DOE/METC-83-8

EXTENSION STRUCTURES

IN THE CENTRAL APPALACHIANS

by Philip S. Berger
West Virginia University
Department of Geology & Geography

United States Department of Energy
Contract number EY-76-C-05-5194

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ABSTRACT

Late tectonic extension faults and extension fractures may form abundantly in steep limbs of detached anticlines. The Devonian Brallier formation of the central Appalachians is the unit that has undergone the greatest amount of extension faulting and fracturing, among Upper Ordovican to Lower Mississippian units exposed and studied on the northwest limb of the Wills Mountain anticline. Gravity data and analogy with exposures on the northwest limb of the Wills Mountain anticline indicate that the Blackwater and Glady anticlines may also have thinned and fractured Brallier on their northwest limbs in the subsurface. Given the underlying dark shale source and updip thickened shales as a seal, these volumes may contain gas in a fractured reservoir.

INTRODUCTION

Fracture is defined by Dennis (1967) as a surface along which loss of cohesion has taken place. Therefore, unless the term is restricted by use of a modifier or used in conjunction with the term fault, fracture as used herein, will include faults, joints, and fissures. When used in conjunction with the term fault, fracture is defined as a surface along which little or no slip has occurred (joint or fissure). When used with a prefix, such as extension fracture, then the term is restricted to those fracture surfaces formed by bed-parallel extension, as defined and characterized in the main text of this thesis.

The purpose of the thesis is to investigate a particular class of bed-extending brittle structures known as extension faults and extension fractures. These formed late during the growth of anticlines in the detached tectonics characteristic of the central Appalachians. A specific goal of this investigation is to develop a predictive tool for the subsurface locations of unusually fractured rock in the Middle and Upper Devonian clastic sequence.

The field area investigated in this thesis was primarily the northwest limb of the Wills Mountain anticline, which extends through Pennsylvania, Maryland, West Virginia and Virginia. Other map-scale anticlines near the Wills Mountain anticline were examined, as well as localities in Pennsylvania that expose outcrop-scale kink folds or kink-bands and which were suggested by Dr. R. T. Faill (written communication, 1978). Conclusions and speculations apply to these areas and to larger parts of the central Appalachians, chiefly in West Virginia.

The thesis is composed primarily of four short manuscripts, each having separate title, authors and abstract, and each building on or extending the last. The references and acknowledgements are a combined list from all four manuscripts. There is a separate section containing conclusions based on the work reported in the four manuscripts. Each manuscript has been or will be submitted for publication.

Each manuscript addresses the problem of predicting where the greatest amount of bed extension occurs. The first manuscript introduces the concept of fold-related brittle extension structures. It presents a relative time sequence for typical central Appalachian faulting and associated fracturing, and describes general field criteria for differentiating between earlier faults and the later extension faults. This manuscript refines a three stage model originally conceived by Dr. William J. Perry, Jr., of the U. S. Geological Survey, Reston, Virginia. Field work was done by all three authors, but most of the manuscript is based on a term paper written by the senior author and his field work at Pinto, Maryland.

The second manuscript develops an approximate mathematical model using measured values of horizontal shortening to estimate the minimum bed dip at which extension structures will occur. This information places geographical constraints on areas which may be investigated for gas-bearing porosity or permeability produced as extension fractures or extension faults. The equations in this manuscript were developed by Dr. Russell. L. Wheeler.

The third manuscript concentrates on extension faults, as described and defined in the first manuscript, and represents the result of field investigations in the area delineated in the second manuscript. The third manuscript describes the stratigraphic distribution of extension faults in and near the Devonian clastic sequence and especially relates abundance of extension faults to structural position. It is also concerned with the relationship between the extension faults and large fold-related splay faults rising from deeper detachments. The manuscript estimates the contribution of extension faulting to anticlinal growth and estimates relative amounts of extension in fold limbs so that limbs may be selected which are most likely to contain gas in fractured reservoirs. The equations in the manuscript were derived by Dr. Russell L. Wheeler.

The fourth manuscript uses a kink-band model of fold development to estimate the amount of extension in the kink-band or in the more rotated limb of a kink fold. The extension can occur as porosity and permeability producing fractures or faults as described in the first and third manuscripts. The kink-band model of fold development is probably the best explanation for the observed geometric form of major folds in the central Appalachians. Field criteria are given for differentiating between folds of kink-band origin and those of buckling origin. The map distribution of gravity lows is used to explain the distribution of extension faulting from manuscript three in a kink fold model. Drilling areas are suggested for fractured reservoirs.

Three-Stage Model of Brittle Deformation in the Central Appalachians

Philip S. Berger[1], William J. Perry, Jr.[2], and Russell L. Wheeler[1]

ABSTRACT

Recent fieldwork in the central Appalachians shows the applicability of a three-stage model of brittle deformation. Distinct minor structures form at specific times during overall horizontal shortening and map-scale folding. Stage I contraction faults shorten horizontal or gently dipping beds and are closely associated with wedging. Stage II uplimb thrust faults shorten folded beds hingeward. Stage III extension faults and extension fractures lengthen steeply dipping to overturned beds. Opening or shearing directions of fractures and crosscutting relationships of faults, other fractures, and stylolites allow distinction of the three stages. Stage III may have produced fracture porosity and permeability on steep limbs of anticlines. We propose a three- stage model of outcrop-scale fault and other fracture development for steeply dipping to overturned beds in

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eastern West Virginia and adjacent areas. Less rotated beds may show at least two of the stages. Kinematic analysis of faults and other fractures allows differentiation of the three stages.

Structural Style

The central Appalachian foreland has many detached anticlines but few outcropping major thrust faults. The anticlines are interpreted as active features generated mainly by duplication of strata by ramping of underlying thrust faults, by splay faults, and by ductile flow of shale-rich intervals into anticlinal crests (Perry and de Witt, 1977, Perry, 1978, Wheeler, 1975). The synclines are regarded as passive features resulting from anticlinal growth in adjacent rocks, rather than from active downbuckling (Gwinn, 1964).

We deal with the sequence of outcrop-scale structures that formed at specific times during overall southeast-northwest horizontal shortening and vertical extension. We assume that each individual structure formed in an orientation that allowed it to accommodate some of that horizontal shortening, or vertical extension, or both.

The first-order anticlines (Nickelsen, 1963, p. 16) of the western Valley and Ridge and the Allegheny Plateau provinces formed predominantly by flexural slip folding

(Faill, 1969), rather than by passive or flexural flow folding (Gair, 1950). Interlayering of shales and evaporites with sandstones, siltstones or limestones allows slip between units of low relative ductility. Stage I structures form before folding, or early during folding, when the angle between bedding and the maximum principal compressive stress is less than about 10 degrees. Experiments at room temperature and confining pressures to 2000 bars (equivalent to about 6 km in depth) suggest that slip parallel to bedding is possible when the angle between the maximum principal compressive stress and bedding ranges from 10 to 60 degrees (Price, 1967). Stage II structures form in this range of limb dips (see below). Stage III structures form in response to additional horizontal shortening, late in or after folding, when steep limb dips preclude further slip parallel to bedding. These structures form when the angle between bedding and the maximum principal compressive stress exceeds about 60 degrees. These conclusions are reinforced by the finite-element derived stress-history of folding (Dieterich and Carter, 1969).

Price (1967) described extension and contraction structures from the Canadian Rocky Mountains using the fault terminology of Norris (1958). We report the same types of structures from the central Appalachians, placing them in a relative time sequence and describing simple field criteria for differentiating among stages.

STAGE I STRUCTURES.—Cloos (1964) first recognized and named the prefolding wedges common in the central Appalachians. Wedges are the wedge-shaped ends of small fault blocks in which the bounding contraction faults (Norris, 1958) form angles of 30 degrees or less to bedding. Examples provided by Cloos (1964, figs. 2, 3, 4 and 6) involve brittle layers (sandstone and limestone) which have been "sheared, wedged, and telescoped together" in a more ductile medium (shale). This process of contraction faulting shortened the stratigraphic section in a northwest-southeast direction and thickened it perpendicular to bedding prior to or early during folding. Such contraction faults can form dipping in either direction (northwest or southeast).

Concerning the Canadian Rockies, Price (1967) writes that if layering was planar and inclined at a low angle to the maximum principal compressive stress, that stress' trajectories would tend to parallel layering, and subsequent failure would take the form of contraction faults acting to shorten layers. The resulting geometry of brittle beds in a more ductile matrix is that of Cloos' wedges. Compound wedges involving a series of brittle beds which have been telescoped together are shown by Cloos

(1964, fig. 7) and Perry and de Witt (1977, fig. 10). Price (1967) used the term contraction faults to include all faults that produce a shortening in the plane of the bedding, thus including uplimb thrust faults (see below).

In wedged beds later rotated to vertical by folding, the bounding contraction faults record normal-fault separation at a low angle to bedding. In cratonward-facing folds (asymmetric to the northwest in the Appalachians), such stage I faults normally show a downlimb sense of displacement (fig. 1). They did not form as normal faults in their present orientation, because (1) the necessary northwest-southeast horizontal extension is inconsistent with central Appalachian structural style, and (2) their formation in a contractional regime is indicated by drag features and absence of associated extension features.

Prefolding fractures have been recognized in the central Appalachians (Dean and Kulander, 1977). These fractures may predate the contraction faults, because such fractures are offset by these faults and by bed-parallel compaction stylolites that are inferred to have formed under overburden stress equal to maximum principal compressive stress, when bedding was horizontal.

Prefolding calcite-filled fractures can be recognized if orientations of calcite fibers record shearing on the fractures. If the fractures are rotated to their original

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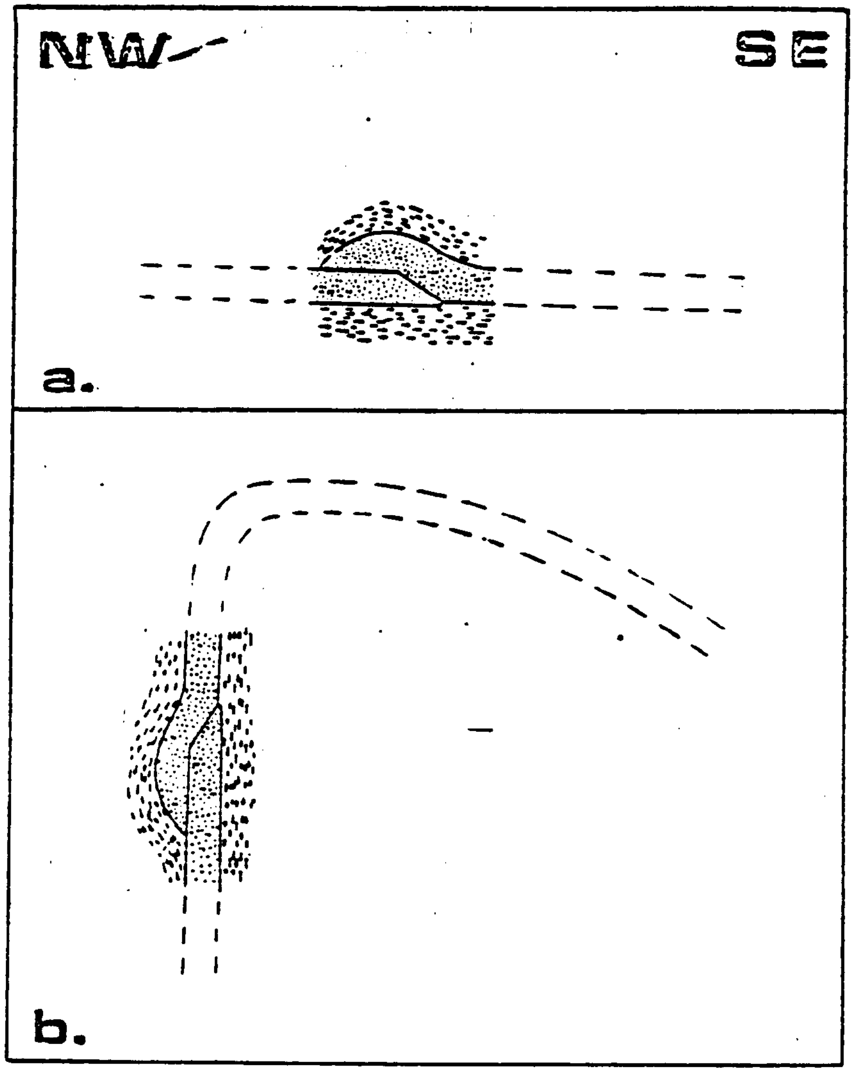

Figure 1. Contraction fault (stage I) formed wedges prior to folding: (a) prefolding attitude, (b) postfolding attitude on northwest limb of west-facing anticline. The angles between the bedding and fault planes and the size of the associated fold are variable.

(prefolding) orientations, the net growth direction of the fibers should be about 30 degrees from the (horizontal) maximum principal compressive stress.

