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Abstract

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STRATIGRAPHIC CHARACTERIZATION AND CORRELATION OF VOLCANIC FLOWS WITHIN THE CATOCTIN FORMATION, CENTRAL APPALACHIANS

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ABSTRACT

The Catoctin Formation of the Central Appalachians consists of a sequence of metabasalts interbedded with volcanic breccias and thin metasedimentary layers. Despite greenschist facies metamorphism, a mappable stratigraphy is still preserved. Two traverses have been mapped southeast of the town of Luray on the west limb of the Blue Ridge Anticlinorium. One traverse, located near Big Meadows Campground in Shenandoah National Park, contains ten distinct volcanic flows. The other traverse, located on Hawksbill Mt., approximately 5 km to the northeast, contains nine distinct flows. Flows 6 and 7 at Big Meadows correlate lithologically with flows 7 and 8 at Hawksbill Mt. The missing flow at Big Meadows is probably one of the lower flows, reflecting the pre-volcanic irregular erosion surface of the Precambrian basement. Absence of erosional unconformities between flows and absence of thick metasedimentary interbeds suggest eruption time was short, probably on the order of 3-5 Ma, similar to the Deccan Traps and Columbia River Basalts. Thin (1 m) interbeds of purple phyllite are interpreted to have an igneous origin, implying felsic volcanism synchronous with, but not necessarily genetically related to, Catoctin magmatism.

GEOLOGIC SETTING

The Appalachian Mountains in central Virginia are divisible into the Valley and Ridge, Blue Ridge and Piedmont terranes (Figure 1). Within the Blue Ridge province, rock types include granulite to upper amphibolite facies paragneisses intruded by charnockite, anorthosite, quartz monzonite and alkalic granites (Bartholomew and others, 1981; Herz and Force, 1984; Bartholomew and Lewis, 1984; Reed and Clarke, 1989) comprising the Precambrian Blue Ridge Basement Complex exposed in the core of the Blue Ridge Anticlinorium in central Virginia (Sinha and Bartholomew, 1984). The predominant rock type underlying the area of study is charnockite of the Pedlar Formation. Intrusive into the Blue Ridge Basement Complex are several alkalic granodiorites collectively known as the Crossnore Plutonic Suite (Rankin and others, 1973). These rocks have been dated by Odom and Fullagar (1984) at 710-680 Ma using Rb-Sr isotopic data and by Tollo and others (1991) at 729-640 Ma using U/Pb isotopic data from zircons.

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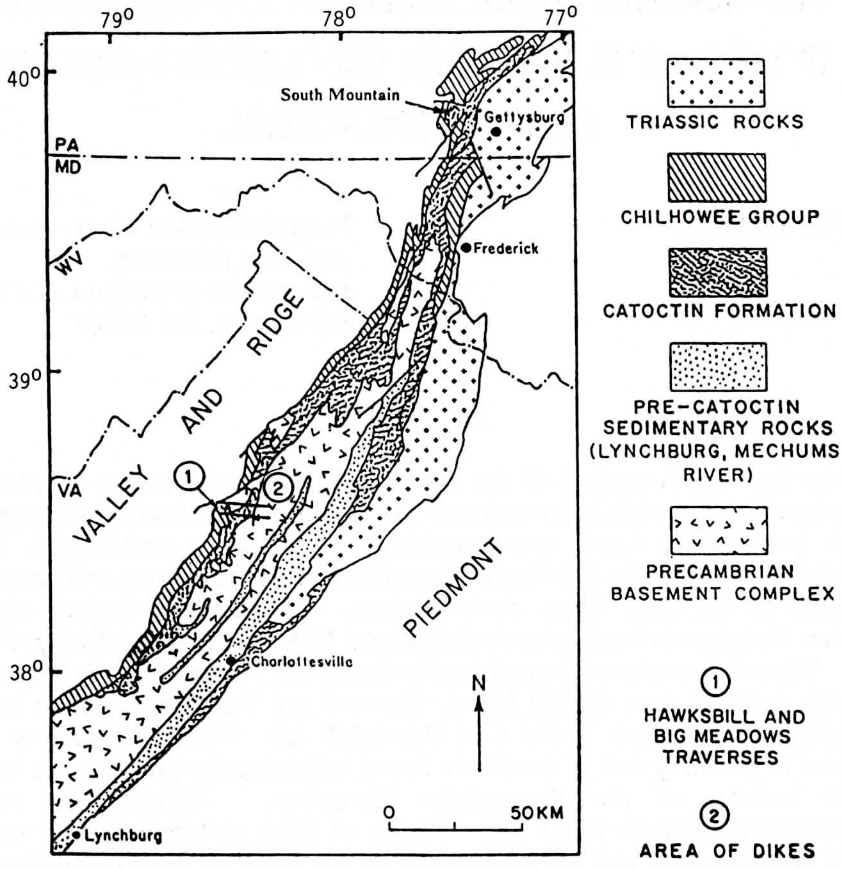


Figure 1. Index map of the central Appalachian Mountains showing location of the Catoctin volcanic province, along with geographic locations of the Blue Ridge Province, Valley and Ridge, and Piedmont. Location of study area near Luray, Va. is shown, along with the general location of many of the mafic dikes. Adapted from Reed and Morgan (1971).

On the east flank of the Blue Ridge Anticlinorium, the basement complex is unconformably overlain by the late Precambrian Lynchburg Formation in central Virginia and its northern Virginia equivalent, the Mechums River Formation, a thick sequence of dominantly marine clastic metasediments (Schwab, 1974; Wehr, 1983; Evans, 1984; Reed and Clarke, 1989), interpreted as a rift sequence (Wehr and Glover, 1985). At the base of the Mechums River Formation, Mose (1981) and Lukert and Banks (1984) have reported cobbles and boulders of granite texturally and lithologically similar to the Robertson River pluton, one of the Crossnore Group, implying its uplift and erosion prior to deposition of Mechums River and Lynchburg lithologies. Overlying and interbedded with the top of the Lynchburg Formation is the Catoctin Formation (Brown, 1958; Reed, 1964; Rankin, 1975).

On the west flank of the Blue Ridge Anticlinorium, the Swift Run Formation, a thin discontinuous basal conglomerate, and the Catoctin Formation rest either unconformably upon, or are in fault contact with the Precambrian Basement Complex (Gathright and others, 1977).

In the central Appalachians, the Catoctin Formation is a sequence of subaerial and subaqueous basaltic lava flows that flank both limbs of the Blue Ridge Anticlinorium. To the north, in Maryland and southern Pennsylvania where the formation wraps around the nose of the anticlinorium, basalts are interlayered with felsic volcanics. To the south, in southern Virginia, the formation pinches out, but it is uncertain whether its absence is controlled by stratigraphic or structural constraints. The entire volcanic sequence has been interpreted as the remnants of a continental flood basalt, erupted during Late Precambrian opening of the Iapetus Ocean (Rankin, 1975, 1976; Blackburn and Brown, 1976; Davis, 1977; Bland, 1978; Badger, 1986). Recently, a major suture has been proposed to exist between the east and west limbs of the Blue Ridge Anticlinorium (Rankin, 1988). This model suggests the east limb sequence of basalts may have been part of an exotic accretionary prism that was allochthonous to the Blue Ridge province, and therefore may not be directly related to the west limb sequence. However, stratigraphic evidence suggests arkosic metasediments within the Catoctin Formation had an eastern, Grenville type source (Kline, 1989) and trace element geochemical similarities between east and west limb sequences (Badger and Gottfried, in progress) do not support this exotic terrane model.

Overlying the Catoctin on the west flank of the Blue Ridge is the Chilhowee Group, a sequence of shallow water clastic metasediments, and on the east limb is the Evington Group, a deeper water equivalent to the Chilhowee Group (Brown, 1970). The trace fossil Rusophycus was recently identified in basal units of the Chilhowee Group, placing an age constraint on the Catoctin Formation as pre-earliest Cambrian (Simpson and Sundberg, 1987). This is consistent with a recent Sr isotopic age obtained for Catoctin magmatism at 570 Ma (Badger and Sinha, 1988).

Mafic dikes, ranging in width from a few centimeters to 15 meters, intrude the Precambrian Blue Ridge lithologies. These dikes are especially abundant in the vicinity of Old Rag Mountain and in the basement terrane west of Old Rag Mt. (see maps of Reed, 1969; Gathright, 1976; and area shown in Figure 1). Major element chemical similarities to Catoctin flows suggest the dikes were feeder dikes to Catoctin flows (Reed and Morgan, 1971), but this correlation has yet to be demonstrated through trace element or isotopic analysis.

The regional extent of the Catoctin Formation, assuming continuity across the Blue Ridge Anticlinorium, would have been at least 11,000 km^2. As such, it is the largest flood basalt province in eastern North America. Highly speculative time and spatial correlations with volcanics of the Camels Hump Group in Vermont (Coish and others, 1985) and Lighthouse Cove Volcanics of Newfoundland (Williams and Stevens, 1974) may indicate aerial extent of Late Precambrian flood basalt volcanism on the same order of magnitude as the more well known and better exposed Columbia River, Deccan Traps and Parana flood basalt provinces.

The entire Blue Ridge terrane has undergone regional low grade metamorphism of probable Taconic age (Butler, 1972; Robison, 1976; Mose and Nagel, 1984; Pettingill and others, 1984). Syn- and postmetamorphic thrusting has resulted in the telescoping of the Blue Ridge terrane along a series of west verging thrust faults (Bartholomew and others, 1981), several of which transect the Catoctin Formation (Gathright, 1976). The most recent structural event, the major overthrust of the entire Blue Ridge terrane, involves rocks of Mississippian age, suggesting correlation with the regional Alleghanian event.

STRATIGRAPHY

Within the area of study, the Catoctin Formation consists of dark green, massive metabasalt flows containing irregular pods of light green epidosite (quartz-epidote rock), interbedded with green to reddish brown metavolcanic breccia, thin beds of gray to purple phyllitic tuff and siltstone, and green to pink arkosic metasandstone. Although a pervasive schistosity as defined by chlorite ± actinolite is present, individual basaltic flows of 1-50 meters thick are recognized by columnar jointing, porphyritic units, vesicular or amygdaloidal margins of flows and stratigraphic separation by volcanic breccia or metasediments.

Three transects across the Catoctin Formation on the west limb of the Blue Ridge Anticlinorium have been mapped in order to determine the physical stratigraphy. The purpose of this paper is to present a detailed description of the lithology of the Catoctin Formation, to present the detailed stratigraphy for two of the transects, located approximately 5 km apart, and to test for stratigraphic continuity between them along strike. The two sections are located at Hawksbill Mt. and near Big Meadows Campground in Shenandoah National Park southeast of the town of Luray, Virginia in the Big Meadows, Va. 7 1/2' quadrangle (Figures 2 and 3) and were chosen because of good exposure. This area was previously studied by Reed (1955) and it was his work that prompted selection of this area for study. A stratigraphic section for the third transect, located along Interstate 64 west of Charlottesville, Va., has already been published (Badger and Sinha, 1988).

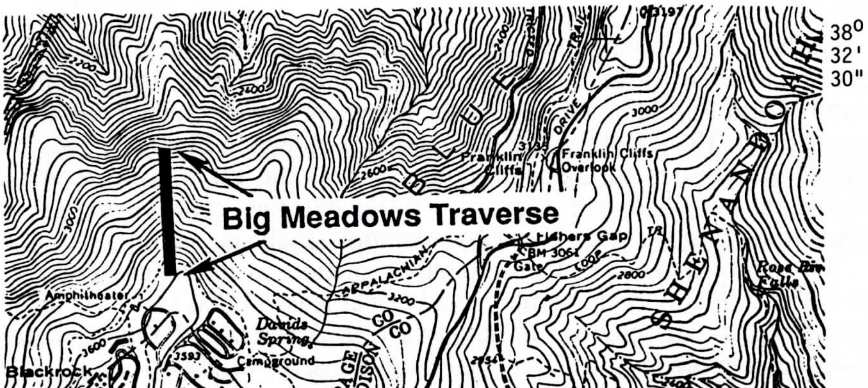


Figure 2. Location of Big Meadows traverse, shown on portion of Big Meadows 7 1/2' quadrangle.

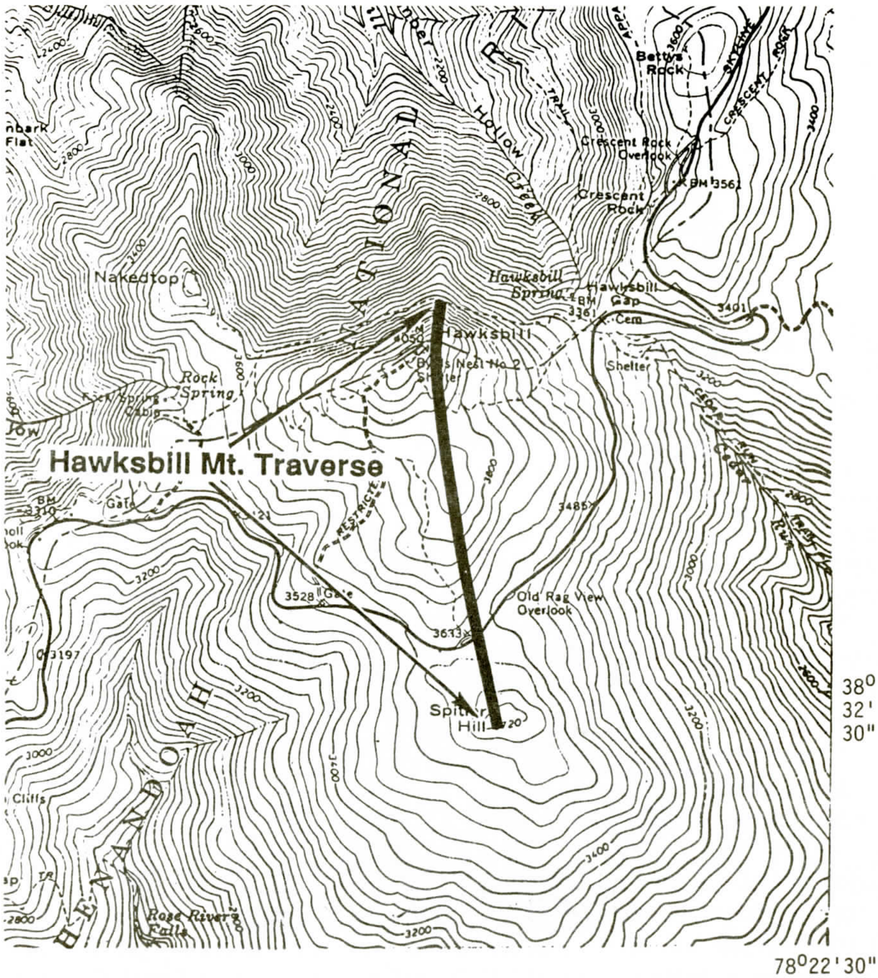
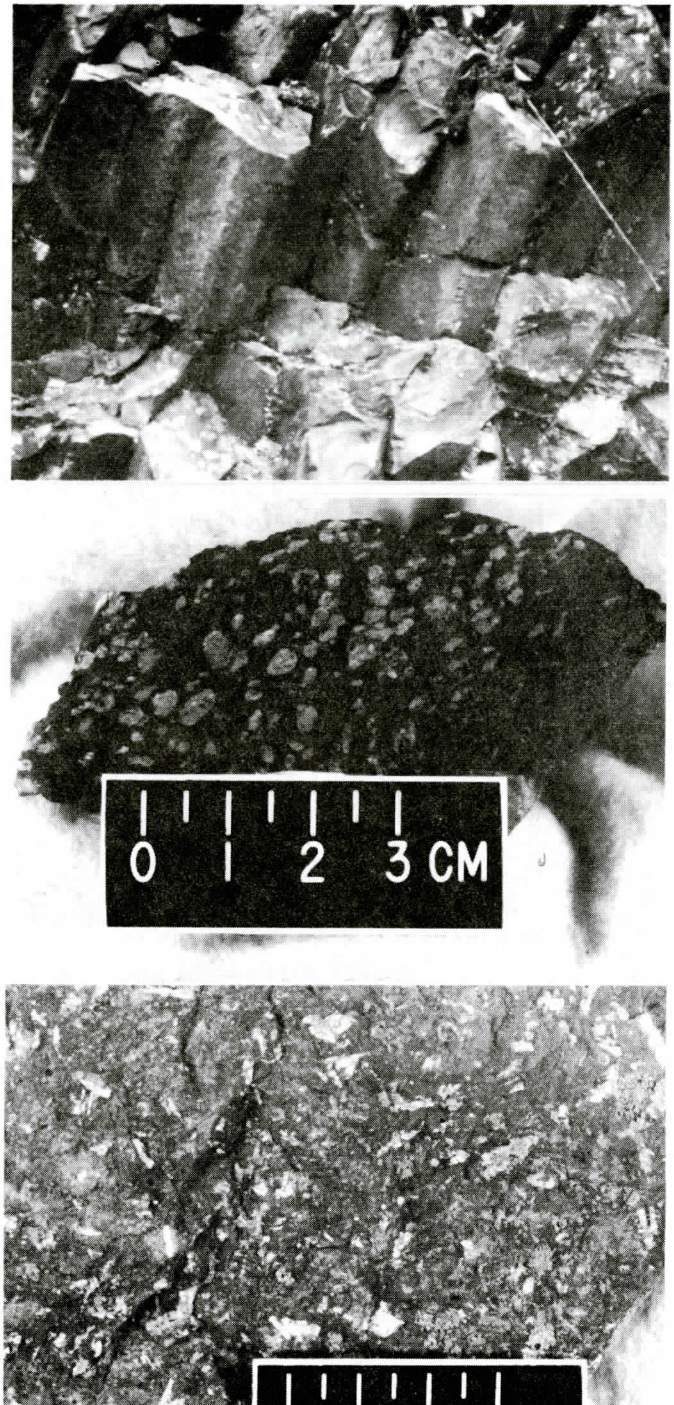


Figure 3. Location of Hawksbill Mt. traverse, shown on portion of Big Meadows 7 1/2' quadrangle.

