changes in the subsurface, and unresolved structural problems in the thrust belt, Devonian black shales are more difficult to correlate regionally than lighter colored and coarser grained clastic rocks.

SURFACE STRATIGRAPHY

Stratigraphic nomenclature of the Devonian black shales of the Appalachian basin evolved slowly as geologists in the 19th century completed local geologic studies within the several States of the basin and sought to extend local stratigraphic nomenclature regionally. Hass (1956) reviewed the general history of black-shale stratigraphic nomenclature for much of the west half of the basin in his discussion of the age and correlation of the Chattanooga Shale. Hass' report indicates that disagreements evolved as stratigraphers sought to correlate the Devonian black shales around the edges of the basin.

By the 1960's, the stratigraphy of outcropping black shales was well fixed. The Chattanooga Shale with its two members, the Dowelltown below and the Gassaway above (Hayes, 1891; Hass, 1956; Conant and Swanson,

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1961, p. 21-40), was recognized in Tennessee and in south-central Kentucky along the drainage of the Cumberland River. The Chattanooga also was recognized locally in southwest Virginia (Roen, Miller, and Huddle, 1964). Near Big Stone Gap, Wise County, Huddle confirmed Hass' observation (1956) that the upper member of the Chattanooga Shale, there identified as the Big Stone Gap Member (Roen, Miller, and Huddle, 1964), contained a Late Devonian conodont fauna equivalent to the fauna in the Cleveland Member of the Ohio Shale in the lower part and an Early Mississippian conodont fauna equivalent to the fauna in the basal part of the Sunbury Shale of Ohio in the upper part (see figs. 8 and 9). These data confirmed the extent of the Sunbury Shale in the subsurface of Ohio, West Virginia, and eastern Kentucky as shown by Pepper, de Witt, and Demarest (1954, fig. 56, p. 92). In northeastern Kentucky and Ohio, the Devonian black shale sequence was recognized as the Ohio Shale with two members, the Huron below and the Cleveland above (Pepper, de Witt, and Demarest, 1954, p. 14). Throughout this area, the younger Lower Mississippian black Sunbury Shale is separated from the Devonian age Ohio Shale by a deltaic wedge of lighter gray clastic rocks. The black Devonian shales plunge under cover of younger rock in northeastern Ohio and are not exposed in northwestern Pennsylvania.

In much of western and central New York, six extensive black shales are recognized in outcrops of the Middle and Upper Devonian rocks. In ascending stratigraphic order these units are the Marcellus Shale, the Geneseo Shale Member of the Genesee Formation (de Witt and Colton, 1959, p. 2816; 1978, p. A7), the Middlesex Shale Member of the Sonyea Formation (Colton and de Witt, 1958), the Rhinestreet Shale Member of the West Falls Formation (Pepper, de Witt, and Colton, 1956), the Pipe Creek Shale Member of the Java Formation (de Witt, 1960, p. 1933), and the Dunkirk Shale Member of the Perrysburg Formation (Pepper and de Witt, 1951). All of the units are well represented in outcrops in the western half of New York. However, to the east, the Upper Devonian black shales thin and grade laterally into lighter colored, coarser grained clastic rock, and only the black Marcellus Shale of Middle Devonian age is present in sections near Albany and southward along the foot of the Catskill Mountains in the Hudson River valley.

In eastern and central Pennsylvania, in outcrops east of and along the foot of the Allegheny front, the black Marcellus Shale is well exposed at many places in the Valley and Ridge fold belt. The younger Burket Shale Member of the Harrell Shale (Butts, 1918; de Witt, 1975), which directly overlies the Tully Limestone in north-central Pennsylvania, is not as well exposed as the Marcellus Shale but is areally almost as extensive. As the Marcellus and the Burket Member are traced to the south across Maryland (de Witt and Colton, 1964) and into northern West Virginia and adjacent Virginia, the medium gray rocks of the intervening Mahantango Formation (de Witt and Colton, 1964, p. 44) become darker and grade laterally into black mudrock and black shale. The Burket Member and Marcellus lose their identities in a thick unit of black shaly rock. Butts (1940) gave the name Millboro to this 1,000-ft thickness of black shale in the vicinity of Millboro Springs, Bath County, in west-central Virginia. The black shale of the Millboro, which had been previously identified simply as "black shale" or "Devonian black shale" by geologists working in the Valley and Ridge belt of Virginia and Tennessee, has been traced southwest in outcrops in Tennessee where Hayes (1891) named it the Chattanooga Shale.

SUBSURFACE STRATIGRAPHY

In contrast to surface exposures, subsurface Devonian black shales were difficult to identify and correlate with data at hand. Although more than half a million wells have been drilled by cable-tool methods in the Appalachian basin and at least half this number were drilled into or through the Devonian shale sequence, black shales generally were not recorded by well drillers except in the Big Sandy gas field of eastern Kentucky and adjacent West Virginia where Devonian black shales are both source bed and reservoir rock for natural gas. In this part of the basin, many drillers referred to the Sunbury Shale as the coffee shale and referred to the upper and lower tongues of the Ohio Shale as the little and big cinnamon shales, respectively. However, we found that the drillers were not consistent in their identification of the units; where the upper tongue of Ohio Shale was absent, drillers gave the name little cinnamon to the Ohio Shale and identified the older and subjacent black Rhinestreet Shale Member of the West Falls Formation as the big cinnamon shale. Consequently, even in areas of concentrated drilling with the black shales as targets, correlations based on drillers' logs were suspect.

The excellent subsurface geologists C. R. Fettke in Pennsylvania and J. H. C. Martens in West Virginia tried to correlate the Devonian black shales by studying drill cuttings from deep exploratory wells. Unfortunately, neither was able to obtain a sufficient number of closely spaced sets of drill cuttings to identify and trace named surface units widely in the subsurface with confidence, and they abandoned their efforts.

With the advent of a deep exploratory drilling campaign in the 1950's, drilling companies shifted rather quickly from cable-tool to rotary drilling. As the program developed, electric well logging, particularly gamma-ray and sonic density logging, became relatively widespread in the basin.

Fortunately for geologists studying the Devonian black-shale sequence, the organic matter in the black shales, which imparts the characteristic black color, also has an affinity for accumulating uranium whose gamma-rays are recorded on the gamma-ray log. Because the black shales contain more organic matter and consequently more uranium than do gray shales, the black shales are identified by their strong positive deflections on the gamma-ray log. Also, because the low-density organic matter in the black shales considerably decreases the total density of the rock, black shales commonly show moderate to strong negative deflections on the compensated sonic density log. The electric logs, used in combination with electric logs of drill cores or with scintillation meter logs of outcrops (Ettensohn, Fulton, and Kepferle, 1978), readily facilitate correlating black shales in adjacent wells and correlating named surface units with their subsurface extensions in adjacent wells.

Joseph Schwietering (1970) demonstrated the usefulness of gamma-ray logs in combination with sample logs derived from the description of drill cuttings when he used the logs to trace and correlate Devonian black shales widely in the subsurface of the Appalachian basin. He showed that the Huron Member of the Ohio Shale extended eastward in the subsurface from Ohio outcrops to Steuben County in south-central New York where the member lies in the base of the Canadaway Group (Schwietering's usage); this corresponds to the stratigraphic position of the Dunkirk Shale Member of the Perrysburg Formation (Pepper and de Witt, 1951). Thus he demonstrated the equivalency of the Huron to the

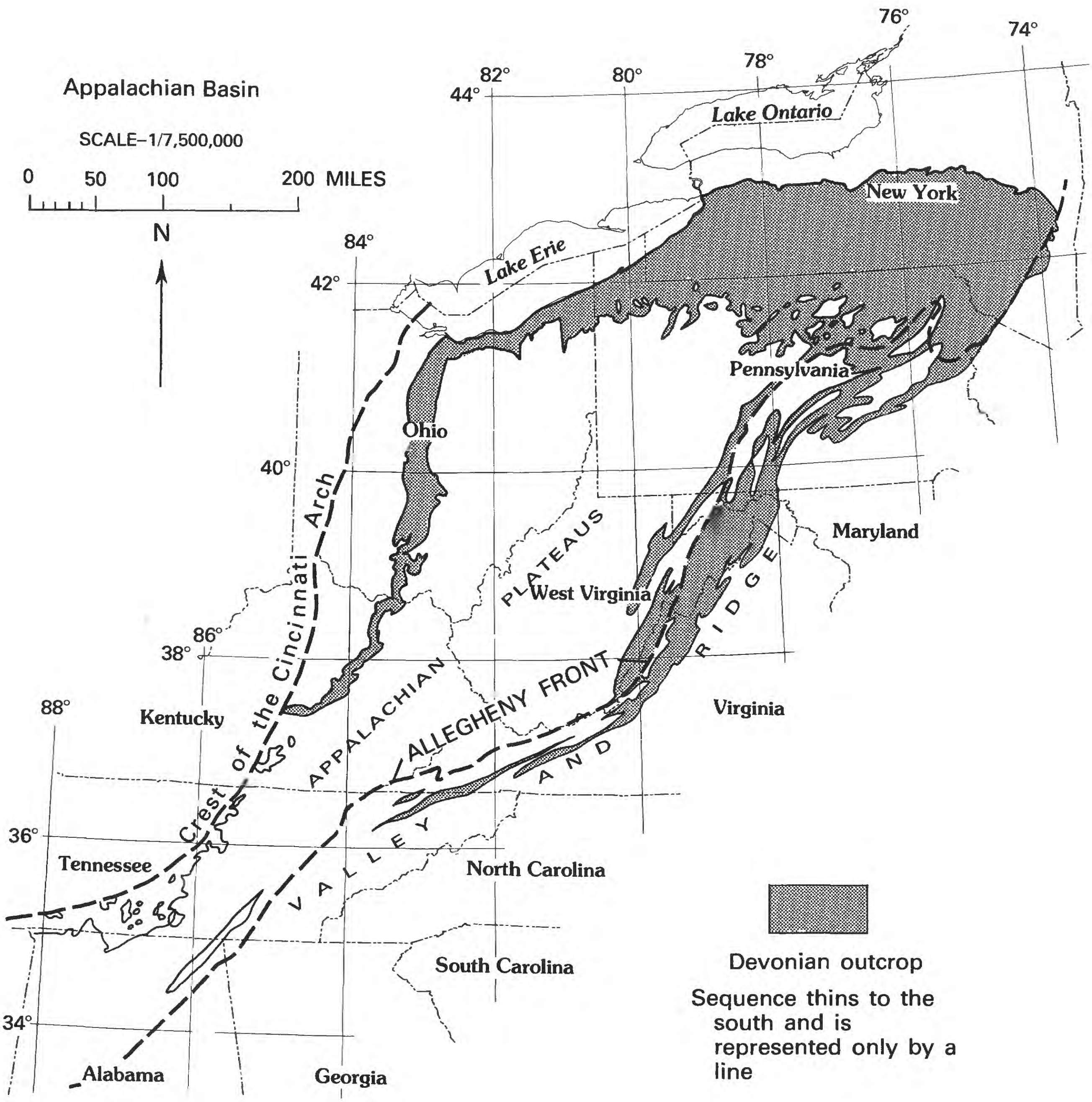

Figure 1. Outcrops of the Devonian rocks in the Appalachian basin in relation to the Appalachian Plateau and the Valley and Ridge segments of the basin.

Dunkirk. Similarly his regional cross sections showed that his key bed "A," a thin bed of black shale in the upper part of the Olentangy Shale of central and northern Ohio, extends to the east into the basal bed of the Java Formation in southwestern New York. Schwietering's bed "A" is the Pipe Creek Shale Member of the Java Formation. The Java Formation's geographic extent in the subsurface is determined by the extent of the Pipe Creek Member. Although they are not present in the central Ohio outcrops because of thinning out against a regional unconformity, black shales of the Rhinestreet Shale Member of the West Falls Formation, the Middlesex Shale Member of the Sonyea Formation, and the Geneseo Shale Member of the Genesee Formation were identified by Schwietering (1970, plates V and VI) in the subsurface of eastern Ohio, extreme northern West Virginia, western Pennsylvania, and in parts of western and central New York where these units crop out and had been described by Pepper, de Witt, and Colton (1956), by Colton and de Witt (1958), and by de Witt and Colton (1959).

In 1976 the U.S. Geological Survey, in cooperation with other groups, began a detailed geologic, geochemical, and geophysical study of the black gas-productive Devonian shales of the Appalachian basin to determine, among other things, the gas-in-place resource of these rocks for the U.S. Department of Energy. Essential to the study was the synthesis of a regional stratigraphic framework for the black shales. From such a framework, geologists could determine the extent and geometry of individual black shales and arrive at volumetric data necessary for resource evaluation. The U.S. Geological Survey took a lead role in coordinating the efforts of geologists from the State Geological Surveys of New York, Pennsylvania, Ohio, West Virginia, Kentucky, and Tennessee, as well as geologists from several universities and research groups.

By correlating many gamma-ray logs with outcrop data, well cuttings, and wells cored by the U.S. Department of Energy for geologic and geochemical data and by relating strongly conspicuous positive anomalies on the gamma-ray logs to named black shales in outcrop, we compiled a regional subsurface stratigraphic framework for the Appalachian basin in which all of the extensive black-shale units are defined in terms of existing surface stratigraphic nomenclature. Corroboration of Schwietering's identification of subsurface black-shale units with named surface units in New York was made by L. V. Rickard (written commun., 1976) of the New York Geological Survey from his detailed work on the Devonian in New York.

At several meetings of all stratigraphers involved in the U.S. Department of Energy's Devonian shale study, the group evaluated the subsurface correlations, discussed existing nomenclature, and agreed to use the surface nomenclature of the New York Devonian as widely as possible in the subsurface of the eastern and central part of the basin. Ohio's surface nomenclature was to be used in much of the subsurface of the western part of the basin. In southern Kentucky and Tennessee, where the Chattanooga Shale had been a long-established formation, the Chattanooga's nomenclature was to be used. By using this nomenclatural framework, we showed the stratigraphic and geographic extent of specific black-shale units on the following U.S. Geological Survey oil charts for parts of the Appalachian basin: OC-80 (Wallace, Roen, and de Witt, 1977), OC-82 (Roen, Wallace, and de Witt, 1978), OC-83 (Wallace, Roen, and de Witt, 1978), OC-85 (Kepferle, Wilson, and Ettensohn, 1978), and OC-86 (West, 1978). These oil charts presented our regional correlations in preliminary form and were intended for review and comment by geologists familiar with the extent and correlation of the Devonian shales and with the Appalachian basin. To date, only minor modifications to the correlation and extent of some units have been suggested, and these changes have been included in the discussion of the geographic extent of specific units that follows. The areal extent of the usage of the formation names Ohio Shale, Chattanooga Shale, and New Albany Shale in south-central Kentucky was revised by de Witt (1981). His changes have been incorporated in this report.

In this paper, stratigraphic names are applied to large areas of the Appalachian basin subsurface. We realize that the subsurface extension of stratigraphic units must be based upon criteria somewhat different from those commonly used for surface geographic extensions. The subsurface criteria may be somewhat limited by the types of well logs available for each well and by the density of subsurface data, which is a factor of the number of wells drilled in a specific area. In some areas, the distal fringe of a thick, named unit may project as a recognizable bed into the body of another well-established surface or subsurface unit. In such instances, the thinning bed of black shale is recognized as a formally named bed in the thicker unit. For example, the Rhinestreet Shale Member of the West Falls Formation, which underlies much of the west-central part of the Appalachian basin (see fig. 5), thins from about 200 ft in western New York to a bed about 5 ft thick in the basal part of the Dowelltown Member of the Chattanooga Shale in central Tennessee. The Rhinestreet Shale Bed is formally recognized here as a unit in the basal part of the Dowelltown Member of the Chattanooga Shale throughout the extent of the Dowelltown Member in the southwest part of the Appalachian basin.

The Marcellus Shale (fig. 2) underlies much of southern New York, Pennsylvania, eastern Ohio, western West Virginia, western Maryland, and eastern Kentucky. In northern Virginia and adjacent West Virginia, the Marcellus and the younger Burket Shale Member of the Harrell Shale merge and lose their identities in the thick Millboro Shale. Thus in west-central Virginia and contiguous eastern West Virginia, the name Millboro is appropriate for the total Devonian black-shale sequence. In eastern Tennessee along Clinch Mountain, the basal 80 ft of Devonian black shale heretofore believed to be Marcellus (Dennison and Boucot, 1974, p. 84) has been shown by its conodont fauna to be predominantly the Rhinestreet Shale Member of the West Falls Formation of Late Devonian age and not the Marcellus Shale of Middle Devonian age.