STAGE II STRUCTURES.--Stage II structures formed during folding. They record relative reverse movement toward the anticlinal hinge as part of the Appalachian fold's internal adjustments to southeast-northwest horizontal shortening, and to fold growth by flexural mechanisms. Perry and de Witt (1977) describe and define uplimb thrust faults, the hanging walls of which move up the limb of the anticline and away from the axis of the adjacent syncline (figure 2). Similarly, Gair's (1950) out-of-syncline thrusts, and Gwinn's (1964) symmetrical thrust faults are northwest- and southeast-dipping reverse faults that resolve space problems in cores of anticlines. In concentric folding, upward and inward motion on anticlines' flanks, and flexural slip above ductile rocks, thrust strut-like brittle beds of the limbs over passive anticlinal crests where flexural slip is inhibited. Perry (1971) mapped southeast-dipping apparent normal faults at a low angle to bedding in vertical beds on the northwest limb of the Wills Mountain anticline in Pendleton County, West Virginia. He interprets these as originally northwest-dipping uplimb thrust faults (Perry, 1971, 1978), later rotated by continued growth and asymmetric

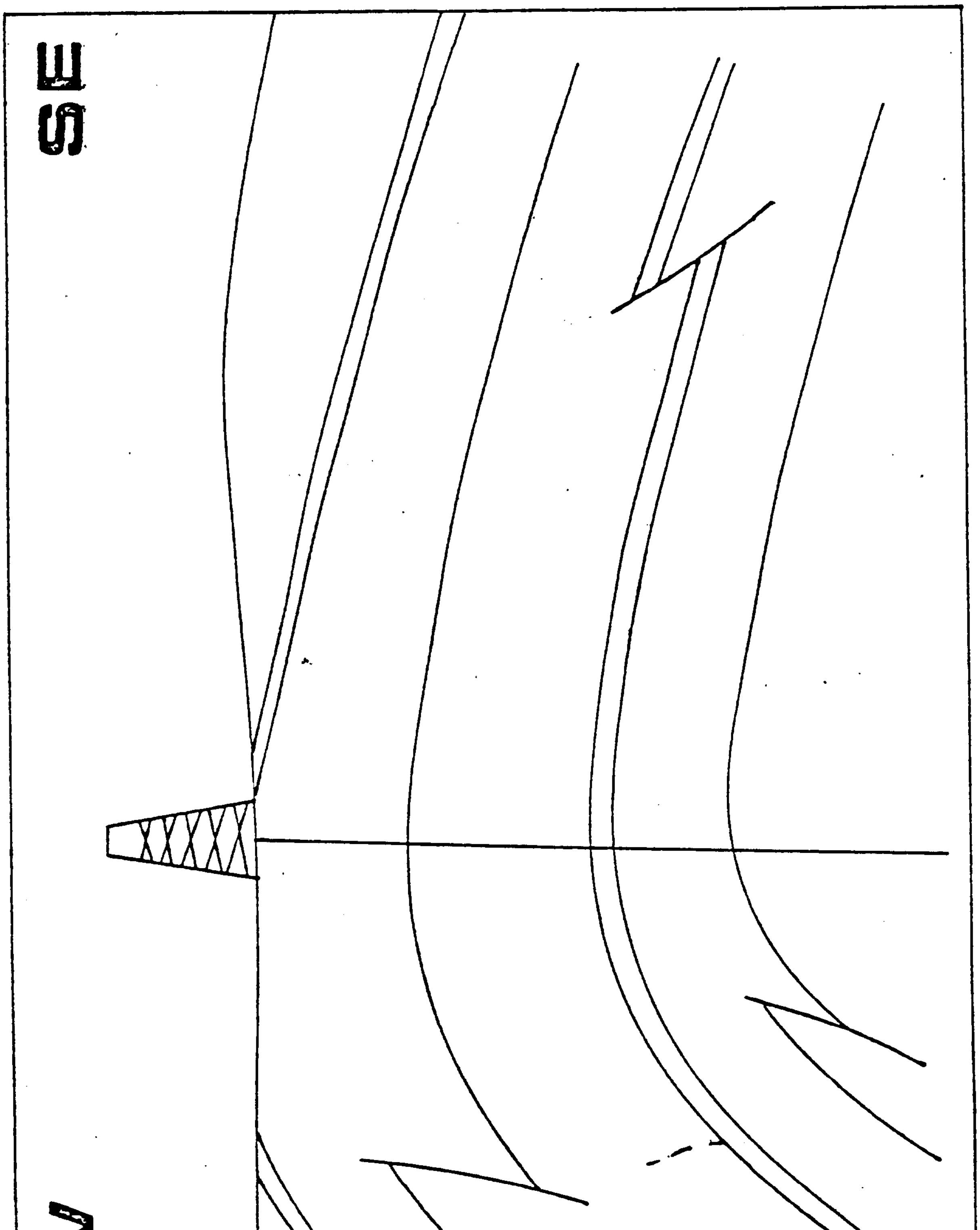

Figure 2. Southeast- and northwest-dipping uplimb thrust faults (stage II).

development of the anticline. Rotated uplimb thrust faults are also present on the nearly vertical northwest limb of the Cacapon Mountain anticline in Maryland (Perry and de Witt, 1977. p. 32).

Stage II structures shorten beds and facilitate anticlinal growth. Both can occur contemporaneously in a planar, mechanically anisotropic medium if net transport is mostly toward the anticlinal hinge. Some rotation of stage II structures occurs if they form early, at low limb dips (Root, 1973). Uplimb thrusts also show a wedge-like geometry and may be difficult to distinguish from stage I wedges. Uplimb thrusts tend to cut many beds and die out in shale flowage (Gair, 1950), folds (Gwinn, 1964), or bed parallel slip (Price, 1964). Wedges tend to be small because they represent the adjustment to shortening of a single or a few strut-like beds. Wedges can show a downlimb sense of displacement on northwest limbs of anticlines. Uplimb thrusts tend to cut more beds because they represent adjustment to shortening of the entire fold. Uplimb thrusts always show hingeward displacement of the upper (or outer) fault block.

As uplimb thrust faults on limbs steepen during growth of the anticline, the normal stress across the thrust surface and the resulting friction increase until the fault locks (Gwinn, 1964). Further tightening of the anticline can produce another fault or faults on the limbs. Continued growth of the anticline, beyond that which flexural slip is capable of relieving, causes onset of stage III.

STAGE III STRUCTURES.--Stage III structures begin to form when beds rotate so far towards the vertical that the effect of vertical extension exceeds that of horizontal shortening. Further growth of the anticline can then only occur by bed-parallel extension. Stages I and II both involve bed-parallel contraction, and so can overlap in time. Because the change from bed-parallel contraction to bed-parallel extension is a discrete event, stages II and III are unlikely to overlap, and their structures should be readily distinguishable. Stage III structures are most easily recognized on and are especially characteristic of steeply dipping to overturned beds.

Norris (1958, 1964) defined extension faults as those that result in elongation in the plane of the layering. He reported extension faulting in otherwise ductile beds (carbonaceous shales) dipping 10 to 25 degrees. Price (1964, 1967), Perry (1971, 1978) and this paper describe extension faults in brittle, steeply dipping beds. Norris and Price noted that if the layering rotated externally in the course of flexural-slip folding until it was at a high angle to the maximum principal compressive stress, then the trajectories of that stress would tend to become perpendicular to the layering and subsequent brittle failure would take the form of extension faults. This rotation of the stress trajectories is shown by the finite-element modelling of Dieterich and Carter (1969). Price (1967) has found that extension faults tend to intersect bedding at about 70 degrees.

Extension faults show net bed-parallel lengthening unique in the kinematic history of a fold, thus allowing easy recognition. Characteristically, extension faults are low-angle, northwest- or southeast-dipping reverse faults on steep or overturned limbs (figure 3). Extension fractures which show bed-parallel lengthening and form normal to bedding are stage III structures. Extension faults have not undergone a significant amount of later rotation. If they formed prior to folding, they would have formed as high angle normal faults, which are inconsistent with Appalachian contractional deformation. Specifically, extension faults cut contraction faults of stages I and II (Perry, 1971, Perry and de Witt, 1977) and are therefore later features. In extreme cases, map-scale extension faults can lead to overthrust west limbs of anticlines (Dennison, written communication, 1978).

With the possible exception of some systematic joints, stage III structures are the latest recognizeable

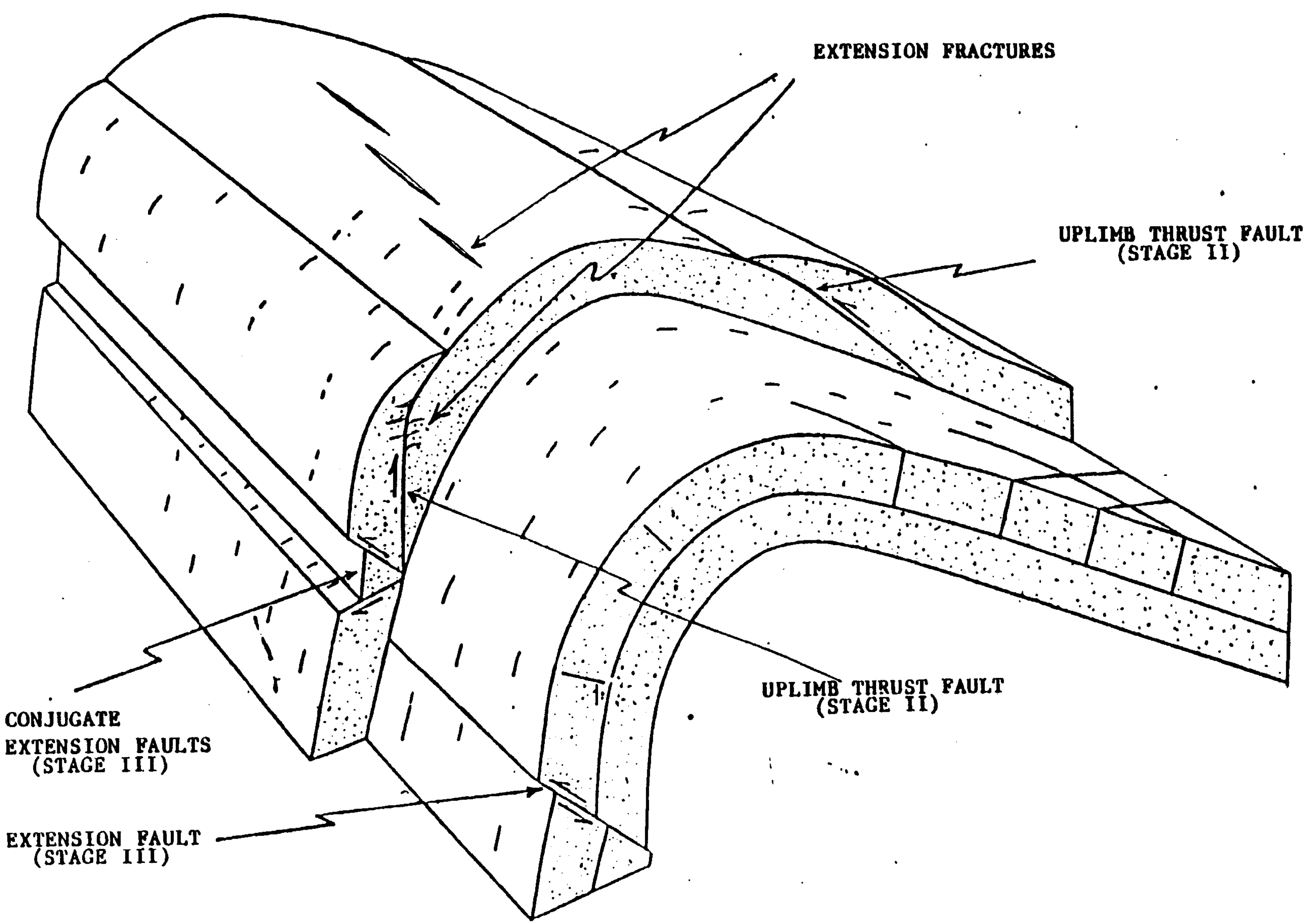

Figure 3. Relationships of stage II uplimb thrust faults to stage III extension faults on an asymmetric fold (modified from Perry and de Witt, 1977, after Price, 1967). Extension fracturing may be associated with both stages II and III.

effects of brittle, fold-related deformation in the central Appalachians. Because they are least likely to be filled by vein material or closed during later deformation, they are a possible target for gas exploration in fractured rock.

Field Applications

Recent fieldwork has shown the applicability of our three-stage model of brittle deformation in steeply dipping beds in the central Appalachians. Nine exposures of Ordovician through Mississippian limestones, sandstones, siltstones, and shales of West Virginia, Maryland and Virginia have been interpreted in terms of the model to yield a sequence of events consistent with Appalachian structural style. Slip-senses of minor faults are usually apparent and show either stage I contraction faulting, stage II crestward slip, or stage III bed parallel extension, depending on the angle between beds and horizontal shortening when failure occurred. Directions of fracture openings are ambiguous unless the fractures are filled with fibrous calcite, cut across compaction or tectonic stylolites, or are slickensided. Durney and Ramsay (1973) showed that crystal fibers in calcite fracture-fillings track the direction of opening of the fracture. If the fracture opened by simple dilatancy, then succesive orientations of incremental extension during fiber growth can be determined. If the direction of fracture opening is parallel or at a small angle to bedding, then the fracture is unequivocally a stage III feature. Fractures whose opening directions have components of crestward shearing can be either stage I or stage II.

An excellent exposure showing examples of all three stages of brittle deformation is the cut in Silurian limestones along the Baltimore and Ohio Railroad, on the northwest flank of the Wills Mountain anticline at Pinto, Allegany County, Maryland.

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Western Limit of Extension Fracturing in West Virginia

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Abstract: Extension fractures can extend beds and create porosity, permeability or both where dips on anticlinal limbs are relatively steep (greater than 45 degrees if there is no strain parallel to hinge lines). Simple calculations and reasonable assumptions permit the conclusion that including the effect of hinge line parallel strain does not significantly alter the minimum bed dip at which extension fractures will open. Therefore the minimum conditions necessary for extension fracture formation are unlikely to occur west of Gwinn's (1964) high plateau, with the exception of the Burning Springs anticline, in Wood and Wirt Counties, West Virginia, where limb dips reach a maximum of 68 degrees.

INTRODUCTION

Detachment faults in ductile beds are the primary mechanism for map-scale deformation in the central Appalachian allochthon, and specifically for the Valley and Ridge and most of the Plateau provinces in West Virginia (Gwinn, 1964; Rodgers, 1963). The location of the western limit of detachment in the central Appalachians is a matter of conjecture, although it may be significant in explaining the distribution or occurrence of gas production from the Middle and Upper Devonian clastic sequence (mostly shales). The purpose of this paper is to determine the western limit in West Virginia of detachment-related fracture porosity and permeability formed as folding rotates beds from low limb dips, where they are contracted, to high dips, where they are extended.

STRUCTURAL STYLE

The West Virginia allochthon has many detached anticlines but few outcropping major thrust faults. The major anticlines are interpreted as active features generated mainly by duplication of strata by ramping of underlying detachment faults, by splay faults, by kink band folding, and by ductile flow of shale rich intervals into anticlinal crests (Gwinn, 1964; Faill, 1969; Wheeler, 1975). The major synclines are regarded as passive features resulting from anticlinal growth in adjacent rocks, rather than from active downbuckling (Gwinn, 1964).