Within the area of the Luray traverses, the stratigraphy is upright, as indicated by amygdaloidal zones at the tops of flows, relatively unsheared columnar basalts and graded bedding in the basal conglomerate beneath the Catoctin lavas. Total thickness of the volcanic sequence in the two measured sections is 200-220 meters. Exposed units within the Luray area are from the lower portion of the Catoctin Formation, overlying either the Swift Run metaconglomerate or unconformably upon the Blue Ridge Basement Complex. Upper portions of the formation have been removed by erosion and by faulting. The predominant rock type consists of 6-40 meter thick beds of basalt composed of albite + chlorite + epidote + magnetite + sphene ± hematite ± actinolite ± quartz ± relict clinopyroxene. Columnar jointing is common in most flows (Figure 4A). A poorly developed, discontinuous fabric, defined by chlorite ± actinolite, is observed throughout most of the formation. Over the bottom 2 to 3 meters of a flow, the foliation is commonly moderately to well developed, but poorly exposed. In basalts of the Columbia River Group, Swanson (1967) also recognized a subhorizontal fabric developed near the base of flows, which is interpreted as a flow fabric. Although the fabric



A

B

C

Figure 4. (A) Columnar jointing, common in many of the basaltic flows. (B) Amygdaloidal portion of a flow. (C) Porphyritic basalt, collected from near Cedar Run Trail, east of Hawksbill Mt. traverse.

at the base of Catoctin flows is defined by the alignment of a metamorphic mineral assemblage, it may merely be a reactivated flow fabric. The basal flow of the Hawksbill Mt. traverse has a basal fabric that fans upward into the plane of the columnar joints, similar to descriptions of fabric at the base of Columbia River flows that curves "upward in vertical bundles which nearly obliterate columnar structure" (Swanson, 1967, p. 1083). Many flows of the Catoctin Formation locally retain columnar jointing, with column diameters in the range of 25-80 cm.

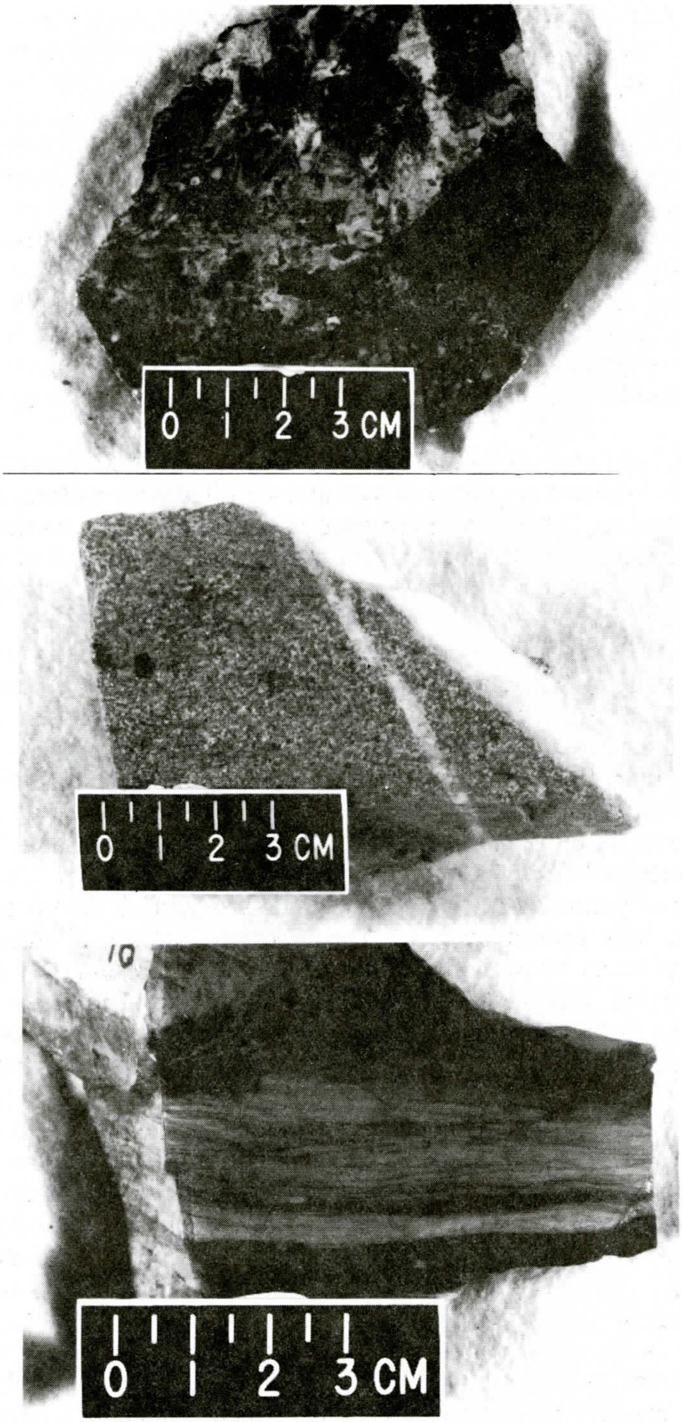
Irregular pods of epidosite (epidote + quartz) up to a meter or two in diameter are ubiquitously present within the flows and along flow margins. These have been interpreted as metamorphic segregations within the basalts during regional metamorphism (Reed and Morgan, 1971), and because of their occurrence adjacent to flow margins, particularly margins adjacent to sedimentary interbeds, and their sometimes vein-like occurrence, are interpreted as forming in zones of high fluid interaction.

Many of the flows are amygdaloidal over the top 1 to 3 meters of the flow (Figure 4B). Amygdules are round or oval and are filled with epidote, albite, quartz, calcite and hematite. These are interpreted as the vesicular or scoriacious tops of flows that were subsequently mineralized by aqueous fluids. All of the flows are aphyric except for one that contains distinctive phenocrysts of albitized plagioclase, up to 1 cm long (Figure 4C).

Several of the flows are separated by 1-10 meter thick zones of volcanic breccia consisting of angular basalt blocks, contorted lenses of purple phyllite, silicified siltstone and metasandstone, with interstices filled with amygdaloidal epidote and quartz (Figure 5A). The basalt blocks are interpreted as fragments of basalt from outer flow surfaces that fell from and were overridden by the advancing lava front, or as parts of an early crust that brecciated as the inner and still molten part of the lava flow moved (Reed, 1955; Gathright, 1976). Any underlying unconsolidated material such as stream deposits, felsic tuffs or weathered upper portions of a previous lava flow would be incorporated into this breccia as the lava flowed over it. The brecciated nature of these flows later (perhaps much later) allowed the influx of fluids that resulted in epidotization. This epidotization has rendered the breccias resistant to erosion, allowing individual units to be traced as marker horizons sometimes for hundreds of meters.

A few metasedimentary units are present in the Luray area, but none in the two detailed traverses presented below. These metasediments consist of tan to reddish brown sandstones, red siltstones and gray quartz phyllites (Figures 5B, 5C). Some graded bedding on a scale of 1-2 cm is locally evident within sandstone and siltstone beds. One notable metasediment is a 6-7 meter thick quartz phyllite located between flows 2 and 3 at Crescent Rocks (Big Meadows Quadrangle). Less than a kilometer away, on Hawksbill Mt., this metasedimentary unit is missing, limiting its value as a marker horizon. Volcanic breccias between flows frequently contain rip-up fragments of metasediments (Figure 5C), which are interpreted as lacustrine or fluvial deposits ripped up by the advancing lava.

Also present locally are thin beds of fine grained purple phyllite. Mineralogically these consist of sericite + magnetite + sphene + epidote + quartz ± chlorite. A few purple phyllites contain relict vitroclastic textures, preserving flattened pumice lapilli and flattened vesicles now filled with sericite (Figure 6A). These have been interpreted as felsic ash-flow tuffs or ignimbrites (Gathright, 1976; Gathright and others, 1977; Bartholomew, 1977). Where vitroclastic textures are not preserved, the purple phyllites are interbedded with metasediments, suggest-



A

B

C

Figure 5. (A) Volcanic breccia, collected from between flows 4 and 5, Big Meadows traverse. (B) Metasandstone, commonly found as thin interbeds between flows. (C) Fragment of crossbedded siltstone, located within a zone of basaltic breccia. Sample collected from between flows 1 and 2, Big Meadows traverse.

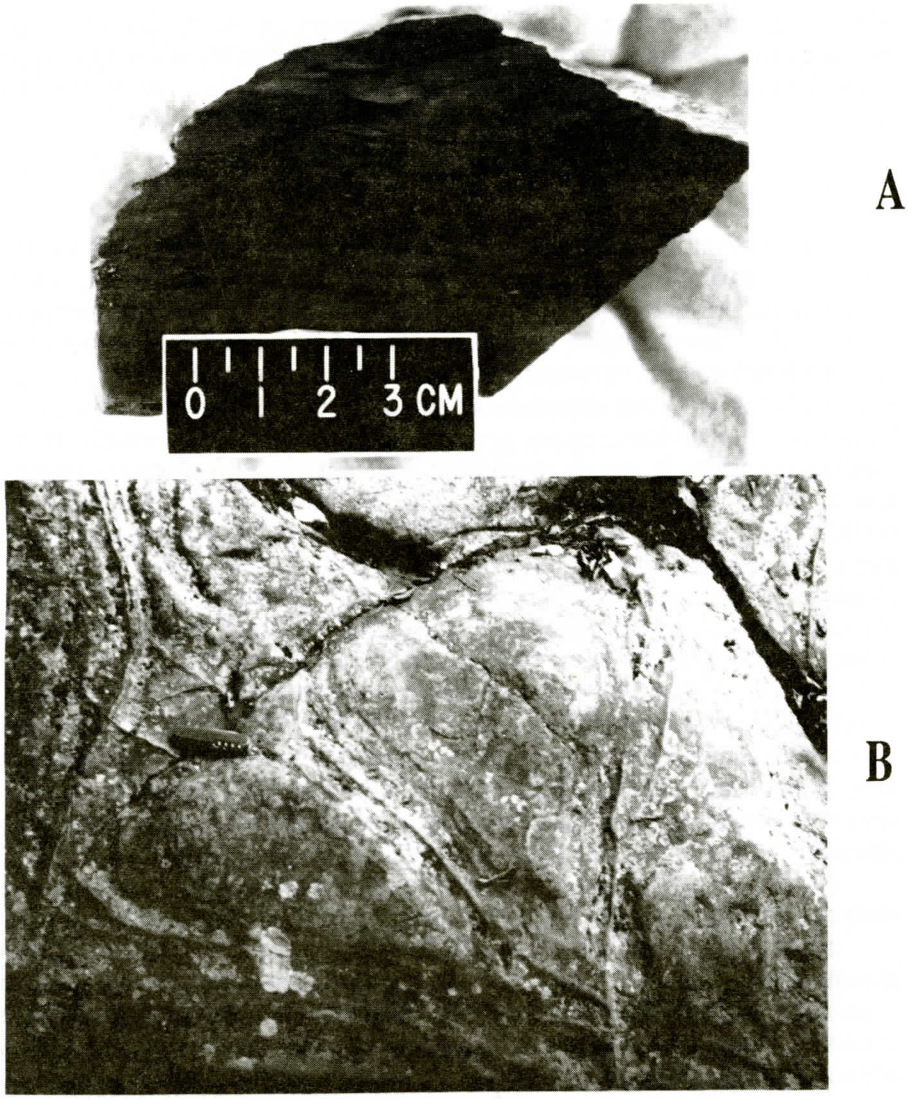


Figure 6. (A) Purple phyllite, locally found as thin interbeds between flows, and interpreted as felsic ash-flow tuffs or ignimbrites. (B) Pillow lava, located at Dark Hollow Falls, east of Big Meadows traverse.

ing reworking of the felsic tuffs by water. The presence of felsic tuffs is similar in occurrence to those reported for the Columbia River province, where an air-fall, vitric tuff, derived from eruption in the Cascade Mountains west of the Columbia Plateau, occurs directly below the Pomona Member of the Saddle Mountains Basalt (Schmincke, 1967). Explosive felsic volcanism therefore must be associated in time with eruption of the Catoctin magmas, but no genetic relationship may exist.

Only one 7-8 meter thick sequence of pillow lavas has been observed in the Luray area, located at Dark Hollow Falls about 1 km east of the Big Meadows traverse (Figure 6B). Due to the preponderance of near-by subaerial columnar jointed basalts, these pillows are interpreted to be the result of lava flowing into a lake or ponded stream and therefore strictly a local feature. A similar interpretation has been reached for flows of the Columbia River Group containing pillow lavas at the base and columnar jointing at higher levels of the flow (Swanson, 1967).

Big Meadows Traverse

The section at Big Meadows is located north of Big Meadows campground, accessed from the Appalachian Trail about 200 meters north of the Big Meadows Amphitheater (Figure 2.) The section is approximately 210-220 meters thick, and consists of 10 flows ranging in thickness from 6 to approximately 40 meters (Figure 7). The basal contact between the Catoctin Formation and rocks of the underlying Pedlar Formation is not exposed, but can be located to within 5-6 meters at several locations. No metasediments crop out along the contact which, from near-by exposed outcrops, is clearly discordant, indicating the pre-Catoctin erosion surface was quite irregular at this location. This observation is consistent with the erosion surface recognized by Reed (1955; 1969). The basal flow has a variable thickness, due to the underlying topography, of 10-20 meters. It is overlain by 1-2 meters of volcanic breccia containing irregular basalt blocks and fragments of silicified siltstone up to 30 cm long (Figure 5C). Graded bedding can still be observed in one siltstone fragment, indicating its sedimentary origin.

Flow 2 is about 30 meters thick, locally containing relict columnar jointing displayed in prominent ledges. A small 3 meter wide break in slope about half way up does not appear to indicate a break in stratigraphy as no vesicles indicating tops or bottoms of flows, and no volcanic breccia or metasediments, are evident. On top of flow 2 is a 6-7 meter thick zone of schistose volcanic breccia containing abundant lenses of red chert and siltstone.

Flow 3 also displays columnar jointing. It has a total thickness of about 15 meters. Overlying it is a 3-5 meter thick zone of volcanic breccia.

Flow 4 crops out only sporadically along a moderate slope, and is only 6-7 meters thick. The sparse outcrops display faint columnar jointing. Above flow 4 is an approximately 10 meter thick volcanic breccia containing contorted silicified siltstone, along with fragments of purple phyllite, tan epidotized sandstone and angular basalt blocks.

Flow 5 displays local columnar jointing, and also contains local zones of volcanic breccia within the flow. The volcanic breccia consists of angular basalt blocks, a result of the formation of a crust on the outer, cooler edges of the flow, that was broken up and incorporated into molten portions as the flow advanced. This type of feature is well documented for other volcanic provinces such as Hawaii (Richter and Moore, 1966) and the Columbia River Basalts (Swanson, 1967). Thickness of the flow is about 20 meters.

Flow 6 is the porphyritic unit, containing phenocrysts of albitized plagioclase up to 1 cm long; total thickness is approximately 15 meters. The basal 1-2 meters of the flow are amygdaloidal, with some of the amygdules unfilled, but not flattened. No preferred orientation of phenocrysts has been observed. In outcrop the flow crumbles rather readily, and so does not form prominent ledges. The unit can be traced sporadically for several kilometers, and thus provides a useful marker bed for mapping, as first used by Reed (1955). Float of purple phyllite near the unexposed top of the flow indicates there may be a thin metasedimentary interbed overlying this flow.

Big Meadows Traverse

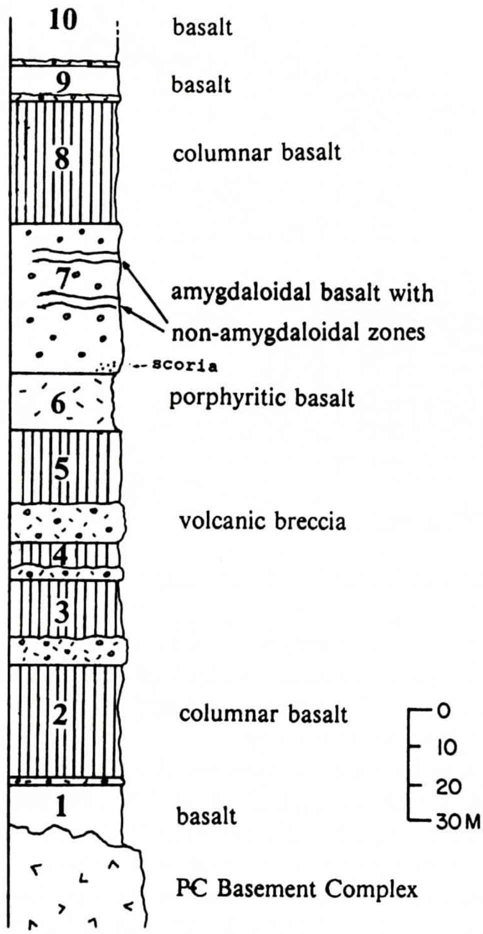


Figure 7. Big Meadows stratigraphic column; flows are labeled 1-10. See text for discussion of individual flows.

Flow 7 consists of a massive amygdaloidal basalt containing amygdules filled with quartz, epidote, chlorite and hematite. At one outcrop near the base of the flow, a zone of scoria a few centimeters thick has been preserved (Figure 8). At this location, interconnected vesicles comprise approximately 50% of the rock, and apparently permeability was so high that if metamorphic fluids moved through this portion of the flow, they were never trapped long enough to precipitate minerals. Adjacent to this zone, vesicle density is slightly less, and vesicles have been mineralized to form amygdules. Density of amygdules varies considerably over the width of the entire flow with some zones nearly amygdule free. The formation of multiple vesicular (and hence, amygdaloidal) zones within a flow has recently been described by McMillan and others (1987) and Long and McMillan (1988) as caused by upward migration of aqueous vapor bubbles that get trapped by overlying zones of slightly higher viscosity. The fact that this flow is distinctive at Big Meadows and also on the flanks of Hawksbill Mt., 5 km to the north, as well as

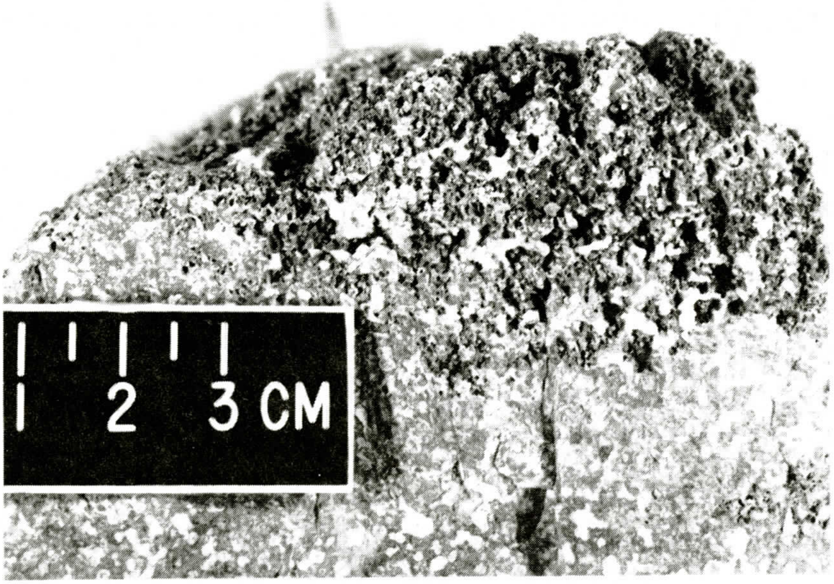


Figure 8. Sample of scoria, collected from base of flow 7, Big Meadows traverse.

at several locations in between, suggests that it is not a local feature.