The Geneseo Shale Member of the Genesee Formation (fig. 3), which overlies the Tully Limestone in central New York and north-central Pennsylvania, is present in the subsurface of New York, western Pennsylvania, eastern Ohio, western Maryland, and part of northern West Virginia. The Burket Shale Member of central Pennsylvania overlies the Tully along the Allegheny front and occupies the same stratigraphic position as the Geneseo Shale Member. The Burket Member is here shown as a correlative of the Geneseo Member although the relation of the two units has not been corroborated by conodont faunas.

The Middlesex Shale Member of the Sonyea Formation (fig. 4) underlies southwestern New York, much of western Pennsylvania, extreme western Maryland, a small segment of southwestern Ohio, and much of northern and central West Virginia. The black shale of the Middlesex Member grades eastward into lighter gray rocks and therefore does not crop out along the Allegheny front in Pennsylvania, Maryland, or West Virginia.

The Rhinestreet Shale Member of the West Falls Formation (fig. 5) is one of the more extensive Devonian black shales. We have traced it in the subsurface under parts of southwestern New York, western Pennsylvania, eastern Ohio, western West Virginia, eastern Kentucky, extreme western Virginia, and a small segment of eastern Tennessee along Clinch Mountain in the Valley and Ridge fold belt. In parts of south-central Kentucky and most of Tennessee, the lower black-shale part (Rhinestreet Bed) of the Dowelltown Member of the Chattanooga Shale (Hass, 1956; Conant and Swanson, 1961) is a distal fringe of the main body of the Rhinestreet Member.

The Pipe Creek Shale Member of the Java Formation (fig. 6), although generally less than 25 ft thick, is characterized throughout its extent in the subsurface by a very strong positive deflection on the gamma-ray log and by a corresponding markedly negative deflection on the compensated sonic-density log. The distinctive log signature of the Pipe Creek Member had been used for

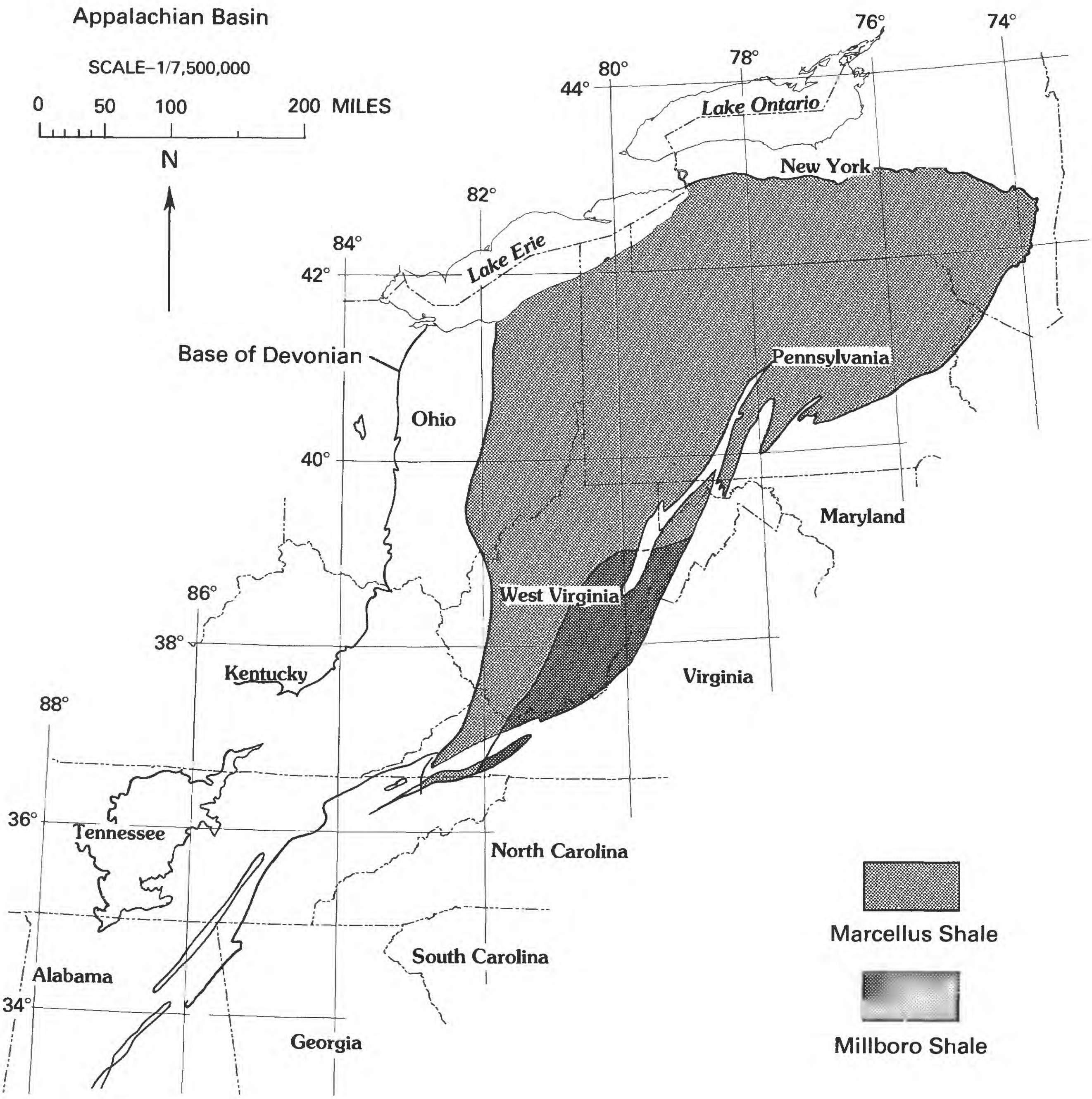

Figure 2. Geographic extent of the Marcellus and Millboro Shales in the Appalachian basin.

local subsurface well-log correlation by some Appalachian basin geologists before we determined the identity of the unit. Our studies demonstrated that the Pipe Creek Shale Member underlies much of western New York, western Pennsylvania, eastern Ohio, western West Virginia, southwestern Virginia, eastern Kentucky, and northeastern Tennessee. Although the Pipe Creek has not been identified in the folded and faulted outcrops of southwestern Virginia to date, it is present in wells in the Greendale syncline of Scott and Washington Counties, in the second thrust sheet southeast of the Allegheny front in the Valley and Ridge. Presumably the Pipe Creek Shale Member would be found if outcrops were adequate for its exposure.

The Dunkirk Shale Member of the Perrysburg Formation (fig. 7) is the eastern part of the most extensive sheet of black shale in the Devonian shale sequence. Other correlatives of this black-shale unit are the Huron Member of the Ohio Shale and the lower part (Huron Bed) of the Gassaway Member of the Chattanooga Shale. We have determined that the Dunkirk Member, which is the youngest of the extensive Upper Devonian black shales in

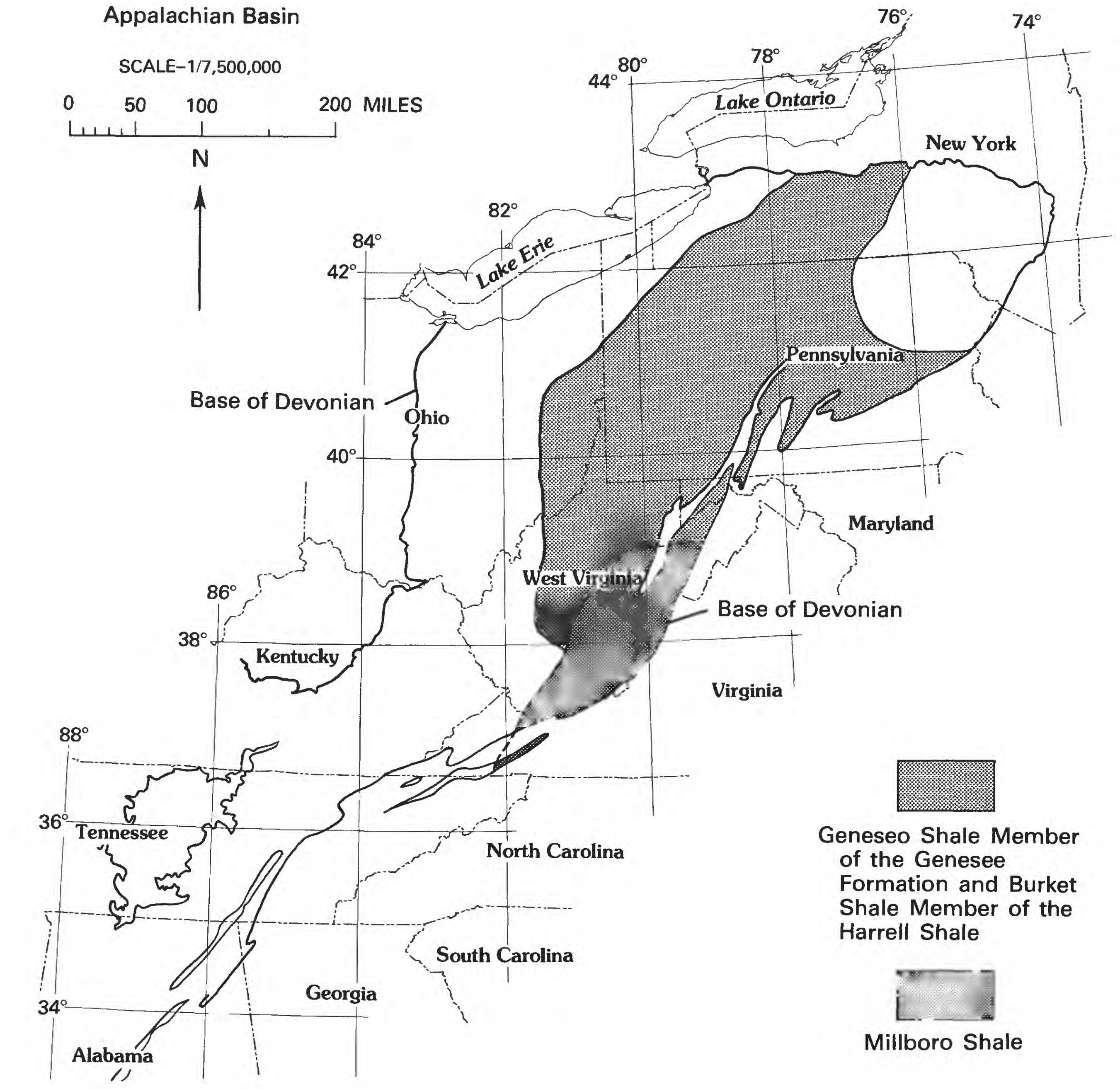

Figure 3. Geographic extent of the Geneseo Shale Member of the Genesee Formation, the Burket Shale Member of the Harrell Shale undivided, and the Millboro Shale in the Appalachian basin.

outcrops in New York, underlies much of western New York, western Pennsylvania, and parts of adjacent eastern Ohio and western panhandle of West Virginia.

The Huron Member underlies most of eastern Ohio, a small part of western Pennsylvania, much of West Virginia, and much of eastern Kentucky (fig. 7). It is represented locally in a small part of southwestern Virginia. In southern Kentucky, Tennessee, northwestern Georgia, and northeastern Alabama we recognize the Huron Bed in the lower part of the Gassaway Member of the Chattanooga Shale. The Huron Bed is here a reduction in rank and a geographic extension of the Huron Member. The Chattanooga with its two members, the Dowelltown and the Gassaway, is present in much of the subsurface of Tennessee. Locally the Dowelltown is absent by nondeposition on shoal areas in the Chattanooga Sea (Conant and Swanson, 1961, p. 18).

The Cleveland Member of the Ohio Shale (fig. 8) is restricted largely to the western part of the basin. It underlies much of central Ohio and eastern Kentucky. In southern Kentucky, it becomes the Cleveland Bed of the Gassaway Member of the Chattanooga Shale. It

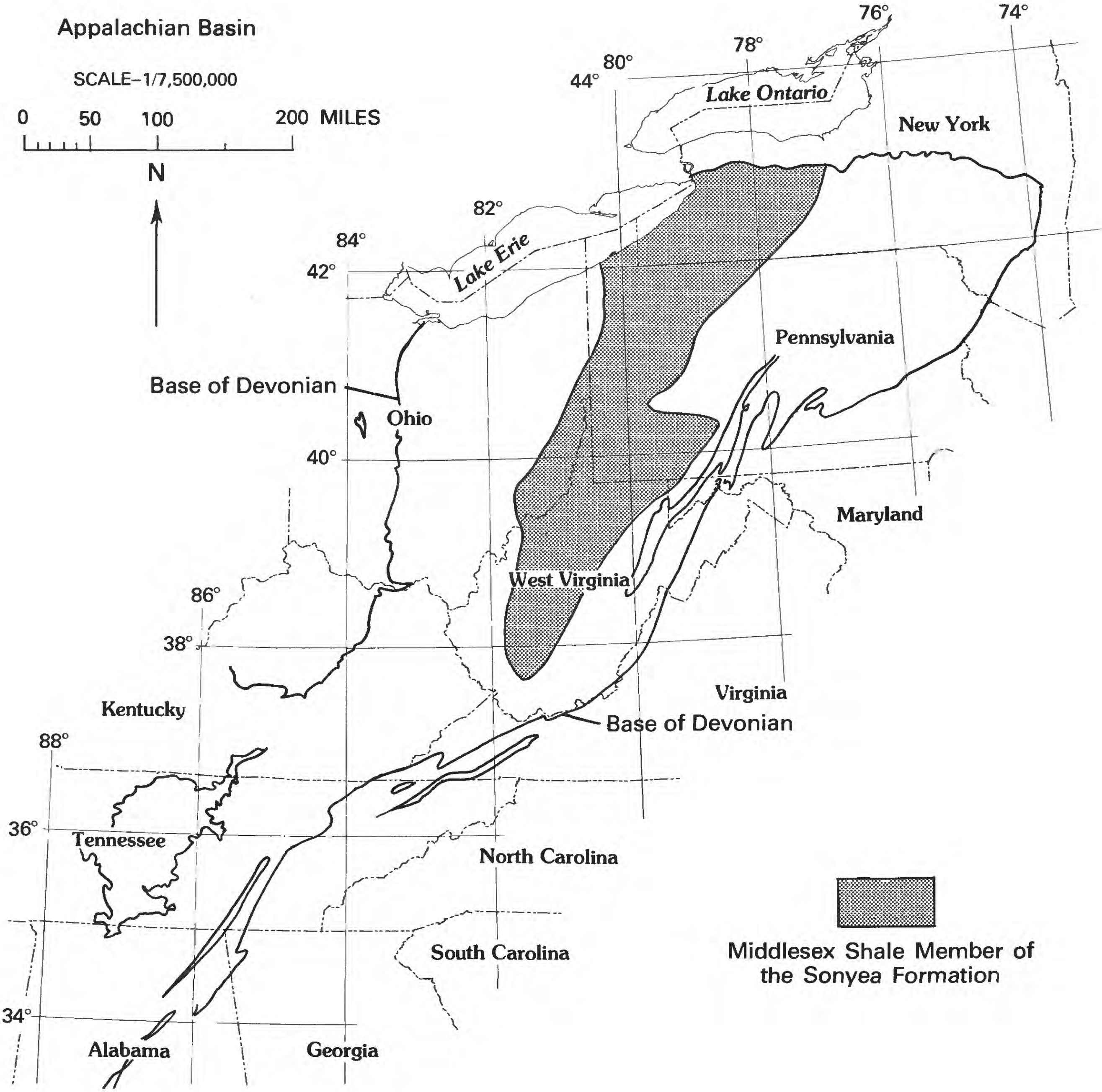

Figure 4. Geographic extent of the Middlesex Shale Member of the Sonyea Formation in the Appalachian basin.

coalesces with the lower Huron Bed to form the Gassaway Member. The Cleveland Bed is here a reduction in rank and a geographic extension of the Cleveland Member. The Gassaway Member underlies the Cumberland River Valley in south-central Kentucky and all of Tennessee except for the fault slices adjacent to the Virginia border in the Valley and Ridge.