Location of the western limit of allochthonous anticlines involving the Devonian clastic sequence is ambiguous, because the folds generally become less distinctive towards the west. Limb dips and closure decrease westward. In some of the folds of central and western West Virginia, such as the Arches Fork and Wolf Summit anticlines, closure is apparent below the Middle and Upper Devonian clastic sequence, on the Middle Devonian Onesquethaw Group (Cardwell, 1973), but not above the clastic sequence, on the Middle Mississippian Greenbrier Group (Haught, 1968). The limestones, sandstones, cherts, and thin shales of the Onesquethaw Group and the underlying Oriskany Sandstone, Helderberg Group, and Tonoloway Formation commonly form a single stiff unit about 600 to 1000 feet (180 to 300 meters) thick (Cardwell and others, 1968) in the mechanical stratigraphy of West Virginia. If the folds in the Onesquethaw are part of the allochthon, then a detachment below the Silurian Tonoloway and the thick Devonian clastic sequence (2600 to 7800 feet; 800 to 2400 meters) has absorbed the deformation by flow (Cardea, 1956) or upper detachment (Dahlstrom, 1969a). On the other hand, some folds show closure in the Mississippian Greenbrier Group but not in the Devonian Oriskany Sandstone, from which we infer that a detachment is above the Oriskany and probably in the ductile Middle Devonian black shales. Perry and Wilson (1977) describe an example of this on the Mann Mountain anticline.

We believe that extension fractures may form a significant part of gas-bearing fracture porosity and permeability. Extension fractures that show bed-parallel extension are interpreted to be late tectonic, formed when folding beds rotated to such steep dips that the rocks were within the extensional, rather than contractional field of the strain ellipsoid (Berger, Perry and Wheeler, ms. in review). These extension fractures are unlikely to be filled by vein material or closed by later deformation and thus may produce gas. We attempt to locate the western limit of extension fracturing in the subsurface as part of a program to predict the location of more highly fractured rock for gas exploration.

In this paper we use the analytical and graphical techniques of Ramsay (1967) to determine the angle (θ) between the X strain axis (the direction of greatest lengthening: vertical) and the surface of no finite longitudinal strain. The surface of no finite longitudinal strain marks the boundary between the contractional and extensional fields in the strain ellipsoid and its dip increases with the horizontal contractional strain. Assuming that the maximum principal compressive stress and the Z strain axis (direction of greatest contraction) are horizontal and perpendicular to strike (Perry, 1971), then the minimum bed dip at which extension fractures can open is $90 - \theta$ degrees. Using published values of shortening or values measured on published cross-sections, we calculate the critical (minimum) bed dip at which extension fracturing will begin. Comparison of predicted values of critical dips with observed or estimated values of limb dips or shortening in anticlines allows us to predict the distribution of extension fracturing and related fracture porosity and permeability in West Virginia.

METHODS

Ramsay (1967, p. 128) determined the equation for the angle (θ) between the surface of no finite longitudinal strain and the X strain axis, assuming constant volume and no hinge line parallel strain ($e_2 = 0$)

$$\theta = \pm\cos^{-1}\sqrt{\frac{\lambda_3' - 1}{\lambda_3' - \lambda_1'}} \qquad (1)$$

where $\lambda_i' = 1/\lambda_i$

and $\lambda_i = (1+e_i)^2,\ i = 1,2,3$ (2)

Therefore to determine the minimum bed-dip at which extension fractures will begin to open for a given anticline and its estimated shortening value, we need values of λ_1 and λ_3.

λ_3 measures length change parallel to the X (contractional) strain axis. λ_2 measures length change parallel to the hinge line of the fold (intermediate strain axis: Y) and is assumed to be equal to 1. λ_1 measures strain parallel to the Z (extensional) strain axis. λ_3 is found by measuring the amount of shortening for a given anticline by the sinuous bed or equal area method, after removal of synfolding or postfolding slip on faults that changed bed length but did not rotate beds. We shall regard folding of a stiff layer within a volume of softer rock as grossly approximating overall homogenous nonrotational strain. We assume that the volume of rock containing the rotating stiff bed is not cut by through-going detachments and thus has not been deformed by shear on or near the detachment. That assumption is reasonable within a single anticline (Kulander, oral communication, 1978). Then by the sinuous bed method after restoration of fault slip, the shortening strain is

$$e_z = (\ell-\ell_o)/\ell_o \qquad (3)$$

where ℓ_o = the arc length of a stiff unit like the Onesquethaw—Tonoloway sequence, and

ℓ = the linear distance between the two inflection points of the anticline (the distance into which ℓ_o has shortened).

The sinuous bed method is valid for stiff units that have buckled, kinked, or faulted, but have not flowed internally or pressure-dissolved. The values of shortening in this paper are determined by the sinuous bed method but are consistent with values determined by the equal area method (Gwinn, 1970), which does not assume that there has been no internal flowage or mass-removing pressure solution.

Our values of shortening need not include shortening by layer parallel penetrative strain (Engelder and Engelder, 1977) or by formation of pressure solution cleavage (Geiser, 1977), because most of this shortening appears to have formed early, prior to folding (Geiser, 1970; Nickelsen, 1976, 1978; Berger, Perry and Wheeler, ms. in review).

λ_1 can be calculated as follows. For a unit sphere deforming at constant volume to a strain ellipsoid

$$V(\text{sphere}) = \tfrac{4}{3}\pi r^3 = \tfrac{4}{3}\pi$$

$$V(\text{ellipse}) = \tfrac{4}{3}\pi(1+e_1)(1+e_2)(1+e_3)$$

Substituting equation (2) and setting V(sphere) equal to V (ellipsoid)

$$\lambda_1 = 1/\lambda_2\lambda_3 \qquad (4)$$

Therefore to determine λ_1 we need only λ_3, which we estimate from shortening, and λ_2, which we assume is equal to 1. Therefore

$$\theta = \pm\cos^{-1}\sqrt{\frac{1/\left[\left(1+\frac{l-l_0}{l_0}\right)^2 - 1\right]}{1/\left[\left(1+\frac{l-l_0}{l_0}\right)^2 - \left(1+\frac{l-l_0}{l_0}\right)^2\right]}} \qquad (5)$$

If we take into account hinge line parallel strain, then λ_2 will be close to but not equal to 1. We assume that e_2 will be no more than ±10 percent of e_3. This figure is arbitrary but is probably in excess of the true value in most places, especially where large folds are straight in map view. There is no doubt that strain parallel to hinge lines occurs. The existence of cross joints in folded rocks demonstrates hinge line parallel extension, although hinge line parallel contraction may be equally likely in folded rocks. The arcuate trends in parts of the central Appalachians may help determine whether hinge line parallel contraction or extension has occurred. Where the Appalachians are convex toward the craton, as in central Pennsylvania, hinge line parallel extension may be more likely. Where the Appalachians are concave cratonward, as in southern West Virginia and western Virginia, hinge line parallel contraction is more likely.

Using our values of shortening, and reasonable estimates of λ_2, the angle (θ) between the vertical X strain axis and the surface of no finite longitudinal strain is determined graphically using a Mohr diagram for three-dimensional strain (Ramsay, 1967, p. 152). The points along the line $\lambda' = 1$ on the Mohr diagram yield values of points on the surface of no finite longitudinal strain, in degrees from the X and Y strain axes. These values can be plotted in equal area projection to determine the shape of the surface of no finite longitudinal strain: conical with hinge line-parallel strain, planar without (Ramsay, 1967, fig. 4-21). The minimum bed dip at which extension fractures will open is measured along the east-west axis of the projection.

RESULTS

Table 1 lists the predicted bed dips at which extension fractures will begin to open for given values of shortening and $\lambda_2 = 1$. These values were calculated using equation (5) and are the complements of the angles (θ) between the X strain axis and the surface of no finite longitudinal strain. Some of the predicted dips were checked by Mohr diagrams for three-dimensional strain. With no change in length parallel to the hinge line, the minimum bed dip at which extension fractures will begin to open is 45 degrees. That finding is consistent with the results of finite-element modeling of viscous layers (Dieterich and Carter, 1969), and other work summarized by Perry (1978, p. 524). Where there is $\pm .1(e_3)$ hinge line parallel strain, the minimum bed dip at which extension fractures will open does not differ significantly from the values in Table 1.

Table 2 lists measurements of shortening for anticlines in West Virginia, determined from cross-sections by the sinuous bed method or from estimates in the literature. With the exception of the Burning Springs anticline, in Wood and Wirt Counties, West Virginia, no anticlines west of Gwinn's (1964) high

TABLE 1

PERCENT SHORTENING	MINIMUM LIMB DIP
1%	45 degrees
2%	45 degrees
3%	46 degrees
4%	46 degrees
5%	46 degrees
10%	48 degrees
15%	50 degrees
20%	51 degrees
25%	53 degrees
30%	55 degrees
35%	57 degrees
40%	59 degrees
45%	61 degrees
50%	63 degrees

plateau, which have structural relief less than 300 feet (90 meters), should show extension fracturing unless faulting in the subsurface has created abnormally high limb dips. Fault imbrication in the subsurface of the Burning Springs anticline has caused dips as steep as 68 degrees (Shockey, 1954). The Arches Fork and Wolf Summit anticlines may also show extension fracturing, since Gwinn (1964) indicates that they have between 300 and 800 feet (90 and 240 meters) of relief, intermediate between anticlines of the high plateau and those of western West Virginia.

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This work was aided by material comments and computer help from E. Werner, B. Kulander and T. Wilson. R. Shumaker and M. T. Heald reviewed the manuscript. The manuscript was prepared with a modification of WYLBUR, a text editing system developed at the Stanford University Computation Center.

Table 2.

Anticline	Location	Shortening	Maximum Limb Dip	Source
RELIEF GREATER THAN 800 FEET (240 METERS)				
Burning Spgs.	Wood/Wirt Co.	2%	68 degrees	Shockey (195)
Briery Mtn.	Preston Co.	6%	45 degrees	Cardea (1959)
Chestnut Rdg.	Monongalia Co.	3%	12 degrees	Mitchell (19
Wills Mtn.	Pendleton Co.	20%	overturned	Perry (1971)
Browns Mtn.	Pocahontas Co.	16%	overturned	Kulander and Dean (1972)
RELIEF BETWEEN 300 and 800 FEET (90 and 240 METERS)				
Wolf Summit	Lewis Co.	<1%	5 degrees	Milner (1968
Hiram	Harrison Co.	<1%	<1 degree	Cardwell (19
Arches Fork	Doddridge Co.	<1%	<1 degree	Cardwell (1973)
Warfield	Logan Co.	<1%	<1 degree	Cardwell (19
RELIEF LESS THAN 300 FEET (90 METERS)				
Mann Mtn.	Fayette Co.	<1%	<1 degree	Perry and Wilson (197)

Late-Tectonic Extension Faulting in the Central Appalachians

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Abstract: Examination of outcrops in parts of the western Valley and Ridge and eastern Allegheny Plateau provinces of Maryland, Pennsylvania, Virginia and West Virginia located 227 bed-extension faults of which 115 had measurable separation at 24 stations. Previous work showed the faults formed late during folding when beds were steeply dipping, and not by early normal faulting when beds were horizontal. The distribution and amount of extension faulting have implications bearing on tectonic development and gas potential. First, outcrop-size and larger extension faults contribute significantly to anticlinal growth in more brittle units when external rotation steepens fold limbs. Second, stratigraphic localization of extension faulting suggests that reported thinning of brittle units on steep limbs of anticlines may be partly due to late-tectonic extension faulting, rather than wholly to splay faults rising from detachment surfaces. Third, measured values of bed-parallel extension for packets of especially brittle rocks indicate

significant amounts of fracture potentially suitable for gas reservoirs.

INTRODUCTION

The western Valley and Ridge and eastern Allegheny Plateau provinces of Maryland and West Virginia are characterized by anticlines with the greatest amounts of structural relief within their respective provinces. In the Valley and Ridge province, the Wills Mountain anticline (Fig. 4) has approximately 4000 m of structural relief (Perry, 1975). In the Plateau province, the Browns Mountain anticline may have structural relief as great as 2500 to 3000 m (Kulander and Dean, 1972). The Elkins Valley anticline also exposes vertical beds and may have structural relief of almost 3000 m (Rodgers, 1970). Northwest limbs of these folds are steeply dipping and locally overturned. Other eastern Plateau anticlines are characterized by more steeply dipping rocks in the subsurface than at the surface, at least to the level of the Lower Devonian (Rodgers, 1970), and may be asymmetric with more steeply dipping southeastern limbs at least to the level of the Oriskany (Gwinn, 1964).

The steeply dipping limbs of map-scale anticlines in the central Appalachians are due primarily to external rotation of stiff beds (mostly sandstones and limestones) during folding. The map-scale folding is caused by stepwise westward rise (ramping) of detachment surfaces or by splay faulting from nearly horizontal detachment surfaces (Rodgers, 1963; Gwinn, 1964). In most places in the central Appalachians, the detachment surface or splay fault does not reach the surface. As bed rotation and accompanying horizontal shortening and vertical extension continued, anticlinal growth in stratigraphic units above the detachment involved flow of shaly sequences out of anticlinal limbs and into crests (Wheeler, 1975) and bed-parallel extension in steeply dipping to overturned beds (Perry, 1971).

This study is based on outcrop examination of bed-parallel extension structures, primarily faults, which begin to form when beds rotate sufficiently toward the vertical that the effect of vertical extension exceeds that of horizontal shortening. Extension faults can form at any bed orientation, depending on the attitude of the local maximum principal compressive stress (Norris, 1958, 1964). However in the central Appalachians, analysis of structural style and field investigation both indicate that extension faults form when the dip of beds exceeds 45 degrees (Berger and Wheeler, in review), and mostly as reverse faults, serving to thin the beds. Cloos and Broedel (1943) and Root and Wilshusen (1977) described faults with similar structural style in other parts of the central Appalachians. Perry and de Witt (1977), Perry (1978), and Berger, Perry and Wheeler (in press) recognize these reverse faults as extension faults, analogous to those described from the Canadian Rockies by Norris (1958, 1964) and Price (1964, 1967).