Flow 8 forms the most prominent ledges of columnar basalt in the traverse. The unit is approximately 30-35 meters thick and contains a 2-3 meter thick zone of epidotized volcanic breccia at the top.

Flow 9 crops out in low ledges between the top of the cliffs marking flow 8 and the flat area upon which the Appalachian Trail is located, for a thickness of 8-10 meters. There are several good exposures at lookouts along the Appalachian Trail just north of the Big Meadows amphitheater. Float of amygdaloidal breccia in the flat area between flows 9 and 10 indicates there probably is a 1-2 meter thick zone of volcanic breccia on top of flow 9, but it is not exposed.

Flow 10 crops out in low ledges behind the Big Meadows Campground amphitheater. Only about 6-8 meters is exposed, so either the upper part has been eroded, or it was a thin flow.

Hawksbill Mt. Traverse

The section at Hawksbill Mt. is approximately 5 km northeast of the section at Big Meadows. The traverse begins on the north side of Hawksbill Mt., accessed from the Appalachian Trail about 1 km east of Hawksbill Gap (Figure 3). The traverse climbs the steep side of the mountain to the top (flow 5), then gently descends along the southeast side. The southeast slope is only slightly gentler than the southeast dip of the rocks in this area, so it nearly represents a dip slope. Absent are the prominant ledges of outcrop, as seen along the northwest side of the mountain, so observation of stratigraphic features and estimation of bed thicknesses are difficult.

The section is approximately 200 meters thick and consists of 9 flows ranging in thickness from 10 to approximately 40 meters (Figure 9). Underlying the

Hawksbill Traverse

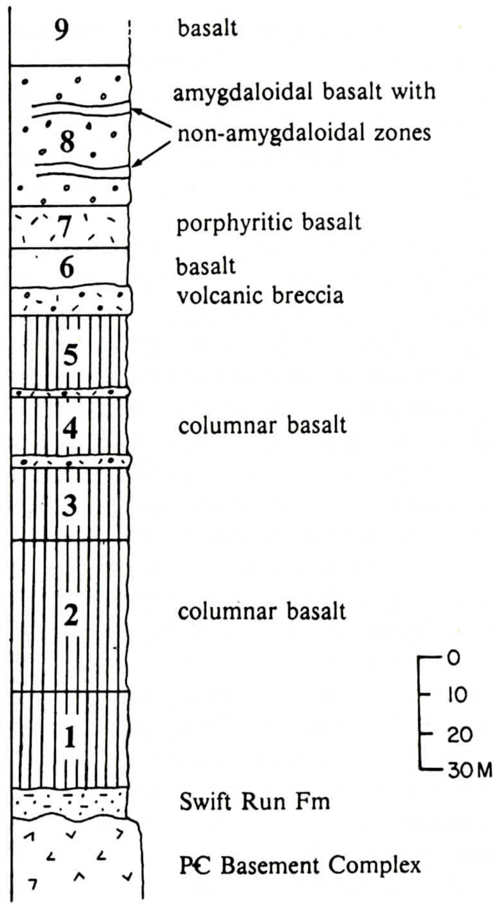


Figure 9. Hawksbill Mt. stratigraphic column; flows are labeled 1-9. See text for discussion of individual flows.

Catoctin Formation are 6-7 meters of gray phyllite and conglomerate of the Swift Run Formation. The metasediments rest unconformably upon the Pedlar Formation of the Precambrian basement complex. The upper 1-2 meters of the conglomerate contain thin stringers of greenish gray phyllite. The basal contact of flow 1 with the Swift Run Formation is irregular, reflecting an irregular erosion surface and perhaps zones where the unconsolidated sediments were gouged out by the advancing lava flow. The bottom 6 meters of the flow is schistose, containing flattened chlorite-filled vesicles and reflects a fanning cleavage. Above this schistose zone the flow contains columnar jointing. About half way up the flow is a 1 meter thick pronounced schistose zone which does not appear to mark a flow boundary. Total thickness of flow 1 is 25-30 meters. The top of the the flow is marked by a break in slope along a flat ledge.

Flow 2 has a 1 meter thick schistose zone at its base. The flow consists of massive columnar basalts with column diameters generally 30-80 cm and column axes plunging 65° due west. This is the best example of columnar jointing in the sequence. The unit is about 40 meters thick. Its top is marked by an intensifying schistosity over the top 1 meter and topographically is marked by a flat ledge and sharp break in slope.

Flow 3 only crops out in sporadic 2-8 meter high ledges along a moderate slope. Columnar jointing is present, and numerous pods of epidosite, up to 2 meters in diameter, are prominent. Total thickness is approximately 25 meters. The top of the flow is marked by a meter of shaley greenstone. Overlying flow 3 are 3-5 meters of volcanic breccia consisting of epidotized and amygdaloidal basalt blocks mixed with contorted red chert and gray phyllite. Amygdules are filled with quartz, albite, epidote, calcite and hematite.

Overlying the volcanic breccia, flow 4 is very similar to flow 3, consisting of sporadic ledges of massive columnar basalt. The basal 1 meter has a pronounced schistosity while the top is marked by a break in slope and a 2-3 meter thick zone of volcanic breccia. Total thickness is about 20 meters.

Flow 5 consists of 25 meters of massive columnar basalt forming impressive cliffs at the top of Hawksbill Mt. Measured diameters of columns range from 25 to 85 cm. The metabasalts are slightly magnetic, reflecting the presence of abundant metamorphic magnetite. Overlying flow 5, just east of the peak on Hawksbill Mt., are abundant outcrops of volcanic breccia composed of oxidized amygdaloidal basalt breccia and epidosite mixed with fragments of epidotized sandstone, purple phyllite and silicified red siltstone. Some of the siltstone fragments still retain graded bedding. This breccia zone crops out extensively along the top of the east dipping slope, and locally has a total thickness of at least 6 meters. The presence of this thick bed of very resistant rock on top of Hawksbill Mt. may be the reason that this is the highest point in Shenandoah National Park.

Flow 6 occurs as sparse, small, flat lying outcrops along the south flank of Hawksbill Mt. The top meter contains flattened chlorite filled vesicles. Total thickness of the flow is approximately 10-12 meters.

Flow 7 contains plagioclase phenocrysts, now altered to albite, up to 1 cm in length. On a weathered surface, the albite lathes stand out as white phenocrysts and, as such, provide a useful marker bed that can be traced sporadically for several kilometers. This marker bed is useful to help correlate flows between Hawksbill Mt. and the section at Big Meadows campground, approximately 5 km to the southwest. Total thickness of the porphyritic unit is 10-12 meters. No preferred orientation of phenocrysts has been observed, so flow direction can not be determined from them.

Flow 8 is approximately 30-40 meters thick, and is amygdaloidal throughout. Some zones contain abundant amygdules, filled with quartz, albite, epidote, calcite and hematite, while in other portions amygdules are sparse. Amygdaloidal zones frequently contain a few small albite phenocrysts. Lithologically, this flow is similar to flow 7 at Big Meadows, which crops out immediately above the porphyritic unit, as does this flow. Exposure of the flow is sparse along the east flank of Hawksbill Mt., but the unit appears to extend all the way to upper Hawksbill parking area along Skyline Drive, and across the road towards Spitler Hill.

Flow 9 crops out on top of Spitler Hill, as small ledges of greenstone. A few small phenocrysts of plagioclase can be seen locally and in thin section, but this is not the second porphyritic unit shown on the map of Reed (1955; 1969), which is further to the east. Scarcity of outcrop does not permit the continuation of the section to the east to include the second porphyritic unit.

Stratigraphic Correlation

The two sections at Luray were mapped approximately 5 km apart in order to test for stratigraphic continuity between them (Figures 7, 9, 10). Because the physical stratigraphy is not continuous, the sections cannot be directly correlated. However, the distinctive porphyritic flow (Figure 4C), 6th flow from the bottom at Big Meadows and 7th from the base at Hawksbill Mt., provides a useful marker bed upon which to base correlations, as first used by Reed (1955). The amygdaloidal flow located directly above the porphyritic flow in both traverses also provides a good marker unit. Below the porphyritic flow, correlation is more tenuous (Figure 10), but most likely occurs sequentially at least for the next three flows down through the section. The flow absent at Big Meadows is most likely one of the lower flows, probably the lowest, reflecting the pre-Catoctin erosion surface. No evidence is seen to suggest the absence of one or more flows may be due to syn- rift faulting or subsequent thrusting. Further evidence of a pre-volcanic erosion surface is provided elsewhere in Shenandoah National Park, southeast of Fisher's Gap (Big Meadows Quadrangle) and at the lower (east) end of White Oak Canyon (Old Rag Mt. Quadrangle) by two granitic monadnocks that can be shown to have volcanic flows truncated against them (see map and Figure 3 of Reed, 1969, and map of Gathright, 1976). Also, about 200-300 meters west of Spitler Hill (Big Meadows Quadrangle) at outcrops along the Appalachian Trail, a well sorted coarse sandstone on top of the granitic basement complex probably represents a former stream bed on the pre-volcanic erosion surface.

DURATION AND REGIONAL EXTENT OF VOLCANISM

Absence of major erosional unconformities between flows and absence of thick sequences of sedimentary interbeds suggest eruption time for the Catoctin magmas was probably short. Many flows show flow-on-flow contacts, whereas other flows are separated by zones of volcanic breccia containing contorted fragments of the metasediments, indicating the sediments were still unconsolidated when overridden by the next flow and incorporated into the breccia. Where flows can be traced and correlated, no unconformities are observed, and thicknesses of flows remain remarkably constant, suggesting time intervals between flows were too short to permit significant erosion. This is consistent with short eruption times for other flood basalt provinces. For example, the majority of Deccan Traps volcanism occurred during a 5 million year time span, 65-60 Ma (Mahoney, 1984), and 95% of the Columbia River Plateau basalts were generated over a mere 3 million years, 17-14 Ma (Swanson and Wright, 1981).

Individual basalt flows of the Columbia River Province have been traced over areas of hundreds of square kilometers (Waters, 1961; Swanson and Wright, 1981). In the Blue Ridge province, the structural configuration and erosion make correlation of units extremely difficult. The longest distance a single flow can be traced is approximately 8 km for the porphyritic flow at Luray. While trying not to contradict this contention that some flows may have had a regional extent of several tens or even hundreds of square kilometers, it is also observed that south of the traverses at Luray, sedimentary interbeds are more abundant and volcanic flows are generally thinner and have a smaller lower limit in thickness (Badger and Sinha,

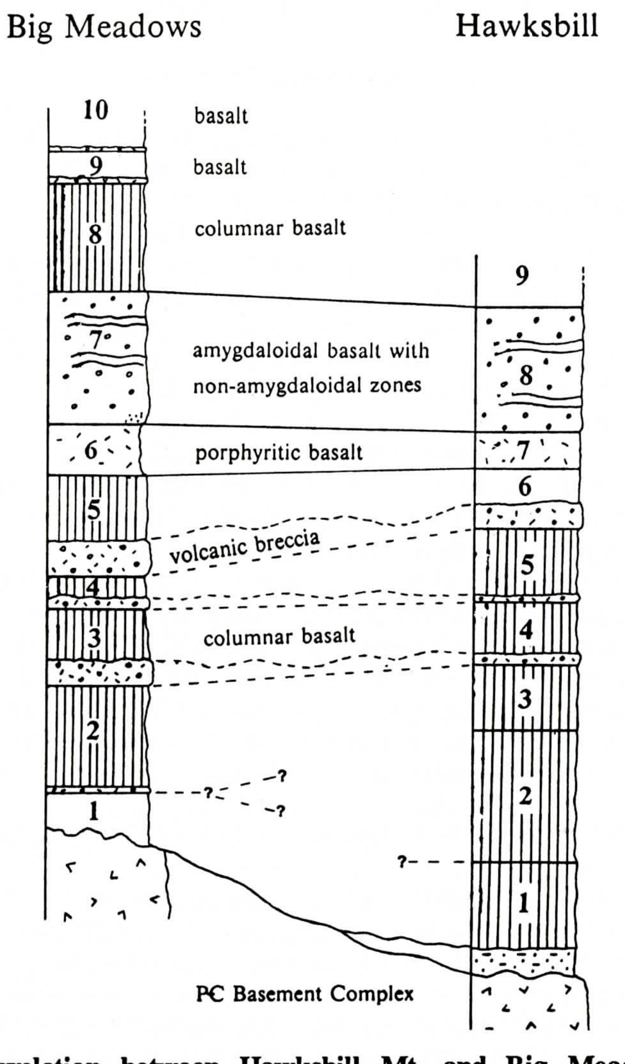


Figure 10. Correlation between Hawksbill Mt. and Big Meadows sections. Correlation between porphyritic and amygdaloidal flows is probable. Other correlations more tenuous. See text for discussion.

1988), than for the traverses at Luray. Figure 11 compares thickness of flows between the Hawksbill Mt. traverse and part of the traverse along I- 64 (previously published in Badger and Sinha, 1988). The inference is that the area south of Luray may have been farther from the locus of volcanic activity, and that only the distal edges of some flows reached the area of the I-64 traverse. This would be consistent with a stratigraphic pinching out of the Catoctin Formation in southern Virginia. If dikes observed in the Blue Ridge Basement Complex were indeed feeder dikes to Catoctin flows, it is interesting to note that considerably more dikes are observed north and east of the Luray area, particularly in the dike swarm around Old Rag Mt., than in the Blue Ridge Basement Complex south of the Luray area.

Presence of thin beds of purple phyllite, interpreted as felsic tuffs (Furcron and Woodward, 1936), is not compatible with a tholeiitic province. However, as previously noted, a similar occurrence has been reported within the Columbia

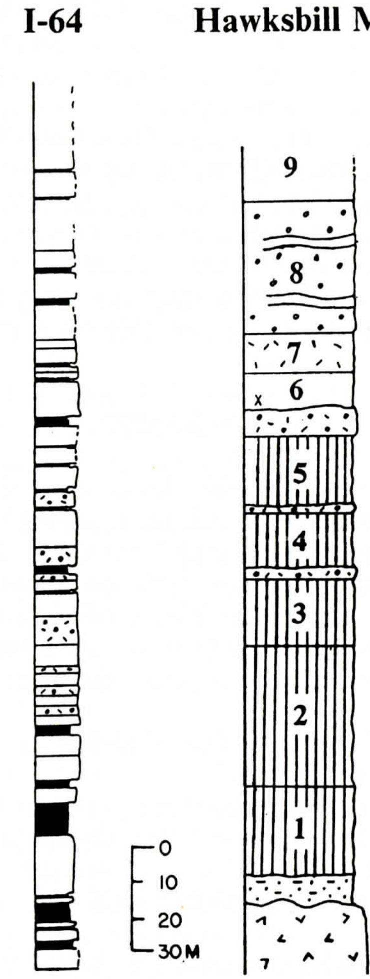


Figure 11. Comparison of thickness of flows between I-64 and Hawksbill Mt. I-64 section from middle of the sequence (see Badger and Sinha, 1988); solid pattern = metasedimentary interbeds. Both sections drawn to same scale, showing flows at I-64, on the average, are considerably thinner than for the Luray traverses, suggesting the section at I-64 may be further from the source.

River Province, and has been interpreted as an air-fall vitric tuff derived from eruption in the Cascade Mountains, over 200 km to the west (Schmincke, 1967). The nearest felsic volcanism to the Catoctin Formation in central Virginia that may have been contemporaneous is at Mount Rogers, 300 km to the southwest in southwestern Virginia, or at South Mountain, 200 km to the north in southern Pennsylvania. Peralkaline affinities for the Mount Rogers volcanics suggests its correlation with peralkaline granites of the Crossnore suite that yield Rb-Sr isotopic ages of 710-680 Ma (Odom and Fullagar, 1984) and U/Pb zircon ages of 729-640 Ma (Tollo and others, 1991), and a recent U-Pb zircon age for a rhyolite of the Mount Rogers Formation yields an age of 759 Ma (Aleinikoff and others, 1991). However, an Rb-Sr age for Catoctin volcanism has been calculated to be 570 Ma (Badger and Sinha, 1988). This latest Precambrian age for Catoctin magmatism is consistent with stratigraphic and paleontological evidence dating the base of the overlying Chilhowee Group metasediments as earliest Cambrian (Simpson and Sundberg, 1987). Magmatism at Mount Rogers therefore appears to be 100-200 Ma older, and could not be the source of vitric air-fall tuffs between Catoctin flows. Rhyolites at South Mountain, however, appear to be interbedded with basalts mapped as part of the Catoctin Formation (Fauth, 1973) and therefore may be comagmatic. A recent U-Pb zircon age for a rhyolite at South Mountain yields an age of 565 Ma (Aleinikoff and others, 1991). This provides strong evidence that mafic flows of the Catoctin Formation in central Virginia may have been synchronous with felsic magmatism at South Mountain, Pa. Regardless of a genetic relationship between the two magmatic events, their proximity in age provides a source and explanation for the thin felsic tuffs found between mafic flows in the study area.

CONCLUSION

Recognition of individual basaltic flows in conjunction with interbedded volcanic breccias and sedimentary interbeds permits the documentation of a physical stratigraphy within the Catoctin Formation. This physical stratigraphy provides a basis upon which to discuss correlation of units, duration of magmatism, source and possible flow direction. The physical stratigraphy also provides a framework upon which to construct a chemical stratigraphy (Badger and Gottfried, in progress) in order to study the evolution of the magmas.