The Sunbury Shale of Early Mississippian age (fig. 9) is the youngest black shale in this sequence of marine Appalachian black shales. Throughout much of the western part of the basin, the Sunbury Shale is separated from the older Cleveland Shale by the Bedford Shale and the overlying Berea Sandstone (Pepper, de Witt, and Demarest, 1954). It underlies much of eastern Ohio, western West Virginia, and eastern Kentucky. In southern Kentucky, the gray rocks of the Bedford Shale and Berea Sandstone deltaic sequence feather out below the Sunbury, and locally the Sunbury correlates with the upper part of the Gassaway Member of the Chattanooga Shale. Because the Sunbury correlative is lithologically indistinguishable from the remainder of the Gassaway Member, its presence can be detected only by the

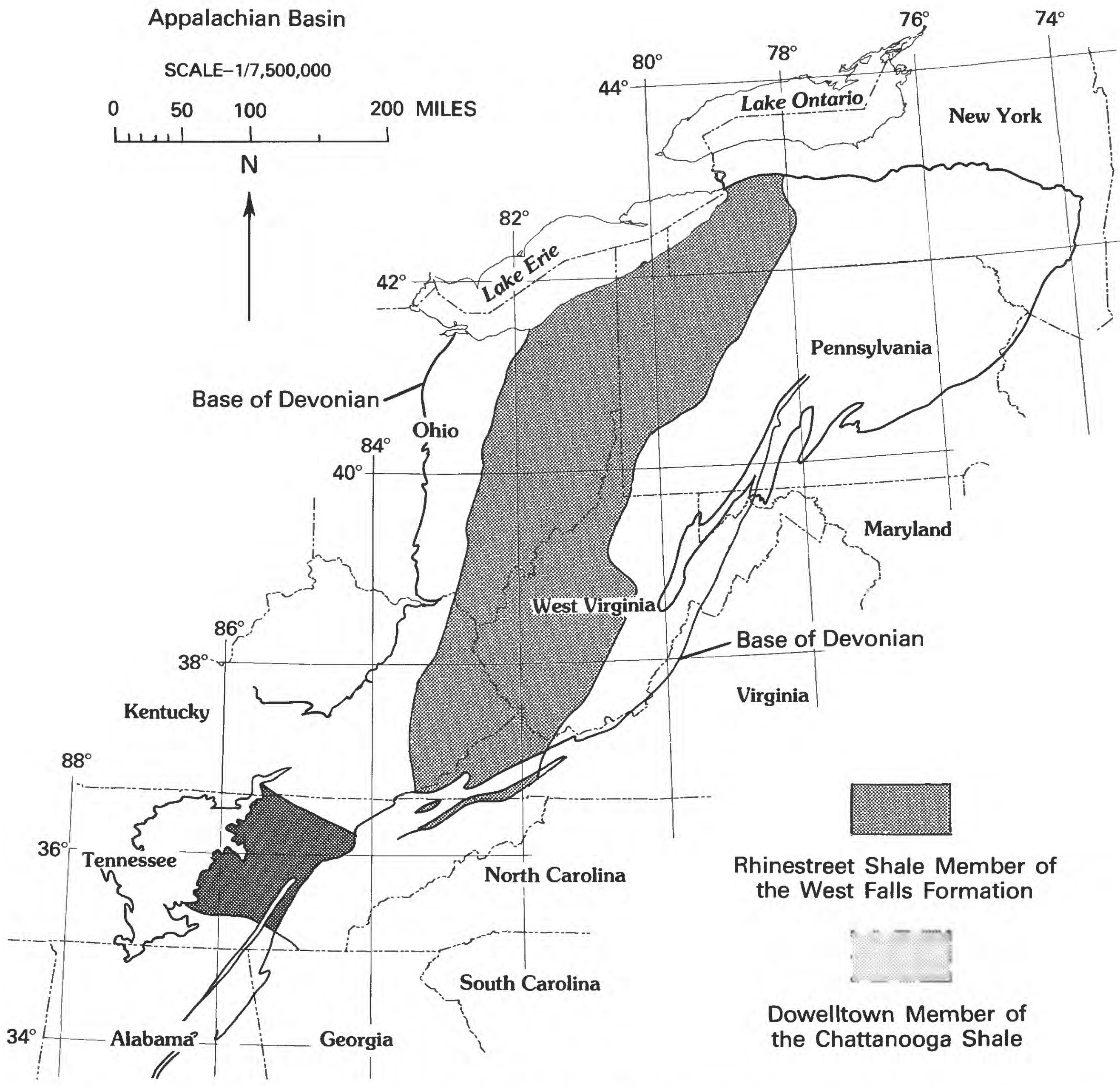

Figure 5. Geographic extent of the Rhinestreet Shale Member of the West Falls Formation and extent of the Rhinestreet Shale Bed of the Dowelltown Member of the Chattanooga Shale in the Appalachian basin.

occurrence of a conodont fauna characterized by many species of *Siphonodella* and by a slightly more positive response on the gamma-ray log.

CONCLUSIONS

In general, the subsurface correlation of black shales is now well established for much of the basin. As drilling activities spread eastward into the basin, particularly into the Valley and Ridge, and as more closely spaced data points become established along the area of intertonguing black and gray rocks, local correlations may be revised as unit boundaries are better delineated. Resolution of structural complexities in the Valley and Ridge must precede solution of some local problems in the black-shale stratigraphy of this province.

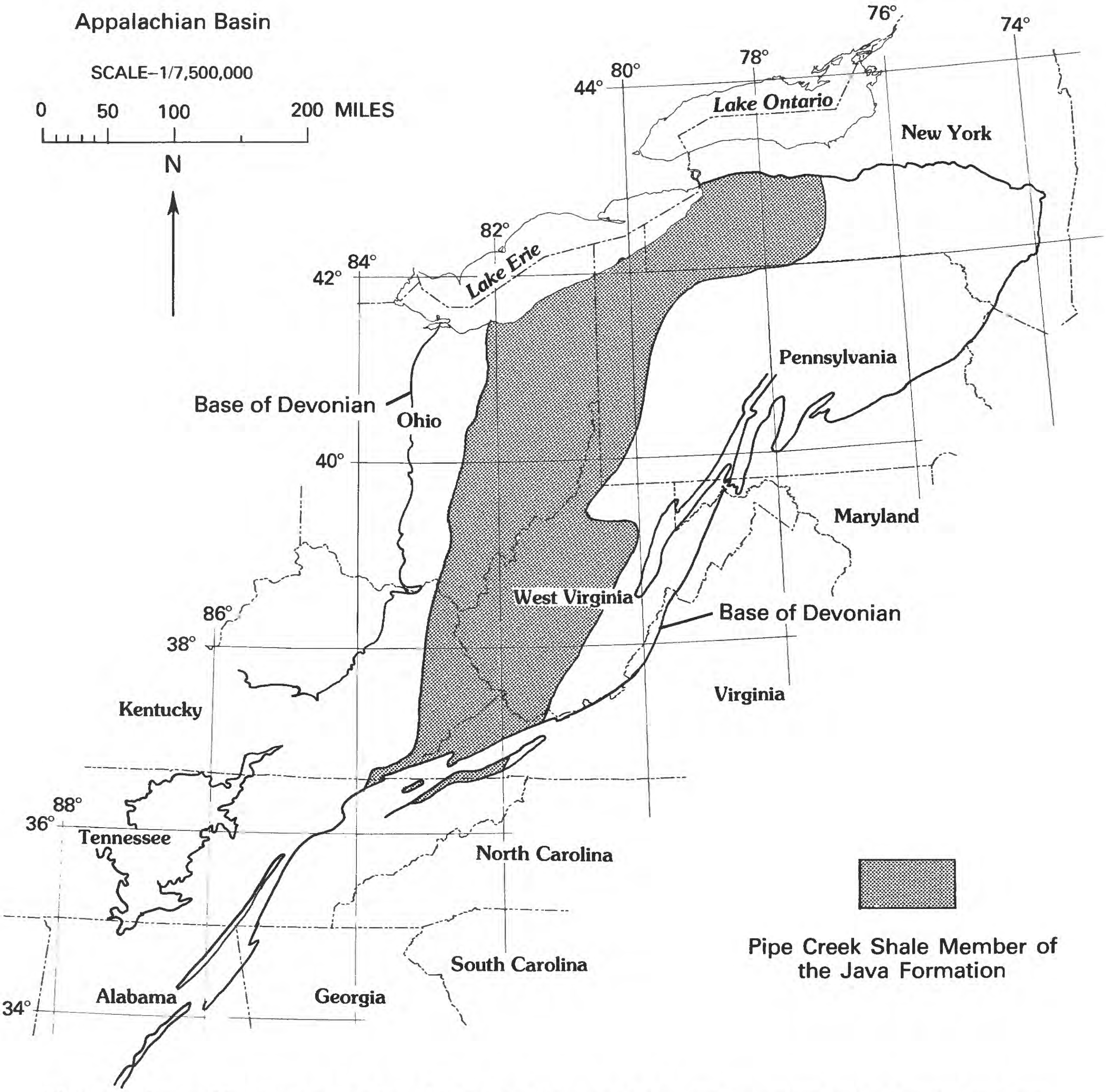

Figure 6. Geographic extent of the Pipe Creek Shale Member of the Java Formation in the Appalachian basin.

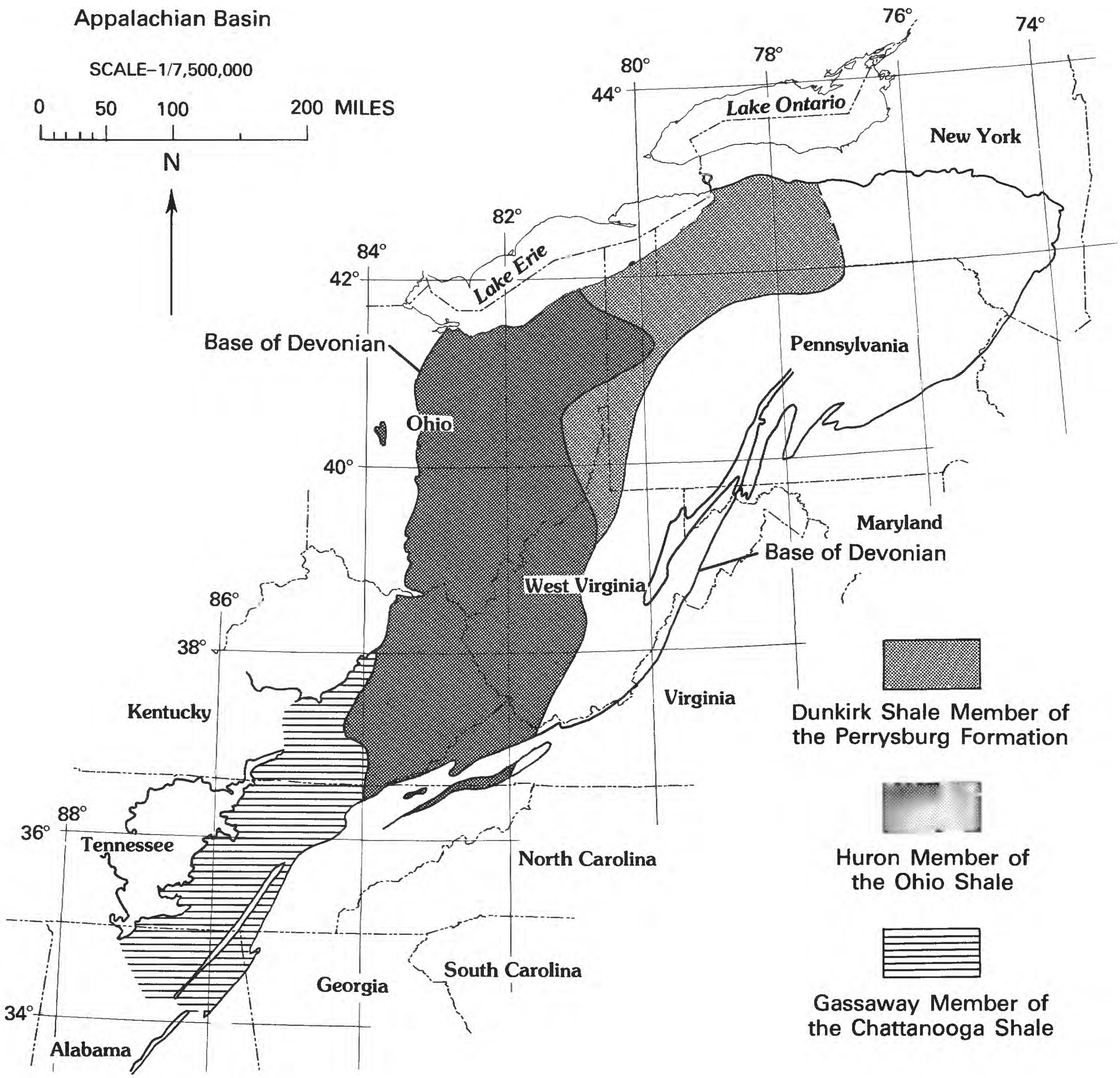

Figure 7. Geographic extent of the Dunkirk Shale Member of the Perrysburg Formation, the Huron Member of the Ohio Shale, and the Huron Bed of the Gassaway Member of the Chattanooga Shale in the Appalachian basin.

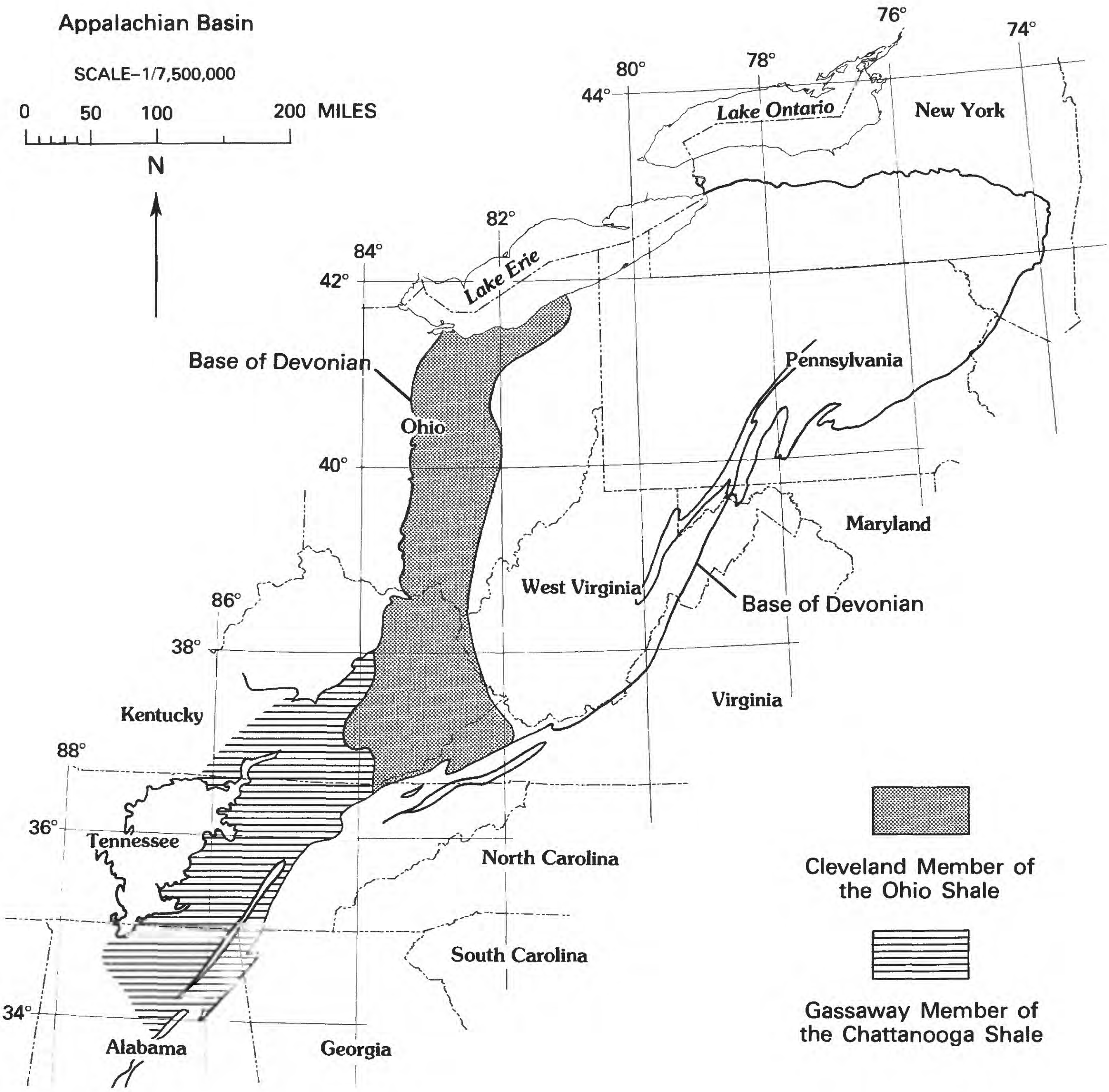

Figure 8. Geographic extent of the Cleveland Member of the Ohio Shale and extent of the Cleveland Bed of the Gassaway Member of the Chattanooga Shale in the Appalachian basin.