With the possible exception of some systematic joints, these extension faults are the latest recognized effects of brittle but fold-related deformation in the central Appalachians (Berger, Perry and Wheeler, in press). Because they are formed late during deformation, they are least likely to be filled by vein material or closed during later deformation. If sufficiently numerous and extensive they are an excellent target for gas exploration assuming the presence of a gas-bearing source and reservoir seal. In particular, they may form permeable fractured reservoirs in the Middle and Upper Devonian clastic sequence. This paper attempts to evaluate that hypothesis.

FIELD METHODS

More than 100 large exposures of steeply dipping or overturned beds were examined in the western Valley and Ridge and eastern Allegheny Plateau provinces in parts of Maryland, Pennsylvania, Virginia and West Virginia (Fig. 4). Of these exposures, 24 were intersected by at least one measurable extension fault. Each fault was identified

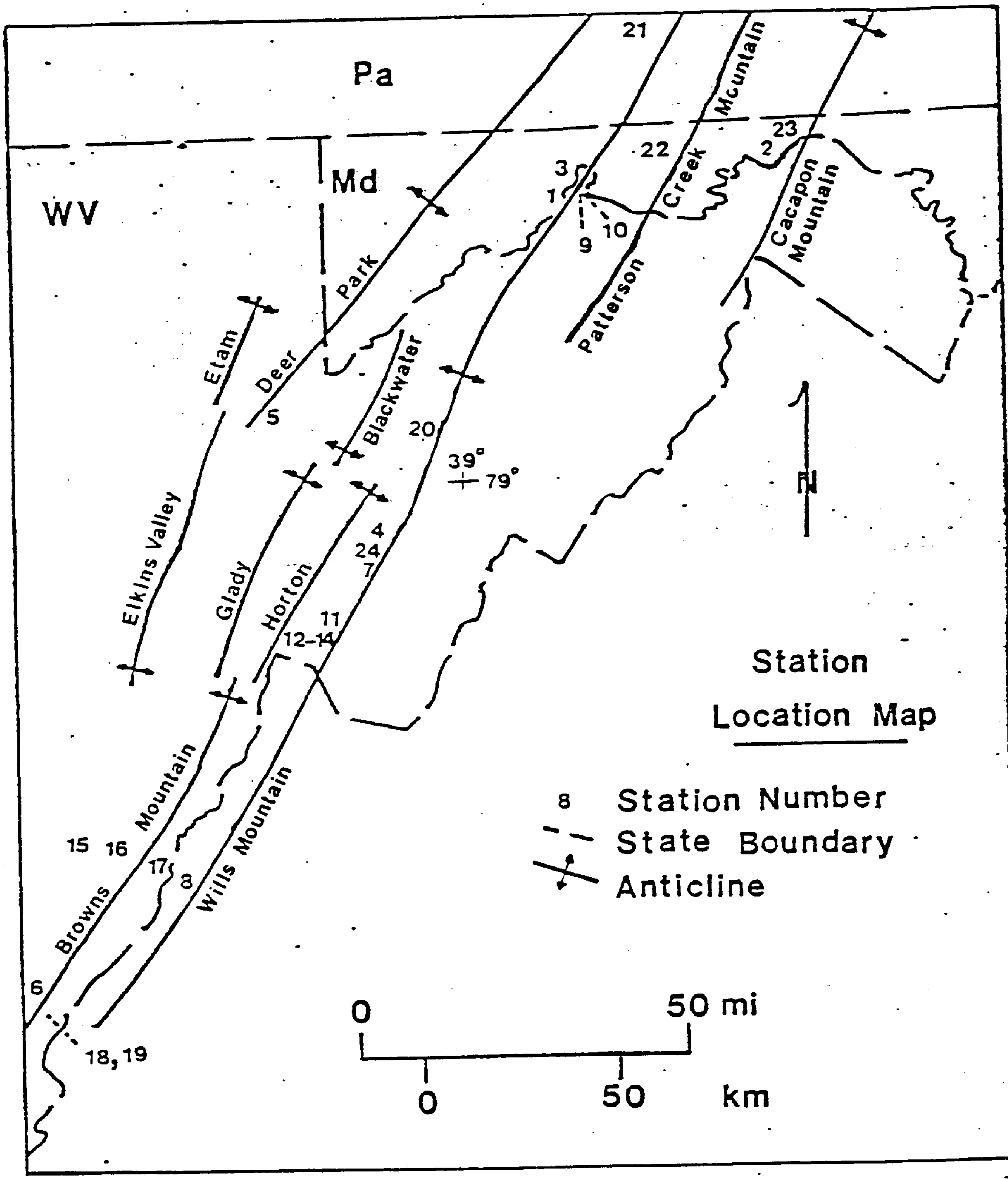

Figure 4. Map showing major anticlines in the study area. Numbers show locations of stations listed in Table 4.

either by the senses of asymmetry of associated drag folds or by matching of beds across the fault. The following were measured or calculated for each fault: vertical separation, horizontal separation, dip separation, bed-parallel extension (separation measured parallel to the beds, not defined by Dennis, 1972), and the thickness of the packet of beds affected by the fault. For each station, the exposure length and thickness of each stratigraphic unit present were measured. In exposures with variable dip, only beds dipping more than 45 degrees were included, because only there are extension faults expected (Berger and Wheeler, in review).

At a few places in the exposures, zones of en echelon filled feather fractures with sigmoidal habit were presumed to mark incipient extension faults. The dip separation was calculated as the sum of the zone-parallel thicknesses of the fracture fillings and the other separation values were calculated accordingly. Styles of terminations of the fault surfaces were noted, as were spatial relations to folds and to contraction faults (faults that shortened beds: Norris, 1958, 1964).

STRUCTURAL STYLE

Perry (1971) recognized 31 extension faults on the northwest limb of the Wills Mountain anticline in Pendleton County, West Virginia. He (unpublished notes, 1965-1967) found extension faults in the Silurian Tuscarora quartzite, Rose Hill formation, Keefer sandstone, Williamsport sandstone and Tonoloway limestone and the Devonian Oriskany sandstone. Table 3 details the number of extension faults recognized in our study. Extension faults were found in many of the same units in the study area in which de Witt and Dennison (1972) noted minor faulting and Perry (1971) noted minor extension faults. Extension faults were found as well in other units.

Outside the study area, Cloos and Broedel (1943, their Plate 1) show 27 extension faults in 180 m of outcrop in the Hamilton Group near Harrisburg, Pennsylvania. Not all the reverse faults reported by Cloos and Broedel (1943) are extension faults. Many are fractures with unmeasurable amounts of offset or are contraction faults. However, the number and distribution of faults and other fractures (843 fractures in 12 stations) led Cloos and Broedel (1943) to conclude that reverse faults in steeply dipping beds are widespread and characteristic of Appalachian folding (p. 1388).

Some extension faults occur singly. Others are conjugate pairs with associated synthetic and antithetic extension faults. Numerous other fractures were observed, especially in the shaly parts of the Brallier Formation,

Table 3

STRATIGRAPHIC UNIT	NO. OF EXTENSION FAULTS	NO. OF MEASURABLE EXT. FAULTS	CUMULATIVE THICKNESS OF SECTION: OUTCROPS WITH MEAS. EXT. FAULTS	FAULTS PER m OF SECTION
Mississippian				
Pocono Formation	9	4	266 m	0.034
Devonian				
Hampshire Formation	1	0	--	—
Chemung Group	8	6	308 m	0.026
Brallier Formation	104	43	1251 m	0.063
Millboro Formation	1	1	75 m	0.01
Marcellus Formation	4	0	--	--
Silurian				
Tonoloway Formation	35	24	200 m	0.175
Wills Creek Formation	1	1	159 m	0.006
Bloomsburg Formation	7	2	94 m	0.074
McKenzie Formation	27	17	148 m	0.182
Rochester Shale	14	14	139 m	0.101
Keefer Sandstone	2	1	6 m	0.3
Rose Hill Formation	4	2	7 m	0.6
Tuscarora Formation	7	4	82 m	0.08
Ordovician				
Juniata Formation	3	2	40 m	0.08
Total	227	115		

many with slickenside, slickenline, or apparent dip orientations similar to those of nearby extension faults but not showing drag folding or matchable beds. The number of extension faults recognized, as well as the numerous undifferentiated fractures, indicates to us that extension faults are an important and little recognized component of, and are indeed characteristic of, certain stratigraphic units where those units are rotated to steep dips. Thus field observations indicate that in the Devonian clastic sequence, the Brallier Formation may form unusually many extension faults when bed dips increase into the extensional field.

Field observations indicate that the extension faults may terminate against each other, or against a thick bed of siltstone or limestone. In some places the thick bed is broken by a fracture zone along the continuation of the extension fault, and the fractures may show small amounts of slip. An extension fault may die out as a single fault or by distributing its slip among several splays in shales. At many places the extension faults curve upwards or downwards into surfaces of bedding-plane slip (Price, 1964).

Extension faults may form and have been recognized on limbs of folds ranging in size from typical outcrop-scale to map-scale. Therefore they represent a common response to the local maximum principal compressive stress, which tends to be roughly parallel to beds until dips reach 45 to 60 degrees, and then tends to be perpendicular to the beds (Dieterich and Carter, 1969). Previous workers have shown that most extension faults form at high angles to beds, although they can form at lower angle (Norris, 1958; Price, 1967; Perry, 1971).

Price (1964) writes that in less steeply dipping beds in the Canadian Rockies, extension faults tend to show a uniform dip direction and commonly begin and terminate in a bedding-plane slip surface. With increased bed rotation, he noted that extension faults form conjugate sets. We have not found this to be generally true in the central Appalachians. An extension fault that terminates in a bedding-plane slip surface may be part of a conjugate set of extension faults. Overturned beds in the central Appalachians show both single faults and conjugate sets of extension faults.

Because some extension faults are found to begin or terminate in bedding planes, earlier flexural slip folding may still have been active at the time of extension faulting. We propose that extension faulting can occur during flexural-slip folding if the contacts within a packet of beds become locked, perhaps around small folds, so that the shear strength at a bed contact exceeds the tensile strength of the bed or beds. The slip surface may then cut across the locked beds, forming an extension fault. Some subsequent external rotation of those extension faults by further folding is likely (Price, 1964), especially where beds are now overturned.

Other extension structures found include boudinage of thin siltstone or limestone beds, abundant lenticular extension fractures oriented roughly perpendicular to bedding, and three normal faults of small displacement. There is a strong positive association between exposed fault length and amount of dip separation (for 119 observations, Spearman's correlation coefficint (Siegel, 1956) gives a significance level of .0001.

RESULTS

Table 4 summarizes the results of our field investigations for the 24 stations that showed measurable separation. Below, we suggest that useful conclusions of regional applicability can be reached by extrapolating our data into the subsurface. Inferences about fold growth and tectonic thinning of the limbs can be made from compilation of the vertical and horizontal separation data, respectively. Estimates of gas potential are based on the amount of bed parallel extension.

It must be emphasized that measurements reported here are minimum values. Only definite extension faults

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Table 4.

STATION NO. AND NAME	E/R	T/R	F/R	LOCATION	STRATIGRAPHIC UNITS(1)
1 Pinto, Md	27.0%	1.6%	30.0%	NW limb, Wills Mt. ant.	Rose Hill/Keyser
2 Woodmont, Md	2.9%	0.4%	3.7%	NW limb, Cacapon Mt. ant.	Mahantango/Brallier
3 La Vale, Md	4.0%	0.5%	3.0%	NW limb, Wills Mt. ant.	Mahantango/Mar[illegible]
4 Kettermans Knob, WV	40.0%	0.1%	14.4%	NW limb, Wills Mt. ant.	Brallier
5 Parsons, WV	2.6%	3.5%	3.6%	axial region, Deer Park ant.	Chemung
6 White Sulfur Springs, WV	40.0%	0.1%	14.4%	NW limb, Browns Mt. ant.	Chemung/Pocono
7 Judy Gap-a, WV	<.1%	0.2%	-----	NW limb, Wills Mt. ant.	Tonoloway
8 Ryder Gap, VA	23.0%	3.7%	25.0%	NW limb, Wills Mt. ant.	Brallier/Chemung
9 Cedar Cliffs, Md	1.0%	0.6%	1.3%	SE limb, Wills Mt. ant.	Bloomsburg/McK[illegible]
10 Rome Hill, Md	0.5%	0.2%	0.2%	SE limb, Wills Mt. ant.	Rose Hill/McKenzie
11 Teeter Hollow, WV	<.1%	0.1%	<.1%	NW limb, Wills Mt. ant.	Rochester
12 Snowy Mt. Rd.-a, WV	2.0%	0.4%	2.4%	NW limb, Wills Mt. ant.	Rochester/Keef
13 Snowy Mt. Rd.-b, WV	3.0%	2.9%	2.3%	NW limb, Wills Mt. ant.	Rochester/Keefer
14 Snowy Mt. Rd.-c, WV	1.0%	1.9%	2.3%	NW limb, Wills Mt. ant.	Rochester/Keef
15 Huntersville, WV	0.1%	<.1%	0.6%	NW limb, Browns Mt. ant.	Bloomsburg
16 Rte. 39, WV	3.4%	<.1%	0.6%	axial region, Browns Mt. ant.	Tuscarora
17 Minnehaha Springs, WV	0.5%	0.3%	0.4%	SE limb, Browns Mt. ant.	Juniata/Tuscar[illegible]
18 I-64-a, WV	<.1%	0.1%	<.1%	SE limb, Browns Mt. ant.	Brallier
19 I-64-b, WV	<.1%	0.2%	<.1%	axial region, Browns Mt. ant.	Harrell/Brallier
20 North Fork Gap, WV	0.2%	0.3%	<.1%	NW limb, Wills Mt. ant.	Tuscarora
21 Wolf Camp Run, Pa	<.1%	0.1%	<.1%	NW limb, Wills Mt. ant.	Mahantango/Brallier
22 Martin Mt., Md	<.1%	0.3%	<.1%	NW limb, Patterson Ck. Mt. ant.	Tonoloway
23 Hancock, Md	<.1%	0.1%	<.1%	SE limb, Cacapon Mt. ant.	Bloomsburg
24 Judy Gap-b, WV	31.0%	3.1%	34.8%	NW limb, Wills Mt. ant.	Brallier

E = anticlinal growth; T = tectonic thinning; F = amount of extension; R = structural relief were measured. All measurements were taken in 2 to 3 m high strips along the accessible basal segments of exposures. The larger the displacement along the fault the less the chance of matching beds across the fault. The extrapolations into the subsurface could be maximum values because they assume the same intensity of faulting throughout the vertical extent of the fold limb as that measured in our selected exposure. However, because it seems unlikely that the present erosion level preferentially exposes unusually extended rocks, the true values may approach our extrapolated values. Thus the results reported here are conservative (minimum) but probably roughly accurate values for the amount of fold growth by extension faulting.