ACKNOWLEDGEMENTS

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TECTONIC SIGNIFICANCE OF CROSS-STRIKE FAULTS IN THE CENTRAL APPALACHIAN GREAT VALLEY OF MARYLAND AND WEST VIRGINIA

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ABSTRACT

Structural analysis of strike-slip and dip-slip cross-strike faults in Cambrian and Ordovician carbonate rocks of the Great Valley of Maryland and West Virginia indicates a previously unrecognized phase of deformation. Although fault surfaces are rarely exposed, the fault orientations can be inferred from offsets of stratigraphic units, preexisting folds and faults, and linear topographic features. Orientation analysis of calcite veins, joints, and tension gashes (en echelon and sigmoidal) indicates two distinct maximum-principal-stress directions. The first direction is associated with a N10° E structural trend, and the second, with a N30° E trend. Offset cleavage, sigmoidal tension gashes, and mesoscopic faults suggest that a continuum of deformation occurred during the Alleghanian orogeny that progressed from ductile-brittle to brittle. The cross-cutting relationships of mesoscopic structures and the resulting maximum-principal-stress orientations demonstrate that the preexisting N10° E trend of structures was overprinted by the later deformation phase. Strike-slip and dip-slip cross faults formed simultaneously during the final stage of brittle deformation and are related to the second maximum-principal-stress orientation.

INTRODUCTION

Cross-strike faults have been recognized throughout the Appalachian Valley and Ridge province (Butts, 1933; Bick, 1960; King and Ferguson, 1960; Page and others, 1964; Root, 1973; Root, 1978; Nickelsen, 1979; Wheeler and others, 1979); however, most of these faults have been mapped in the Great Valley of Maryland and West Virginia. Sando (1957) and Dean and others (1987) mapped strike-slip and dip-slip faults that cross regional structural trends at high angles in Cambrian and Ordovician carbonate rocks of the Great Valley of Maryland and West Virginia (Figure 1). Cross faults have been variously mapped in the Valley and Ridge as small tear faults that offset the leading edge of thrust sheets (Donaldson and others, 1964), as wrench faults related to changes in deformation mechanisms (Nickelsen, 1979), and as faults that accommodate structures having different wavelengths and different amounts of shortening (Root, 1973; Root and Hoskins, 1977; Schasse, 1979).

The purpose of this research is to determine the direction of fault movement and the kinematics of faulting and to refine methods of determining fault traces in the Great Valley of the central Appalachians. For this study, two cross faults were mapped in detail: (1) the Rocky Marsh strike-slip cross fault located west of

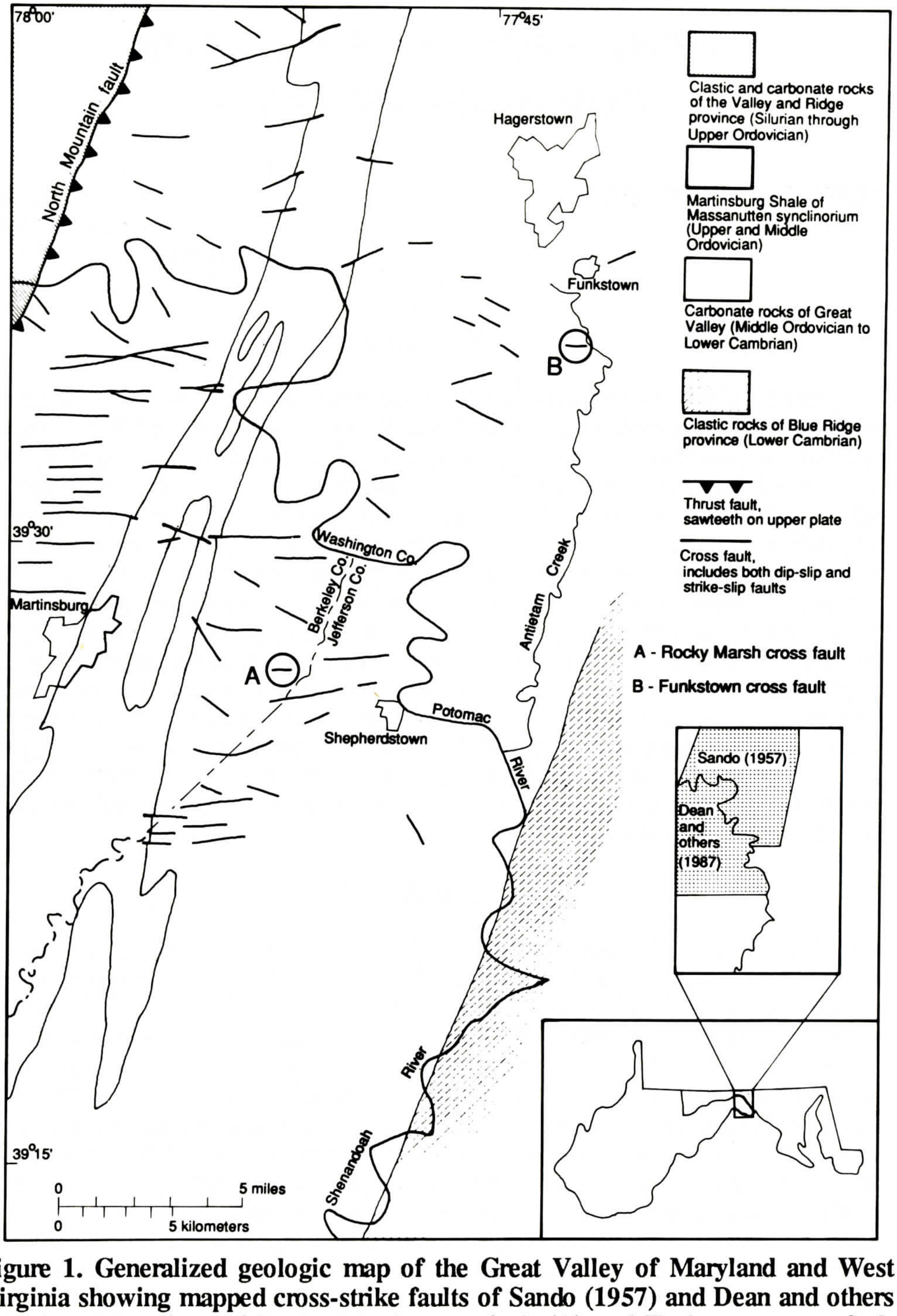


Figure 1. Generalized geologic map of the Great Valley of Maryland and West Virginia showing mapped cross-strike faults of Sando (1957) and Dean and others (1987) and locations of Rocky Marsh cross fault (A) and Funkstown cross fault (B).

Shepherdstown, West Virginia and the Funkstown dip-slip cross fault south of Funkstown, Maryland (Figure 1). Both cross faults occur in the Cambrian and Ordovician Conococheague and Stonehenge Limestones. Mesoscopic-scale joints, faults, and tension gashes (en echelon and sigmoidal) were studied to define fault kinematics and to quantify orientations of principal stresses during deformation. In addition, 23 other cross faults were checked in the field and examined on aerial photographs. These cross faults were compared to predicted fault patterns, which were in turn based on regional-scale Great Valley structural trends.

The Cambrian and Ordovician carbonate rocks of the Great Valley of the central Appalachians is characterized by asymmetric folds and thrust faults. These rocks occur west of the Blue Ridge province and east of the North Mountain thrust fault (Figure 1). The major structure within the Great Valley is the Massanutten synclinorium that extends northeast and southwest of the study area.

The lack of outcrops in the central Appalachians makes mapping the precise orientation of fault planes difficult. Several methods were used to define the orientation of the cross faults. For example, the orientation of mesoscopic faults, shear sets, and joints was compared to the predicted orientation of structures determined from the maximum-principal-stress directions inferred from regional and local structures. In the field, cross faults could only be inferred by brittle structures (fractures, mesoscopic faults, anomalous structural trends) and offset stratigraphy. Joints and calcite-filled veins were measured to compare with cross fault orientations. Data on joints include orientation, exposed length and shape, and relationship to other fractures. Data were collected in a 1-square meter area at 15 randomly selected stations on outcrops near the cross faults.

DETAILED GEOLOGY OF CROSS FAULTS

Rocky Marsh Cross Fault

The Rocky Marsh cross fault zone is located 5.6 kilometers (3.5 miles) west of Shepherdstown, West Virginia, on State Highway 45, just west of Rocky Marsh Run, Shepherdstown 7-1/2 minute quadrangle. One right-lateral strike-slip cross fault (previously mapped by Dean and others, 1987) offsets a topographic ridge and anticlinal strike belt of Lower Ordovician carbonate rocks (Figure 2). A second right-lateral strike-slip cross fault offsets the anticline exposed in a road cut. The structure cut by the cross faults is a tight anticline having the Stoufferstown Member of the Stonehenge Limestone in its core. Spaced solution cleavage is prominent in these units and is generally vertical to steeply northwest dipping.

Detailed mapping (1:1000 scale) shows that the ridge of the Stoufferstown Member has 53 meters (175 feet) of right-lateral offset (Figure 2). Various orientations of cleavage exist in the zone of faulting (Figure 3). The Stoufferstown Member is exposed on both sides of State Highway 45. There is little indication of topographic offset at this locality, however, the anticlinal hinge has 8 meters (26 feet) of right-lateral offset.

Joints, calcite veins, mesoscopic faults, and tension gashes (en echelon and sigmoidal) (Figure 4) are abundant in exposures near the faults. Joint sets are oriented N75° W and N40° W and are nearly vertical. The majority of the sigmoidal tension gashes trend N65° E, and the shear zone containing the gashes trends N70° W and has left-lateral displacement. The conjugate right-lateral

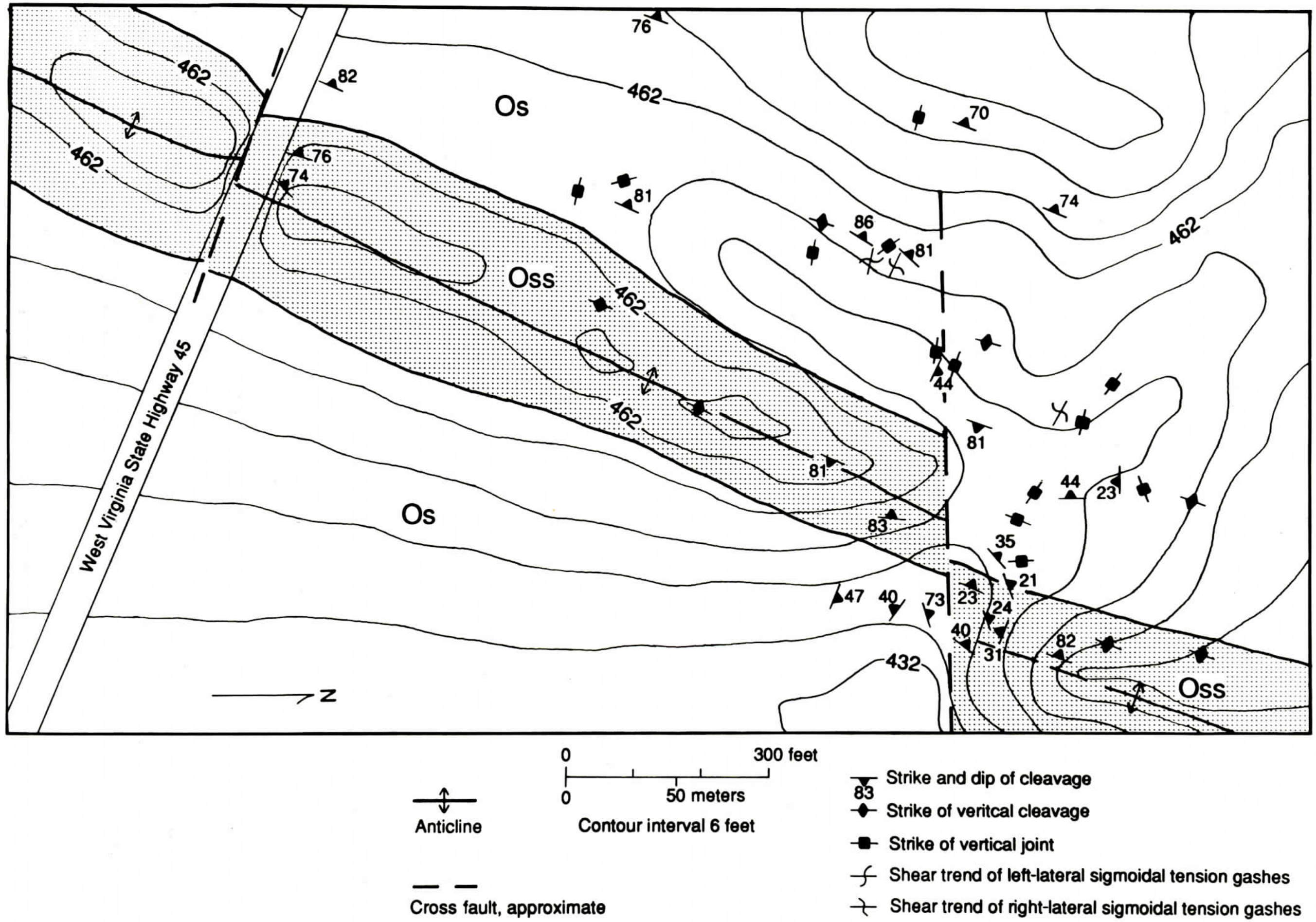
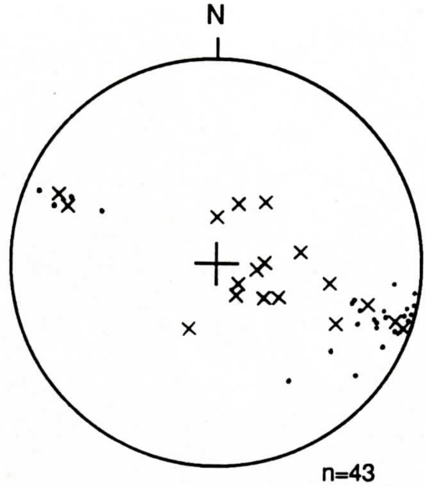


Figure 2. Detailed geologic map of the Rocky Marsh cross fault zone, Berkeley County, West Virginia (A on figure 1). Os, upper member of the Stonehenge Limestone (Lower Ordovician); Oss, Stoufferstown Member of Stonehenge Limestone.



× Measurements from cross fault zone

• Measurements from area near cross fault

n, number of measurements

Figure 3. Equal area stereographic projection (lower hemisphere) of poles to cleavage for the Rocky Marsh cross fault zone, Berkeley County, West Virginia. Note the scattered poles reflecting the various cleavage orientations in outcrops in the cross-fault zone.

component for this set (set 1) is exposed infrequently, but a set of right-lateral sigmoidal tension gashes (set 2) trend in the same direction as the left-lateral shears of set 1. Also, right-lateral mesoscopic faults trend N70° W and have centimeter-scale offsets.

Funkstown Cross Fault

The Funkstown cross fault is located 3.2 kilometers (2 miles) south of Funkstown, Maryland, Funkstown 7-1/2 minute quadrangle, on Garris Shop Road. This fault was first recognized by Sando (1957). The study area was mapped at a scale of 1:2500 and includes exposures along Antietam Creek and exposures in a pasture to the west. A north plunging asymmetric syncline occurs in this area (Figure 5). The upper member of the Stonehenge Limestone is exposed in the core and the Stoufferstown Member of the Stonehenge Limestone and the Conococheague Limestone are exposed on the limbs. The eastern limb of the syncline is faulted by a high-angle thrust that places the Conococheague Limestone over the middle part of the Stonehenge Limestone. The cross fault offsets stratigraphic units and truncates the Stoufferstown Member, which is absent north of the cross fault on the eastern limb. Joints, mesoscopic faults, and tension gashes (en echelon and sigmoidal) are common.

The relationship of the high-angle thrust to the cross fault is difficult to determine because exposures are limited where the two faults intersect. It is impossible to determine if the cross fault continues eastward within the Conococheague Limestone or terminates at the thrust fault. The cross fault probably has relative dip-slip movement down to the north (Figure 6) and erosion

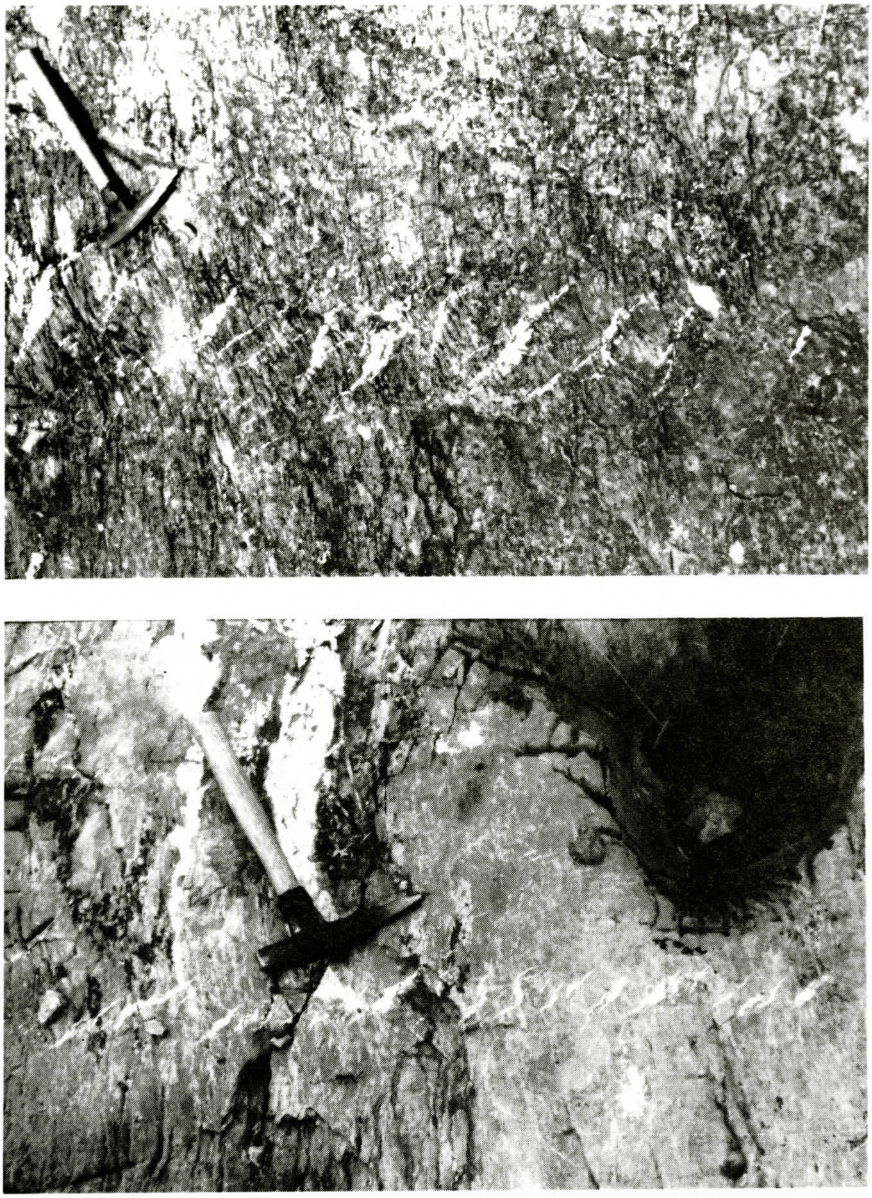


Figure 4. Photographs of set 1 sigmoidal tension gashes on horizontal pavement outcrops in the Rocky Marsh cross fault zone. Direction of shearing is left lateral on both sets. Handle of rock hammer points north.

has exposed the Stoufferstown Member on the south side of the cross fault.