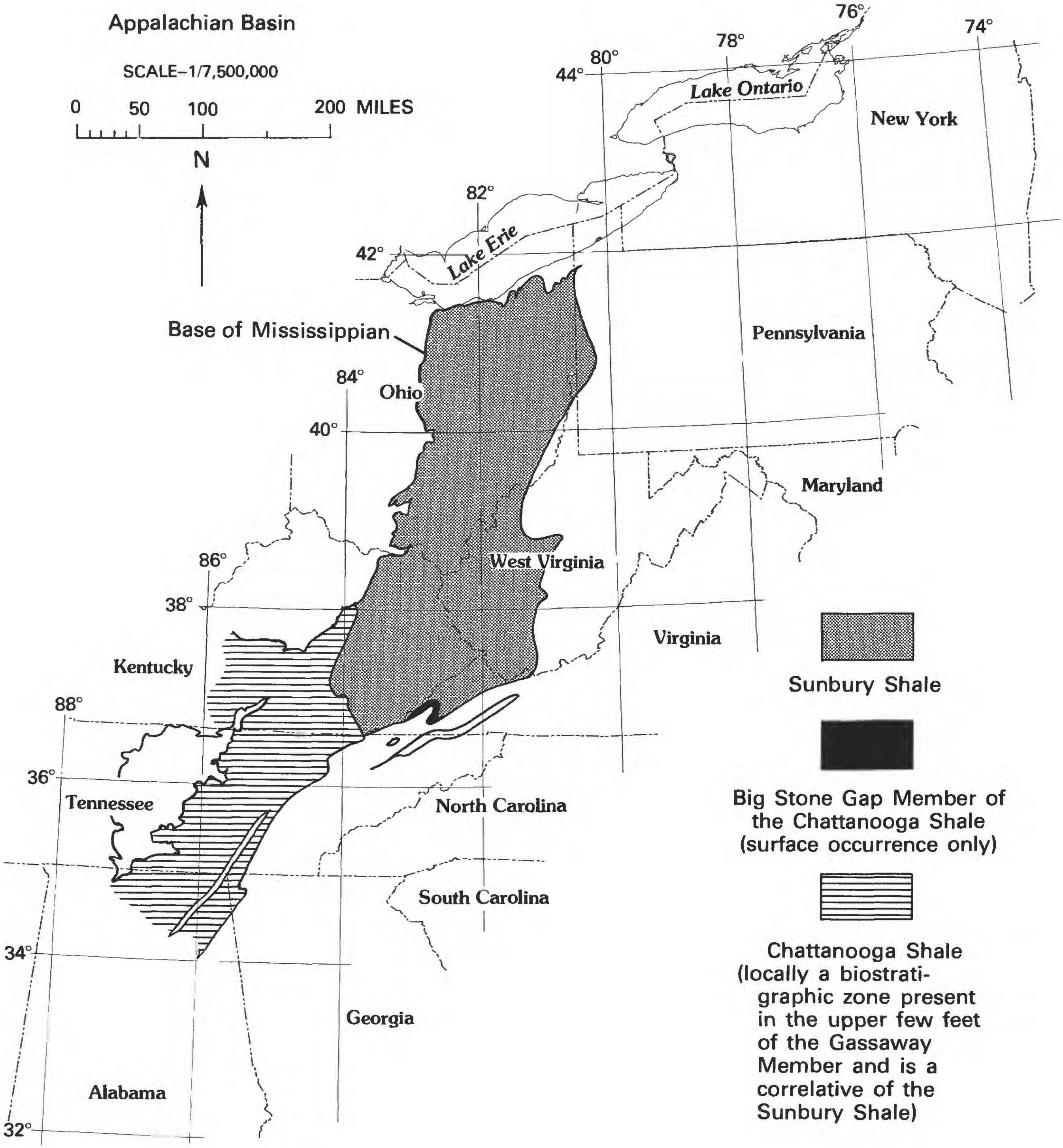

Figure 9. Geographic extent of the Sunbury Shale and of the Gassaway Member of the Chattanooga Shale in the Appalachian basin. A biostratigraphic zone in the top of the Gassaway is the local correlative of the Sunbury.

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STRATIGRAPHIC NOMENCLATURE IN THE RICHARD RUSSELL AND HELEN THRUST SHEETS, GEORGIA AND NORTH CAROLINA

By Arthur E. Nelson[1] and Kenneth A. Gillon[2]

Abstract

Mapping in the Richard Russell and Helen thrust sheets in the northwest part of the Greenville $1^{\circ} \times 2^{\circ}$ quadrangle, Georgia, South Carolina, and North Carolina, has shown the necessity to adopt and to name five units of Late Proterozoic to early Paleozoic age for usage by the U.S. Geological Survey. A large part of the Helen thrust sheet is underlain by formations of the Helen Group, which consists of the Nacoochee, Horton, and Robertstown Formations. The names Helen Group and Nacoochee Formation are herein adopted, and the Horton Formation and Robertstown Formation are herein named for use in Georgia and North Carolina. In addition, much of the Richard Russell thrust sheet is underlain by the Richard Russell Formation, which also is herein adopted for use in Georgia and North Carolina.

INTRODUCTION

This report formally adopts for use by the U.S. Geological Survey names given by Gillon (1982) to several metamorphic rock units that form significant parts of the Richard Russell and the Helen thrust sheets in the northwest part of the Greenville $1^{\circ} \times 2^{\circ}$ quadrangle, Georgia, South Carolina, and North Carolina (fig. 1).

Rocks underlying the eastern Blue Ridge of North Carolina were considered to be parts of the Carolina and Roan Gneisses of Archean age (Keith, 1904, 1907). Jones (1909), LaForge and Phalen (1913), Pardee and Park (1948), and Crickmay (1952) extended the usage of the names Carolina and Roan Gneisses to crystalline rocks that underlie a large part of northern Georgia. Until Gillon's work, the study area had not been mapped in detail, but, along strike to the southwest, rocks of the Dahlonega Gold Belt correlate with rocks named by Gillon in the Helen thrust sheet (fig. 1). Crickmay (1952) included rocks of the Dahlonega Gold Belt (Yeates and McCallie, 1896; Jones, 1909) as the Wedowee-Ashland belt. Hurst (1973) included much of the Wedowee-Ashland belt as part of the Ashland Group. In the area west and southwest of Canton, Ga. (fig. 1), Abrams and McConnell (1981) designated rocks that I believe to be in the Helen thrust sheet and that were formerly assigned to the Wedowee-Ashland belt as the New Georgia Group.

Gillon (1982) gave the name Helen Group to rocks that constitute the Helen thrust sheet in the vicinity of Helen, Ga. Some of the rock units of the Helen Group probably correlate with rock units of the New Georgia Group and appear to be in the same thrust sheet. Since the geology between the two groups is incompletely known, however, stratigraphic correlations between the Canton vicinity and the Helen area are uncertain.

Some of Gillon's named rock units, which constitute much of the Helen and Richard Russell thrust sheets in Georgia and North Carolina, are here adopted for usage by the U.S. Geological Survey. These units are discussed below.

HELEN GROUP

Gillon (1982) called rock assemblages between the Dahlonega and Shope Fork faults in the vicinity of Helen, Ga., the Helen Group. The name Helen Group is here adopted for these rock sequences present in the Helen thrust sheet (Nelson, in press) in Georgia and North Carolina. Gillon recognized and named three formations within the Helen Group—the Nacoochee, Chattahoochee, and Unicoi Park Formations. The name Chattahoochee is here changed to Horton Formation, and the name Unicoi Park is here changed to Robertstown Formation. These names are adopted for use by the U.S. Geological Survey.

The thickness of the Helen Group is unknown. The group is fault bounded; that is, the southeast part of the Nacoochee Formation is truncated by the Dahlonega fault, and the northwest part of the Robertstown Formation is truncated by the Shope Fork fault. In addition, the rocks have been isoclinally folded. The age of the Helen Group is uncertain, but most of the group probably correlates with the New Georgia Group. Therefore, the group is tentatively assigned a Late Proterozoic to early Paleozoic age (McConnell and Abrams, 1984). Formations of the Coweeta Group (Hatcher, 1979) are exposed in the Helen thrust sheet near the North Carolina border. These rocks appear to overlie the rocks of the Helen Group, but the nature of the contact is uncertain.

Nacoochee Formation

The Nacoochee Formation was named by Gillon (1982) for fresh and weathered exposures of amphibolite and aluminous schist along the Chattahoochee River at Nacoochee, Ga., in the Helen 7 1/2-min quadrangle. The name Nacoochee Formation is here adopted for usage in Georgia and North Carolina. Rock exposures along the Chattahoochee River that extend for 0.3 mi from the Nacoochee bridge (Ga. Hwy. 17, White County, Ga.) are designated as the type locality (fig. 2). The Nacoochee is gradational northwestward into the adjacent Horton Formation. The southeast side of the formation is truncated by the Dahlonega fault.

Gillon describes the Nacoochee as an interlayered sequence of generally thick-layered amphibolite and graphitic-aluminous schist, with lesser amounts of micaceous metasandstone, feldspathic metasiltstone, epidote quartzite, and biotite schist.

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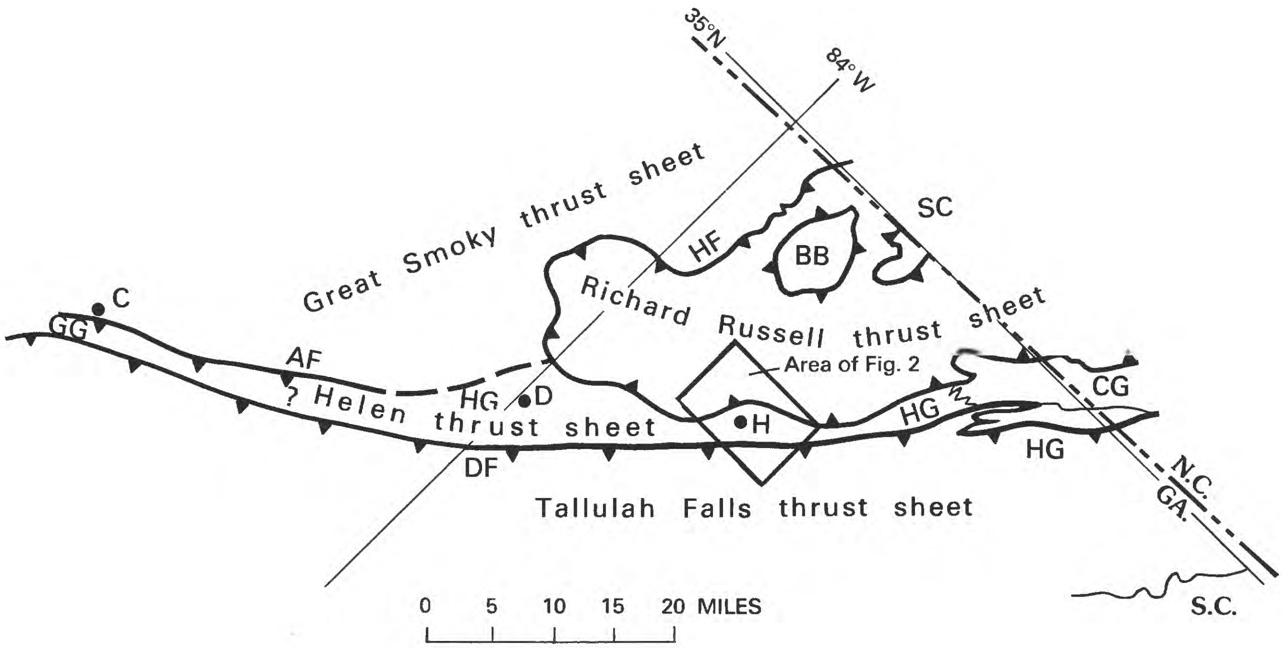

Figure 1. Generalized tectonic map of the northwest part of the Greenville quadrangle and adjacent areas. C, Canton; D, Dahlonega; H, Helen; BB, Brasstown Bald window; SC, Shooting Creek window; HF, Hayesville thrust fault; SF, Shope Fork fault; AF, Allatoona fault; DF, Dahlonega fault; GG, New Georgia Group, undivided; CG, Coweeta Group, undivided; HG, Helen Group, undivided. Allatoona fault dashed where implied.

Horton Formation

Gillon (1982) named an interlayered sequence of metasandstone and metasiltstone, exposed along the Chattahoochee River, the Chattahoochee Formation. This name is here changed to the Horton Formation, after Horton Creek, which flows into the Chattahoochee River south of Robertstown, for usage in Georgia and North Carolina. Exposures along the Chattahoochee River, in the city of Helen, White County, Ga., are designated as the type locality (fig. 2). The southeast part of the formation is conformable and gradational into the adjacent Nacoochee Formation, and its northwest part is conformable with the adjacent Robertstown Formation.

The Horton consists of an interlayered sequence of up to 70 percent argillaceous and feldspathic metasandstone and metasiltstone (locally sulphidic), 20 percent mica schist, and 10 percent aluminous schist and amphibolite. Amphibolite and sulphidic metasiltstone are most abundant in the southeastern part of the formation.

Robertstown Formation

Gillon (1982) named the metasandstone and schist exposed at Unicoi State Park, the Unicoi Park Formation. This name is here changed to the Robertstown Formation, after nearby Robertstown, for usage in Georgia and North Carolina. Exposures along a road on the west side of an unnamed lake at Unicoi State Park, White County, Ga., are designated the type locality (fig. 2). The southeast side of the formation is conformable to the adjacent Horton Formation, but the northwest side of the formation is truncated by the Shope Fork fault.

Gillon indicated that the formation consists of as much as 75 percent thick to thinly layered, coarse-grained, two-mica feldspathic metasandstone interlayered with about 20 percent thinly layered, coarse-grained mica schist and 5 percent amphibolite and pebbly quartz metasandstone. Thin, aluminous schist partings between metasandstone beds characterize the formation.

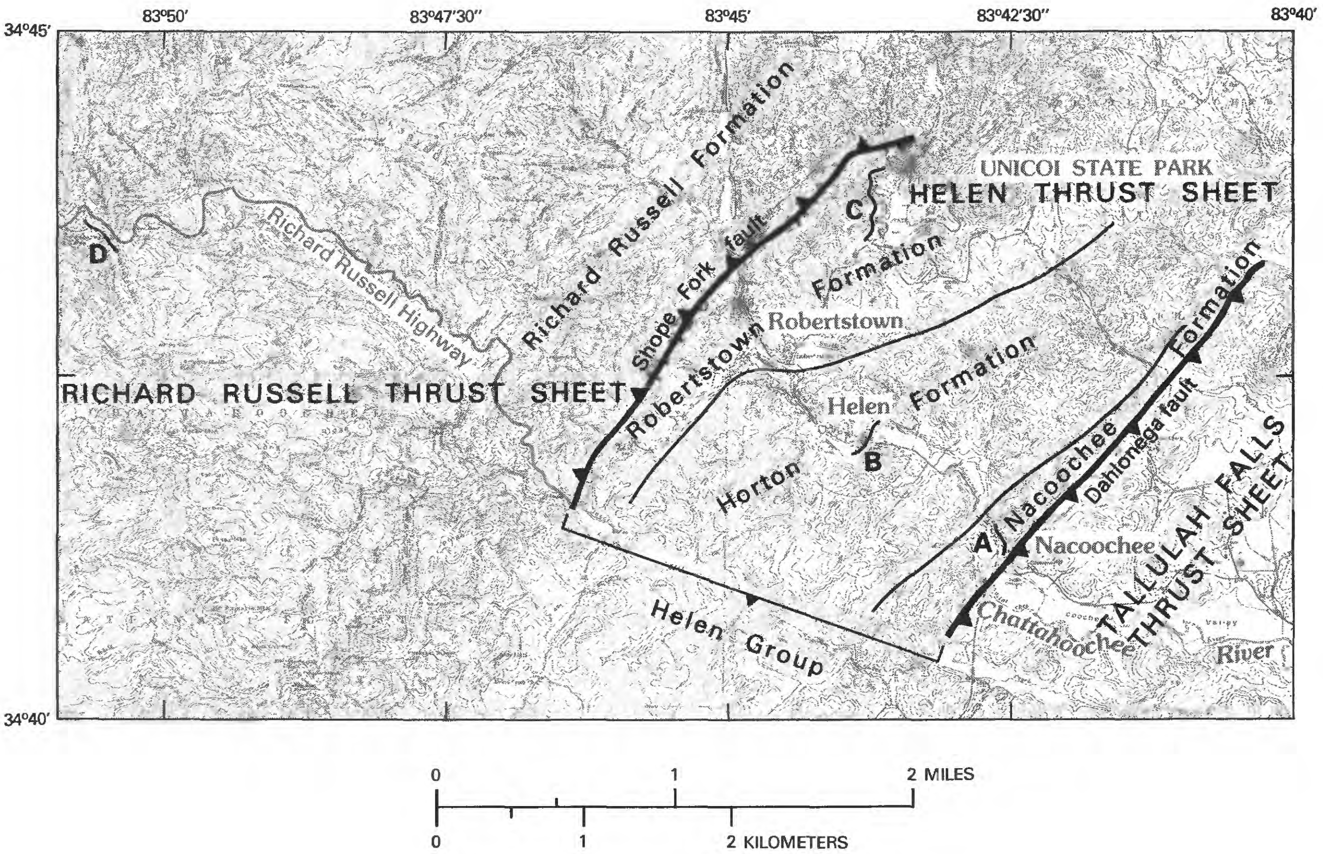

Figure 2. Generalized geology of the Helen area and positions of the type localities for the formations named in this report. A, Nacoochee Formation; B, Horton Formation; C, Robertstown Formation; D, Richard Russell Formation.

