Journal of Structural Geology, Vol. 8, Nos. 3/4, pp. 361 to 371, 1986
Printed in Great Britain

0191-8141/86 $03.00 + 0.00
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Cleavage duplexes in the Marcellus Shale of the Appalachian foreland

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(Received 1 July 1984; accepted in revised form 29 August 1985)

Abstract—Cleavage duplexes are zones of platy, spaced cleavage, that either parallel bedding or ramp through shale sections. Examples are 2–60 m thick. The cleavage within the zones is generally subperpendicular to the zone boundaries but sigmoidally dragged against floor, roof and internal thrusts. Between the well-developed floor thrust and the more obscure roof thrusts of the duplexes the shale has been shortened >30% perpendicular to cleavage. The thrusts are sharp strain discontinuities because both overlying and underlying shale is uncleaved and less deformed (<10% layer parallel shortening). Cleavage in the duplexes was initiated perpendicular to bedding by pure shear in front of a propagating tip line. Cleavage halos at the ends of associated stiff carbonate concretions are small scale models illustrating initiation of cleavage. The cleavage is a primary crenulation cleavage with dissolution of limbs indicated by clay carbon partings. Transfer of thrusting from the floor to the roof of the cleavage duplex proceeded incrementally toward the foreland, imposing the simple shear that led to sigmoidal cleavage traces as the cleavage continuously evolved. Environmental conditions for formation of cleavage duplexes during the Alleghanian Orogeny are estimated as temperature 200–250°C, and pressure 1.1–1.3 kb, based upon conodont coloration (CAI 4), fluid inclusions and restored stratigraphic thicknesses. Cleavage duplexes or similar fold duplexes have been recognized in suitable black shales of Ordovician to Carboniferous age. They are manifestations of the progressive transfer of slip from floor to roof through a disturbed zone that serves as a shear boundary between large, more internally passive, thrust sheets.

INTRODUCTION

PREVIOUS illustrations of duplexes have shown stiff stratigraphic units (carbonates and sandstones) imbricated on subsidiary thrust faults that coalesce downwards to a floor thrust and upwards to a roof thrust (Dahlstrom 1970, fig. 23; Boyer & Elliott 1982, fig. 19). The floor and roof thrusts are parallel to the bedding of ductile units such as shale. An essential feature of these duplexes is a strain discontinuity between the rocks within the duplex and those above and below the roof and floor thrusts (Boyer & Elliott 1982, p. 1200). As slip is transferred progressively from the floor thrust to the roof thrust during foreland-directed footwall imbrication, large layer-parallel shortening strains accumulate within the duplex. Natural shortening strains due only to imbricate thrust faulting, in duplexes listed by Boyer & Elliott (1982, table 1), may range from −34 to −92% (Cooper *et al.* 1983, p. 145). Additional layer-parallel shortening by penetrative deformation that thickens section has been demonstrated to yield total shortening, locally, of −123% within a duplex (Cooper *et al.* 1983). However, they also showed that the general bulk shortening of the lower duplex of the Basse Normandie duplexes was −49%, partitioned between thrusting (−22%) and internal deformation (−27%). They speculated that the cleavage front or tip line of penetrative strain propagated ahead of the imbricate thrusts, in a manner similar to that observed in some shales of the Appalachian foreland.

This paper is a description of thin zones of slight to intense rock cleavage confined between roof and floor thrusts within black shale sections. These zones of spaced, platy cleavage that record at least 30% layer-parallel shortening are named *cleavage duplexes* because they exhibit features that are characteristic of fault duplexes: (1) a thin, layer-parallel zone of shortening strain that is discontinuous with the virtually unstrained rock above and below; (2) incremental propagation toward the foreland by repeated floor thrusting and layer-parallel shortening at the tip line and (3) progressive transfer of dominant slip from the floor thrust to the roof thrust by imbricate thrusts that dip backwards and merge with the roof and floor. A cleavage duplex is dominated by the penetrative layer-parallel shortening strain component (as expressed by rock cleavage) but includes imbricate thrusts that sequentially transfer shear from the floor to the roof during the propagation of the zone. It is unique among duplexes in that imbricate thrusting (ramping) through the cleavage duplex occurs as a consequence of the change in rock properties induced by the creation of the cleavage. Cleaved shale within the duplex behaves like a stiff stratigraphic unit while the uncleaved shale above and below retains properties typical of the bed-parallel decollement horizons that constitute the treads in the well-known flat-ramp trajectories of foreland thrusts.

REGIONAL SETTING

Stratigraphically bounded zones of intense strain occur within the Ordovician–Lower Carboniferous section of the middle Appalachians. These enigmatic and commonly poorly exposed zones were initially called strain discontinuities (Nickelsen 1981) or disturbed zones (Pohn & Purdy 1982). Examples have been described by Wheeler (1978), Bosworth (1984), Pierce

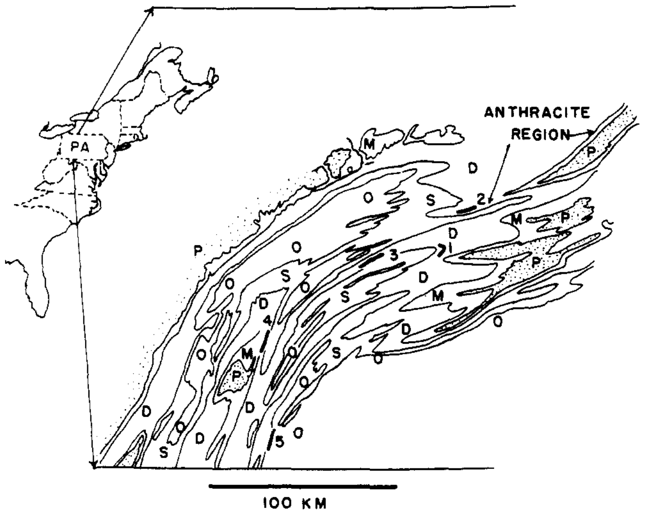

Fig. 1. Map of the Appalachian foreland (Valley and Ridge Province) in Pennsylvania. 1, 2, 3, 4 and 5 are cleavage duplex localities. 1, Selinsgrove Junction, Marcellus Formation; 2, Mausdale, Harrell Formation; 3, Bannerville, Harrell Formation; 4, Mapleton Depot, Marcellus Formation; 5, McConnellsburg, Reedsville Shale; O, Ordovician; S, Silurian; D, Devonian; M, Mississippian and P, Pennsylvanian.