In the area of the Funkstown cross fault, there are abundant calcite-filled veins, mesoscopic faults, and tension gashes (en echelon and sigmoidal). Two sets of sigmoidal tension gashes show nearly equal numbers of left- and right-lateral shearing. The trends of the set 1 left-lateral and set 2 right-lateral tension gashes are N70° W. In outcrops of the middle Stonehenge Limestone, joints and calcite-filled veins are prevalent.

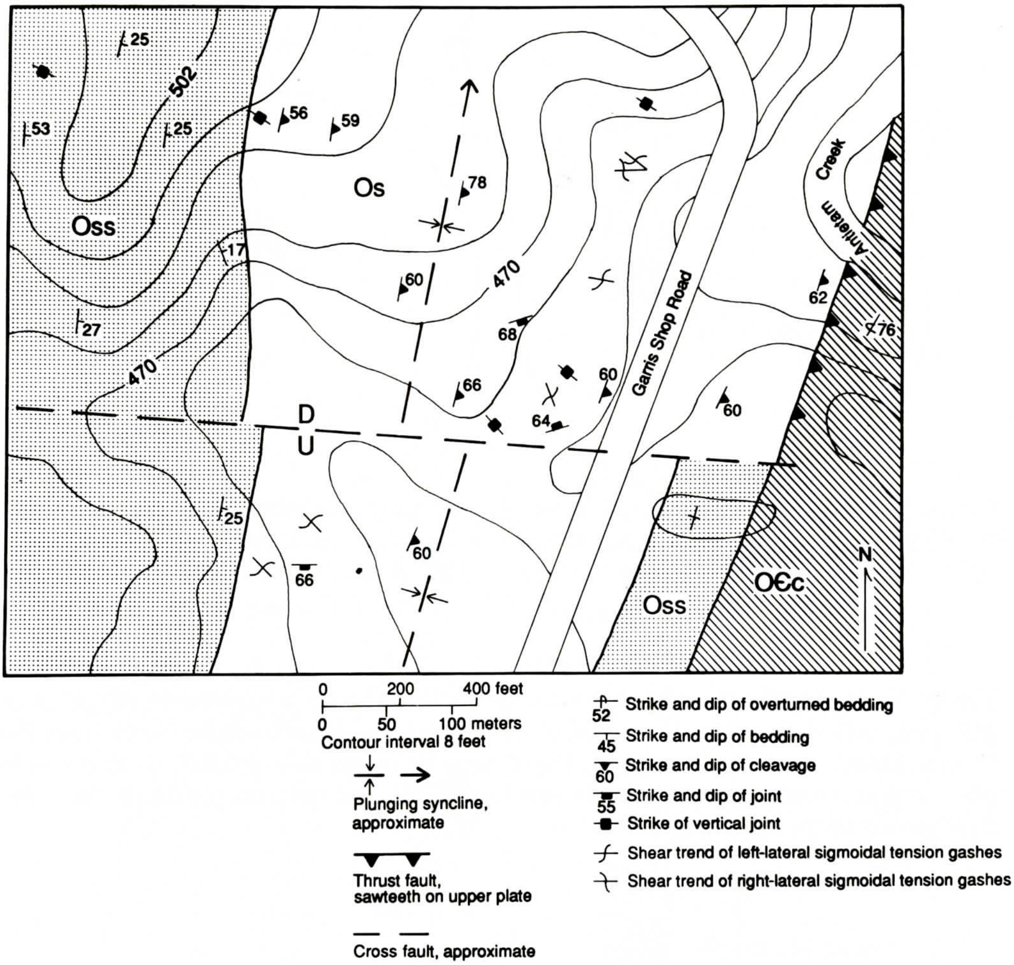


Figure 5. Detailed geologic map of the Funkstown cross fault, Washington County, Maryland (B on figure 1). OЄco, Conococheague Limestone (Upper Cambrian and Lower Ordovician); Os, Stonehenge Limestone (Lower Ordovician); Oss, Stoufferstown Member of Stonehenge Limestone.

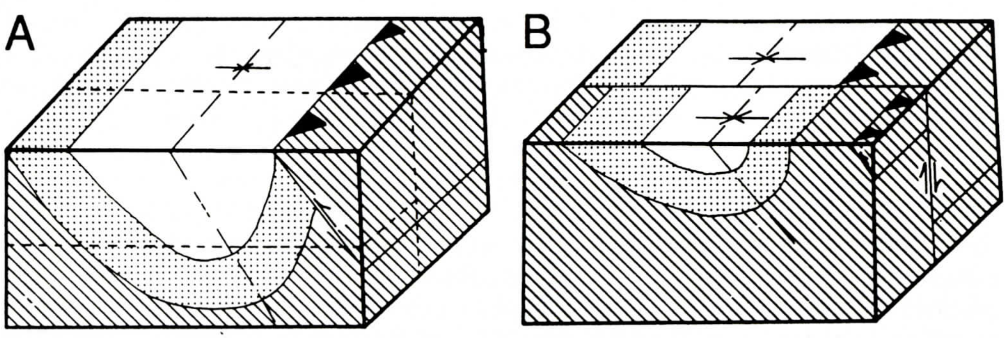


Figure 6. Block diagram showing the possible evolution of the Funkstown cross fault. A, before cross faulting; B, after cross faulting and subsequent erosion.

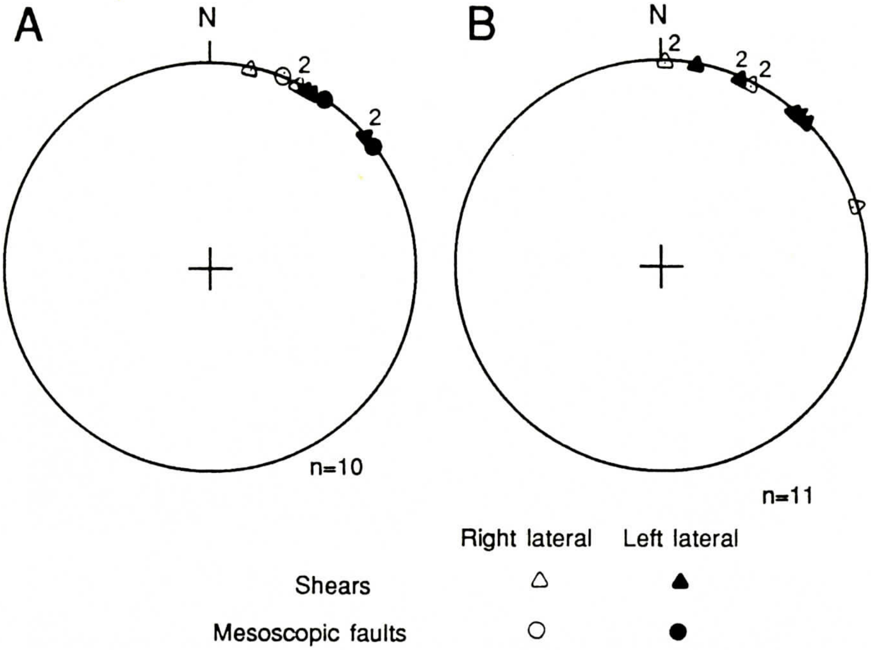


Figure 7. A, Equal area stereographic projection (lower hemisphere) of poles to planes of en echelon and sigmoidal tension gashes and mesoscopic faults from the Rocky Marsh cross fault area. B, Equal area stereographic projection of poles to planes of en echelon and sigmoidal tension gashes and mesoscopic faults from the Funkstown cross fault area.

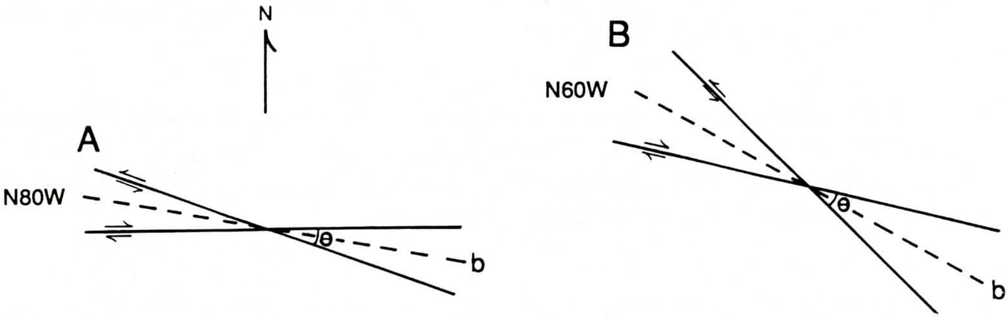


Figure 8. Diagrammatic representation of orientation of set 1, A and set 2, B conjugate shear zones and their interpreted strain ellipses. θ, dihedral angle; b, acute bisector.

ANALYSIS OF STRUCTURES

Both the Rocky Marsh cross fault zone and the Funkstown cross fault have two sets of conjugate sigmoidal and en echelon tension gashes; left-lateral shear planes of set 1 and right-lateral shear planes of set 2 are oriented approximately parallel (Figures 7 and 8). The dihedral angle of set 1 is between 10 and 15 degrees, and its bisector is oriented approximately N80° W (Figure 8). Set 2 has a dihedral angle between 20 and 30 degrees, and its bisector is oriented approxi-

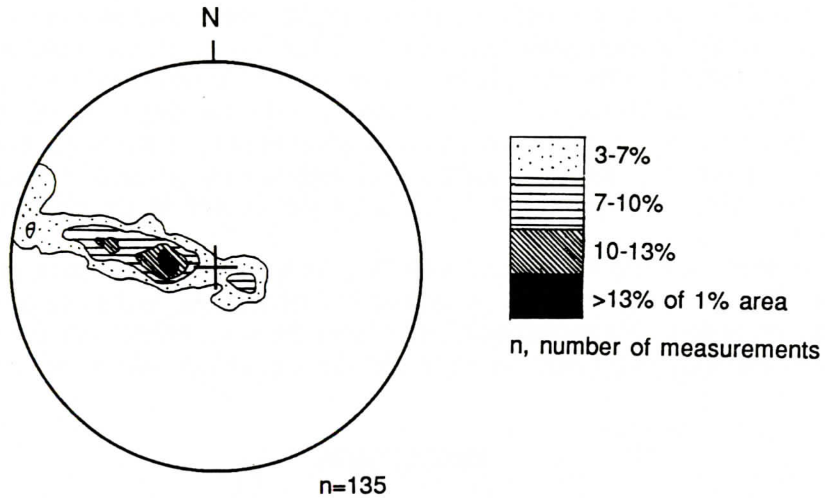


Figure 9. Equal area stereographic projection of poles to cleavage from Cambrian and Ordovician carbonate rocks of the study area and surrounding region of Washington County, Maryland, and Berkeley and Jefferson Counties, West Virginia. Contours are at 3, 7, 10, and 13 percent of 1 percent area.

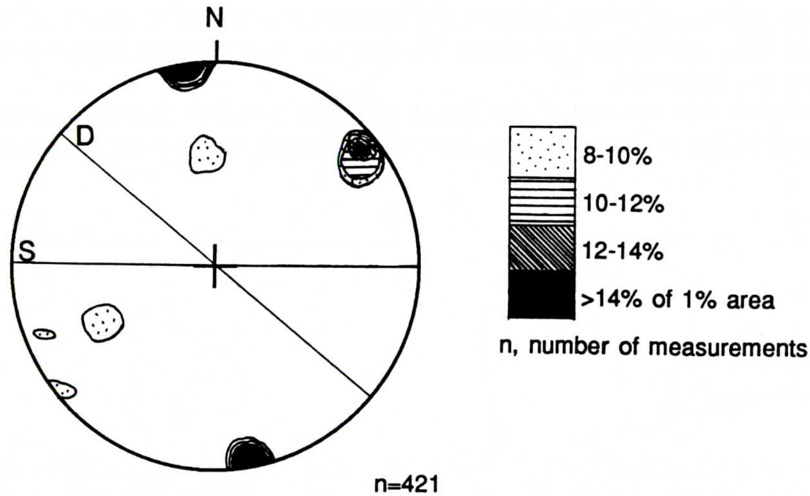


Figure 10. Equal area stereographic projection of poles to calcite-filled veins and joints from the Funkstown cross fault area. Contours are at 8, 10, 12, and 14 percent of 1 percent area. D, great circle of dip-slip fault plane, estimated from figures 13 and 14. S, great circle of right-lateral strike-slip fault plane, estimated from figures 13 and 14.

mately N60° W. The two directions of maximum principal stress indicated by these conjugate fractures are approximately N80° W and N60° W. The right-lateral component of shear in set 1 is more abundant than the left-lateral component. Set 1 has a compressive stress direction consistent with trends of Alleghanian folds and cleavage (Figure 9); compression directed N80° W is perpendicular to the strike of fold axial surfaces and cleavage planes indicating that set 1 conjugate shears may be related to folding. Set 2 has a compressive stress oriented N60° W and is probably not related to the major folds in this area.

Some of the tension gash arrays in the Great Valley of Maryland and West Virginia are related to mesoscopic faults that are within or are subparallel to the shear. These mesoscopic faults, that have right-lateral displacement, cut left-lateral shears of set 1. Since these mesoscopic faults have the same orientation and sense of offset as the set 2 right-lateral shears, they probably formed at the same time. Set 2 overprints set 1 indicating the change in the principal-stress direction.

The nearly vertical N80° E and N40° W joint sets of the Funkstown cross fault (Figure 10) were compared to the orientation of mesoscopic faults and other cross faults in the region. No mesoscopic faults have the same orientation as the major joint sets, and only two faults are within 10 degrees of the N40° W maxima joint set.

DISCUSSION

Hancock (1985) and Sylvester (1988) described structures related to strike-slip fault zones (Figure 11) including Reidel and "Reidel-within-Reidel" structures. Several conjugate shears can be related to one compressional deformation (Hancock, 1985). Although these sets differ in their dihedral angle, the orientation of the bisector of the acute angle is the same for all sets. In the two shear sets in the Great Valley, however, because the orientations of the bisectors of the acute angles of each set differs by about 20 degrees, two distinct compressive stress directions are indicated.

Different orientations of maximum principal stress in Paleozoic rocks of the Valley and Ridge of Pennsylvania have been recognized by Nickelsen (1979) and

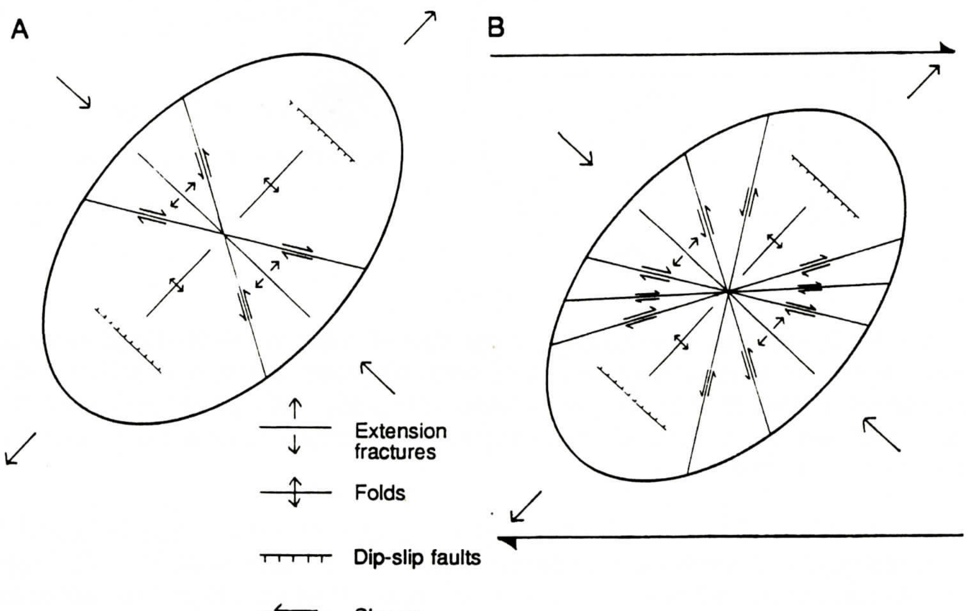


Figure 11. Strain ellipse diagram of, A, structures characteristic of pure shear, and, B, structures characteristic of simple shear (modified from Hancock, 1985, and Sylvester, 1988).