RICHARD RUSSELL FORMATION

In the Richard Russell thrust sheet, exposures of biotite gneiss, gneissic metasandstone, metasandstone, and schist along the Richard Russell Highway (Ga. Hwy. 348) were named the Richard Russell Formation by Gillon (1982). This name is here adopted for usage in Georgia and North Carolina. Exposures extending for 0.3 mi along the Richard Russell Highway northwest from Hogpen Gap in Union County, Ga., are designated as the type locality (fig. 2). The thickness of the formation is unknown; the layering is transposed into the regional foliation, and the rocks of the formation are recumbently folded. The formation is bounded on the northwest by the Hayesville fault and on the southeast by the Shope Fork fault (fig. 1).

Gillon reports that the formation contains approximately 65 percent thinly layered, migmatitic biotite gneiss, 15 percent metasandstone, 15 percent aluminous biotite schist and biotite-muscovite schist, and 5 percent calc-silicate granofels, amphibolite, tonalite gneiss, and ultramafic schist. These rock types are variably interlayered. Aluminous schist and amphibolite usually occur at lower elevations, whereas biotite gneiss and metasandstone generally form ridges. Age data are not available for the rock units in the Richard Russell Formation, but the formation correlates with rocks mapped as late Precambrian age (Hadley and Nelson, 1971) in North Carolina. Rocks of the Richard Russell Formation, therefore, are probably of Late Proterozoic age.

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THE ELLENTON FORMATION IN SOUTH CAROLINA— A REVISED AGE DESIGNATION FROM CRETACEOUS TO PALEOCENE

By David C. Prowell[1], Lucy E. Edwards[2], and Norman O. Frederiksen[2]

Abstract

The Ellenton Formation at its type section in the western South Carolina Coastal Plain is about 80 ft (24 m) thick and is composed of clayey sand and silty clay. The presence of a marine palynoflora and beds with abundant carbonaceous matter reflects nearshore marine deposition. Analyses of dinoflagellates and pollen from cored wells near the type section indicate that the age of the Ellenton is Midwayan (Paleocene) and not Cretaceous as was previously suggested.

INTRODUCTION

The Ellenton Formation is a 50- to 80-ft- (15- to 24-m-) thick deposit of clayey sand and silty clay first recognized by Siple (1967) in the subsurface of the updip western South Carolina Coastal Plain. The formation is named after the abandoned town of Ellenton, which was located within the boundaries of the Savannah River Plant, also known as the SRP, prior to Federal aquisition of the land. The plant, located northeast of the Savannah River about 20 mi (32 km) southeast of Augusta, Ga. (fig. 1), produces strategic nuclear materials, and security measures have limited independent investigation of the local subsurface geology. As a result, the name Ellenton Formation was not widely used outside the plant boundaries. Recent studies have shown, however, that the Ellenton is present in adjacent areas in western South Carolina, as well as eastern Georgia. (Faye and Prowell, 1982; Prowell and others, 1985).

The geology and hydrology of the strata beneath the plant area have been discussed in a variety of site-specific reports. A summary of the local geology is presented in Siple (1955, 1967), Christl (1964), Petty and others (1965), Bechtel Corp. (1972, 1973, 1982), Daniels (1974), Marine and Siple (1974), Marine and Root (1978), Marine (1979), Faye and Prowell (1982), and Prowell and others (1985). Many of these reports applied formation names from the Gulf coast region and other Atlantic coast areas to geologic units in this area, but such characterizations are commonly inappropriate because of differences in lithology. To avoid possible confusion, Siple (1967) proposed the name Ellenton Formation to denote lithologically distinct strata of unknown age situated above sand and clay of presumed Cretaceous age and below fossil-dated Eocene strata.

The Ellenton Formation is not known to crop out in the Savannah River Plant area and has only recently been recognized in small stream cuts north of the plant boundary near the community of Hollow Creek, S.C. (fig. 1). Siple (1967, p. 28-31) established the type section in test well 52-C located near the center of the plant (fig. 1). This well was drilled in 1952, and cuttings and geophysical logs from it were used to describe the lithologic character of the Ellenton. Siple (1967) had no direct way of dating the Ellenton, but he suggested a Late Cretaceous or Paleocene age on the basis of stratigraphic relations.

Our study of the Ellenton Formation began with our efforts to make geologic and paleontologic correlations from western South Carolina to central Georgia (fig. 2, modified from Prowell and others (1985)). The geologic section in Prowell and others (1985) shows the stratigraphic equivalence of the Ellenton with the lower part of the Huber Formation in central Georgia (see Buie, 1978, 1980; Tschudy and Patterson, 1975), and their fossil data suggest a correlation with the Rhems Formation of the Black Mingo Group of Sloan (1908) as revised by Van Nieuwenhuise and Colquhoun (1982) in eastern South Carolina. Prowell and others (1985) also suggest that the lower Huber Formation is part of a delta-dominated, fluviomarine sedimentary environment, whereas the Black Mingo strata reflect open marine sedimentation. Our evaluation of the Ellenton type section adds to existing evidence of depositional history in the region.

Samples from test well 52-C on the Savannah River Plant are no longer available for study; therefore, our reevaluation of the Ellenton required that we examine cores from other test wells near the type section. A north-trending geologic section by Siple (1967, pl. 3) shows the distribution of the Ellenton and other formations in various "area" wells within the plant boundaries. The various cluster facilities in the area have alphabetic designations as opposed to formal names. Hence, well 52-C is well number 52 in "C-area." Siple's line of section passes near several wells in the "F-area," approximately 3 mi (5 km) north of the type section in test well 52-C (fig. 3A,B). Siple (1967) reported the presence of the Ellenton strata in test well 21-F in this area, and the cored test wells of the FC-series in this same area provide our lithologic and paleontologic control. Two FC-series wells were projected onto Siple's line of section (fig. 3A) to establish stratigraphic correlation. Samples and geophysical logs from wells FC-3A and FC-5A (fig. 3A,B) were used to obtain the lithologic and paleontologic data described in the following text.

LITHOLOGY

The Ellenton Formation of the Savannah River Plant area generally can be subdivided into a lower clayey sand phase and an upper clay phase. The lower Ellenton phase is about 30 ft (9 m) thick in the F-area and constitutes about half of the total thickness of the formation. The unconformity at the base of the Ellenton is typically marked by a thin bed of very coarse sand and (or) gravel but is most easily recognized on the basis of the appearance of the underlying Cretaceous strata, as defined by Cooke (1936). The uppermost Cretaceous in the area is typically a dense, sticky, sandy clay with extensive red or orange staining in the upper 20 ft (6 m).

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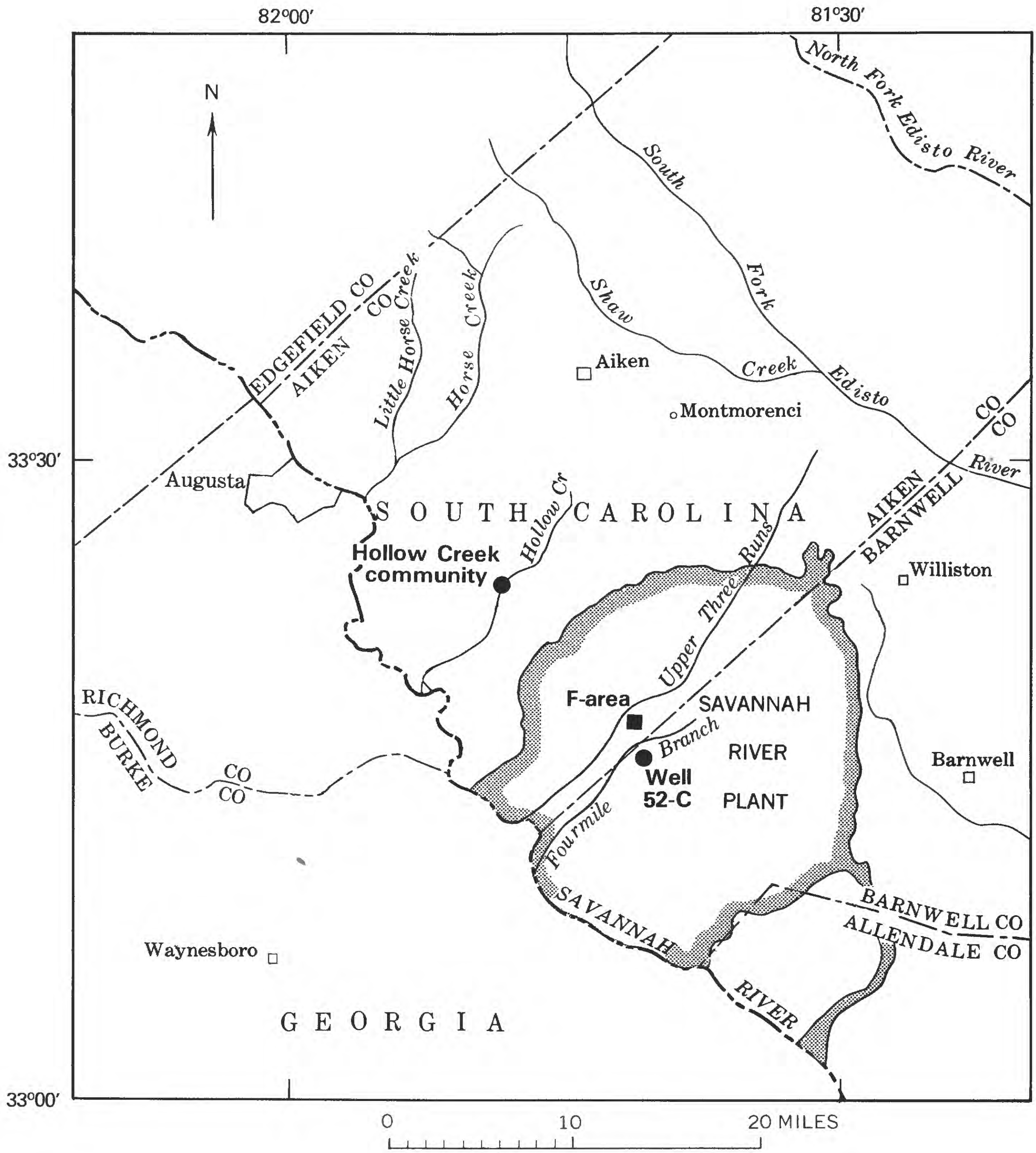
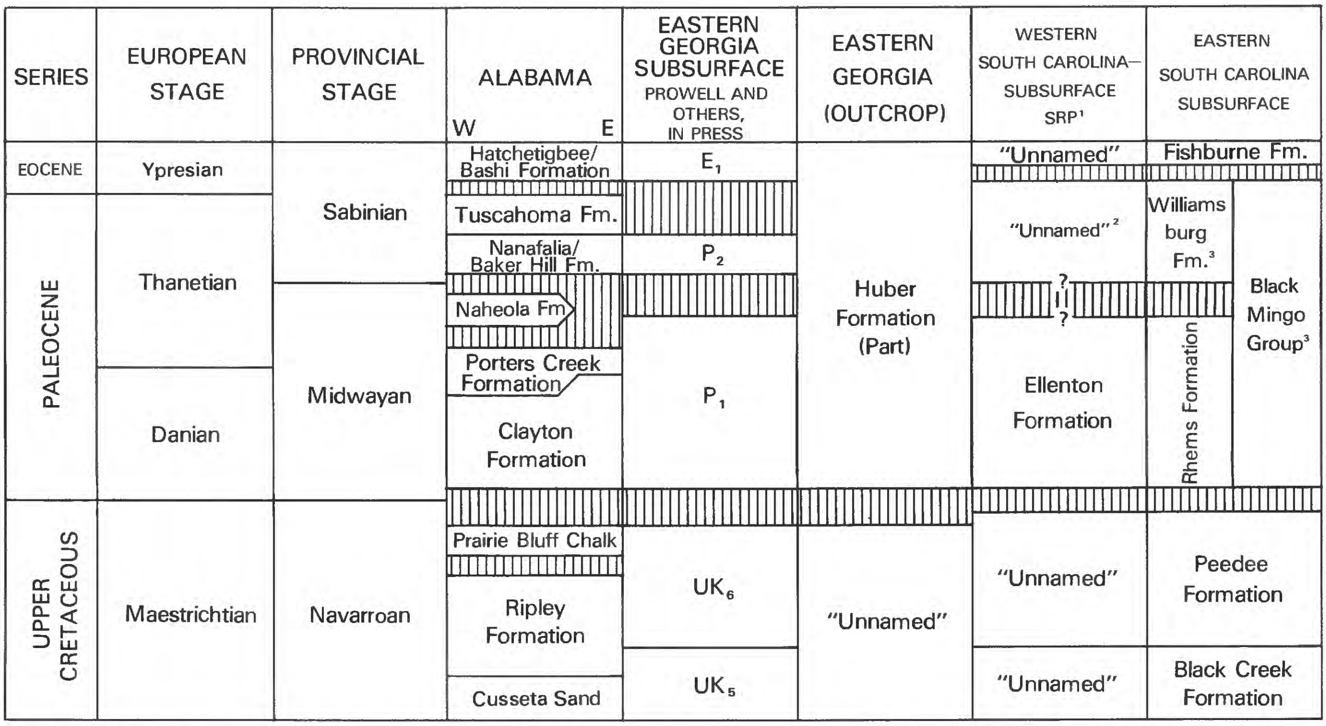


Figure 1.--Location of the Savannah River Plant, F-area, and test well 52-C.

The staining, which provides a distinct contrast with the pale coloration of the Ellenton strata, is presumably the result of subaerial weathering and oxidation of iron-bearing minerals during the late Maestrichtian (Cretaceous) hiatus recognized throughout the Atlantic Coastal Plain. In places where this oxidized clay is absent, the basal sands of the Ellenton can be difficult to distinguish from underlying Cretaceous sand. In these areas, Cretaceous sand is commonly reworked into the lowermost Ellenton, and any abrupt change in lithology is obscured.

The lower Ellenton is composed of medium to coarse sand consisting of subangular quartz in a light-gray to off-white clay matrix. Light-gray to blue-gray quartz forms the majority of the sand fraction and gives the formation a characteristic pale-gray appearance. Finely disseminated carbonaceous debris and fragments of lignite are common. Secondary minerals include muscovite, feldspar, iron-bearing minerals (typically pyrite, marcasite, and siderite), garnet, rutilated quartz, and various unidentified dark heavy minerals. Siple



[1]SRP, Savannah River Plant   [2]Locally referred to as Black Mingo Formation   [3]Of Sloan (1908) as revised by Van Nieuwenhuise and Colquhoun (1982)

Figure 2.--Correlation chart of geologic units (modified from Prowell and others, 1985).

(1967) reported finding gypsum and minor chlorite in the Ellenton type section. These minerals, however, have not been identified in strata in the F-area.

The upper Ellenton phase is largely sandy to silty kaolinitic clay having some thin beds of clayey coarse sand. Mica is typically abundant in the clay layers, and X-ray analysis of sandy clay samples indicates significant amounts of cristobalite. Some clay and clayey silt beds in the upper Ellenton are highly carbonaceous and are dark gray to black. These carbonaceous clay beds are typically less than 5 ft (1.5 m) thick and are well laminated. Some sandy beds contain fragments of lignite in the coarse size fraction, but these beds are not very common. Finely disseminated carbon probably contributes to the characteristic pale-gray color of the upper Ellenton described by Siple (1967). The Ellenton is typically off white in the absence of carbonaceous material, but it is light green or pale yellowish green in layers where cristobalite has been recognized.

The top of the Ellenton Formation can be difficult to define, depending on the nature of the overlying upper Paleocene (Sabinian) or Eocene strata. Most of the units overlying the Ellenton were deposited in open marine environments and are distinguished by their better sorting, trace fossils, and stratification. Upper Paleocene (Sabinian) strata can be easily distinguished from the Ellenton outside the Savannah River Plant, but the two are lithologically similar in the vicinity of the type section (test well 52-C and in the F-area). In these places, the top of the Ellenton must be defined by its characteristic pale-gray to green color or by the localized red, yellow, and purple stains caused by postdepositional weathering at the contact. In the eastern part of the area, the top of the Ellenton is marked by several feet (1 m) of dense black clay that is easily distinguished from the carbonaceous, low-density, fine-grained sandstone of the overlying upper Paleocene (Sabinian) strata.