(1966) and Kepferle *et al.* (1981). All are in black carbonaceous shales. They are manifested as a strain disharmony, expressed as either spaced cleavage or small-scale folding, that is sharply discontinuous with overlying and underlying less strained rock. The best exposure of these features is in the Middle Devonian Marcellus Shale between major thrust complexes below and above (Fig. 1, locality 1). The underlying blind Cambro-Ordovician fault duplex has been illustrated by Perry (1978, fig. 10) and Boyer & Elliott (1982, fig. 26), whereas the overlying imbricate fan in the Carboniferous of the Anthracite Region has been described by Wood *et al.* (1969) and Wood & Bergin (1970).

Cleavage duplexes are smaller scale structures. In the Marcellus Shale at Selinsgrove Junction, Pennsylvania, two are exposed, a thin (<2 m) and a thick (60 m) cleavage duplex. The thin cleavage duplex can be seen at three outcrops on the limbs and crest of a second-order

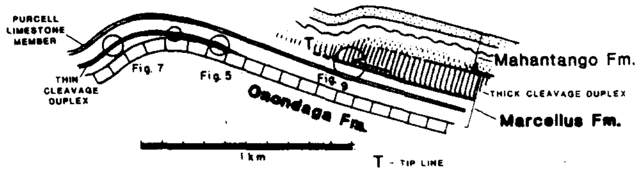

Fig. 2. Interpretative diagram of Selinsgrove Junction second order anticline, showing the stratigraphic setting of the thin and the thick cleavage duplexes and the locations of other figures.

anticline. It serves as a model for interpreting the thick cleavage duplex that is exposed at the tip line of a thrust on the southeast limb.

A geologic map and sections through the Selinsgrove Junction area were published by Nickelsen & Cotter (1983, pp. 144–145). Generalized structural relations of both the thin and the thick cleavage duplexes are shown in Fig. 2, while their stratigraphic position and associated strain gradients and strain disharmonies appear in Fig. 3. Localities to be discussed are circled in Fig. 2.

THIN CLEAVAGE DUPLEX

On the southeast limb the thin cleavage duplex cuts through an outcrop of limestone and shale that, in general, shows little penetrative deformation (Fig. 4). The cleavage duplex is in the Marcellus Shale 10 m above the Onondaga Limestone. (See Faill & Wells 1974, for stratigraphic descriptions.) The argillaceous Onondaga Limestone contains widely spaced (20 cm) planes of dissolution cleavage as well as intra-bed zones of gash fractures that follow wedge (thrust) faults. However, total layer-parallel shortening is less than 5%. Thin sections of the Marcellus Shale below the duplex show detrital phyllosilicates preserved in bedding-parallel orientations that are not cut or disrupted in any way by rock cleavage. In contrast, the 0.5 to 2 m thick cleavage duplex is identified by transverse, sigmoidally-shaped, space cleavage (Fig. 4). The zone of cleavage is followed by thrust faults along which two layers of carbonate

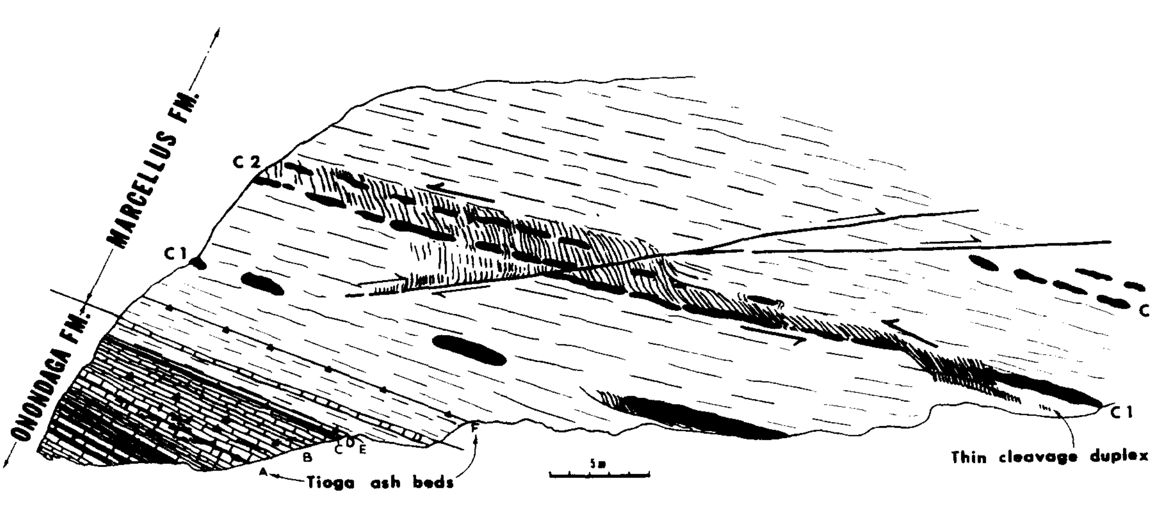

Fig. 4. Thin cleavage duplex on the southeast limb of the Selinsgrove Junction second order anticline. C1 and C2 are horizons of carbonate nodules (with an associated gamma radiation peak) that are duplicated by thrusting along the cleavage duplex. A foreland-dipping thrust displaces the cleavage duplex.