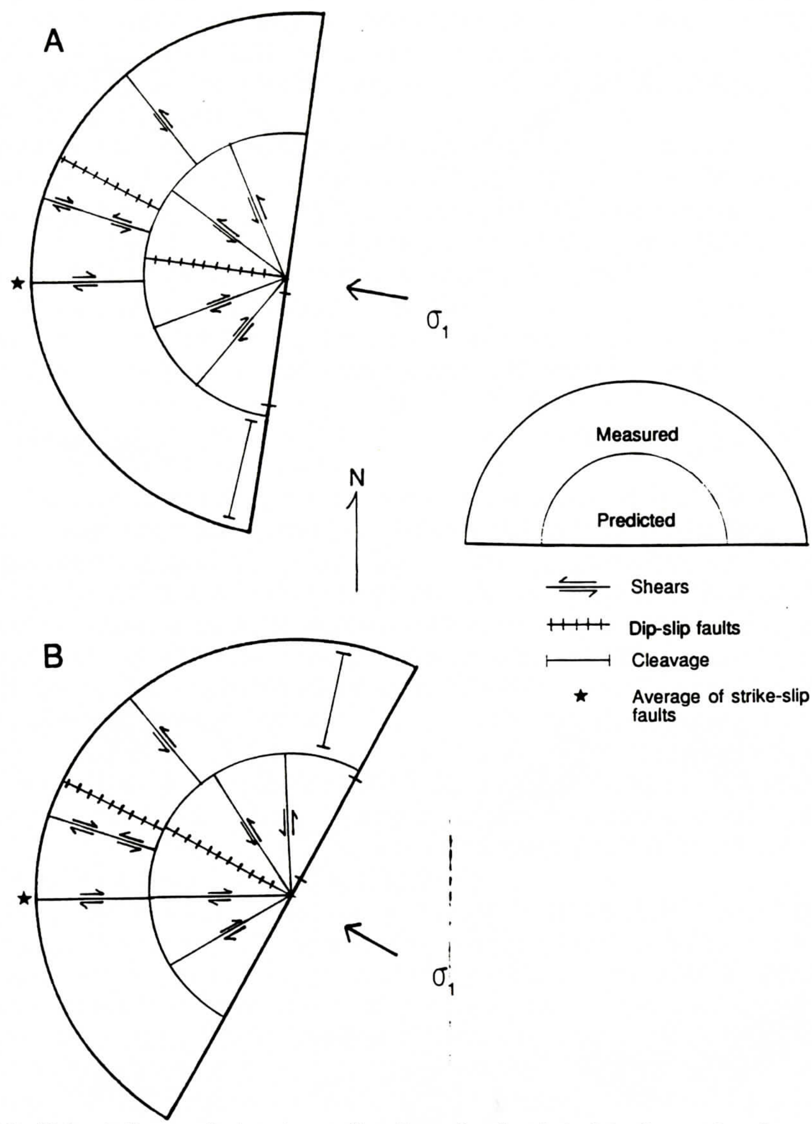


Figure 12. Orientations of structures (in plan view) related to the early, A, and later, B, maximum- principal-stress orientations in the Great Valley. Inner area, predicted orientation of structures from figure 11; outer area, observed orientation of structures (observed orientations are the same in A and B).

Geiser and Engelder (1983). Nickelsen (1979) concluded that these orientations were caused by a dextral rotation of strain axes during deformation. Rotation around a vertical axis resulted in overprinting of variously oriented extension joints and conjugate wrench faults (Nickelsen, 1979).

In the case of the opposing sense of offset of shears and mesoscopic faults in the Great Valley, the zone of tension gashes became the location of brittle failure. The mesoscopic faults postdate formation of the shear zones. Evidence for the sequence of deformation in this study consists of (1) the opposing direction of offset for structures having the same orientation (right-lateral faults and left-lateral shear zones) and (2) the brittle mesoscopic faults that cut the tension gashes in

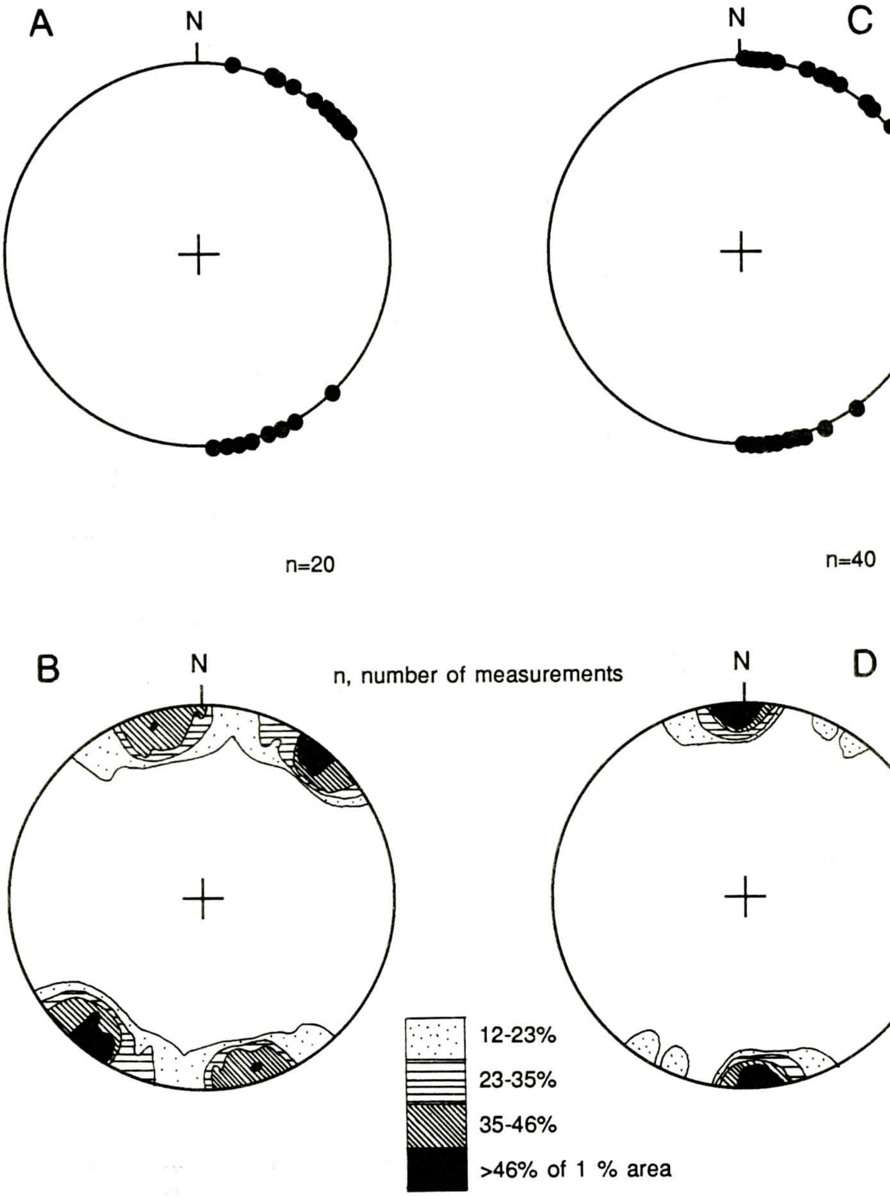


Figure 13. A, Equal area stereographic plot of poles to assumed vertical fault planes of cross-strike faults measured from Sando (1957). B, Equal area stereographic projection of poles to orientations of cross-strike faults measured from Sando (1957) in Washington County, Maryland assuming vertical fault plane. Contours are at 12, 23, 35, and 46 percent of 1 percent area. C, Equal area stereographic plot of poles to assumed vertical fault planes of cross-strike faults measured from Dean and others (1987). D, Equal area stereographic projection of poles to orientations of cross-strike faults measured from Dean and others (1987) in Berkeley and Jefferson Counties, West Virginia assuming vertical fault plane. Contours are at 12, 23, 35, and 46 percent of 1 percent area.

the shear zone. If the right-lateral mesoscopic faults are a component of the set 2 conjugate shear system, then set 2 is younger than set 1. This cross-cutting relationship suggests a clockwise change of the maximum-principal-stress direction during deformation.

Shearing and (or) strike-slip faulting, which exists in the Appalachians, can occur during compression. Cross-strike dip-slip faults are difficult to explain; however, they can form at an angle to the shearing in strike-slip fault zones, often perpendicular to the minimum-principal-stress direction (Hancock, 1985; Sylvester, 1988). This relationship may explain the combination of both strike-slip and dip-slip cross faults in the Great Valley.

Structures associated with wrench or strike-slip fault systems include strike-slip faults, conjugate shears, extension joints, and normal faults (Figure 11). In plan view, the orientations of theoretical structures associated with the two recognized maximum-principal-stress orientations (Figure 12) show close correlations to the observed structures.

Because the maximum-principal-stress orientations are N80° W and N60° W, conjugate shear orientations can be expected to trend between southwest and nearly due north depending on the dihedral angle. Therefore, strike-slip cross-fault orientations fall within this same interval. For the Rocky Marsh strike-slip cross fault, the fault orientation determined by outcrop structures suggests a trend of approximately N85° E; this trend is within 5 degrees of the expected orientation based on the second maximum-principal-stress orientation. Examination of trends of strike-slip cross faults mapped by Sando (1957) and Dean and others (1987) in the Great Valley indicates that the orientations of 9 (21 percent) cross faults are within 10 degrees of the predicted N70° E orientation (Figure 12), and 19 (44 percent) cross faults are within 10 degrees of the predicted due east orientation (Figure 12); 8 (20 percent) cross faults trend between N60° W and N50° W, and 7 (19 percent) have other orientations (Figure 13).

Cross-strike dip-slip faults can be related to strike-slip faults (Figure 11). For the Great Valley of Maryland and West Virginia, these extensional cross faults are expected to be oriented at approximately N80° W and N60° W (Figure 12), which correspond to the two maximum-principal-stress directions. Of 23 dip-slip cross faults mapped by Sando (1957) and Dean and others (1987), 10 (44 percent) cross faults are within 10 degrees of the predicted N60° W trend for the second maximum- principal-stress direction, and 2 (9 percent) cross faults are within 10 degrees of the N80° W orientation for the first maximum-principal-stress direction. Of the remaining 11 (35 percent) extensional cross faults, 8 trend between N70° E and N85° E.

A comparison of the major orientations of strike-slip and dip-slip cross faults recognized in the Great Valley to the predicted orientations (Figure 12) shows that both strike-slip and dip-slip cross-fault orientations are related to the second phase of deformation in this area. Right-lateral strike-slip cross faults are oriented approximately east-west, left-lateral strike-slip cross faults are oriented north-northwest, and dip-slip cross faults are oriented northwest (Figure 14).

The cross-cutting relationships of mesoscopic and macroscopic structures reveal the relative timing of cross faulting within a sequence of Alleghanian deformation in the Great Valley. This sequence is a transition from ductile-brittle and brittle (development of cleavage, folds, set 1 tension gashes, and thrust faults) to brittle deformation (development of set 2 tension gashes and cross faults) (Figure 15). The two sets of shear-generated tension gashes document a change

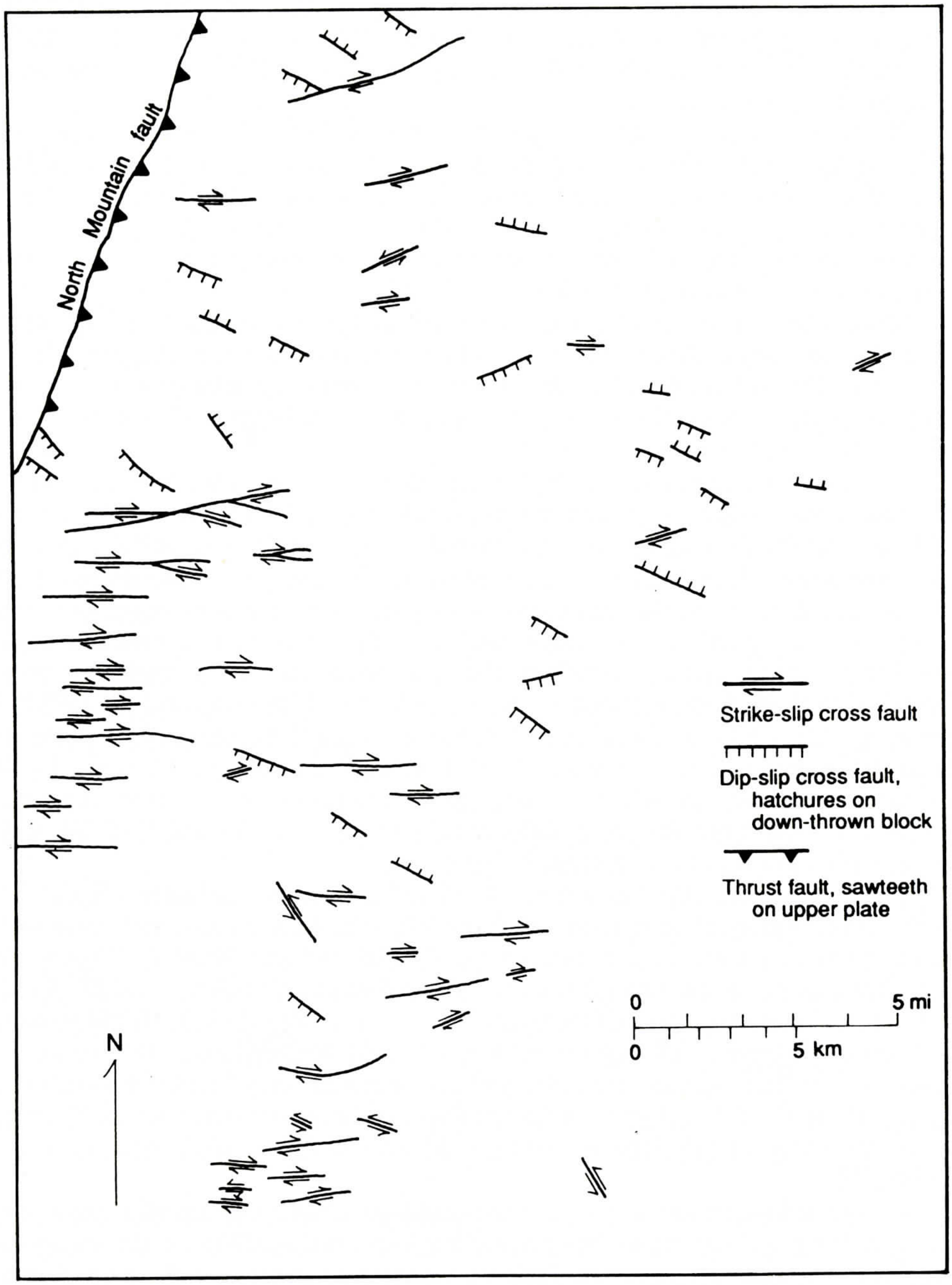


Figure 14. Interpreted sense of movement on cross faults based on field observations and predicted orientations in the Great Valley of Maryland and West Virginia. See figure 1 for location of cross faults.

in the maximum-principal-stress direction. Cross faulting is the latest phase of brittle deformation. Strike-slip and extensional faults probably developed coevally, since both shearing and extension can occur in a compressive regime (Hancock, 1985; Sylvester, 1988), and cross-fault orientations are consistent with the second maximum-principal-stress orientation.

Late Brittle	Cross-faulting strike-slip and dip-slip Set 2 shears
Early Ductile-brittle and Brittle	Thrust faulting Set 1 shears Folding Development of cleavage

Figure 15. Sequence of Alleghanian deformation for the Great Valley of Maryland and West Virginia.

REGIONAL IMPLICATIONS

The change in structural trends in the Great Valley, the presence of cross-strike faults, and the interpreted change of the compressive stress direction may be related to the formation of the Pennsylvania salient. The Pennsylvania salient has been interpreted to be the product of an inherited Precambrian continental margin (Rankin, 1976; Thomas, 1977), oroclinal bending from a Paleozoic orogeny (Kent and Opdyke, 1978; Kent, 1988; Lefort and Van der Voo, 1981; Lefort, 1989), and shearing related to early Mesozoic rifting (Milici and Bayer, 1988).

If the Pennsylvania salient formed during a Paleozoic orogeny, horizontal bending about a vertical axis of the orogen may have caused shearing and extension that resulted in the formation of cross faults. However, the cratonward bending of the orogen suggests a counterclockwise rotation of the maximum-principal-stress direction to the south of the salient; this rotation is opposite to that recognized in the structures observed in the Great Valley of Maryland and West Virginia.

Another orocline in the central Appalachians occurs south of the study area, best recognized by the change in trend of the Blue Ridge near Bluemont, Virginia, and in the Great Valley to the west (Figure 16). The Great Valley region of Maryland and West Virginia is located, therefore, between two oroclines. It is possible that the cross faults and two principal-stress orientations may be a local response to the bending associated with the two oroclines.

CONCLUSIONS

Investigation of cross-strike faults and associated mesoscopic structures in the Great Valley of Maryland and West Virginia suggests that compressive deformation during the Alleghanian orogeny consisted of two separate events and that the maximum-principal-stress orientations underwent clockwise rotation. Cross-cutting conjugate shears suggest that the N10° E structural trend in the Great Valley was produced from an early Alleghanian tectonic event. Strike-slip and dip-slip cross faults that can be predicted in this structural model are interpreted to

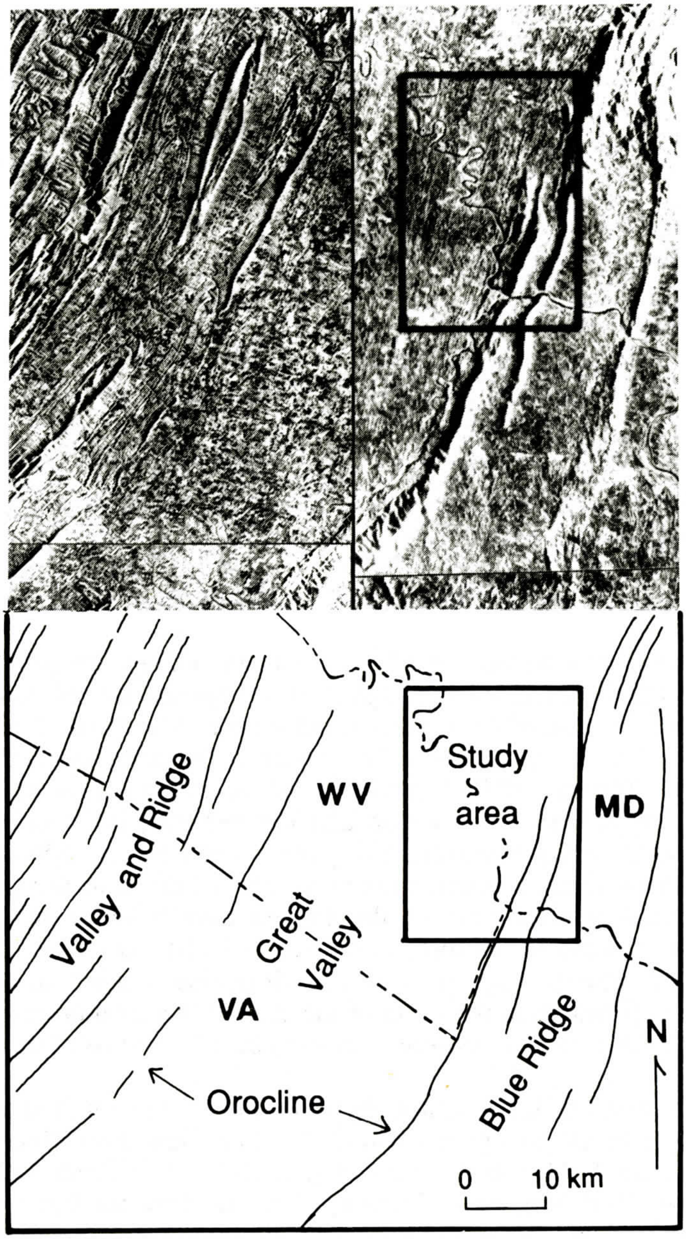


Figure 16. Side-Looking Airborne Radar (SLAR) image mosaic (Southworth, unpublished data) showing the study area and oroclines.

be the result of a later deformation event.