FOLD GROWTH

The major anticlines of the central Appalachians are active structures generated mainly by duplication of strata by ramping of underlying detachment faults, by splay faulting, and by ductile flow of shale or evaporite sequences into anticlinal crests (Gwinn, 1964; Faill, 1969; Wheeler, 1975). The major synclines are passive structures resulting from anticlinal growth in adjacent rocks, rather than from active downbuckling (Gwinn, 1964). We propose that extension faulting late during folding provides an important contribution to anticlinal growth.

At Station 1, near Pinto, Maryland on the northwest limb of the Wills Mountain anticline (Fig. 4, Table 4), the amount of growth by bed-parallel extension may be as much as 27% of the entire structural relief of the fold. This value is based on data from a single exposure on the limb of the anticline and is extrapolated into the subsurface by

$$E = V(BPT/ET)(R/2.5) \qquad (1)$$

where E = the amount of anticline growth by extension, in meters,

V = the sum of the amounts of vertical separation of measured extension faults,

BPT = the sum of the thicknesses of the packets of beds containing the extension faults,

ET = the present (post-extension) length of the exposure measured perpendicular to beddinig,

R = the structural relief of the anticline, and

2.5 = the height of accessible exposure in meters.

Estimates of structural relief are: Wills Mountain anticline - 4000 m (Perry, 1971), Browns Mountain anticline - 3000 m (Kulander and Dean, 1972), Deer Park anticline, 3000 m (Gwinn, 1964), Cacapon Mountain anticline - 4000 m (Perry and de Witt, 1977), and Patterson Creek Mountain anticline - 2500 m (Cardwell, 1975). The amount of structural relief of the Wills Mountain anticline in Maryland and Pennsylvania is equal to or greater than the amount measured by Perry (1971) in Pendleton County, West Virginia, because although the anticline appears to lose structural relief northeastward into Maryland (Cardwell and others, 1968), that is compensated by increasing depths to basement from southwest to northeast along the strike of the Wills Mountain anticline (Kulander and Dean, 1978).

Using equation (1), values of E have been determined for each station in which extension faulting was measurable. Similar results have been obtained by Cloos and Broedel (1943), using a similar method of extrapolation. They estimated that the reverse faults contribute 10% to the structural relief of the anticline they studied. Table 4 lists our percentages of anticline growth by extension. The contribution of extension to fold growth is significant in many exposures, especially considering that extension does not begin until late in folding when the limbs are steeply dipping and the fold may have already attained considerable relief. The widespread stratigraphic and geographic distribution of extension faults in steeply dipping beds (Tables 3 and 4,

Fig. 4) supports our suggestion that our results have regional applicability.

TECTONIC THINNING

Tectonic thinning of measured stratigraphic sequences has been reported in the study area (Dennison and Naegele, 1963; Dennison, Travis and Ferguson, 1966; de Witt and Dennison, 1972; de Witt, 1974) and was used as a guide by us in locating exposures that have undergone extension faulting. We hoped to be able to recognize enough extension faults to account for a large proportion of the tectonic thinning. The amount of tectonic thinning (T) is determined using equation (2)

$$T = H/ET \qquad (2)$$

where H = the sum of the amounts of horizontal separation of measured extension faults.

At Station 4, near Ketterman's Knob, Pendleton County, West Virginia (Fig. 4, Table 4), Dennison and Naegele (1963) report that approximately 60 m of the Brallier formation is absent (about 25% tectonic thinning). Outcrop examination at Station 4 indicates a minimum of .1% tectonic thinning by extension faulting. Here Dennison and Naegele (1963) write that tectonic thinning was probably accomplished by minor faulting within a fault zone in the Brallier Formation.

To the north along strike, Dennison, Travis and

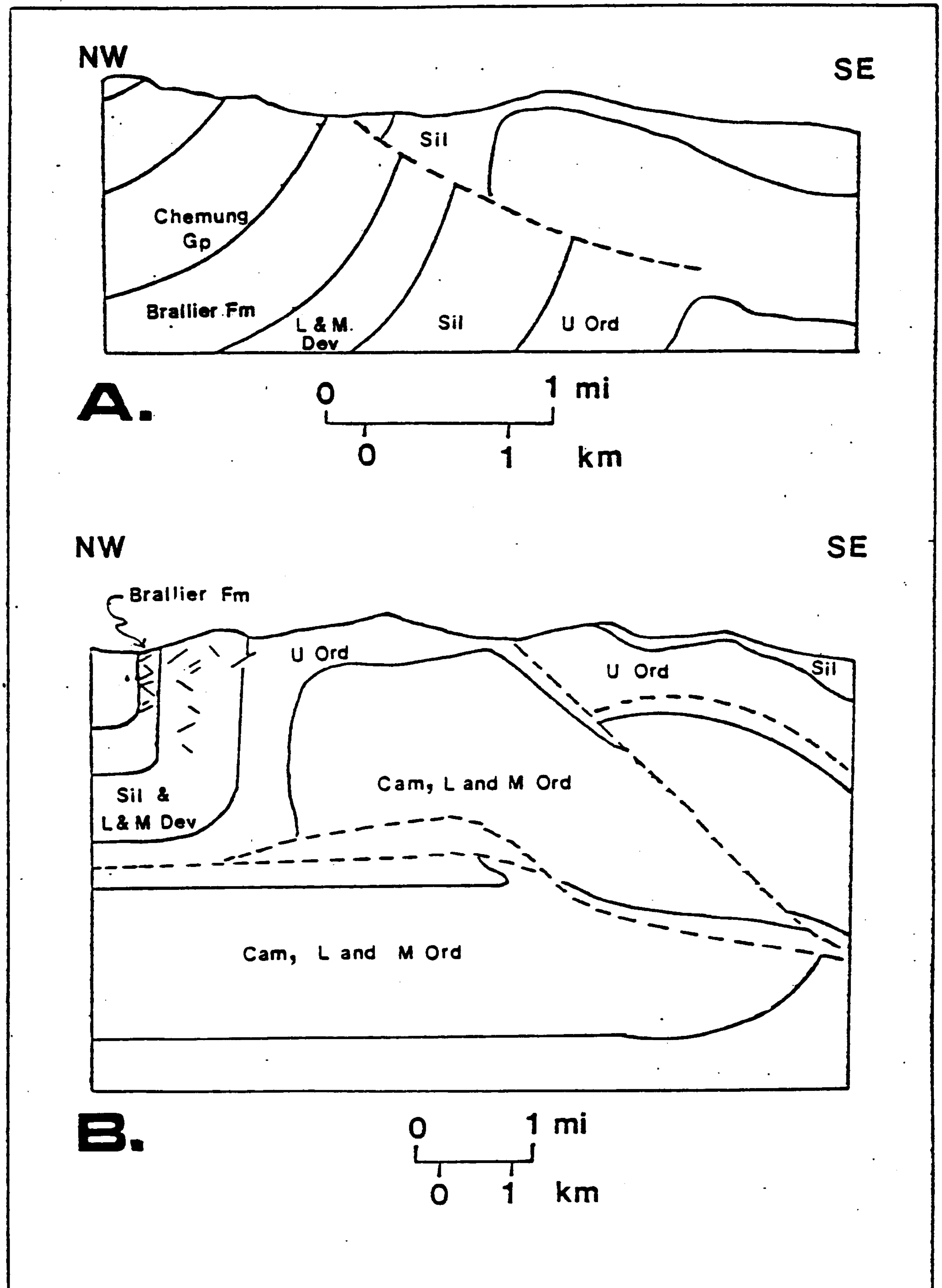

Figure 5. Contrasting models of tectonic thinning by faulting on the northwest limb of the Wi... Mountain anticline. No vertical exaggeration. A. Splay fault interpretation modifi... from Dennison and Naegele (1963). Dennison and others (1966) and de Witt and Denniso... (1972). B. Extension fault interpretation modified from Perry (1971). Dashed lines ... decollement or splay faults. Short solid lines in upper left of sect. B are extensi...

Ferguson (1966) infer the presence of four major fault zones in the Brallier Formation (Kittlelick, Dawson, Cresaptown [station 3], and Hyndman [station 21] faults) largely from geomorphic evidence of a narrow strike valley, but also from measured stratigraphic sections. Dennison, Travis and Ferguson (1966) infer that these fault zones are splay faults rising from a deep detachment, probably located in the Ordovician Reedsville Formation. De Witt (1974 and written communication, 1978) proposes on the basis of seismic data that the Hyndman fault is a large splay fault with associated extension faults in the fault zone, particularly in the vicinity of the village of Hyndman.

We believe that the tectonic thinning reported by Dennison and Naegele at our stations 3 (La Vale) and 4 (Ketterman's Knob) was accomplished more by small and medium size late tectonic extension faults than by the single splay faults envisioned by Dennison and Naegele (1963) and de Witt and Dennison (1972) (Fig. 5). Our arguments are three. First, all four inferred splay faults crop out in and thin the Brallier Formation. This requires a fortuitous coincidence of erosion level with the intersection of the Brallier Formation and the fault surface and must occur for all four faults. If a fault is a splay and thus not confined to the Brallier, then it is about equally likely to crop out stratigraphically above, within, or below that unit. Second, as shown in Table 3, the Brallier Formation shows more extension faults than does any other Devonian unit examined. Third, as noted above under "Field Methods", the Brallier contains more minor faults, many unclassifiable, than does any other unit examined. We suggest a simpler explanation that the Brallier Formation localizes extension faulting in our study area (Fig. 5). These small extension faults may not descend to the detachment in the Reedsville Formation, but could have considerable slip. Perry (1971) has recognized a map-scale extension-like fault on the Wills Mountain anticline (see also de Witt, 1974).

We suggest that the Dawson thrust of Dennison and Naegele (1963) and de Witt and Dennison (1972) is a large extension fault because we observed few minor extension faults in the outcrop belt of the fault zone (station 21 and in exposures near station 1).

The thinning in the Hyndman fault zone is more likely to be due to splay faulting than to the associated extension faults (de Witt, written communication, 1978) on the basis of confidential seismic data available to de Witt. However, the thinning mapped at the surface (de Witt, 1974) may not be clearly connected with the splay fault observed on the seismic section at depth because poor reflections are usually received in such areas of steep dips.

The Kittlelick fault is thought to be a fault zone by Dennison, Travis and Ferguson (1966). We agree but suggest that the thinning along the Kittlelick fault is due to a zone of extension faulting similar to the zone of numerous extension faults observed along the Cresaptown fault and near Ketterman's Knob. We also suggest that thinning of stratigraphic units adjacent to the Kittlelick and Hyndman faults as reported by Dennison and others (1966) and de Witt and Dennison (1972) may also be due in part to extension faulting as well as to branches from splay faults.

PERCENT EXTENSION

Maximum possible amount of extension due to faulting (F) is estimated by

$$F = BPE(BPT/ET)(R/2.5) \qquad (3)$$

where BPE = the sum of the amounts of bed-parallel extension of measured extension faults.

The value (F) is the amount of extension for a limb of a fold based on extrapolation of measured values of bed-parallel extension. The greater the (F) value the greater the possibility of a fold being suitable for gas in fractured reservoirs. This estimate is a maximum because the extension is generated as beds are faulted apart in directions parallel to bedding but the resultant voids are simultaneously reduced by tectonic thinning.

As shown in Table 4, the (F) value due to bed-parallel extension on the Wills Mountain anticline can be locally almost 35% of total structural relief. We have noted that extension faults are always associated with extension fractures. We suggest that qualitative estimates of the relative amounts of fracture porosity and permeability in the subsurface can be made by comparison of bed-parallel extension values (F).

Murray (1968) obtained a value of .05 percent fracture porosity from a geometric analysis of bed curvature in a producing field with very low primary porosity. We believe there occur zones or patches of concentrated fracture porosity, or permeability, or both, depending on the response of a particular stratigraphic unit to the local maximum principal compressive stress. Where many small measurable faults extend the beds, rather than one or a few larger unmeasurable extension faults, then our data will show a higher percent extension. Along the Allegheny Front in southern Pennsylvania, Maryland and northern West Virginia (Dennison and Naegele, 1963), these faulted zones vary in intensity northeast and southwest along strike of the Brallier Formation. It is also possible that faulted zones vary vertically throughout the limb in the subsurface. More detailed stratigraphic work is necessary before drilling targets can be selected. However, estimates of tectonic thinning by Dennison and Naegele (1962), Dennison and others (1966), and de Witt and Dennison (1972), values in our Table 3, and the abundant unclassified fractures observed in the shaly parts of the Brallier Formation all suggest to us that the Brallier may be the most highly extension-fractured and -faulted unit in the subsurface.

Extension faulting is usually restricted to a relatively brittle bed or beds bounded by more ductile shaly beds. Thick-bedded brittle units do not show well developed extension faults, nor do ductile black shales that we examined. Results consistent with these statements have been obtained in experimental work (Griggs and Handin, 1959). We suggest that the 0.3 to 0.7 m thick, shale-bounded siltstones near the base of the Brallier Formation would be the most likely candidates for production controlled by extension faults. They may be the equivalent of the Sycamore sandstone or siltstone (Patchen, 1977), a unit recognized only in the subsurface of central West Virginia and which produces small amounts of gas.

Any patchy distribution of tectonic thinning may be due to the presence or absence of lubrication between beds (see Reches and Johnson, 1976). Where the bed packets are unlubricated, they may lock and fail by extension, as described above. Where they are lubricated, they may undergo flexural slip without extension. The type and distribution of lubrication is beyond the scope of this paper. Alternatively, there exists the possibility that the zones of intense extension fracturing and extension faulting may be due to the presence of a nearby large splay fault.