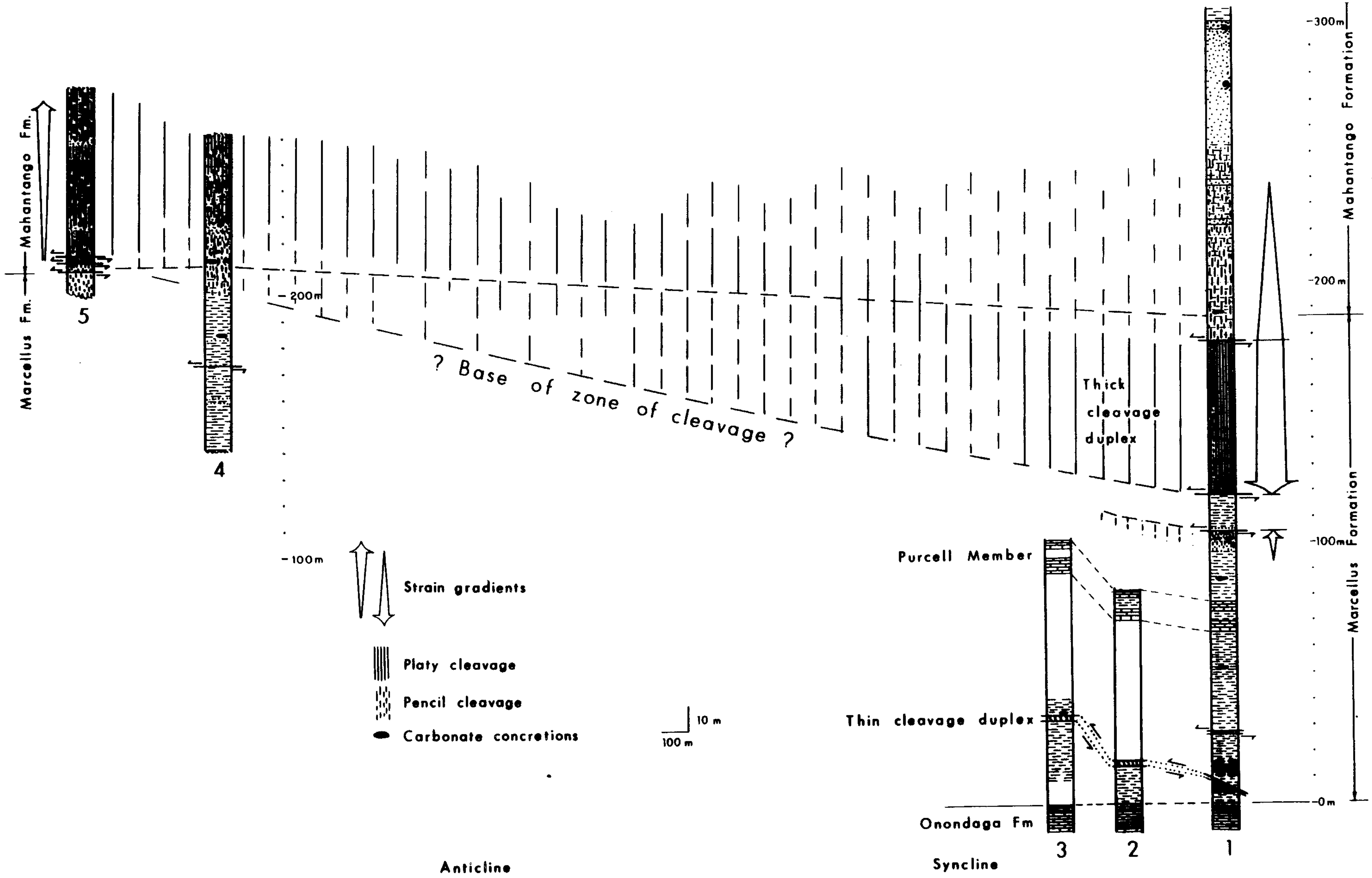

Fig. 3. Correlated stratigraphic sections showing cleavage duplexes, thrust faults, and strain gradients. Sections 1, 2 and 3 are at Selinsgrove Junction (compare Figs. 2 and 4) but sections 4 and 5 are along Route 147 approximately 5 km to the northeast. Section 4 is at a road intersection at 76°48′30″, 40°50′08″ and Section 5 is 0.4 km south of the bridge over Shamokin Creek.

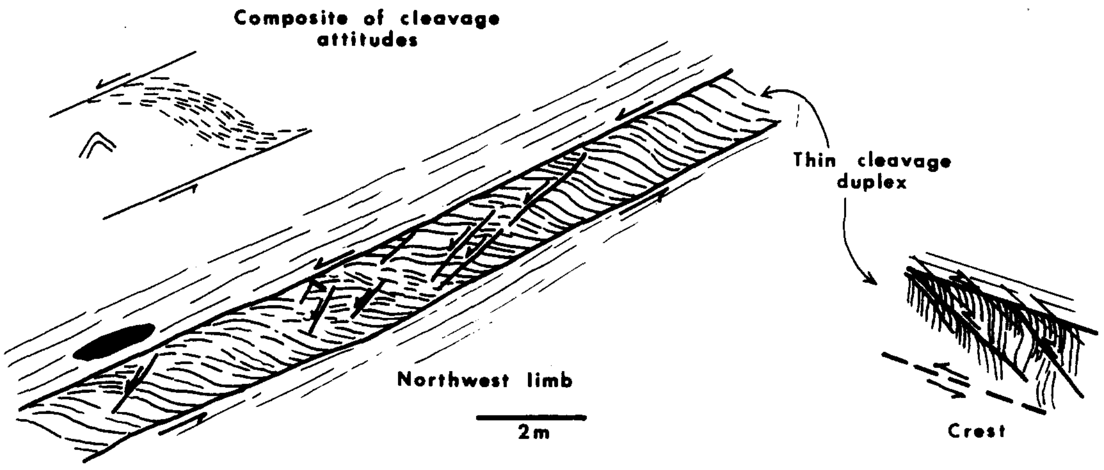

Fig. 5. Thin cleavage duplex on crest and northwest limb of Selinsgrove Junction second order anticline.

concretions (C1 and C2) have been moved 5 m of stratigraphic separation upward toward the northwest. The zone of cleavage and thrust faults is inferred to extend downward at a small angle to bedding off the lower right corner of Fig. 4 until it cuts through C1 and C2 of the lower plate. Duplication of section by thrust faulting is also proven by the repetition of the characteristic gamma radiation profile for the lower Marcellus Shale of the Appalachian Basin (Nickelsen & Cotter 1983, p. 149). Slickensided and slickenlined fault surfaces occur at the base (floor) and the top (roof) of the cleavage duplex, but are best seen at the base. The sigmoidal cleavage within the duplex continued evolution as it was dragged against the thrusts at the floor and roof as the whole zone underwent stepped differential northwestern transport. The thrusts at the floor and the roof are strain discontinuities separating uncleaved rock above and below from the spaced platy cleavage within the duplex. In these calcareous shales and carbonate concretions the cleavage is a primary crenulation cleavage consisting of B domains of minute kink folds alternating with A domains of residual carbon and reoriented phyllosilicates. Wave lengths of crenulations and spacing of A domains is 0.03–0.15 mm. B domains are crenulations of formerly bed-parallel carbonate grains, whereas A domains are residues of pressure solution.

In parts of the outcrop where the cleavage duplex impinges upon stiff carbonate concretions, cleavage halos are formed at the NW and SE end of concretions (Figs. 10 and 11). (See below for discussion of cleavage halos.) A small foreland-dipping thrust fault cuts the outcrop, slightly displacing the cleavage duplex, and forming its own cleavage.

The thin cleavage duplex on the anticlinal crest is shown in Fig. 5. Here the spaced cleavage is cut by several thrust faults that transfer slip from the floor to the roof of the cleavage duplex and it is overprinted by a second crenulation cleavage that cuts through the roof thrust and into overlying shales.