The large number of cross faults and the two distinct principal-stress orientations recognized in the Great Valley of Maryland and West Virginia are probably local structural features related to oroclines to the north and south of this area.

Future geologic mapping in the central Appalachians may demonstrate that cross-strike faults are common in other areas; the lack of exposures and short traces of the cross faults makes study difficult. By investigating the kinematic indicators and interpreted maximum-principal-stress directions, it is possible to predict the type of faulting, strike-slip versus dip-slip cross faulting, as well as fault orientations.

ACKNOWLEDGMENTS

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SHALLOW COMMON-OFFSET SEISMIC PROFILING OF A SUB-PENNSYLVANIAN PALEOVALLEY, WESTERN KENTUCKY

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ABSTRACT

Shallow common-offset seismic reflection data were used to define features associated with a sub-Pennsylvanian paleovalley (Pennsylvanian-Mississippian unconformity) in western Kentucky. The presence of a shallow water table in the study area permitted good energy transmission and allowed the recording of high-frequency reflections from the unconformity surface using a sledge hammer source. The erosional surface of the paleovalley is characterized by discontinuous angular reflections and diffractions. The data have been tentatively interpreted to show a small secondary channel feature located near the projected valley boundary and a series of Upper Mississippian slump blocks. Although data quality was disappointing, the use of common-offset seismic reflection methods could provide a practical way of investigating shallow targets in the Illinois Basin and areas of similar geologic setting.

INTRODUCTION

Throughout the Illinois Basin, the Pennsylvanian-Mississippian unconformity is characterized by numerous northeast-southwest-trending paleovalleys (Figure 1). The economic significance of shallow oil production from valley-fill sequences within the basin has been recognized for many years (Shiarella, 1933; Goudarzi and Smith, 1968; Bristol and Howard, 1980; Greb, 1985; Howard and Whitaker, 1988). Since 1927, a portion of the Madisonville Paleovalley (the study area) in Ohio County, Kentucky, has produced oil from basal Pennsylvanian deposits less than 200 m deep. The objective of this study was to determine the feasibility of using shallow high-resolution seismic reflection methods to identify the unconformable valley boundary and, if possible, define the geometry of the valley fill.

Seismic exploration was first attempted in western Kentucky in 1929 following the discovery of oil in basal Pennsylvanian sand-bar deposits within the Madisonville Paleovalley. Gulf Oil Company was unsuccessful in attempts to define the extent of channeling in the area using seismic methods (Stanonis, 1968). Recent high-resolution seismic experiments have succeeded in identifying features associated with sub-Pennsylvanian channeling (Harris and Street, 1989) in the Madisonville Paleovalley area using common-offset seismic reflection methods.

The "optimum offset" seismic reflection technique (Hunter and others, 1984; 1985) has been successfully used to map shallow bedrock valleys and valley-fill geometries (Pullan and Hunter, 1990). This study demonstrates use of the method in delineating characteristics of an erosional surface which lacks the high acoustic impedance contrast associated with most overburden-bedrock interfaces. Use of the optimum offset technique as described in this paper was primarily intended as a quick, inexpensive reconnaissance tool that could be used to aid small exploration companies and shallow-oil-field operators working in the Illinois Basin and other areas with similar geology.

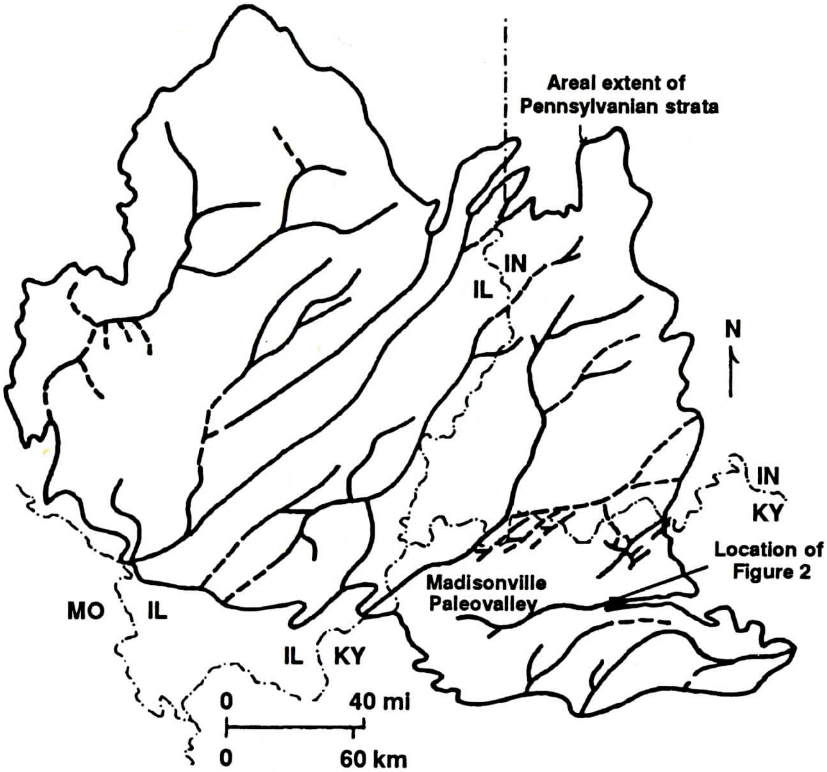

Figure 1. Sub-Pennsylvanian paleovalleys of the Illinois Basin (after Bristol and Howard, 1971; and Greb, 1989a).

GEOLOGIC SETTING

The study area, shown in Figure 2, is located in the southeastern portion of the Illinois (Eastern Interior) Basin. The predominant local structural feature is a section of the east-west-trending Rough Creek Fault Zone that bounds the study area to the south. The fault zone, which appears to have been active throughout the Paleozoic (Sutton, 1971; Smith and Palmer, 1981), is expressed as a series of high-angle normal faults displaying as much as 200 m of offset (Goudarzi and Smith, 1968). Tectonic activity along the fault zone during Early Pennsylvanian time may have influenced the location and orientation of the Madisonville Paleovalley (Goudarzi and Smith, 1968; Greb, 1985, 1989a, 1989b). Examination of local well-log data (Shiarella, 1933; Greb, 1985) has produced a rough outline of the paleovalley surrounding the study area. The primary target for this study was the erosional surface associated with the valley boundary (Pennsylvanian-Mississippian unconformity). The stratigraphic units of interest included the interbedded carbonates and clastics of the Upper Mississippian Chesterian Series and the thick sandstones, conglomerates, and shales of the basal section of the Lower Pennsylvanian Caseyville Formation (Figure 3).

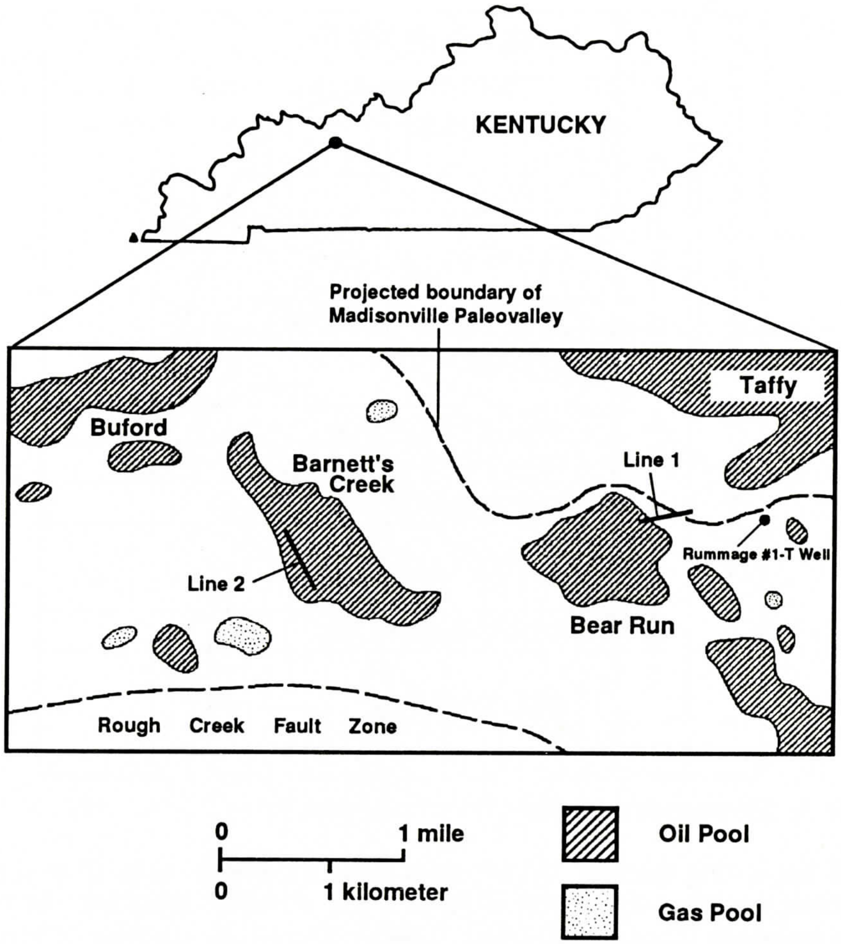

Figure 2. Map of the study area (Taffy-Barnett's Creek oil field and vicinity) in western Kentucky showing the locations of the common-offset seismic lines and the Rummage #1-T well.

FIELD METHODS AND DATA PROCESSING

The choice of a common-offset method to complete this study was dictated in part by the available field equipment (most of the data were collected with only six 100 Hz geophones), data- processing limitations, and a small field crew (two-man operation). Common-offset reflection investigations require minimal field equipment, data processing, and manpower, and, under favorable conditions, are usually fast and cost effective.

Two common-offset seismic reflection lines were shot in the Taffy-Barnett's Creek oil field of northwestern Ohio County, Kentucky (Figure 2). The locations of lines 1 and 2 were chosen because their positions were within the projected boundaries of the Madisonville Paleovalley, the sites were on hard-packed gravel roads with very little topographic variation, and well-log data were available in the area.

Several multichannel test records were shot along each line in order to view the target reflections from a range of source-receiver offsets. After examination

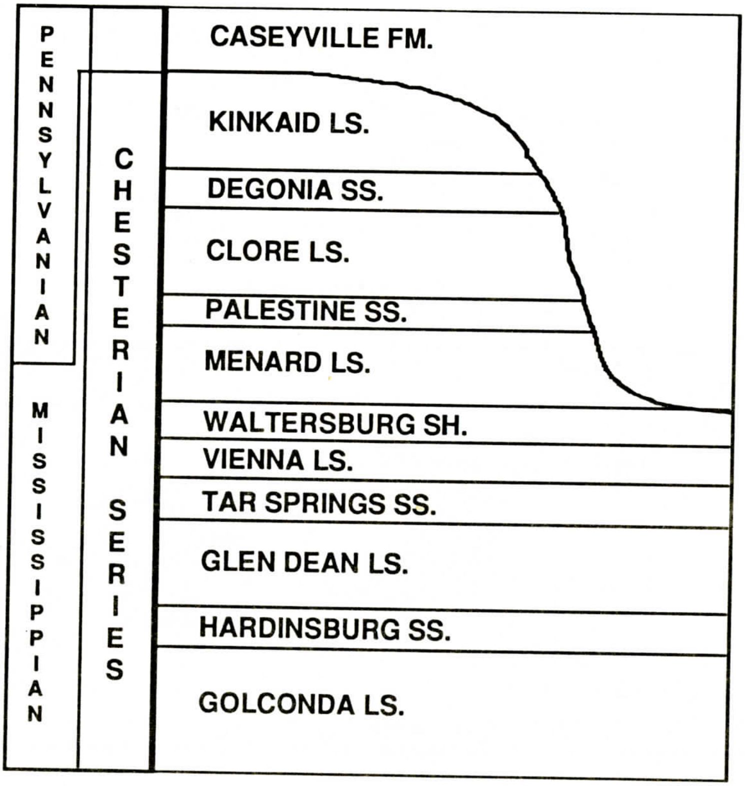

Figure 3. Stratigraphic column of the study area.

of the "expanding spreads", the optimum offset was chosen to be 61 m (Figure 4). The common-offset lines were then shot and recorded following the optimum offset approach (Hunter and others, 1984). The 61 m optimum offset allowed events arriving earlier than 180 ms to be recorded without contamination by the air wave.

Because the optimum offset technique requires the reflection data to be collected one channel at a time, every recorded trace must have its own shotpoint. To make the survey efficient, the energy source must be easy to transport and set up. The source must also be durable because repeated shots are often performed at each shotpoint. Although the total energy output is substantially lower, a 7.3 kg sledge hammer vertically striking an aluminum plate was chosen as the energy source for this study instead of the other available source, a trailer-mounted, weight-drop device, because of its portability, low maintenance, and higher frequency output. For the collection of the reflection data on lines 1 (290 m long) and 2 (365 m long), slightly under 1000 hammer blows were recorded at 216 shot points.

The receivers were single 100 Hz geophones firmly attached to the ground with 7.5 cm spikes at a spacing of 3 m. On line 1, the geophones were planted at the edge of the road surface, while on line 2 a drainage ditch was present and the geophones were planted in the ditch adjacent to the road. The high-frequency air wave appeared much weaker on line 2, probably because the geophones were better coupled and the signal-to-air wave ratio was higher.

The recording system was an EG&G 2415F 24-channel signal enhancement

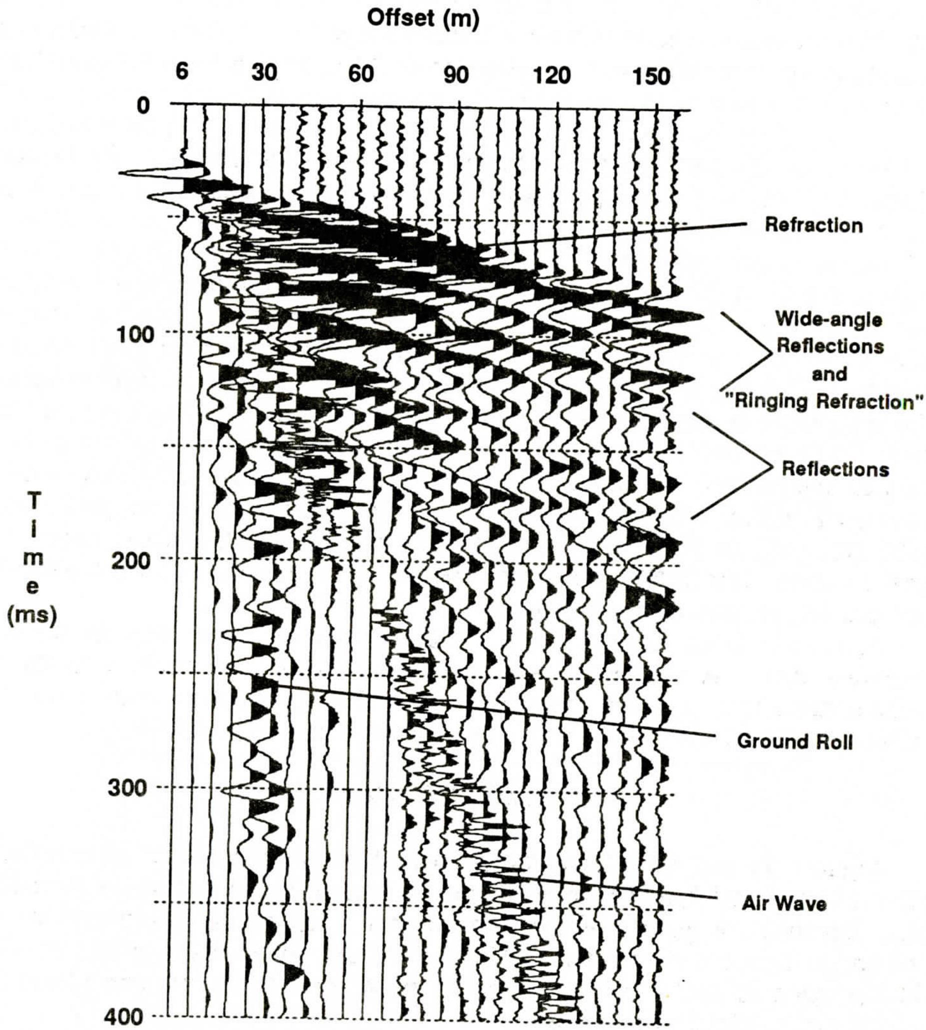

Figure 4. Expanding spread from the study area with various events identified. The source to first receiver offset is 6.1 m and the receiver spacing is 6.1 m. To avoid ground roll and air wave interference, a 61 m source-receiver offset was chosen for recording the common-offset data.

seismograph with a DMT-911 tape deck. Analog-to-digital conversion of the fixed-gain data resulted in an 8-bit digital word that was recorded on tape in SEG-D format. The analog low-cut filters were set at 100 Hz with a slope of 12 decibels/octave and the sampling interval was 1/4 ms.

To begin processing, the data were dumped from the tape deck to an IBM 386-based microcomputer and converted to SEG-Y format for compatibility with the processing software, VISTA 6.0 (Seismic Image Software Ltd., 1989). The 24-channel field files were assembled to form complete sections, and the common-offset field data were plotted for initial analysis.

First arrival data from expanding spreads were examined by X-T analysis (i.e., Dobrin and Savit, 1988, section 11-2) to develop the near-surface velocity structure used in the application of static corrections. Expanding spreads from both ends of each line as well as reversed spreads in the center of each line were used. The unsaturated surface layer velocities range from 270 m/s to 480 m/s and the depth to the water table was calculated to be 2.5 m on line 1 and 1.2 m on line 2. The depth to the base of the weathered layer varies by less than 1 m along each line (12 m deep on line 1 and 9 m deep on line 2) and is consistent with the 9 m depth to the Pennsylvanian seen in the Rummage #1-T well (see Figure 2 for location). Velocities of the sub-weathered layer (Pennsylvanian sands and shales) ranged from 2590 m/s to 3600 m/s. The velocities are consistent with those observed for the Pennsylvanian section in the region (Gochioco and Cotten, 1989). The first arrival at the 61 m common offset is the refraction from the base of the weathered layer, which was used as the datum for static corrections. Although small velocity variations in the near-surface can cause problems in static correction of common-offset data, aligning of refractions from the base of the weathered layer is believed to be acceptable in this case because the base of the weathered layer appears to be relatively horizontal locally, the velocity of P-wave energy traveling in the saturated overburden is laterally consistent (approximately 1500 m/s), and the unsaturated surface material is thin. After static corrections were performed, a digital bandpass filter of 100 Hz to 300 Hz was utilized to further improve the signal-to-noise ratio (S/N). Amplitude spectra of the filtered reflection data show dominant frequencies of approximately 100 Hz for both lines.