AGE

Siple (1967) implied that the Ellenton Formation was probably Late Cretaceous on the basis of lithologic similarity to underlying Cretaceous strata. Alternatively, Siple (1967) suggested that the Ellenton might be Paleocene. To better define the age of the strata, samples from the carbonaceous clay layers of the Ellenton Formation from well FC-3A (USGS Paleobotanical Locality R3038) at elevations of +34 ft (+10 m) and +37 ft (+11 m) MSL (mean sea level) were examined for dinoflagellate cysts, and samples from well FC-5A (Locality R3062) at elevations of +38, +45, +61, and +75 ft (+12, +14, +19, and +23 m, respectively) MSL were examined for dinoflagellate cysts and pollen.

DINOFLAGELLATE ASSEMBLAGE

All samples contained moderately well preserved dinoflagellates. Most samples, however, were dominated

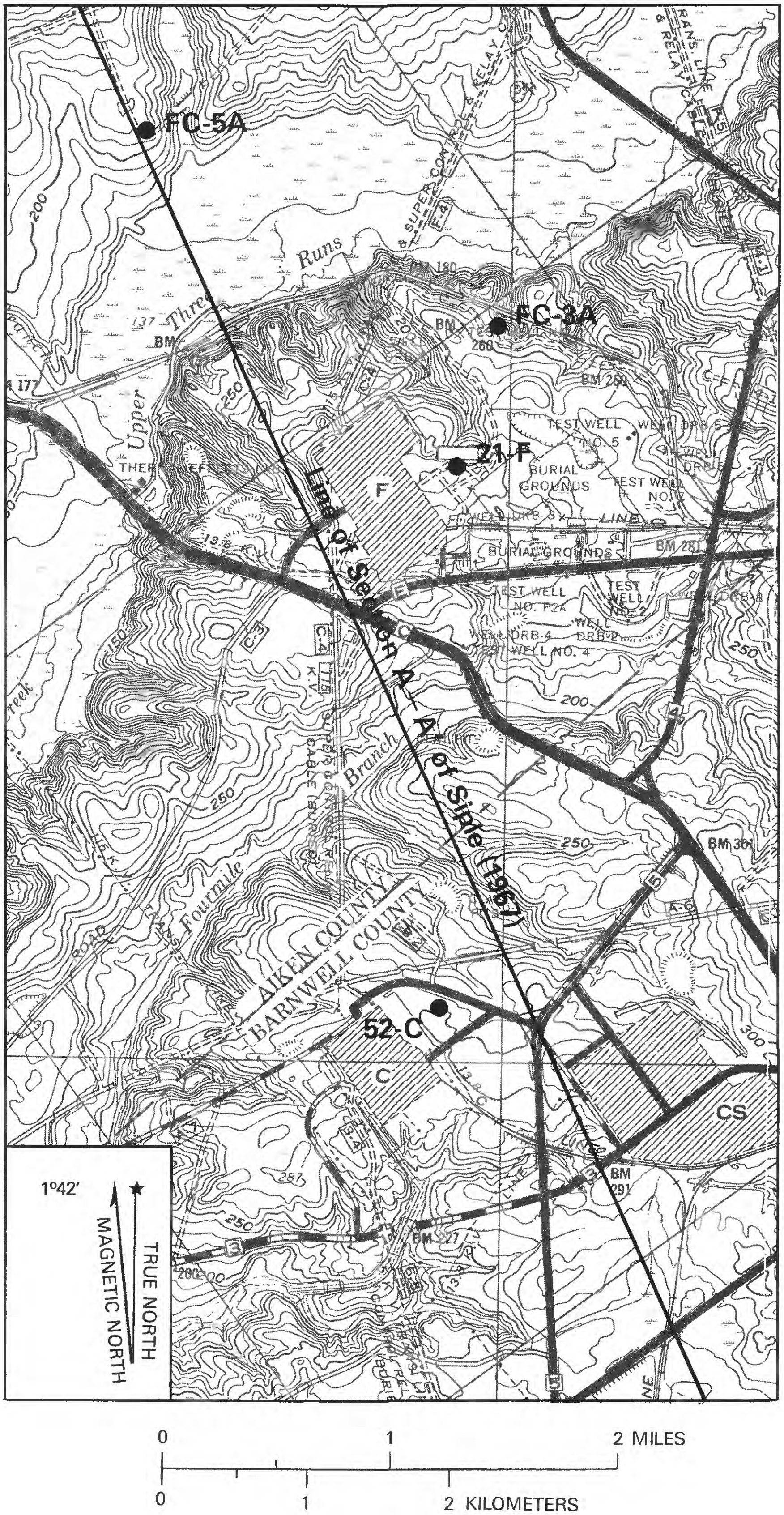


Figure 3A.--Portion of the Savannah River Plant that shows the line of section A-A' of Siple (1967) and location of F-area and critical test wells.

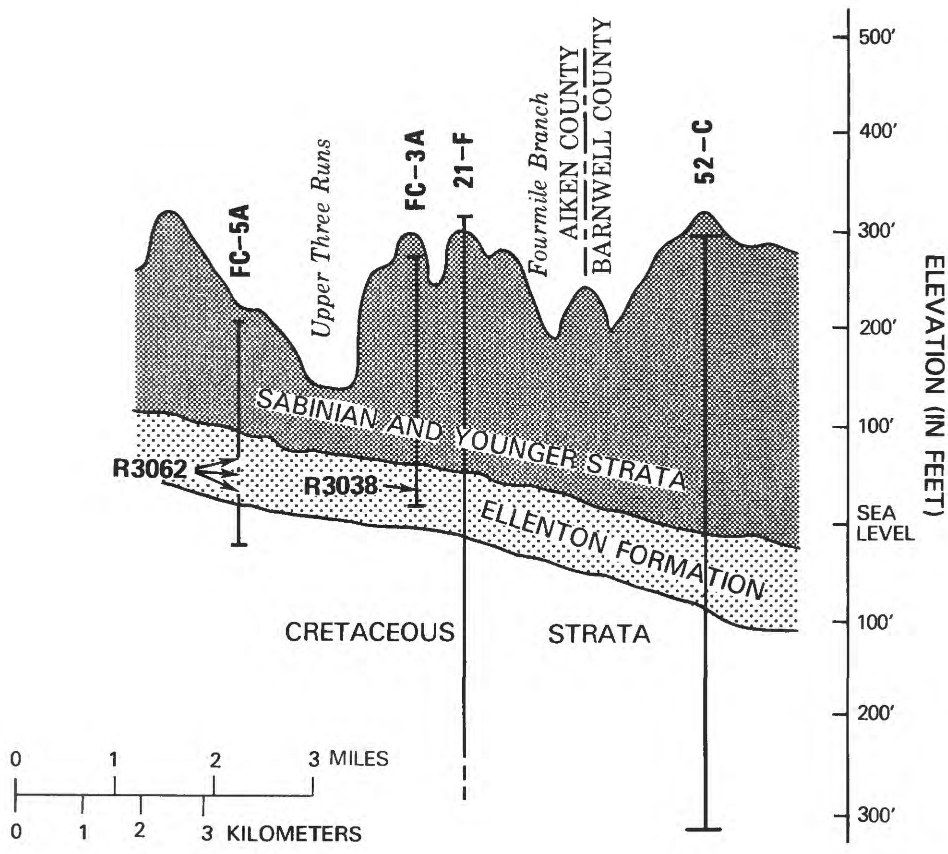


Figure 3B.--Part of geologic section A-A' of Siple (1967, pl. 3) showing critical test wells and USGS Paleobotanical localities R3062 and R3038 in the Ellenton Formation in the Savannah River Plant area.

by a single, morphologically variable form, Peridiniacean cyst sp. B of Edwards (1980). The domination by a single species suggests a brackish or restricted nearshore marine environment. The lower two samples from well FC-5A contain moderately diverse dinocyst floras and probably represent a more normal marine environment.

Dinocysts present in the Ellenton samples include *Adnatosphaeridium* sp., *Andalusiella* sp. aff. *A. polymorpha* of Edwards (1980), *Areoligera* sp., *Cordosphaeridium fibrospinosum* Davey & Williams, *Cordosphaeridium gracile* (Eisenack) Davey & Williams, *Danea californica* (Drugg) Stover & Evitt, *Deflandrea* sp. cf. *D. diebelii* of Drugg (1967), *Deflandrea* cf. *D. pentaradiata* of Benson (1976), *Exochosphaeridium* sp., *Fibradinium annetorpense* Morgenroth, *Hafniasphaera septatc* (Cookson & Eisenack) Hansen, *Palaeoperidinium pyrophorum* (Ehrenberg) Sarjeant, Peridiniacean cyst sp. B of Edwards (1980), *Spinidinium pulchrum* (Benson) Lentin & Williams, and *Spiniferites* spp. *Danea californica* is generally regarded as indicating an early Paleocene age (Danian), but Jan du Chene (1977) has reported specimens comparable to this species from the Maestrichtian (Cretaceous) of Switzerland, and Edwards (1980) has reported this species from the Porters Creek Formation (Midwayan provincial stage, upper Paleocene) and possibly in the Nanafalia Formation (Sabinian provincial stage, upper Paleocene) in the Gulf coast. *Spinidinium pulchrum* has been reported only from the lower Paleocene.

The diverse, well-preserved dinocyst assemblages from the FC-5A well at elevations of +38 and +45 ft (+12 and +14 m, respectively) MSL suggest correlation with the Clayton and (or) Porters Creek Formations in Georgia and Alabama (Edwards, 1980), the Brightseat Formation in Maryland (Benson, 1976; Whitney, 1984), the Rhems Formation of the Black Mingo Group in coastal South Carolina (Edwards, unpublished data), and the P1 unit of Prowell and others (1985) in westernmost South Carolina and eastern Georgia. *Andalusiella* sp. aff. *A. polymorpha* of Edwards (1980), *Deflandrea* sp. cf. *D. diebelii* of Drugg (1967), *Deflandrea* cf. *D. pentaradiata* of Benson (1976), *Palaeoperidinium pyrophorum*, and *Spinidinium pulchrum* are known from the Brightseat but not from the overlying Aquia Formation in Virginia and Maryland. These forms are also found in the Clayton and (or) Porters Creek Formations but not in the overlying Naheola Formation or Gravel Creek Sand Member of the Nanafalia Formation in the Gulf Coastal Plain. Thus, the combined evidence indicates placement in the middle part of the Midwayan provincial stage. The two higher samples in this core contain less diverse floras, and so correlation is less precise. These samples are also of Midwayan age but could be considerably younger than the lower samples.

Samples from the base of the Ellenton Formation commonly contain reworked dinocysts from older formations. For example, the lowest sample in well FC-5A contains poorly preserved specimens of *Chatangiella* sp.

and *Spongodinium delitiense* (Ehrenberg) Deflandre, which are almost certainly reworked from Cretaceous strata. The lowest sample in well FC-3A contains specimens of *Cyclapophysis monmouthensis* Benson, which may have been reworked, although the range of this species is not well established.

POLLEN ASSEMBLAGE

The samples from well FC-5A also were examined for identification of characteristic pollen assemblages. The pollen species identified from these samples include *Alangiopollis cribellata* (Srivastava) Frederiksen, *Choanopollenites alabamicus* (Srivastava) Frederiksen, *Choanopollenites conspicuus* (Groot & Groot) Tschudy, *Choanopollenites discipulus* Tschudy, and *Pseudoplicapollis limitata* Frederiksen *Choanopollenites alabamicus* is found mainly in the Upper Cretaceous and Midwayan, although it probably ranges to the top of the Paleocene (Tschudy, 1973; Frederiksen, 1979). *Pseudoplicapollis limitata* also ranges throughout the Paleocene but is not known in the Cretaceous. *Choanopollenites conspicuus* has not been found in the lowermost part of the Midwayan (Clayton Formation), and it is not known to range higher than the top of the Midwayan. Tschudy (1973) did not find *C. discipulus* above the top of the Porters Creek Formation (middle Midwayan), but *Alangiopollis cribellata* is not known to range below the Naheola Formation (upper Midwayan). These two species occur together in the +38 ft (+12 m) MSL sample in well FC-5A; this occurrence suggests that this sample may correlate approximately with the Porters Creek-Naheola boundary in Alabama. Alternatively, *C. discipulus* may range higher, or *A. cribellata* may range lower in the section than is presently known. In summary, the pollen data indicate that the Ellenton Formation samples from well FC-5A are Midwayan in age.

The biostratigraphic evidence provided by dinoflagellates and pollen indicates that the Ellenton Formation near its type locality is Midwayan in age and most probably middle Midwayan. Evidence not presented in this report suggests that the strata called Ellenton elsewhere can include sediments of younger Paleocene (Sabinian) age (for example, Cahill, 1982). We are unable to date the very uppermost layers of the Ellenton in the F-area; therefore, we cannot reject the possibility that post-Midwayan strata are present.

CONCLUSIONS

The Ellenton Formation is a distinctive sequence of sand, sandy clay, and clay that forms an 80-ft (24-m) thick geologic unit at its type section in the subsurface at the Savannah River Plant (SRP) in western South Carolina. The lithologic character of the formation and the presence of a marine palynoflora with abundant carbonaceous material suggest that the Ellenton was deposited in a marginal-marine deltaic environment. Analysis of the dinocysts and pollen in six samples indicates that the formation is Midwayan (Paleocene), rather than Cretaceous. The uppermost part of the formation may contain some beds of post-Midwayan age, but we do not favor this possibility.

The Ellenton Formation is broadly correlative with the Naheola, Porters Creek, and (or) Clayton Formations in Alabama and western Georgia; with the Rhems Formation of the Black Mingo Group in eastern South Carolina; and with the lower part of the Huber Formation in central and eastern Georgia. The lithologic character and depositional environment of the Ellenton are most like those of the lower part of the Huber Formation in central Georgia, which suggests that the Midwayan fluviomarine deltaic deposition described by Prowell and others (1985) extends well into western South Carolina.

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REVISION OF SOME STRATIGRAPHIC NAMES IN CENTRAL MASSACHUSETTS

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Abstract

Names, ages, and geographic distributions of many stratified and intrusive rock units in the Bronson Hill zone, the western part of the Merrimack belt, the Connecticut Valley belt, and the Mesozoic basins of central Massachusetts and adjacent States are herein changed. These changes reflect recent radiometric and palynologic age determinations on rocks in Massachusetts and Connecticut and detailed and reconnaissance mapping for the Bedrock Geologic Map of Massachusetts, published in 1983 by the U.S. Geological Survey (Zen, E-an, ed., Goldsmith, Richard, Ratcliffe, N. M., Robinson, Peter, and Stanley, R. S., compilers, U.S. Geological Survey special map, scale 1:250,000). Units discussed are the Dry Hill Gneiss and its Pelham Quartzite Member, Poplar Mountain Gneiss, Mount Mineral Formation, Fourmile Gneiss, Monson Gneiss, Partridge Formation, Glastonbury Gneiss, and Pauchaug Gneiss in the Bronson Hill zone; Coys Hill Porphyritic Granite Gneiss in the Merrimack belt; Putney Volcanics, Belchertown Complex, Prescott Complex, Cooleyville Granitic Gneiss, and Middlefield Granite in the Connecticut Valley belt; and Hitchcock Volcanics, Shuttle Meadow Formation, Granby Basaltic Tuff, Sugarloaf Formation, Deerfield Basalt, Turners Falls Sandstone, and Mount Toby Formation in the Mesozoic basins.

INTRODUCTION

New radiometric and palynologic age determinations on rocks in central Massachusetts and Connecticut and mapping for the new Bedrock Geologic Map of Massachusetts (Zen and others, 1983) have resulted in the reassignment of ages, revision of names, and restatement of the geographic distribution for many rock units. Brief descriptions of these changes in nomenclature are presented for rock units in four lithotectonic assemblages in central Massachusetts. These are the pre-Silurian Bronson Hill zone, the Silurian and Devonian Merrimack and Connecticut Valley belts, and the Mesozoic basins. The distribution of these assemblages in Massachusetts is shown on figure 1. The letter symbols in parentheses following unit names in this report are the symbols used on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983).

BRONSON HILL ZONE

The rocks of the Bronson Hill zone, named for and partly coincident with the Bronson Hill anticlinorium (fig. 2), are Late Proterozoic to Ordovician gneisses, quartzites, schists, and amphibolites intruded by Ordovician plutons and overlain by metamorphosed Middle Ordovician volcanics and black shales. On the west, the Bronson Hill zone is believed to overlie structurally the east edge of the pre-Silurian Rowe-Hawley zone. The contact, concealed beneath the younger Connecticut Valley belt and Mesozoic basins, is the postulated Bristol thrust (Hatch and others, 1984). The eastern boundary is beneath the Silurian and Devonian Merrimack belt, west of the Massabesic Gneiss Complex in the Nashoba zone and east of the easternmost anticline exposing Middle Ordovician rocks of the Bronson Hill zone (Hatch and others, 1984).

Changes are presented for stratified rock units that include the Dry Hill Gneiss and its Pelham Quartzite Member, the Poplar Mountain Gneiss and its quartzite and gneiss members, the Mount Mineral Formation, the Fourmile Gneiss, the Monson Gneiss, and the Partridge Formation and for the intrusive Glastonbury Gneiss and Pauchaug Gneiss.