On the northwest limb, the same cleavage duplex dips parallel to bedding, but has climbed section to 33 m above the Onondaga Limestone (Fig. 5). The distance of the cleavage duplex above the Onondaga Limestone as well as an associated gamma radiation peak indicate that the stratigraphic section has been repeated by thrusting along the cleavage duplex. Within the duplex the sigmoidal shape of the spaced cleavage has been summarized by measuring 50 cleavage attitudes and plotting them in correct relative position between the floor and roof as a composite cleavage attitude. The spaced cleavage does not curve asymptotically into the bounding faults, but rather abruptly terminates against uncleaved shale at the floor and roof as shown in Fig. 6, a drawing of a thin section of cleavage and bedding at the roof. Lenses of fine-grained granular quartz of probable tectonic origin mark the contact. The strain discontinuity at the floor of the cleavage duplex is clearly a thrust fault but here at the roof the abrupt contact cannot be distinguished as a slickensided fault surface. Within the duplex there is a single strong microscopic fabric of phyllosilicates parallel to the mesoscopic cleavage (Fig. 6). In the absence of intersecting S-planes it is difficult to trace the origin or estimate the strain associated with this structure. It could be transposed bedding.

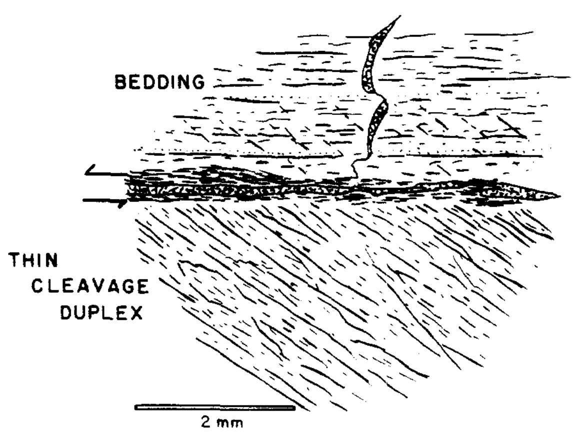

Fig. 6. Drawing of thin section across the roof of the thin cleavage duplex illustrated in Fig. 5. Granular quartz veins are parallel to the contact between bedding and the duplex and also cut bedding. Prominent *S*-planes are bedding, and, within duplex, spaced cleavage or transposed bedding.

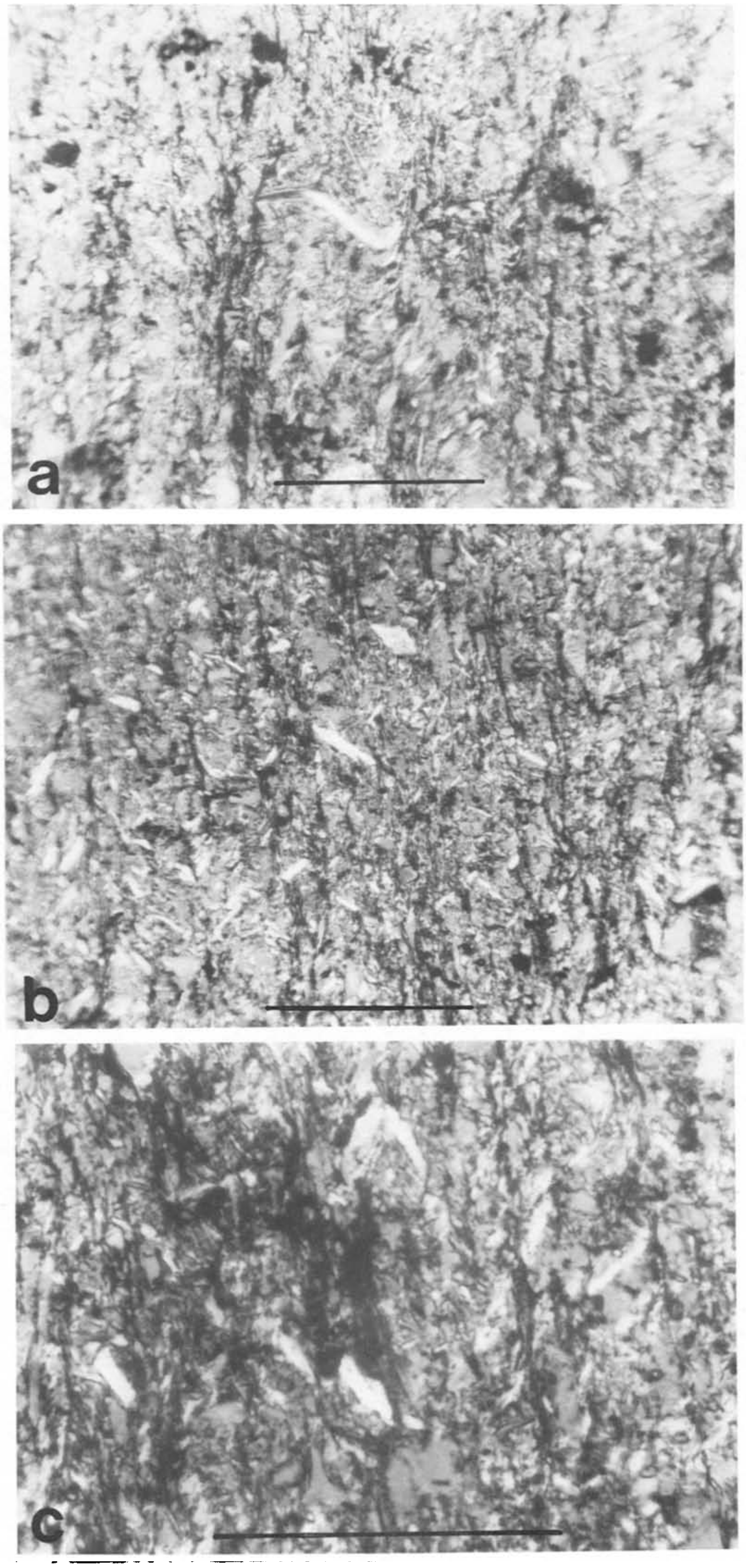

Fig. 7. Crenulated and rotated phyllosilicates from the thick cleavage duplex showing different amounts of rotation from their original bed-parallel orientation. All scale bars are 0.1 mm long. Bedding is E–W, spaced cleavage N–S. (a) Crenulated mica, layer-parallel shortening 13%. (b) Phyllosilicates rotated 40°, layer-parallel shortening 24%. (c) Phyllosilicates rotated 55–60°, one mica above the center has been sharply bent; layer-parallel shortening 40%.