GAS POTENTIAL

Any potential fractured reservoir whose porosity and permeability are partly or wholly due to extension fracturing would be located in folds of the eastern or High Plateau (Gwinn, 1964) or Valley and Ridge province where dips greater than 45 degrees allow extension fracture formation (Berger and Wheeler, ms. in review). If the Brallier Formation is selected as the target gas bearing formation, then the Glady, Horton, Blackwater and Deer Park anticlines would be the best potential sites in the High Plateau because the Brallier is not exposed but is present at shallow depths and may contain steeply dipping beds. In the Valley and Ridge province the Broadtop synclinorium also offers steeply dipping Brallier Formation in subsurface anticlines (Jacobeen and Kanes, 1974). If the source of gas in the Brallier is the underlying dark, kerogen-rich shales, then the prime drilling targets would be off the crests of the anticlines on the steep, extended limbs. Drilling targets on the limbs have not been adequately tested in the past (R. Shumaker, oral communication, 1978). Here thinned, steeply dipping, extended Brallier rocks may overlie dark shales thinned by flowage toward cores of detached anticlines that grew over imbricated stacks of Oriskany Sandstone and older rocks. The presence of gravity lows near the crests of anticlines indicates shale flowage (Kulander and Dean, 1978).

ACKNOWLEDGMENTS

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Extension in Kink-Bands and on the Limbs of Kink Folds

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ABSTRACT: Beds in kink-bands can undergo bed-parallel extension when the angle between the kink surface and the kink-band is less than the angle between the kink surface and the enveloping beds. Field investigations in the Middle and Upper Devonian clastic sequence of the central Appalachians indicate that the kink-band undergoes more extension fracturing and jointing and shows more twist hackle than the enveloping beds. However, comparison of measured values of extension with predicted volume increases in kink-bands indicates that only about 10 percent of the volume change is translated into extension fractures.

The intersection of two inclined kink-bands produces a kink fold that is identical in gross form to a chevron fold formed by buckling, but can be distinguished by associated fold forms. The geometry of second order folds (Nickelsen, 1963) of the central Appalachians can best be accounted for by a kink fold model (Faill, 1969, 1973).

Gravity lows indicate thickened low density shale on the northwest limbs of the Blackwater and Glady anticlines. We suggest that there exist thinned and extended units downdip from the thickened shale sections. Therefore, map-scale kink folds should have the most thinning by extension fracturing and faulting low on the steep limb in the subsurface. There, such folds are more likely to contain gas in fractured reservoirs.

INTRODUCTION

Dennis (1967) defines kink-band as a tabular zone along which foliation is deflected. A kink fold is formed from the intersection of two kink-bands. A kink fold may be in the form of a box fold or may have planar limbs and narrow hinges similarly to a chevron fold (Fig. 6B).

Recent work by Faill (1969, 1973) has shown that the typical map-scale folds of Pennsylvania have planar limbs and narrow hinges that can best be explained by kink-band deformation. The geometric and kinematic similarities between map- and outcrop-scale kink-bands suggest to Faill that the processes and mechanisms that gave rise to the large folds were identical to those that produced the small folds.

In this paper, we use outcrop-scale folds to provide information about relative amounts of bed extension on limbs in map-scale folds. Anticlines with large amounts of extension may contain suitable exploration targets for gas in fractured reservoirs.

KINK-BANDS AND KINK FOLDS

Faill (1969, 1973) is able to explain most attributes of central Appalachian fold style by using the intersection of two kink-bands inclined toward each other with opposite senses of rotation and divergent kink axes. However, Johnson (1977) argued from elastic theory that folds with planar limbs and narrow hinges also can arise from another sequence of development (Fig. 6A). Johnson proposed that hinges can narrow and limbs straighten progressively during folding, thus transforming a concentric-like fold into a chevron fold. Conceivably, tension fractures may form due to initial bending and later straightening of the limb (B. Kulander, oral communication, 1978). These chevron folds can appear almost identical in outcrop to kink folds formed by the intersection of two inclined kink-bands. The kink-bands tend to form when the beds are lubricated (separated by very weak interbeds) and when the maximum principal compressive stress is at a small angle to beds (Johnson, 1977). Chevron folds formed from concentric-like folds are most likely to occur when the beds are not lubricated, when the surrounding medium is softer than the folded beds and when the maximum principal compressive stress parallels the beds. However, Chapple (1978) notes that

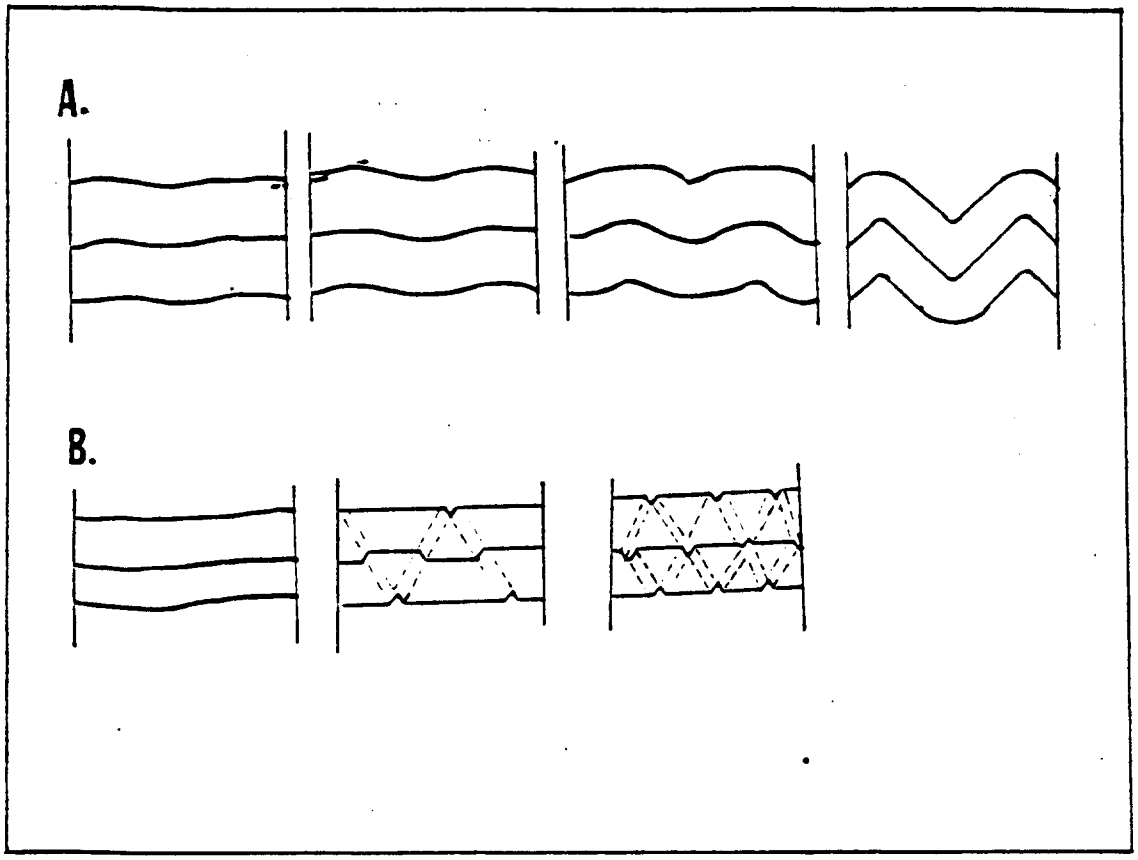

Figure 6. Contrasting models of formation of chevron and kink folds. Horizontal lines at left are selected beds. A. Sequence from sinusoidal to concentric-like to chevron forms. B. Sequence from low, sinusoidal forms to kink forms, with local chevron forms (after Johnson, 1977). Dashed lines denote kink surfaces.

the applicability of Johnson's elastic theory to folding is not entirely clear.

We have tried to discriminate in the field between chevron folds (Fig. 6A) and kink folds formed from the intersection of two kink-bands (Fig. 6B). Johnson (1977) suggests that chevron folds may have vestiges of their earlier concentric-like form (Fig. 6A). The combination of concentric-like and chevron forms causes no room problem in the cores of folds. The folds are similar in style and can extend indefinitely parallel to the axial surface. On the other hand, Berger has observed that many kink-bands and kink folds die out in either direction parallel to the kink surface. If the fold is asymmetric, then it is probably a kink fold formed by the intersection of two kink-bands of unequal width (Paill, 1969). The chevron folds of Johnson (1977) are usually symmetrical. Although these diagnostic criteria are not infallible, they can indicate the affinity of the fold.

Ramsay (1967, p. 450) concludes that less work is expended to shorten a layer by the production of kink-bands or paired conjugate kinks than by the formation of symmetrical chevron folds. Thus it seems logical to conclude that kink folds will be more prevalent in nature. Planar-limbed folds of uncertain affinity are thus more likely to be kink folds.

EXTENSION IN KINK-BANDS

Ramsay (1967, p. 449) uses a geometric model to predict the amount of dilation (Δ) in a kink-band (Fig. 7). The amount of dilation $\Delta = (\sin \beta_2 / \sin \beta_1) - 1$ (1) where β_1 is the angle between the axial surface and the enveloping beds, and β_2 is the angle between the axial surface and the kink-band. If $\beta_1 > \beta_2$, then $\Delta < 0$ and the layers thin and extend parallel to bedding. Kulander and Dean (in review) use this method to predict the amount of porosity in kink bands. They estimate that 10 percent porosity can occur when the beds separate ($\beta_1 < \beta_2$ and $\Delta > 0$). They determine that porosity can also occur as the beds separate in the hinges of kink-bands if $\beta_1 = \beta_2$, but that the amount of such hinge porosity depends on the thickness to length ratio of the beds composing the kink band.

In this study, we are interested in kink folds where $\Delta < 0$, so that one limb is extended. Generalizing this model to central Appalachian folds that have planar limbs and narrow hinges, β_1 is the angle between the axial surface and the shallow limb. β_2 is the angle between the axial surface and the steeper dipping limb. We have investigated several planar-limbed folds to try to observe more extension in the kink-band or in the steep limb. We compare the measured amount of extension with the amount of extension predicted by equation (1), and assume to

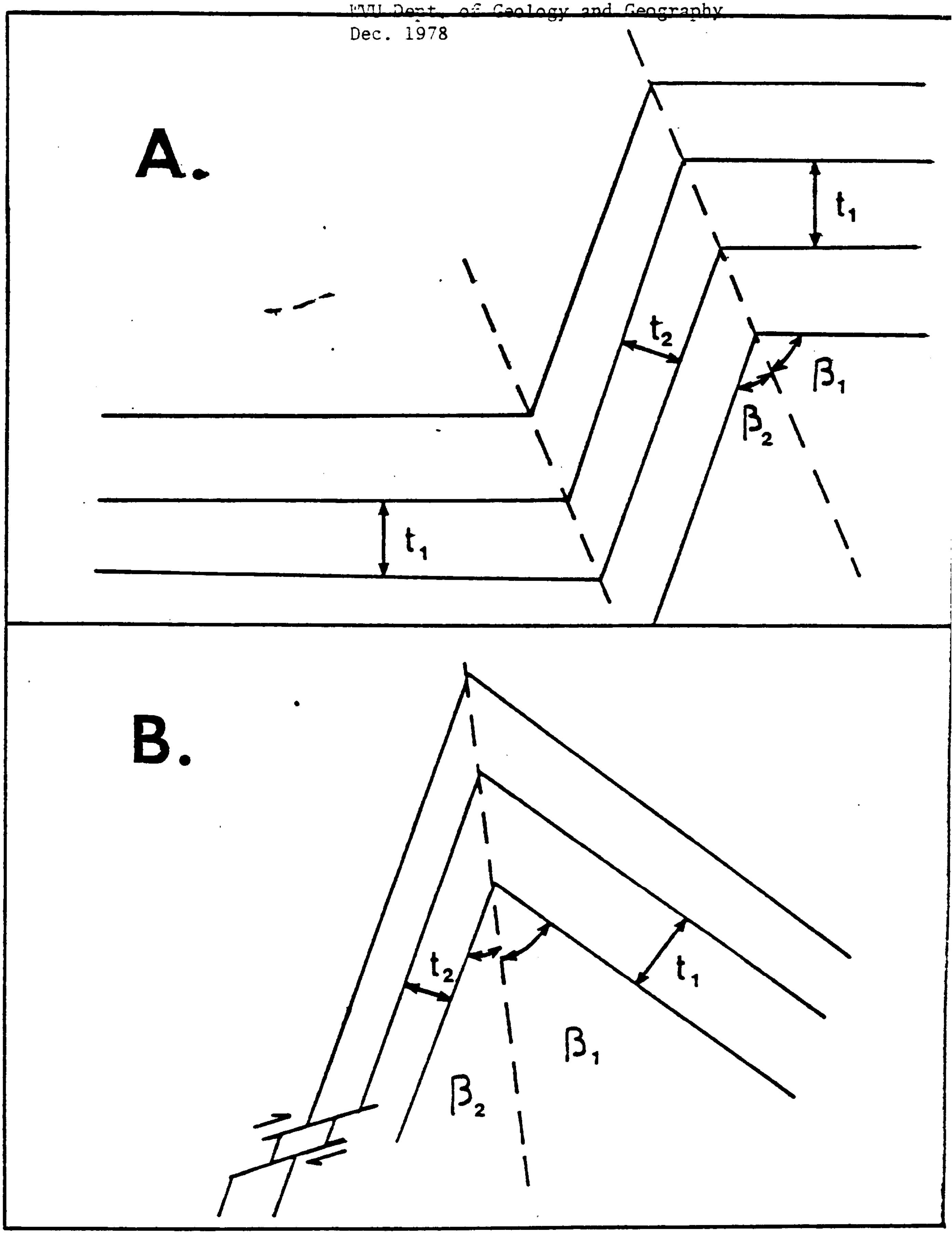

Figure 7. Kink-band model of fold development showing relationship of β_1 and β_2 to the kink surface and thinning and extension of beds in the kink-band (A) or steep limb of a kink fold (B). t_1 is original bed thickness, t_2 is thinned bed thickness.

start that the dilation is completely translated into extension fractures. As we shall see, that assumption is overly optimistic, but that will not change the validity of the following analysis.