Dry Hill Gneiss (Zdh, Zdhs).--The Dry Hill Gneiss was originally defined by Balk (1956) to include most felsic gneisses in the Pelham dome (fig. 2), except for rocks described by him as Poplar Mountain Gneiss. Earlier, the Dry Hill Gneiss had been called Becket Gneiss (Emerson, 1898) and subsequently Pelham Granite (Emerson, 1917). The Dry Hill Gneiss was redefined by Ashenden (1973) to include most of Balk's Dry Hill Gneiss, except for those rocks assigned by Balk to the so-called "border facies" (1956). Ashenden (1973) showed that the rocks of the "border facies" are completely different from the bulk of the felsic gneisses and assigned the "border facies" to the Fourmile Gneiss (see Fourmile Gneiss section, p. A73).

The Dry Hill Gneiss (Zdh) is considered to be a series of metamorphosed rhyolitic tuffs (Ashenden, 1973; Hodgkins, 1983), with minor intercalated layers of sedimentary derivation (Zdhs). The Pelham Quartzite Member of the Dry Hill Gneiss occurs in an area mainly outside Ashenden's study area (1973). The Pelham Quartzite Member is a massive to well-bedded quartzite, locally up to 250 m thick.

The informal name "Rocky Run gneiss" (Robinson and others, 1973) for that part of the Dry Hill Gneiss stratigraphically above the Pelham Quartzite Member is dropped from usage. The age of the Dry Hill Gneiss is changed from late Precambrian to Late Proterozoic (Zartman and Naylor, 1984).

Pelham Quartzite Member (of the Dry Hill Gneiss) (Zdpq).--Emerson (1898) described a number of thick quartzite layers within his Becket Gneiss. Although the layers were not named in his text, his maps and sections show these layers clearly labeled as Pelham Quartzite. Emerson (1917) later reconsidered the origin of these rocks, named them "northfieldite," a supposed igneous form of SiO_2, and so designated them on his preliminary geologic map of Massachusetts and Rhode Island (Emerson, 1917). Detailed mapping in the central and southern parts of the Pelham dome (Ashenden, 1973, map only; Robinson and others, 1973) shows that the quartzite is within rocks mapped as Dry Hill Gneiss. The quartzite commonly contains subordinate plagioclase, microcline,

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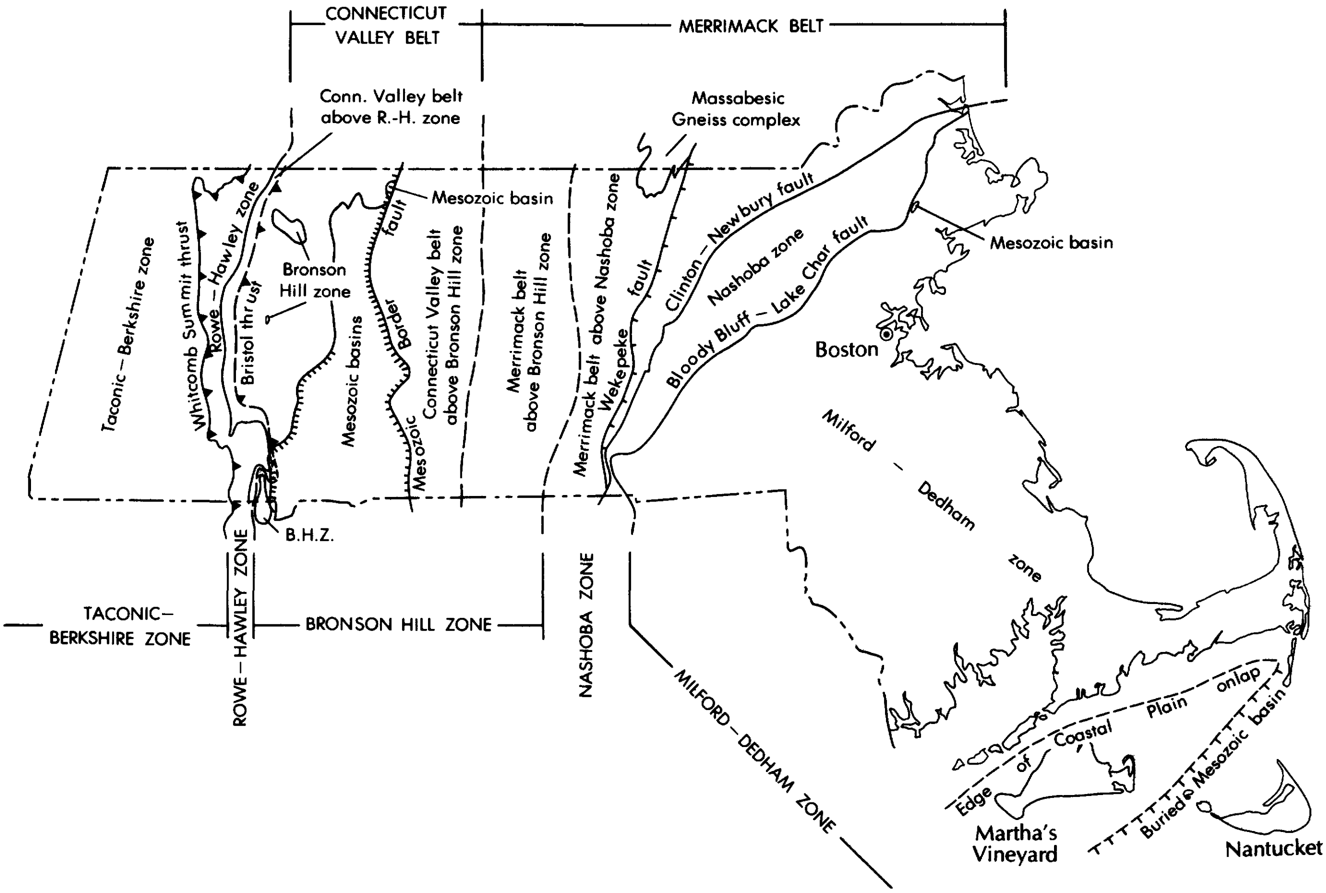

Figure 1. Geographic distribution of the Bronson Hill zone, Merrimack belt, Connecticut Valley belt, and Mesozoic basins in Massachusetts. Adapted from Hatch and others (1984).

and light brown biotite and locally abundant actinolite, which indicates the presence of dolomite cement prior to metamorphism. These quartzite rocks are here redefined as the Pelham Quartzite Member of the Dry Hill Gneiss. The type locality may be considered to be a low ledge, partially buried by sand and gravel, in the front yard of the Town Hall of Pelham, Hampshire County, Mass. (fig. 2). Better exposures occur on Route 202 approximately 1.6 km to the north and 1.6 km to the south of the Town Hall.

Poplar Mountain Gneiss (Zpm, Zpmq, Zpmg).-- The name Poplar Mountain Gneiss was applied by Balk (1956) to distinctive gray quartz-biotite-microcline gneisses formerly included in the Pelham Granite (Emerson, 1917). The type locality is Poplar Mountain, northeast of Millers Falls, Franklin County, Mass. (fig. 2). The Poplar Mountain Gneiss was subdivided by Ashenden (1973) into a basal quartzite member (Zpmq) and a gneiss member (Zpmg). In the area of exposure near Poplar Mountain, the Poplar Mountain Gneiss lies in the core of the Pelham dome, structurally beneath the Dry Hill Gneiss. Here the quartzite member of the Poplar Mountain Gneiss occurs continuously near the contact with the Dry Hill Gneiss, as shown on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983). A second area of Poplar Mountain Gneiss (Zpm) occurs in a thin layer that lies structurally above the Dry Hill Gneiss. In this area also, the quartzite member occurs along the contact with the Dry Hill Gneiss, although here the quartzite member is too thin to map separately. Ashenden's structural interpretation is that the Dry Hill Gneiss occupies the core of a recumbent anticlinal nappe in which one Poplar Mountain sequence with basal quartzite is right side up on the top limb and another sequence with basal quartzite is upside down on the bottom limb. The Poplar Mountain Gneiss on the top limb appears to grade southward along strike into the Mount Mineral Formation. The inverted basal quartzite on the bottom limb plunges beneath the surface south of Millers River, but reappears in a window southwest of Dry Hill and possibly again in a small window exposed in a brook northwest of Mount Lincoln in Pelham (fig. 2). Because the Poplar Mountain Gneiss grades into the Dry Hill Gneiss, because layers of Dry Hill-like gneiss occur within the Poplar Mountain Gneiss, and because the Dry Hill Gneiss contains many quartzite beds, the age of the Poplar Mountain Gneiss is considered to be Late Proterozoic.

Mount Mineral Formation (Zmm, Zmmu).-- The Mount Mineral Formation of Robinson and others (1973) is here adopted by the U.S. Geological Survey. The unit appears on the bedrock geologic map of the eastern part of the Shutesbury quadrangle (Robinson and others, 1973) and is described in the map explanation. The name is taken from Mount Mineral, 5 km north of the village of

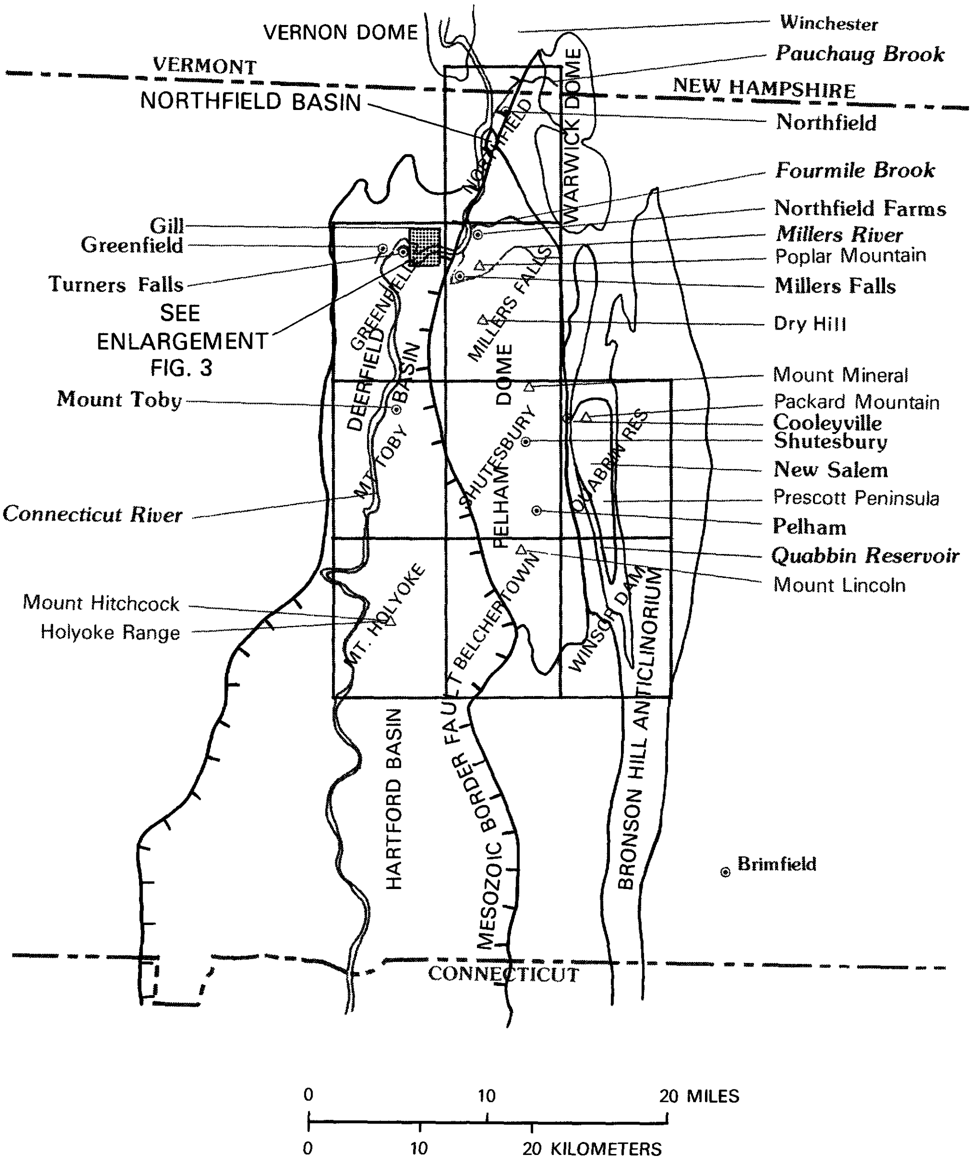

Figure 2. Map of part of Massachusetts and adjacent States showing locations of towns and geographic features described in the text.

Shutesbury, Franklin County, in the northeastern part of the Shutesbury quadrangle where most of the formation is well exposed (fig. 2).

The Mount Mineral Formation (Zmm) contains aluminous schist, amphibolite, and quartzite that are undifferentiated on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983) but have been mapped separately in some areas (Robinson and others, 1973). The schists are locally rich in garnet and kyanite and contain relict sillimanite and orthoclase from pre-Middle Ordovician metamorphism. The formation has lenses of partially serpentinized harzburgite (Zmmu) that contains abundant veins of anthophyllite, including the "Pelham asbestos quarry." Because the Mount Mineral Formation probably correlates with the Poplar Mountain Gneiss and may be a southerly facies of it, the age of the Mount Mineral Formation is considered to be Late Proterozoic.

Fourmile Gneiss (OZfm, OZfmu, OZfmq).— The Fourmile Gneiss (OZfm) of Ashenden (1973) is here adopted by the U.S. Geological Survey. The type locality is the gorge of Fourmile Brook northwest of the village of Northfield Farms, Franklin County, Mass. (fig. 2). The unit corresponds in large part to the rocks mapped by Balk (1956) as the "border facies" of the Dry Hill Gneiss, although the distribution shown on the map as published by the U.S. Geological Survey (Balk, 1956) based on Balk's field work differs considerably from Balk's manuscript map on file at the University of Massachusetts. The best available evidence is that the contact with the underlying Poplar Mountain Gneiss is sharp, so that a Late Proterozoic age is not necessarily implied. Unfortunately, what could have been the best exposed contacts, in the tunnels of the Northfield Mountain Pumped Storage Project at Northfield Farms, are now known to be invaded by a 10-m sill of quartz monzodiorite gneiss of the Devonian Belchertown Complex (Robinson, unpublished data, 1980), thus invalidating some of Ashenden's earlier discussion (1973) of the contact relations. The Fourmile Gneiss (OZfm) consists of biotite-feldspar gneiss and amphibolite. In the southwest corner of the Pelham dome, one lens of ultramafic hornblendite (OZfmu) and three lenses of muscovite quartzite (OZfmq) have been mapped separately. The unit is overlain, with possible unconformity, either by the Ammonoosuc Volcanics or the Partridge Formation. Thus the age is uncertain but must be Late Proterozoic, Cambrian, or Ordovician.

Monson Gneiss (OZmo, OZmou, OZmoa).--The age of the Monson Gneiss (OZmo), named by Emerson (1898), is changed from Ordovician (Leo and others, 1977) to Late Proterozoic, Cambrian, or Ordovician on the basis of uncertainties similar to those regarding the Fourmile Gneiss (see discussion of Fourmile Gneiss). An unconformity between the Monson Gneiss and the overlying Ammonoosuc Volcanics is suggested by regional relations, as well as by one conglomerate lens (Robinson, 1979) and two quartzite lenses (Robinson, unpublished data, 1983) where the Ammonoosuc rests on the Monson. Several lenses of ultramafic rock (OZmou) and hornblende amphibolite (OZmoa) are shown separately on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983).

Partridge Formation (Ops, Opsa, Opsc, Opa, Opv, Opvs, Opau, Opu, Opsg, Opf, Opbg, Dl + Ops).--The Partridge Formation (Billings, 1934), as mapped on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983), includes the typical sulfidic mica schist (Ops) and a variety of other lithic types, among them many but not all of the rocks formerly mapped in Massachusetts and Connecticut as Brimfield Schist (Emerson, 1898, 1917). The rank of Emerson's Brimfield Schist, based on exposures in Brimfield, Mass. (fig. 2) (exposures for which the town itself was named), was raised to group status by Peper, Pease, and Seiders (1975), to include the Bigelow Brook Formation (Pease, 1972), and the Hamilton Reservoir and Mount Pisgah Formations (Peper, Pease, and Seiders, 1975). The Bigelow Brook Formation was later removed from the Brimfield Group and assigned to the Paxton Formation as the Bigelow Brook Member (Goldsmith and others, 1982). The Brimfield Group was not used on the Bedrock Geologic Map of Massachusetts because the unit can be shown to be a complex tectonic interlayering, by both folding and faulting, of a wide variety of strata of Ordovician, Silurian, and Early Devonian age (Field, 1975; Tucker, 1977; Robinson, Field, and Tucker, 1982). Only a fraction of the Brimfield Group was assigned to the Partridge Formation on the Bedrock Geologic Map of Massachusetts; the remainder was assigned to the Silurian Paxton Formation or the Lower Devonian Littleton Formation. In areas of poor exposure and incomplete mapping in northern Worcester County, Mass., and immediately contiguous Cheshire County, N.H., the Partridge is intimately interfolded with the Littleton Formation (Dl) of Early Devonian age (Dl + Ops) and cannot be shown separately on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983).