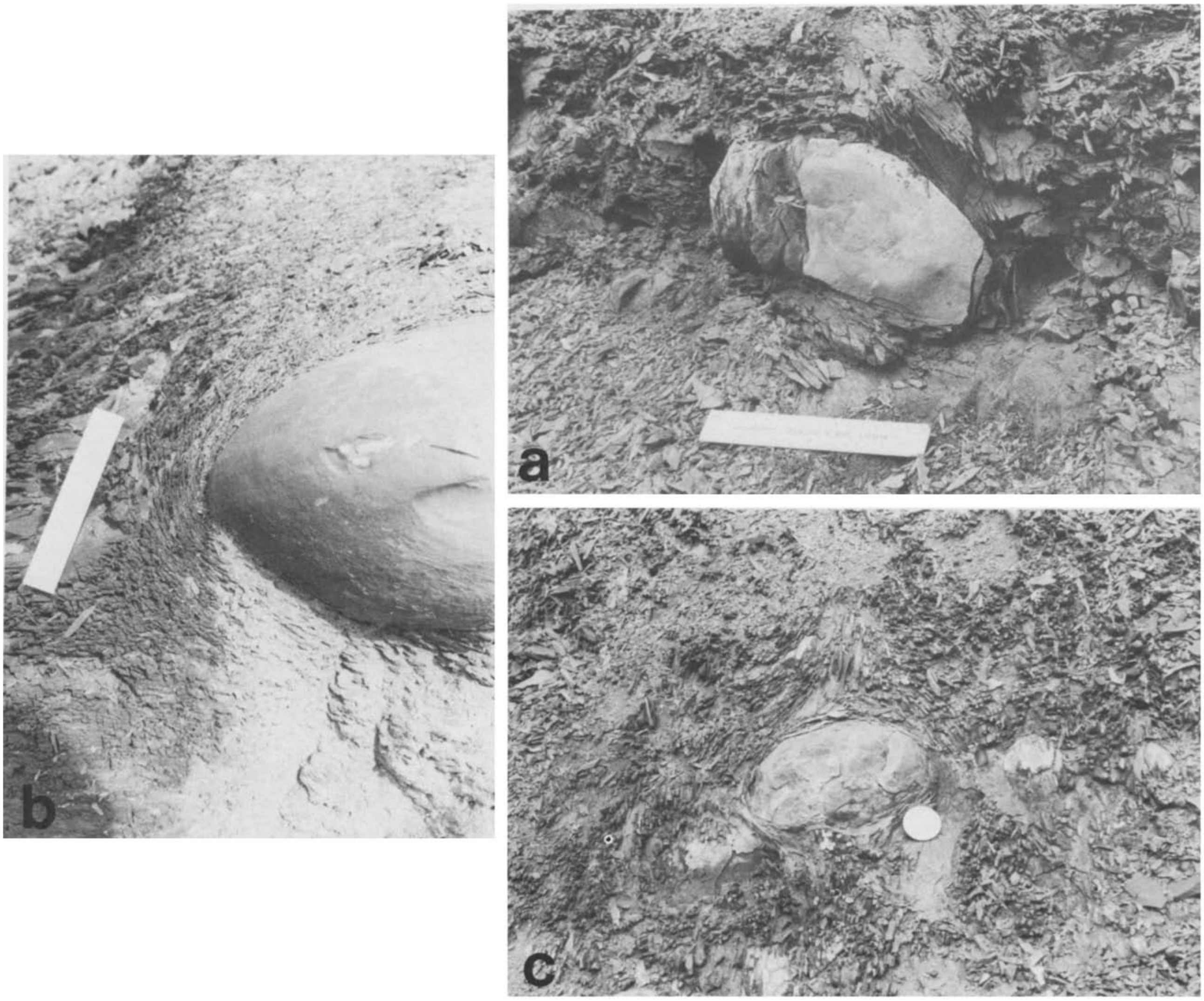

Fig. 11. Photographs of three types of cleavage halos around carbonate concretions. (a) Concretion with kink band or vertical cleavage zone, Type II scale 15 cm. Photographed in center of Fig. 12. (b) Concretion with platy cleavage zones at tips. Surrounding shale has pencil cleavage, Type III, scale 15 cm. Photographed at 210 m in Column 4, Fig. 3. (c) Concretion distorts pervasive spaced platy cleavage, Type IV, coin 23 mm. Photographed at lower right of Fig. 12.

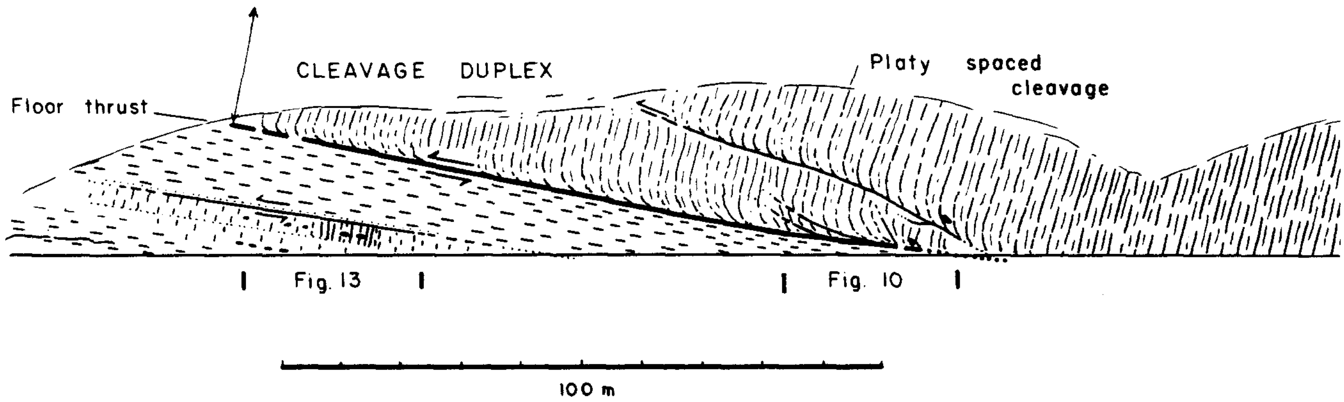

Fig. 8. Lower part of the thick cleavage duplex, showing floor thrust, cleavage drag and imbricate thrusting that displaces cleavage. Bedding is preserved for 12 m below the floor thrust, but is underlain by another minor thrust and strain gradient.

Enigmatic extensional faults originally not thought to be related to the formation of the cleavage duplex, dip more steeply to the north (Fig. 5).

One reviewer suggested that they may represent late ramping back down to the floor thrust as a consequence of sticking on the roof thrust. Since the relative slip sense indicated by the sigmoidal cleavage remains constant across the anticline, it is inferred that thrust transport towards the northwest preceded folding or occurred early in the folding history.