FIELD INVESTIGATIONS

Tipton, Pennsylvania (Fig. 8, 9). This fold is a northeast-plunging kink-band in the steeply dipping Devonian Brallier Formation on the northwest limb of the Nittany Arch. The shallow limb is the kink-band. the steep limb is the enveloping bedding and $\Delta < 0$. As shown in Figure 9, there are more joints, unclassified fractures, extension fractures, and twist hackle (Kulander, Dean and Barton, 1977; Kulander, Barton and Dean, in review) in the kink-band than in the enveloping bedding. Berger measured 1.7 percent extension by fracturing in the kink-band. The amount of extension by fracturing is the sum of the amounts of separation of all fractures in a measured length of a single bed. Equation (1) predicts 19.5 percent extension in the kink-band.

Watts, Pennsylvania (Fig. 8). This kink-band in the Devonian Trimmers Rock Formation (Faill, written communication, 1978) dies upward in the exposure, parallel to the kink surface. In this case, the steep limb is the kink-band and $\Delta < 0$. Berger measured 3.0 percent extension by fracturing in the kink-band. Equation (1) predicts 35.8

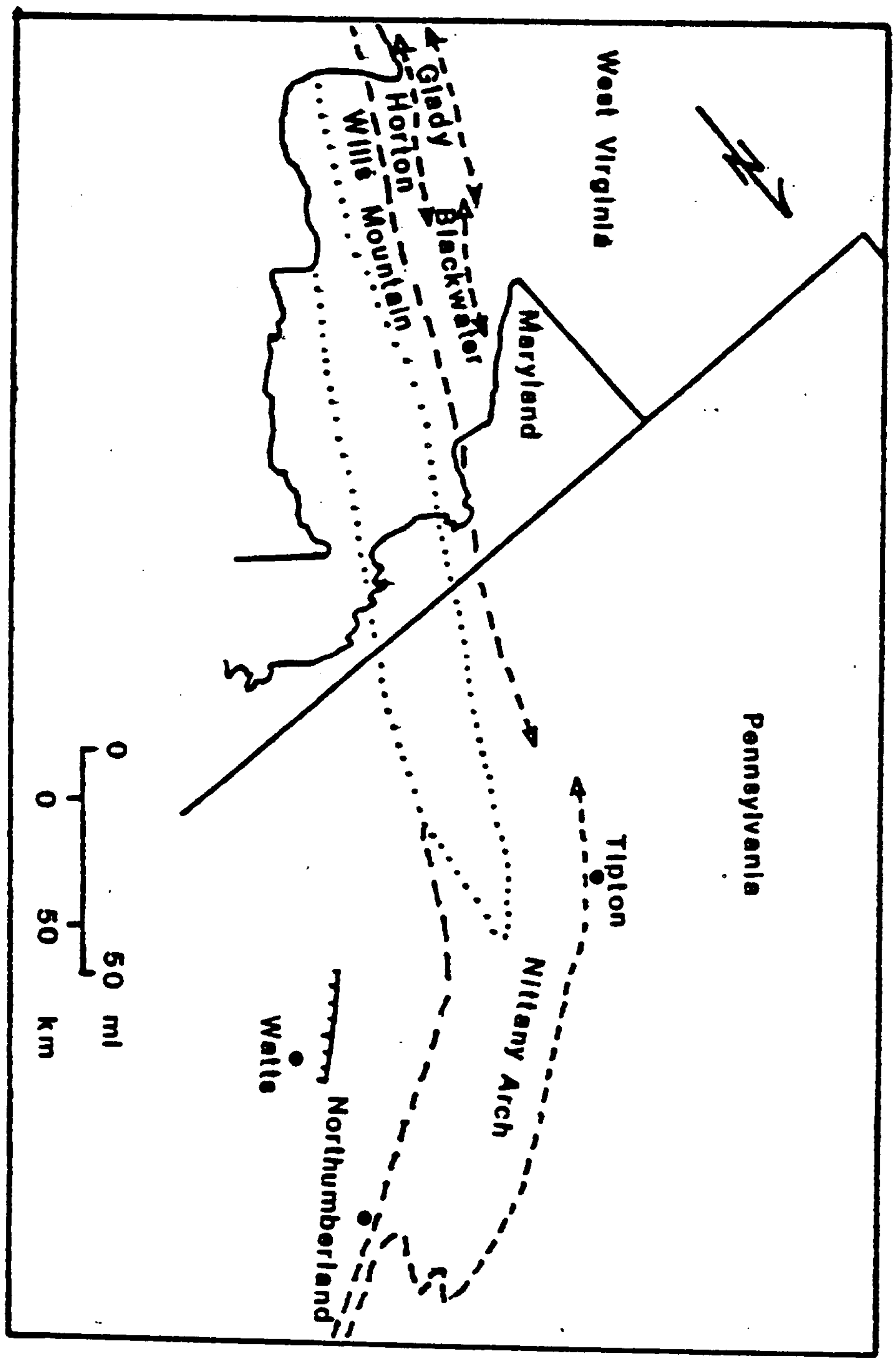

Figure 8. Location map showing exposures and significant tectonic elements. Dashed lines are anticlines, arrows denoting plunge. Dotted line marks boundary of the Broadtop synclinorium. Hachured line is a thrust fault. Circles are exposures.

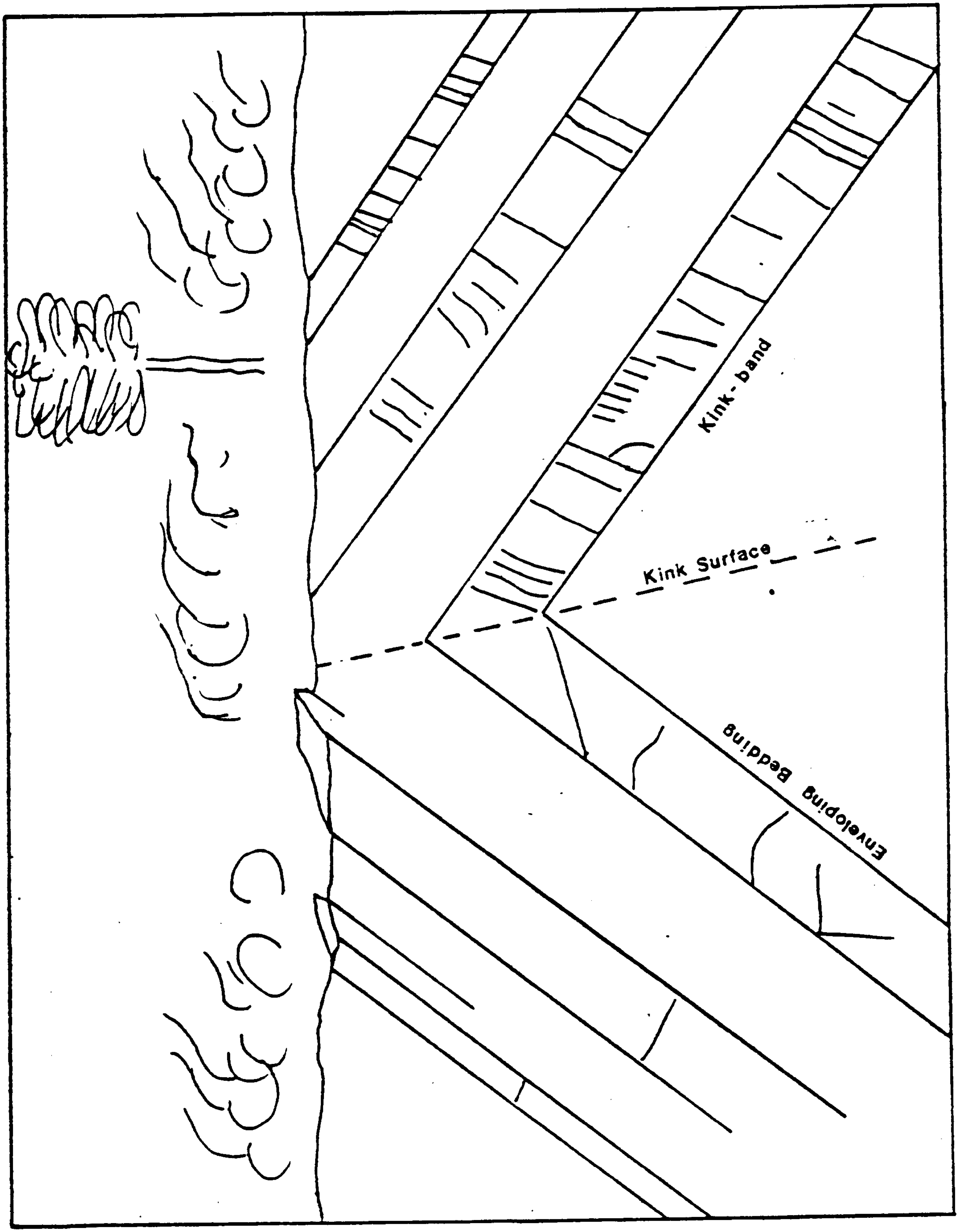

Figure 9. Outcrop at Tipton, Pennsylvania in the Devonian Brallier Formation on the northwest limb of the Nittany Arch. Siltstones show more fracturing in the kink-band than in the enveloping bedding. Schematic sketch from a photograph, looking northeast. Outcrop width about 15 m.

percent extension.

Northumberland, Pennsylvania (Fig. 8). This large kink-band in nearly horizontal beds of the Devonian Catskill Formation (Faill, written communication, 1978) has $\Delta > 0$ so that we would expect shortening and separation of beds rather than extension. However, the kink-band showed excellent extension fractures (Berger and Wheeler, ms. in review) that produced 1.6 percent extension.

The results of our field investigations indicate that extension by fracturing amounting to approximately ten percent of the predicted value is likely in kink-bands. Berger noted as predicted that beds at Tipton, Pennsylvania are more extended in the kink-band than in the enveloping bedding. This is consistent with the Ramsay (1967) model of beds undergoing extension in the kink-band, if $\Delta < 0$. Results more consistent with predicted values might be found in thinner bedded units than we examined, because thicker units are extended a lesser amount, proportionally to their thickness (Ramsay, 1967, p. 451).

MAP-SCALE FOLDING

Anticlines in the central Appalachians can grow by crestward flowage of shale and mudstone units (Wheeler, 1975; Perry and de Witt, 1977; Kulander and Dean, 1978). For more brittle units, Berger and Wheeler (in preparation) show that extension faults can thin and extend beds on the steep limbs of asymmetrical anticlines, providing a significant contribution to anticlinal growth. The presence of gravity lows on the northwest (steeper dipping) limbs of the Wills Mountain, Blackwater, Browns Mountain and Glady anticlines indicates to us one or both of (1) flowage of low density shale from low on the steep limbs to high on those limbs, or (2) thickened or repeated shale sections due to (a) detachment-related, pre-folding thrusting (Kulander and Dean, 1978; stage I of Berger and others, in press), or to (b) vertically dipping low density shales (B. Kulander, oral communication, 1978).

We would expect gravity lows that are detachment-related to extend parallel to the anticlinal axis until the detachment surface changes level along a transverse step. Thus the relatively long, linear gravity lows on the northwest limbs of the Wills Mountain and Browns Mountain anticlines (Kulander and Dean, 1978, plate 2) are probably more the result of fault repetition and thickening of the Ordovician Martinsburg Formation and Devonian shales than to lobate crestward flow of ductile rocks (B. Kulander, oral communication, 1978). In contrast, the circular lows on the northwest limbs of the Blackwater and Glady anticlines are more likely due to lobes of uplimb flowage in the ductile units of the Middle and Upper Devonian clastic sequence. On the northwest limb of the Wills

Mountain anticline, the discontinuous strike-parallel linear gravity lows terminate within a few kilometers along strike of the discontinuous strike-parallel linear belts of thinned Devonian Brallier Formation mapped by Dennison and Naegele (1963). The gravity lows are caused by thickening of Ordovician Martinsburg formation and the vertical beds of the Middle and Upper Devonian clastic sequence and the thinned outcrop belt is caused by extension and other tectonic thinning of the Devonian Brallier formation (Berger and Wheeler, in preparation; Dennison and Naegele, 1963). On the Blackwater and Glady anticlines the inferred thickened shale lobes suggest that there exists a thinned Brallier formation lower on the limb that is subjected to extension faulting (Berger, Perry and Wheeler, in press) in the brittle units and flowage in the ductile units. This thinning would be similar to that observed in outcrop in the Devonian Brallier Formation (Dennison and Naegle, 1963; Berger and Wheeler, in preparation) and could indicate gas potentially in a fractured reservoir. The source beds could be the Devonian black shales (the Marcellus and Harrell formations) and the flowed, thickened, perhaps heavily slickensided shale updip could seal the reservoir. We suggest further exploration for thinned and fractured shales down-dip of the buried, thickened shales that we infer to lie beneath the observed gravity lows (Fig. 10). For steep northwestward dips, the suggested exploration target would lia roughly under the northwestern boundary of a circular gravity low. Additional gravity measurements and modeling of the gravity lows mapped by Kulander and Dean (1978), guided by construction of balanced cross sections (Dahlstrom, 1969b; Perry, 1971), should define those targets more exactly.

The kink fold model of folding also implies that thinning by extension can occur on the northwest (steep) limb (Figure 2B). Folds thickened near the crest and thinned lower on the limb can still have planar limbs and narrow hinges consistent with a kink-fold model of map-scale folds.