Glastonbury Gneiss (Ogl).--The age of the Glastonbury Gneiss, named by Gregory (1906), is changed from Devonian (Snyder, 1970) to Ordovician on the basis of isotopic work reported by Leo and others (1984).

Pauchaug Gneiss (Opc).--The Pauchaug Gneiss of Robinson (1963, 1967) is here adopted by the U.S. Geological Survey. The name was applied by Robinson to the massive intrusive-looking granitic rocks of the core of the Warwick dome, east of the village of Northfield, Franklin County, Mass., and in adjacent New Hampshire (fig. 2). These rocks range in composition from biotite granite to biotite tonalite. The name is used also for the massive granitic gneiss in the core of the Vernon dome in Vermont and New Hampshire (fig. 2). The unit is named for the type locality in the upper reaches of Pauchaug Brook in the southern part of Winchester, Cheshire County, N.H. (fig. 2). Robinson's original use of this name avoided assignment of the rocks to the Oliverian Plutonic Suite (Billings, 1956), which was thought to be of Devonian age. The name Monson Gneiss, as used by Hadley (1949) for the rocks in the Warwick dome, is inappropriate because the rocks are different. On the basis of zircon dating (Zartman and Leo, 1985), the age of the Pauchaug Gneiss is determined to be Ordovician.

MERRIMACK BELT

The Merrimack belt includes Silurian and Lower Devonian metamorphosed sedimentary rocks cut by Devonian intrusive rocks. The Merrimack belt extends from the east edge of the Connecticut Valley belt eastward to the Clinton-Newbury fault. To the west, the belt rests unconformably on the eastern part of the Bronson Hill zone. The eastern portion of the Merrimack belt is locally overlain by Pennsylvanian sedimentary rocks (Hatch and others, 1984).

Coys Hill Porphyritic Granite Gneiss (Dchgr, Dchh).—The Coys Hill Granite, named by Emerson (1898), of Early Devonian age is here renamed the Coys Hill Porphyritic Granite Gneiss (Dchgr) to reflect more accurately the distinctive character of the unit in Massachusetts. The unit is characterized by variously deformed tabular microcline and locally sodic plagioclase phenocrysts 3 to 10 cm long in a finer matrix of quartz, plagioclase, and biotite with or without muscovite, garnet, and sillimanite (Field, 1975). The unit may be continuous (P. J. Thompson, manuscript map, 1983) with the Cardigan pluton of Kinsman Quartz Monzonite, named by Billings (1956), in New Hampshire. The Kinsman also should be called granite under currently adopted igneous rock terminology. Lenses of hornblende-pyroxene gneiss (Dchh) within the Coys Hill are mapped separately on the Bedrock Geologic Map of Massachusetts (Zen and others, 1983).

CONNECTICUT VALLEY BELT

The Connecticut Valley belt includes Silurian and Lower Devonian sedimentary strata and Devonian intrusive rocks. The belt unconformably overlies both the Rowe-Hawley zone to the west and the Bronson Hill zone to the east and extends eastward to the edge of the Merrimack belt (Hatch and others, 1984).

Putney Volcanics (Dpv).--The age of the Putney Volcanics, named by Hepburn (1972) and revised by Trask (1980), is changed from Devonian to Early Devonian (Robinson and others, 1984). Correlations preferred by some workers imply a Silurian age.

Belchertown Complex (Dbmd, Dbmdt, Dbmdg, Dbd, Dbh, Dbi, Dbt, Dbp).— The Belchertown Tonalite (Emerson, 1898, 1917), later Belchertown Quartz Monzodiorite (Leo and others, 1977; Ashwal and others, 1979), of Devonian age is here renamed the Belchertown Complex. The complex includes zones of quartz monzodiorite (Dbmd, Dbmdt), quartz monzodiorite gneiss (Dbmdg), pyroxenite (Dbp), hornblendite (Dbh), intrusive breccia (Dbi), tonalite (Dbt), and dacite porphyry (Dbd).

Prescott Complex (Dpgg, Dpgb).-- The Prescott Diorite (Emerson, 1898, 1917) was renamed the Prescott Intrusive Complex by Makower (1964; see also Robinson, 1967) and expanded to include both the original "diorite," redescribed as hornblende gabbro (Dpgb), and surrounding gneissic biotite tonalite, granodiorite, and granite that he named Cooleyville Granitic Gneiss (Dpgg). The unit is here renamed the Prescott Complex. The name is taken from Prescott, Mass., a town that was obliterated when the Quabbin Reservoir was established. All exposures of the Prescott Complex are on the present Prescott Peninsula of the Quabbin Reservoir (fig. 2). The hornblende gabbro (Dpgb) is most prominently exposed on Packard Mountain and two unnamed mountains south of it. The gabbro intrudes and contains inclusions of Middle Ordovician Partridge Formation; it is intruded by and appears as inclusions in the Cooleyville Granitic Gneiss. The age is changed from Late Carboniferous or post-Carboniferous (Emerson, 1917) to Devonian on the basis of the age relations of the Cooleyville Granitic Gneiss (see below).

Cooleyville Granitic Gneiss (of the Prescott Complex) (Dpgg).--The Cooleyville Granitic Gneiss was defined by Makower (1964; see also Robinson, 1967) to include extensive exposures of fine- to coarse-grained gneissic biotite tonalite, granodiorite, and granite that surround the hornblende gabbro on Prescott Peninsula of the Quabbin Reservoir. The name is here adopted by the U.S. Geological Survey. The unit is named for the village of Cooleyville at the western edge of the Town of New Salem, Franklin County, Mass., and typical exposures occur in the hills 0.8 to 3.2 km to the northeast, east, and southeast of the village (fig. 2). The Cooleyville Granitic Gneiss intrudes the Lower Devonian Littleton Formation and contains extensive inclusions of the hornblende gabbro of the Prescott Complex (see section on the Prescott Complex). The foliation and lineation in the gneiss are congruent and have features identical to those found in adjacent gneiss domes formed late in the Acadian orogeny. A preliminary Rb-Sr whole-rock isochron (Naylor, 1970), based in part on deformed aplite dikes within the Cooleyville, suggests a minimum age of 385 m.y. (Devonian).

Middlefield Granite (Dmg).--The age of the Middlefield Granite named by Emerson (1898, 1917) and revised by Hatch and others (1980), is changed from Early(?) Devonian to Devonian.

MESOZOIC BASINS

The Mesozoic rocks of the Connecticut Valley belt consist of unmetamorphosed terrestrial sedimentary rocks and associated basaltic flows, tuffs, sills, and dikes deposited and intruded in Late Triassic to Early Jurassic time in three rift basins.

Age of the rocks

The interbedded sedimentary and volcanic rocks of the Hartford, Deerfield, and Northfield basins (fig. 2) were long thought to be of Late Triassic age. Recent palynoflora studies by Cornet, Traverse, and McDonald (1973) suggest an early Liassic or possibly a late Rhaetic age for the Shuttle Meadow Formation in the Hartford basin. Reptilian evidence in older strata of the Hartford basin suggests that the Triassic-Jurassic boundary may be just below the Shuttle Meadow Formation (Cornet and Traverse, 1975). The lowermost units, the New Haven Arkose in the Hartford basin in Connecticut and Massachusetts and the continuous and lithically similar Sugarloaf Formation in the Deerfield basin, are of Late Triassic and Early Jurassic age. The boundary is arbitrarily drawn through clastic rocks of similar type below gray mudstone containing a Lower Jurassic palynofloral zone. The overlying rocks, including the Hitchcock Volcanics, Shuttle Meadow Formation, Holyoke Basalt, East Berlin Formation, Hampden Basalt, Granby Basaltic Tuff, and Portland Formation in the Hartford basin, the Deerfield Basalt, Turners Falls Sandstone, and Mount Toby Formation in the Deerfield basin, and a conglomerate facies member of the Turners Falls Sandstone in the Northfield basin, are all of Early Jurassic age. The uppermost exposed parts of the Portland and Mount Toby Formations may be of Middle Jurassic age (Cornet, 1977). All of the Triassic and Jurassic units in the Mesozoic basins are part of the Newark Supergroup.

Hartford Basin

Hitchcock Volcanics (Jhv).-- The name Hitchcock Volcanics of Brophy and others (1967) is here adopted by the U.S. Geological Survey. The age is changed from Late Triassic to Early Jurassic. No type locality is designated, but the unit is found only on the north and west slopes of Mount Hitchcock, Holyoke Range, Hampshire County, Mass. (fig. 2). The Hitchcock Volcanics consists of nested cones of basaltic breccia containing abundant fragments of New Haven Arkose, locally intrusive into arkose near the base. The unit underlies the Shuttle Meadow Formation and (or) the Holyoke Basalt.

Shuttle Meadow Formation (Jsm, Jsmc).— The name Shuttle Meadow Formation (Lehmann, 1959) is geographically extended from Connecticut into central Massachusetts. The unit lies just below the Holyoke Basalt and consists of sandstone strata containing one interval of gray mudstone beds of lacustrine origin. The mudstone beds contain a Jurassic palynoflora. The unit grades eastward along strike into a conglomeratic facies (Jsmc).

Granby Basaltic Tuff (Jgb).-- The Granby Tuff, named by Emerson (1898), is here renamed the Granby Basaltic Tuff to reflect the composition of this friable, well bedded tuff containing sediment fragments.

Deerfield and Northfield Basins

Sugarloaf Formation (TRs, Js, Jsc).-- The Sugarloaf Formation, named by Emerson (1891, 1917), of Late Triassic and Early Jurassic age includes all sedimentary strata in the Deerfield basin below the Deerfield Basalt or its projected horizon. The Late Triassic-Early Jurassic boundary is arbitrarily drawn through clastic rocks consisting of coarse-grained, locally conglomeratic arkose interbedded with siltstone and sandstone (TRs, Js, Jsc) below a Lower Jurassic palynofloral zone in gray mudstone immediately below the base of the Deerfield Basalt

Deerfield Basalt (Jdb).--The Deerfield Diabase, named by Emerson (1891, 1917), is here renamed the Deerfield Basalt to reflect the extrusive origin of the unit. Evidence for extrusive origin includes (1) extensive exposures at the base in Greenfield and Gill of pillows and pillow breccias with hyaloclastite matrix, and (2) extensive exposures of the vesicular pahoehoe flow top along the west bank of the Connecticut River opposite Turners Falls in the Greenfield quadrangle (fig. 2).

Turners Falls Sandstone (Jt, Jtc).—The Turners Falls Sandstone (Jt), named by Willard (1951), in the center of the Deerfield basin, is here redefined to include only strata between the top of the Deerfield Basalt and the slump zone unconformity defined by Cornet (1977) on palynological grounds in the vicinity of Barton Cove on the north bank of the Connecticut River, southeast of the village of Riverside, Franklin County, northern Greenfield quadrangle (fig. 3). On the Bedrock Geologic Map of Massachusetts (Zen and others, 1983), this unconformity between the Turners Falls and the overlying Mount Toby Formation is mislocated on the northwest side of the Barton Cove peninsula; the unconformity should be located along the southeast side. Elsewhere on the map, it is more or less correctly shown. The uppermost part of the Turners Falls Sandstone, forming extensive outcrops on Barton Cove peninsula, consists of an angular shale breccia that seems to have formed by massive slumping of previously lithified lacustrine beds, possibly triggered by earthquake activity (Handy, 1976). Palynofloral data and fossil fish from the shale breccia at the lily pond on the Barton Cove peninsula indicate that the lake beds must be older than the Shuttle Meadow Formation of the Hartford basin (Cornet, 1977, p. 217-218). Strata immediately above the unconformity consist predominantly of fanglomerate up to 78 m thick, having angular to rounded clasts of schist and granite up to 1.8 m in diameter (Handy, 1976; Bain, 1932). Within undisturbed strata just above the fanglomerate, a gray mudstone bed contains a palynoflora that can be no older than the lower part of the Portland Formation (Cornet, 1977, p. 218). The hiatus represented by the unconformity thus is equivalent to a considerable portion of the section of the adjacent Hartford basin and includes the time of eruption of the Holyoke Basalt, the Hampden Basalt, and the Granby Basaltic Tuff. Cornet (1977, p. 217-218) postulates that the shale breccia at Barton Cove was produced at the time of emplacement of the overlying fanglomerate above the unconformity, whereas further south in the basin, the shale strata beneath the fanglomerate appear to have been relatively undisturbed.

To the south of the Barton Cove peninsula, the slump zone unconformity moves progressively closer to the top of the Deerfield Basalt, so that the Turners Falls Sandstone thins from about 1,180 m near Barton Cove (Cornet, 1977, p. 219) to a feather edge close to the southern point of pinch-out of the Deerfield Basalt (Handy, 1976). Northeastward from the Barton Cove peninsula, the unconformity extends along strike into a conglomeratic facies through which the contact is projected parallel to bedding as far as the Mesozoic border fault. Previously all of the conglomeratic strata were referred to as the Mount Toby Conglomerate (Emerson, 1917; Willard, 1952), but those below the projected unconformity are reassigned to a conglomerate facies

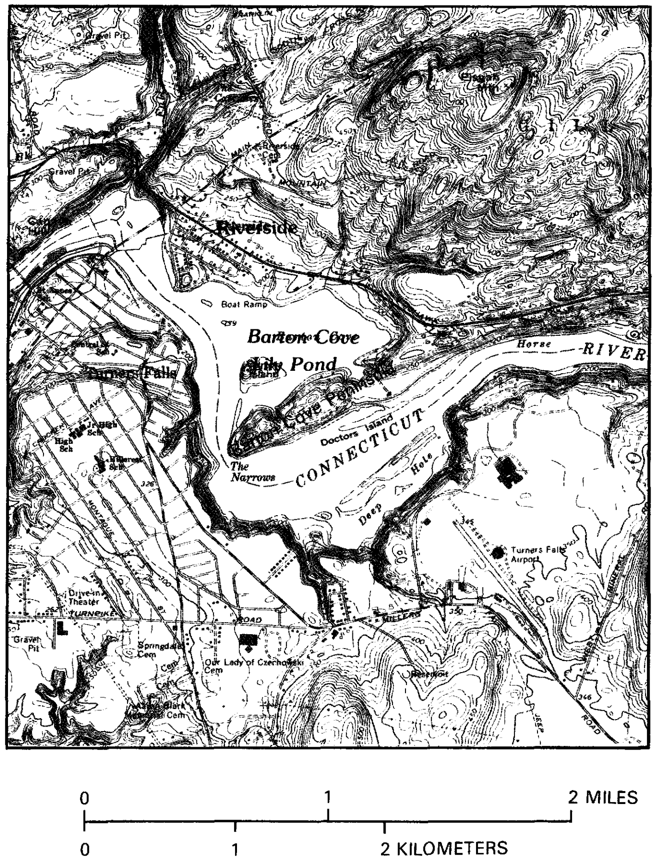

Figure 3. Barton Cove peninsula, northern Greenfield quadrangle, showing location of the lily pond below the slump zone unconformity between the Turners Falls Sandstone and the Mount Toby Formation on the southeast side of the peninsula.

member of the Turners Falls Sandstone (Jtc). This conglomerate member includes all of the Jurassic strata of the small Northfield basin.

Mount Toby Formation (Jm, Jmc, Jma, Jmg).--The Mount Toby Conglomerate (Emerson, 1898) is here renamed the Mount Toby Formation and redefined to include only the sedimentary strata in the Deerfield basin above the slump zone unconformity or its projected equivalent (see section on Turners Falls Sandstone). The Mount Toby thus includes the conglomerates at or near the type locality at Mount Toby (Jmc) (fig. 2), landslide deposits within the conglomerate (Jma, Jmg), and sandstones and lake beds above the slump zone unconformity (Jm) formerly included in the Turners Falls Sandstone. Rocks mapped as Mount Toby Conglomerate by Emerson (1898) in the Hartford basin, the Northfield basin, and the Deerfield basin below the slump zone unconformity have all been assigned to the Sugarloaf, Turners Falls, and Portland Formations (Willard, 1951, 1952; Leo and others, 1977; Peper, 1977). These changes are consistent with the suggestion of Cornet (1977, p. 219) that "The Turners Falls Sandstone below the proposed disconformity at the slump zone could therefore be treated as a separate formation from the strata above the slump."

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