THICK CLEAVAGE DUPLEX

The thick cleavage duplex occurs in the upper Marcellus Shale, one hundred meters above the thin cleavage duplex (Figs. 2 and 3). Its floor is a thrust fault that places shale with a well-developed spaced cleavage upon shale containing only a bedding fabric (Fig. 8). The floor thrust is parallel to the underlying bedding and is a zone of sheared shale containing euhedral crystals and cataclastic fragments of quartz that were originally precipitated in gash veins or fault-step pressure shadows and then later milled and broken during continuing fault movement. Above the floor thrust the well-developed spaced cleavage, extending upward for at least 60 m, completely obliterates bedding (Fig. 3, Section 1).

In thin sections, phyllosilicates that were parallel to bedding have been crenulated (B domains) and on the limbs of crenulations, spaced 0.02–0.06 mm, there are residual accumulations of clay and carbon (A domains). In a few places >30% longitudinal shortening normal to cleavage is estimated by unfolding phyllosilicates, but no estimate can be made of the additional shortening associated with pressure solution. Figure 7 shows crenulated or rotated phyllosilicates from three different thin sections of the thick cleavage duplex. Bedding is E–W and spaced cleavage, marked by clay–carbon partings, is N–S. Detrital micas that were originally parallel to bedding have been rotated through an angle up to 60° (Fig. 7c) resulting in layer-parallel shortening from −13% (Fig. 7a) to −24% (Fig. 7b) to −42% (Fig. 7c). (Layer-parallel shortening strain equals cos ∡ of rotation − 1). This cleavage is generally perpendicular to the bedding that is preserved above and below the duplex and also to the faults that bound the duplex, but within 4 m of the floor thrust it has been rotated counter clockwise through an angle of 60° toward the thrust (Figs. 8 and 9).

Imbricate thrusts and kink bands rise off the floor thrust and cut or distort the spaced cleavage within the duplex (Fig. 9). The superposition of cleaved upon uncleaved Marcellus Shale suggests that cleavage that was initiated earlier, probably by pure shear perpendicular to bedding, continued to evolve in attitude and aspect during deformation by simple shear near the base of the thick cleavage duplex. Progressive deformation is thus demonstrated by the rotation of cleavage against the floor thrust and by the later imbricate thrusts and kink bands that rise from the floor thrust and cut the cleavage (Figs. 8 and 9).

In contrast with its base, the top of the thick cleavage duplex is defined by a gradient of decreasing strain in which platy cleavage changes upward to pencil cleavage through a stratigraphic interval of 80–100 m (Fig. 3, Section 1). Most of the change in cleavage type occurs above 60 m. Reks & Gray (1982) have shown elsewhere in the Valley and Ridge Province that pencil cleavage is associated with >9 to <26% layer parallel shortening, which agrees with my estimate. Consequently the change in cleavage type probably means an upward decrease in layer-parallel shortening strain from >30% to <10%. At one place, 60 m above the floor thrust, there is a strain discontinuity at a small thrust which is manifested as an abrupt upward change in the quality of cleavage—platy to pencil. This small thrust has been

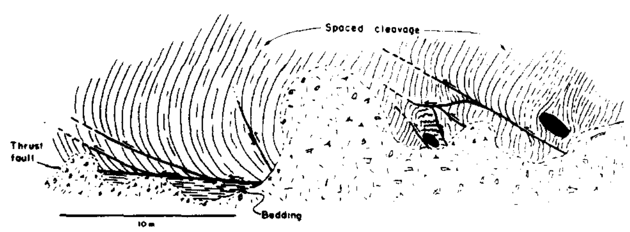

Fig. 9. Detail of cleavage drag against the floor thrust of the thick cleavage duplex and of imbricate thrusting that drags or displaces cleavage. See Fig. 8 for location.

marked on Figs. 2 and 3 (Section 1, 180 m) as the roof thrust because it defines the top of the 60 m zone of well-developed platy spaced cleavage. But there is no observed change in cleavage attitude due to fault drag against this roof thrust, so it is less important than the floor thrust, serving only to separate zones of different amounts of layer-parallel shortening strain. The strain gradient above the roof thrust occurs through a stratigraphic interval of changing rock type as the Marcellus Shale grades into the overlying, more sandy, Mahantango Formation. In the absence of strain markers it is difficult to assess the role that changing lithology may play in affecting the appearance of cleavage. Is it possible that the presumed upward decreasing strain suggested by a change in cleavage type may be a manifestation of the gradually increasing sand content of the coarsening-upward Mahantango deltaic cycle? An indirect answer to this question is provided at another outcrop 3700 m to the northwest where the transition from the Marcellus Shale to the Mahantango Formation is well-exposed and the polarity of the cleavage (and strain?) gradient is reversed (Fig. 3, Section 4). Here the gradient of change in cleavage types is directed upward and it is possible to walk from uncleaved Marcellus Shale upward into Marcellus Shale with pencil cleavage and cleavage halos at the tips of carbonate concretions. In the sandy Mahantango Formation above the Marcellus Shale the cleavage is platy and better developed than in the shale below. This gradual improvement in the quality of cleavage upward into the Mahantango which occurs despite an increase in sand content suggests that the quality of cleavage is controlled by factors other than lithology. In Fig. 3, Section 5 there is an abrupt change in cleavage type from pencil cleavage below to platy cleavage above which can be correlated with a zone containing at least three small thrusts that were not observed in Section 4. At each of the thrusts there are slight changes in cleavage attitude. In Section 5 the distribution of cleavage types and associated strains is more like that in the cleavage duplex observed to the southeast. My interpretation is that the good cleavage in the upper part of Sections 4 and 5 is in front of the tip line of the thick cleavage duplex. It is in a 'channel' of layer-parallel shortening strain that would have been accreted to the thick cleavage duplex if that duplex had continued propagating toward the northwest.

In the interbedded sandstone and shale and massive sandstone higher in the Mahantango Formation, layer-parallel shortening is accommodated by 3rd and 4th order folds, intrabed wedge faults, and wrench faults, as well as by pressure-solution cleavage (Figs. 2 and 3). Finite strain cannot be estimated but is probably less than that below in the Marcellus Shale. Above and behind the tip line of the thick cleavage duplex there is a fold in the Mahantango Formation that results from layer-parallel shortening and vertical thickening behind the tip line of the cleavage duplex in the underlying Marcellus Shale (Fig. 2). This fold also compensates for differences in layer-parallel shortening between the Mahantango and the Marcellus Formations.