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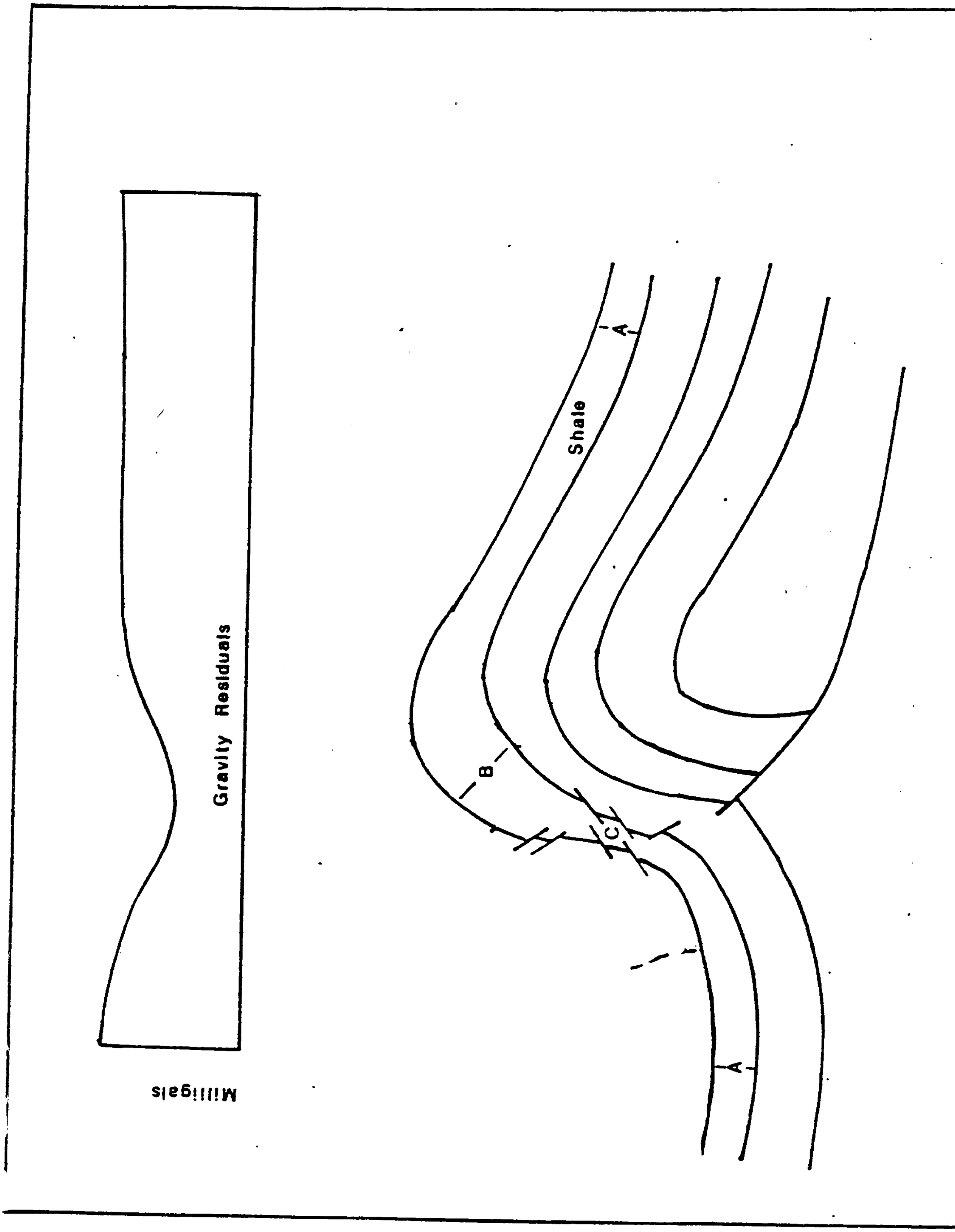

Figure 10. Relationship of gravity low to anticline. A. Original thickness. B. Siltstone and shale thickened by uplimb flow of shale. C. Siltstone and shale thinned by uplimb flow of shale and thinning by extension faulting of siltstones and shales.

CONCLUSIONS OF ENTIRE THESIS

1) Outcrop-scale and map-scale faults that formed during overall horizontal shortening and associated detached folding can be placed in a three-stage relative time sequence.

2) Wedge-shaped contraction faults form early (stage I) when beds are nearly horizontal. These faults may have associated anticlines and may be smaller versions of the detachment and ramp systems characteristic of allochthonous fold belts. In beds later rotated to steep dips by folding, contraction faults will show normal fault separation at a low angle to bedding. In cratonward-facing folds, contraction faults will show a downlimb sense of displacement.

3) Uplimb thrusts form during folding (stage II) and record relative reverse movement toward the anticlinal hinge. These faults serve to raise the limbs of anticlines to solve space problems.

4) Extension faults and extension fractures form late in folding (stage III) when beds rotate so far towards the vertical that the effect of vertical extension exceeds that of horizontal shortening. Extension faults are reverse faults formed at high angles to beds and rotated

little or not at all. Unequivocal extension fractures are lenticular and form normal to beds.

5) Extension fracturing will begin when bed dips reach 45 degrees assuming that there is no more than $\pm .1(e_3)$ hinge line parallel strain.

6) Extension faults and associated fractures provide a significant contribution to: a) anticlinal growth, b) unit thinning, and c) bed extension thus favoring formation of fractured gas reservoirs.

7) The Dawson, Cresaptown, and Kittlelick thrusts of Dennison and Naegle (1963) are more likely zones of extension faulting than single detachment-related splay faults. The Hyndman fault is more likely a splay fault but has associated extension faults.

8) The Brallier Formation is the Devonian unit most intensely deformed by extension faulting and extension fracturing. Given a gas source (the underlying Marcellus and Harrell Formations) and a seal, it may be a shallow gas exploration target on the northwest limbs of the Glady, Horton, Blackwater and Deer Park anticlines.

9) Limbs of kink folds and beds in kink bands can undergo extension parallel to the beds when the angle between the kink surface and the Kink-band is less than the angle between the kink surface and the enveloping beds.

10) Although kink-bands show evidence of more extension than do the enveloping beds, the amount of extension measured is less than the amount predicted. Only about 10 percent of the predicted volume change is translated into extension fractures.

11) Circular gravity lows on the northwest limbs of the Blackwater and Glady anticlines can be due to lobate thickening of shales high on the limbs and thinning low on the limbs. Thinning can occur in the Brallier formation similarly to that measured on the northwest limb of the Wills Mountain anticline by Dennison and Naegele (1963). We suggest exploration under the northwestern boundaries of the mapped gravity lows to intersect the inferred thinned shale sections, which may contain gas in fractured reservoirs.

SUGGESTIONS FOR FUTURE WORK

1) In many instances, contraction faults (stage I) have associated anticlines. These form over smaller versions of the ramps with which major detachment surfaces change stratigraphic level. I have identified many contraction fault localities. These are listed in Appendix IV. Study of these faults as well as those from other localities can give insights into relative ductilities of rocks involved in ramp formation, ratios of length of ramp to height of associated anticline, initiations and terminations of ramps, terminations of associated anticlines, problems of conservation of space, and the nature of the most energy efficient structures.

2) An outcrop study of uplimb thrust faults (stage II) would probably not be feasible because these faults are rarely exposed and are usually so large so as to be unidentifiable on outcrop scale. However, I found one uplimb thrust that is classic in form, but it may also be a shear zone associated with an igneous intrusion. Petrographic work in the fault zone to determine if the cataclastic material had igneous or sedimentary affinities would answer this question. The zone in question crops out in Ryder Gap, Bath County, Virginia (Station 8 of manuscript three, Appendix I).

3) Kink folds are formed by a different process than

are chevron folds (see Manuscript four). Kink folds are products of shearing and chevron folds, of buckling. Examination of strain markers such as burrows, fossils or oolites or analysis of quartz deformation lamellae and calcite twin lamellae can yield information on the origin of the fold.

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APPENDIX I

Location of field stations: "Late-tectonic extension faulting in the central Appalachians"

Station 1 - West side of the railroad tracks, .1 km north of Pinto, Allegany County, Maryland (outcrop described in Swartz, 1923).

Station 2 - North side of the C & O Canal, 10 km south of the intersection of U. S. 40 and Woodmont Road, Woodmont, Allegany county, Maryland. Outcrop is 1 km east of Woodmont Road on Deneen Road and is described in Perry and de Witt (1977), p. 33-39.

Station 3 - North side of U. S. 40 in La Vale, Maryland. Outcrop is east of the La Vale Plaza shopping center and is described by de Witt and Dennison (1972) p. 63-68.

Station 4 - East side of U. S. 33, 4 km north of Riverton, West Virginia. Outcrop is across the road from the stream that drains the south side of Ketterman's Knob and is described in Dennison and Naegele (1963), p. 37-39.

Station 5 - North side of Holly Meadows Road, 5 km north of Parsons, West Virginia at bench mark 1579.

Station 6 - Along the north side of I-64, 8 km west of White Sulphur Springs, West Virginia. First outcrop past west bound entrance to I-64 from U. S. 60 at the bridge over Wolf Creek.

Station 7 - In an abandoned quarry, .1 km east of the intersection of U. S. 33 and W. Va. 28, Pendleton county, West Virginia.

Station 8 - Along Va. 39, Bath County, Virginia, .1 km east of the Virginia/West Virginia border.

Station 9 - West side of the railroad tracks, 3 km south of Cumberland, Maryland, .1 km north of the Amcelle Corp. plant (outcrop described in Swartz, 1923).

Station 10- West side of the railroad tracks, 3 km south of Cumberland, Maryland, .3 km south of the Allegany County Fairgrounds, Allegany County, Maryland (outcrop described by Swartz, 1923).

Station 11- North side of Teter Run, 3 km south of Circleville, West Virginia and .5 km east, up the dirt road into Teter Gap, .1 km east of the first cattle guard.

Station 12 - East side of secondary W. Va. 17, first exposure east (.8 km) of intersection with secondary W. Va. 19, Pendleton County, West Virginia.

Station 13- East side of secondary W. Va. 17, second exposure east (1 km) of intersection with secondary W. Va. 19, Pendleton County, West Virginia.

Station 14- East side of secondary W. Va. 17, third exposure east (1.1 km) of intersection with secondary W. Va. 19, Pendleton County, West Virginia.

Station 15- First outcrop on the south side of W. Va. 39, 1 km east of Huntersville, West Virginia at the spring.

Station 16- North side of W. Va. 39, 1.1 km west of Minnehaha Springs, West Virginia. Outcrop is .1 km east of the bridge over the stream flowing through Buzzard Hollow and is pictured on the cover of Cardwell (1975).

Station 17- North side of W. Va. 39, first outcrop (.3 km) west of Minnehaha Springs, West Virginia.

Station 18- along the south side of I-64, 4 km southeast of White Sulphur Springs, West Virginia. Third outcrop east of the bridge over the railroad tracks (outcrop described by King, 1973).

Station 19- Along the south side of I-64, 3.5 km east of White Sulphur Springs, West Virginia. Fourth outcrop east of the bridge over the railroad tracks.

Station 20- North side of W. Va. 28, 4 km north of Hopeville, West Virginia, .2 km east of Smoke Hole caverns (North Fork Gap). Outcrop figured in Sites (1978).

Station 21- South side of Wolf Camp Run, .3 km west of the intersection of Wolf Camp Run Road and Pa. 96. Outcrop described in de Witt and Dennison (1972), p.72-73. Outcrop is .5 km north of Madley, Pennsylvania.

Station 22- North side of U. S. 40, 1 km east of the crest of Martin Mountain, Allegheny County, Maryland.

Station 23- On the south side of the exit ramp, at the juncture of U. S. 40 and U. S. 522, on the exit ramp of U. S. 40 for eastbound traffic. Outcrop is 1 km west of Hancock, Maryland.

Station 24- West side of U. S. 33, first outcrop (1 km) north of the juncture of U. S. 33 and W. Va. 28, Pendleton County, West Virginia.

APPENDIX II

Location of field stations: 'Extension in kink-bands and on the limbs of kink folds'

Tipton, Pennsylvania - This outcrop is located on the north side of the Pennsylvania secondary road that extends from U. S. 220 northwest through Tipton, Pennsylvania. It is the first west of the town of Tipton.

Watts, Pennsylvania - This outcrop is located on the west side of U. S. 322, 3 km north of Amity Hall, Pennsylvania, .1 km north of the entrance ramp from the secondary road through Watts, Pennsylvania. this is the first entrance ramp north of Amity Hall.

Northumberland, Pennsylvania - This outcrop is located on the west side of U. S. 11, .1 km south of the bridge across the west branch of the Susquehanah River into Northumberland, Pennsylvania. The outcrop is overgrown but is identifiable by the words "Susan + Matt" spray painted on the outcrop.

Appendix III

A note on the size of extension faults.

Map-scale extension faults are large enough that body forces (gravity) become significant relative to the surface forces that form outcrop-scale extension faults. Mechanics of faulting may differ between the two scales. Accordingly, the term extension fault and its companion term contraction fault may be limited to outcrop scale structures (W. Perry, Jr., oral communication, 1978). Although the size at which gravity becomes significant is approximately 50 m (R. Wheeler, oral communication, 1978) I do not restrict usage in this thesis because I am not sure of the significance of this difference in mechanics of faulting.

A note on the stratigraphy and sedimentology of the Brallier Formation

Lundegard and others (1978) characterize the Brallier Formation as part of an overall thickening- and coarsening-upward sequence, containing megasequences from 3 to 51 m thick. They recognize six facies of the Brallier, of which three constitute 80 percent of the areal extent of the formation and contain lithologies I observed to contain extension faults and fractures. These three facies are: A) Alternating beds of siltstone, shale or mudstone; B) Bundles of siltstone and fine-grained sandstone with minor interbeds of shale or mudstone; C) Olive-gray mudstone or shale.

APPENDIX IV

Contraction Fault Localities

1) West side of the railroad tracks, .1 km north of Pinto, Allegany County, Maryland.
2) West side of the railroad tracks, 3 km south of Cumberland, Maryland, .3 km south of the Allegany County Fairgrounds, Allegany County, Maryland (two contraction faults).
3) North side of Holly Meadows Road, 5 km north of Parsons, West Virginia at bench mark 1579.
4) East side of secondary W. Va. 17, 1 km east of intersection with secondary W. Va. 19, Pendleton County, West Virginia (two contraction faults).
5) North side of W. Va. 39, first outcrop west of Minnehaha Springs, West Virginia.
6) North side of W. Va. 28, 4 km north of Hopeville, West Virginia, .5 km east of Smoke Hole Caverns (North Fork Gap).
7) West side of U. S. 33, 4 km north of Riverton, West Virginia. Outcrop is across the road from Station 4.
8) East and west sides of U. S. 322, 3 km north of Amity Hall, Watts, Pennsylvania.

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