Automatic Gain Control (AGC) scaling was the final step in the data-processing flow. A number of AGC window lengths were tried, with the best results coming from a 64 ms operator. By equalizing reflection amplitudes, AGC scaling improved the apparent coherency of the data.

INTERPRETATIONS

Figures 5a and 6a present the processed common-offset seismic reflection data for lines 1 and 2, and Figures 5b and 6b present interpretations of the seismic data. Because the quality of the data was not outstanding, interpretations have been made that are speculative. However, these interpretations are consistent with documented geological conditions in the study area and are considered to be realistic explanations of the data.

No geologic significance has been attached to the high-amplitude shallow events (< 75 ms on line 1 and < 50 ms on line 2). They are either wide-angle reflections from the shallow subsurface, a "ringing refraction" from trapped energy in the weathered layer, or a combination of the two. In any case, the events are not useable.

Line 1

Common-offset line 1 was taken near the proposed northern limit of the Madisonville Paleovalley, in the vicinity of the Bear Run oil pool (Figure 2). The event in the middle of the profile between 110 ms and 150 ms is interpreted to represent a section of the Pennsylvanian-Mississippian unconformity. The discontinuous reflection pattern, marked by A on Figure 5a, implies an irregular surface. The horizontal reflections between 125 ms and 170 ms (B on Figure 5a) disappear into a poor data area at trace 30, probably caused by steep dip on the Mississippian erosional surface. The incoherency of the reflection data is believed to be a result of spatial aliasing in the vicinity of the steep dip. On the right side of the

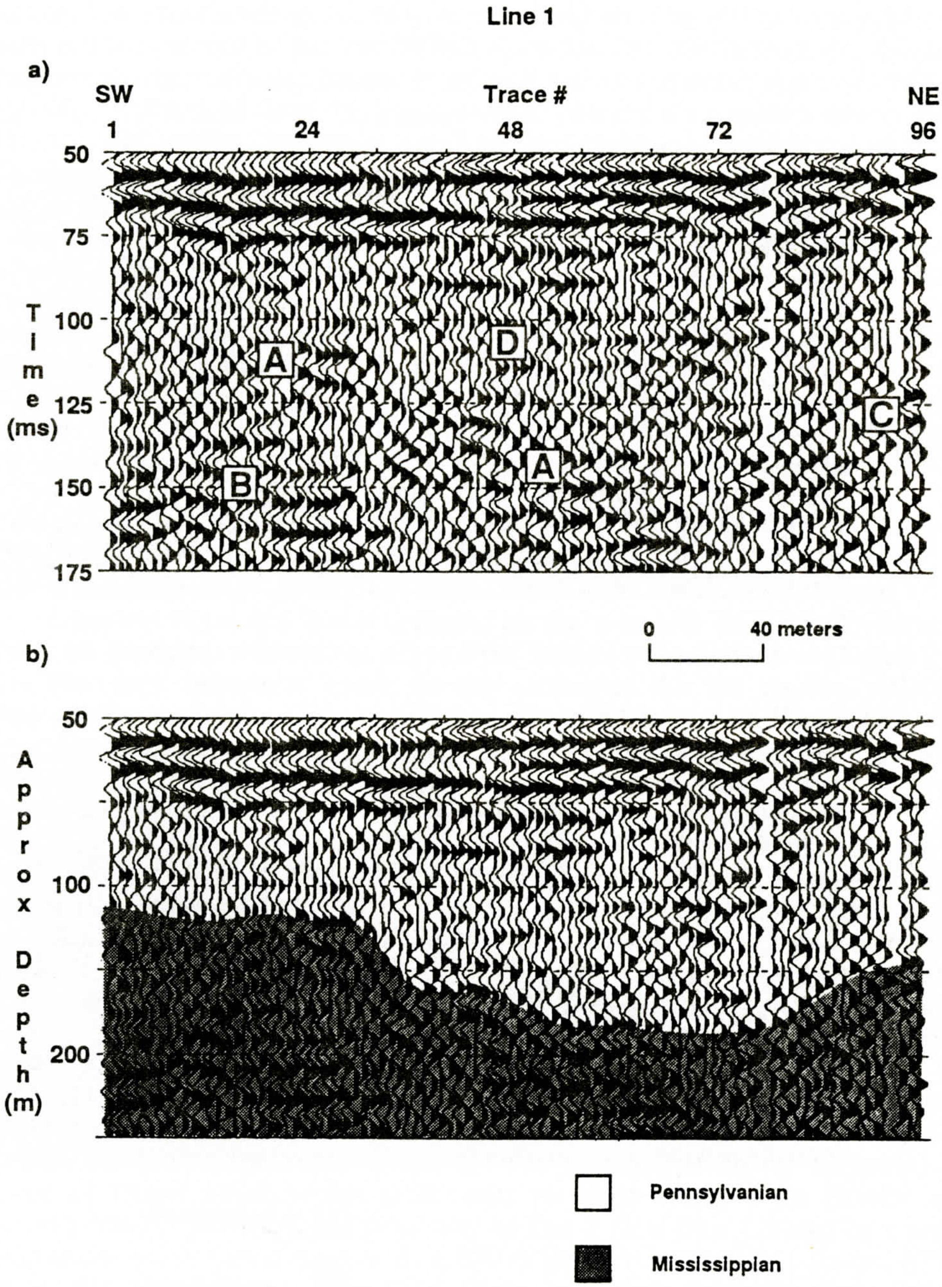

Figure 5. (a) Common-offset line 1 (processed data). The common-offset was 61 m and the geophone spacing was 3 m. The missing traces (78 and 93) are because of narrow bridges on the line. The letters marking various features on the record are referred to in the text. (b) Interpretation of line 1. A small channel feature (approximately 175 m wide) truncating the horizontal Mississippian section has been interpreted. Channel-fill reflections are visible, but weak. A nonlinear depth scale, developed through velocity analysis, has been attached to the interpretation.

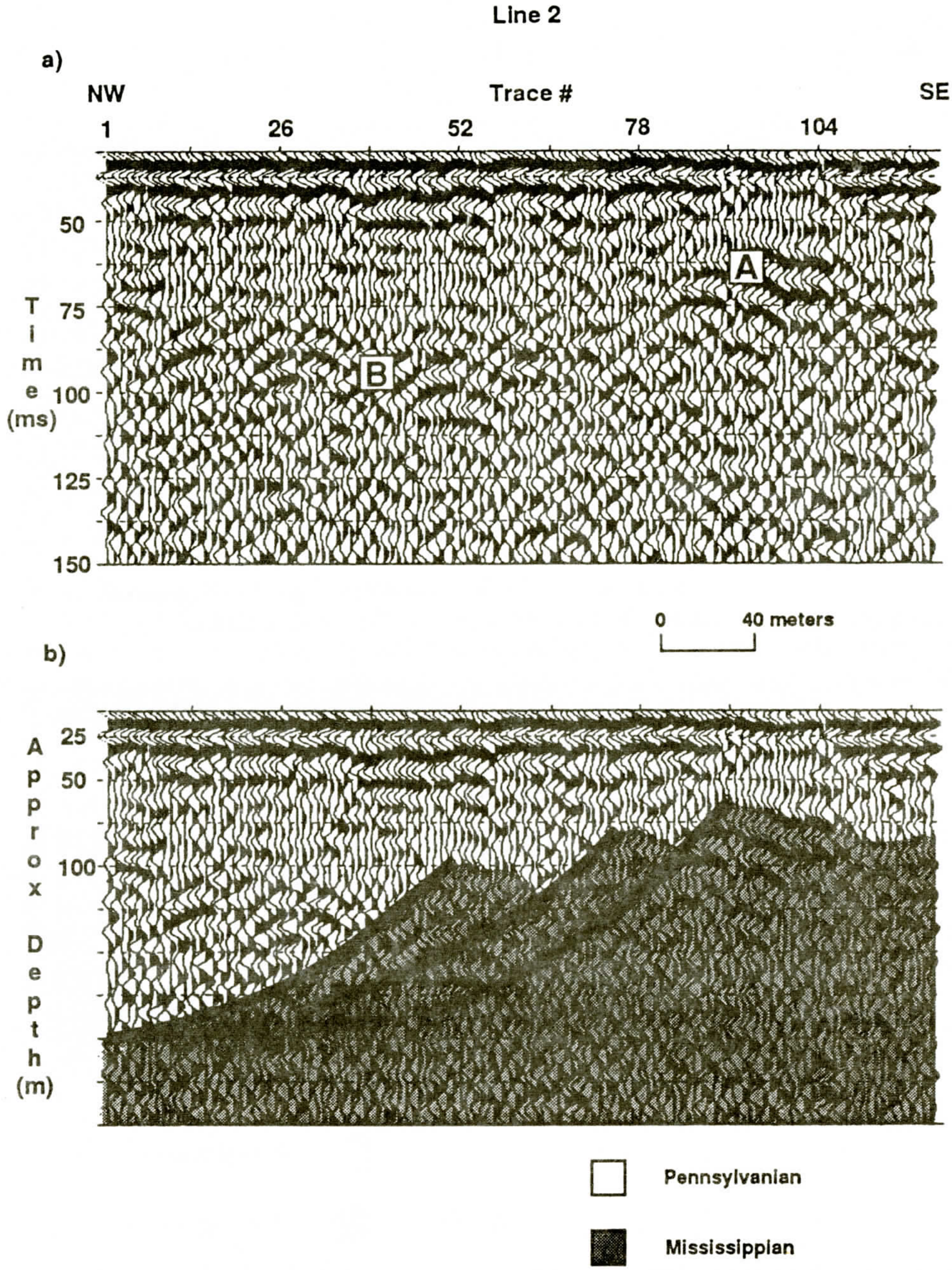

Figure 6. (a) Common-offset line 2 (processed data). The common-offset was 61 m and the geophone spacing was 3 m. The letters marking various features on the record are referred to in the text. (b) Interpretation of line 2. From the chaotic reflection pattern and the presence of diffractions, a series of Upper Mississippian slump blocks has been interpreted. The depth scale is incomplete below 100 m due to the complex nature of the deeper reflections.

section, the unconformity reflection is interfered with by diffractions originating from beyond the end of the line (C on Figure 5a). It appears from the reflection data that the unconformity surface is forming a narrow channel approximately 175 m wide. Reflections from the Pennsylvanian channel-fill sequence (D on Figure 5a) are discernible, but they are low amplitude and variably continuous. Considering its size, seismic characteristics, and proximity to the projected valley boundary, the channel feature is interpreted as a tributary.

An approximate depth scale, derived through X^2-T^2 velocity analysis of expanding spreads (i.e., Dobrin and Savit, 1988, section 7-6), has been attached to the interpreted version of line 1 (Figure 5b). The depth scale is nonlinear: a function of non-zero offset and increasing velocity with depth. Using velocity data from the reflection analysis and a lithology log from the Rummage #1-T well, the depth of the channel was estimated to be approximately 50 m, and the deepest reflection (about 170 ms) is believed to be from the Golconda Formation (Upper Mississippian), a 12 to 15 m thick, crystalline limestone occurring at a depth of 220 m.

Line 2

Common-offset line 2 was collected on the west side of the Taffy-Barnett's Creek oil field near the western edge of the Barnett's Creek oil pool (Figure 2).

The most prominent event on the section is the SE dipping reflection sequence between traces 84 and 105 from 60 to 80 ms (A on Figure 6a). The reflections terminate abruptly at trace 84, suggesting the presence of faulting. From trace 1 through trace 70, the zone between 75 ms and 110 ms shows a group of discontinuous reflectors (B on Figure 6a). Because of the angularity of the group and the presence of diffractions, faulting is again suspected. Due to the complex nature of the faulting, velocity estimates below 75 ms were impossible, resulting in an incomplete depth scale. In view of the repetitive nature of the faulting and the high-amplitude of the reflections, the entire sequence is tentatively interpreted as a series of Mississippian slump blocks (Figure 6b). The presence of Upper Mississippian slump blocks associated with sub- Pennsylvanian channeling has been documented throughout the Illinois Basin (Swann, 1945; Bristol and Howard, 1971, 1980), and the interpretation of a local well log (Rummage #1-T) has suggested the existence of slumping within the study area. Figure 7 presents a reconstruction of Mississippian slumping that has been used as a model for the interpretation of line 2. The proposed slumping mechanism is failure of Upper Mississippian shale units on steep valley walls (Bristol and Howard, 1980). However, the proximity of line 2 (less than 1.5 km) to a major down-to-the-north normal fault with a 200 m throw (Goudarzi and Smith, 1968), the activity of the Rough Creek Fault Zone throughout Paleozoic time (Sutton, 1971; Smith and Palmer, 1981), and the suggested structural control on paleovalley formation (Greb, 1989a, 1989b) indicates possible fault-related influence on nearby slumping.

DISCUSSION

Improving Data Quality

Because the quality of the seismic data obtained in this study was less than expected, confidence in the interpretations obviously suffered. Key factors for high-quality shallow seismic data include adequate S/N in order to distinguish target reflections, and adequate resolution, both lateral and vertical, to identify features associated with the target.

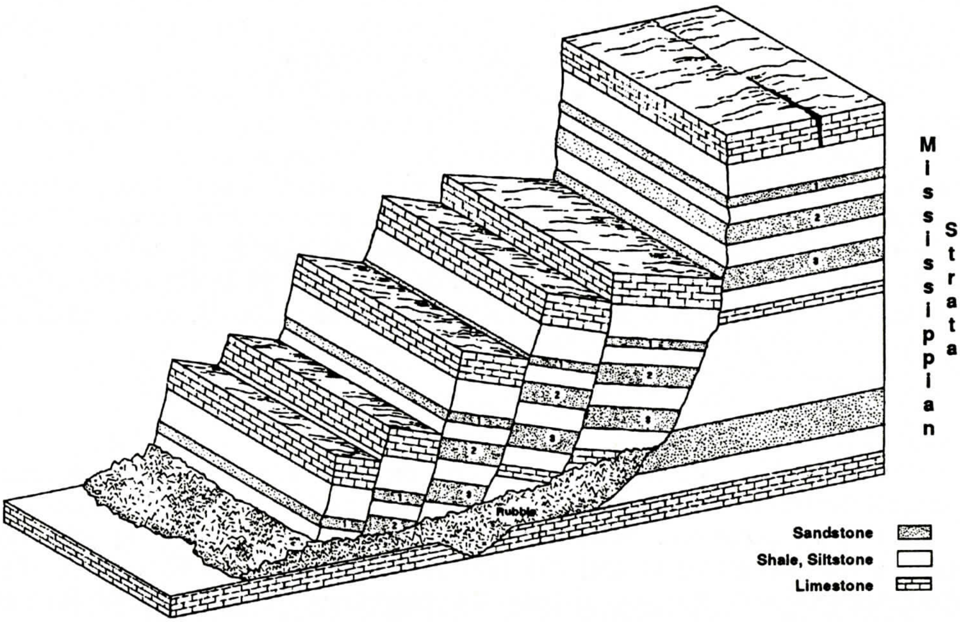

Figure 7. Reconstruction of Mississippian slumping within the Illinois Basin that has been used as a model for the interpretation of common-offset line 2 (after Bristol and Howard, 1980).

An increase in S/N should be expected with the use of common-depth-point (CDP) methods (Knapp and Steeples, 1986) for recording high-resolution seismic reflection data. Although collection and processing is more time consuming and costly, CDP data would probably have been more useful in identifying and characterizing features associated with the geological conditions found in this study. In comparing CDP and optimum offset seismic surveys, Pullan and others (1991) suggested that optimum offset methods are only applicable under ideal conditions and that the improvement of S/N through CDP processing can enhance reflections that are not visible on optimum offset profiles.

The presence of an irregular unconformity surface, faulting, and dipping beds combined with limited, mostly unreversed, receiver spreads, restricted the accuracy of velocity analyses in this study. Because of the increased availability of multichannel seismic records, velocity analyses should also improve through the use of CDP methods.

As mentioned earlier, spatial aliasing resulting from steep dips on the unconformity surface has produced areas of poor data. Lateral resolution should improve by using denser shotpoint and geophone spacings, which would lead to more accurate imaging of the complex structure found in the study area.

The use of a higher energy seismic source, such as the in- hole shotgun described by Pullan and MacAulay (1987), would also have improved the chances of recording higher quality data. The shotgun energy source has a higher frequency output than the sledge hammer (Miller and others, 1986), and undoubtedly would have improved vertical resolution of thin beds within the Pennsylvanian and Mississippian sections.

Neither the adequate field equipment for CDP data collection nor a shotgun energy source were available for this study, but they should be strongly considered for subsequent investigations in similar geologic settings.

Further Use of Shallow Seismic Reflection Methods in the Area

One of the most common applications of shallow seismic reflection techniques is in the delineation of aquifers for groundwater exploration (Geissler, 1989) and monitoring (Miller and others, 1990). Davis and others (1974) reported that several valley sands within the Caseyville formation are major aquifers in the Illinois Basin. Following careful selection of field equipment and parameters, seismic reflection data should be useful in defining groundwater exploration targets and assisting in the placement of recovery and monitoring wells in the region. With increased seismic resolution, thin Pennsylvanian coal beds and paleovalley tar-sand deposits could also be investigated using shallow reflection methods.

CONCLUSIONS

Shallow common-offset seismic reflection data collected over a sub-Pennsylvanian paleovalley in western Kentucky had adequate frequency and amplitude characteristics to allow distinguishable reflections from the valley floor (Pennsylvanian-Mississippian unconformity).

The seismic data suggest a highly irregular unconformity surface characterized by angular reflections and diffractions. On line 1, a narrow channel feature located near the projected valley boundary has been interpreted as a tributary channel. Line 2 has been tentatively interpreted to show a series of Upper Mississippian slump blocks located near the southern boundary of the valley, in close proximity to the Rough Creek Fault Zone. It is suggested that slumping in this area could be fault controlled.

Common-offset seismic data appear to be useful in interpreting features associated with the Pennsylvanian-Mississippian unconformity in western Kentucky. Although data quality is site dependent, the common-offset method should be considered a quick, inexpensive way for small exploration companies to obtain shallow seismic reflection data in the Illinois Basin and other areas with similar geological characteristics.

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