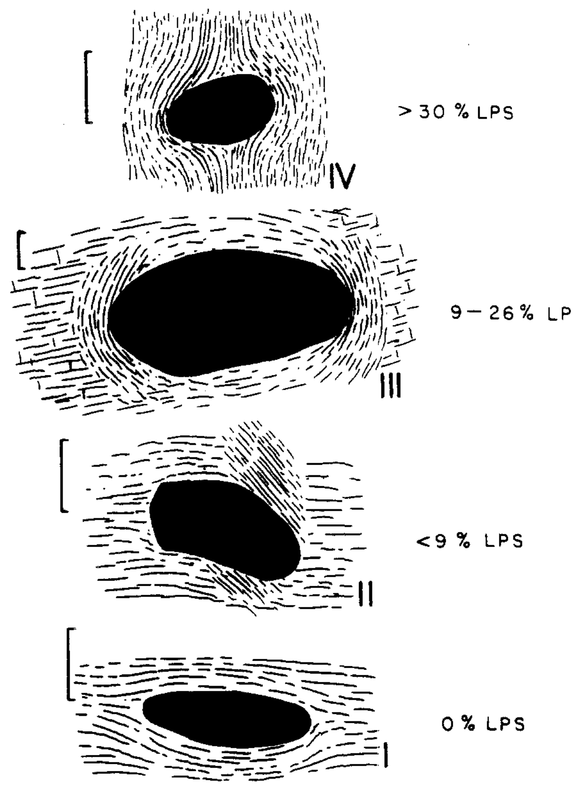

Fig. 10. Cleavage halo types I–IV and associated LPS (layer-parallel shortening). Scale bars are 10 cm long.

STRAIN GRADIENTS AND DISCONTINUITIES RELATED TO THRUST FAULTS AND CLEAVAGE DUPLEXES

Rough estimates of layer parallel shortening strain can be determined from crenulated phyllosilicates (Fig. 7), whereas cleavage type (platy or pencil) and cleavage halos around carbonate concretions provide additional evidence of relative strain. Stiff carbonate concretions remain undistorted during layer parallel shortening while surrounding shales deform and become cleaved. The ductility contrast between shales and carbonate concretions first appears as strain concentrations in shales at the tip of concretions, manifested as local zones of spaced cleavage called cleavage halos. Four classes of relationships between carbonate concretions and the surrounding either deformed or undeformed shales have been recognized (Figs. 10 and 11). Class I relationships show compaction but no layer-parallel shortening strain effects at the tips of concretions (Fig. 10). The Class II relation between concretion and shale is a kink band of bedding or a cleavage zone extending vertically away from one tip of a concretion (Figs. 10 and 11a). Concretions may appear to have rotated in the shale matrix but there is no associated general rock cleavage, and layer-parallel shortening strain is thought to be less than 9%.

Class III cleavage halos are platy cleavage zones at both tips of the concretions, which are surrounded by shales with pencil cleavage (Figs. 10 and 11b). Layer-parallel shortening is estimated to range from >9% to <26%.

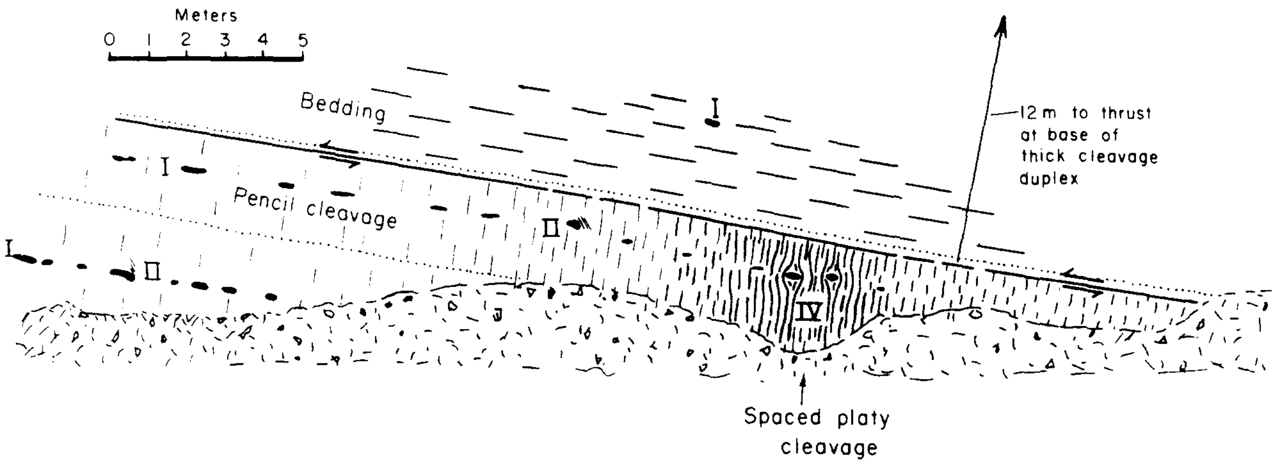

Fig. 12. Drawing of a minor thrust and strain gradients with associated Type I, II and IV cleavage halos and spaced platy to pencil cleavage. See Fig. 8 for location.

Class IV cleavage halos are found in shales where bedding has been obliterated by a platy spaced cleavage. Cleavage is bent around concretions as shown in Figs. 10 and 11(c). Layer-parallel shortening associated with this cleavage is >30%.

Examples illustrated in Figs. 3 and 12 show strain gradients and discontinuities, related to thrust faults and cleavage duplexes, that have been established by cleavage type and cleavage halos.

The first example, at approximately 100 m in Section 1, Fig. 3 is an upward-increasing strain gradient which is abruptly terminated against a minor thrust fault. Figure 12, a drawing of the outcrop, shows concretions with the class of cleavage halo identified by numerals I–IV. Spaced platy cleavage is found with Class IV concretions (Fig. 11c) but this cleavage terminates upward at a thrust fault above which the rock is uncleaved and contains isolated Class I concretions (Fig. 10). This is evidence of an abrupt drop in layer-parallel shortening from >30% to <9% at the thrust fault. Strain also decreases in a gradient toward the left or northwest in Fig. 12 so that the cleavage becomes a pencil cleavage and Class II and I concretions are found. It appears that the thrust is the roof fault at the tip line of a cleavage duplex that is dying out to the left or northwest. The second example, at approximately 200 m in Section 4, Fig. 3 was described previously as a strain gradient increasing upward. The change from no cleavage, to pencil cleavage near the top of the Marcellus Shale, to platy cleavage in the overlying Mahantango Formation is paralleled by observations on cleavage halos. At 180 m in the section no cleavage is present in the Marcellus Shale and a Class I concretion shows no strain effects. Pencil cleavage begins to appear at 200 m and at 210 m there is a zone of Class III concretions (Fig. 10). Figure 11(b) was photographed at this locality. On the basis of cleavage halos and pencil cleavage, up to 26% layer-parallel shortening is suggested. The overlying sandy Mahantango Formation has the best platy cleavage in the section and has been interpreted above as lying in front of the tip line of the thick cleavage duplex. Between Sections 1 and 4–5 of Fig. 3 the base of the zone of cleavage inferred to lie in front of the thick cleavage duplex has risen stratigraphically from the Upper Marcellus Shale to the Mahantango Formation.

ENVIRONMENTAL CONDITIONS DURING FORMATION OF THE SELINSGROVE JUNCTION CLEAVAGE DUPLEXES

Ambient environmental conditions at the time of duplex propagation in the Alleghany Orogeny may affect whether ductile rocks deform as a cleavage duplex or as a fold duplex. The cleavage duplexes discussed in this paper occur at the southwest edge of the Pennsylvania Anthracite Region, where vitrinite reflectance (Levine 1983), coal rank (Damberger 1974, White 1925) and conodont color alteration (Epstein *et al.* 1977) all indicate higher than normal temperatures within the Appalachian foreland. Contours of all paleo-temperature indicators trend perpendicular to strike and demonstrate an increasing temperature toward the east or northeast along strike. In the cleavage duplexes, incipient recrystallization or new growth of phyllosilicates is apparent in grains that are bent on the hinges of microcrenulations.

The Onondaga Limestone beneath the thin cleavage duplex yielded conodonts with CAI index of 3.5–4 (personal communication, A. Harris 1981). Conodonts in Lower Devonian limestones nearby had CAI index of 4, and the No. 4 CAI isograd for Silurian through Devonian carbonate rocks (Harris *et al.* 1978, Sheet 2) passes close to the outcrops. CAI data thus indicate that these rocks were heated for a period in their history to 190–300°C. Maximum temperatures attained were probably in the lower half of this range, due to estimated length of time of heating (Epstein *et al.* 1977).

Fluid inclusions containing H_2O, CH_4 and minor amounts of CO_2 are present in the quartz that was tectonically deformed during thrusting at the base of the thick cleavage duplex (Fig. 8). Most water-rich, CH_4 saturated, fluid inclusions homogenized between 205 and 234°C with an extreme range of 134–296°C. In

CH_4–CO_2 inclusions that are estimated to have a composition of 95% CH_4 and 5% CO_2, temperature of freezing of CO_2 ranged from −95 to −91°C and temperature of homogenization of CH_4 ranged from −114 to −94°C. The thoroughly fractured and refractured quartz contains numerous secondary inclusions representing a more complex history than the quartz in Alpine fissures that was studied by Mullis (1979) and evaluated by Frey *et al.* (1980). Applying these data to phase diagrams provided by N. Orkan and B. Voight (personal communication 1983), ambient pressures at the time of filling of fluid inclusions are estimated to have been 1.1–1.3 kb. The mean thickness of the stratigraphic column above the cleavage duplexes at the time of Alleghanian deformation is estimated at a minimum of 7.7 km, which suggests confining pressure of 1925 bars and pore pressure, if hydrostatic, of 770 bars. The mean of confining pressure and pore pressure derived from estimated depth of burial is 1.3 kb. Temperatures that may be inferred from depth of burial assuming a geothermal gradient of 30°C km^{-1} are similar to those derived from conodont and fluid inclusions.

In summary, data from different sources indicate ambient temperatures in the 200–250°C temperature range and confining pressures of 1.1–1.3 kb during the formation of cleavage duplexes containing spaced cleavage. This agrees with Epstein (1974) who believed that a CAI index of 4–4.5 is the lower thermal threshold for slaty cleavage. The fold duplexes described by Wheeler in West Virginia (1978) are associated with CAI index <3.5. I believe that these parameters (T 200–250°C, P 1.1–1.3 kb, CAI 4) bracket the environment in which cleavage duplexes in black shales may form.

ORIGIN OF A CLEAVAGE DUPLEX

Figure 13 shows the inferred sequence of propagation of a cleavage duplex. Unlike fault duplexes (Boyer & Elliott 1982, fig. 19) cleavage duplexes propagate behind a frontal zone of cleavage that incrementally advances ahead of the tip line.

(1) The sequence is initiated by layer-parallel shortening and pure shear thickening in front of either a stiff layer or a propagating cleavage zone embedded in the ductile rock from which it is forming. As layer parallel shortening increases locally from 5% to greater than 30%, pencil cleavage and then spaced platy cleavage form and the section is locally thickened. Less ductile rocks above the cleavage zone are folded to accommodate the differential shortening and thickening as is shown in Fig. 2.

(2) Continued propagation of the cleavage zone initiates a floor thrust that follows the cleavage in the direction of transport. At this stage, although the cleavage terminates downward in a thrust as shown at 120 m in Fig. 3, Section 1, it dies out gradually upward in a gradient of diminishing strain as shown between 180 and 240 m in Fig. 3, Section 1. Propagation of cleavage ahead of the tip line initially produces strain gradients such as

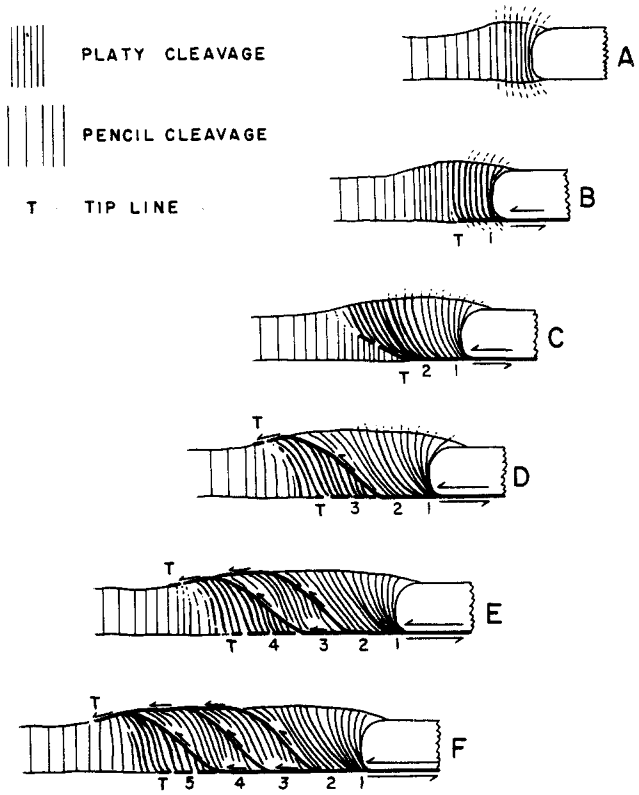

Fig. 13. Sequential development of a cleavage duplex. See text for explanation. The object at the right end of each panel (A, B, C, D, E and F) represents either the tip of a stiff layer or a propagating cleavage zone.

the vertically increasing strain in the zone above 190 m in Fig. 3, Section 4. The floor thrust of the approaching cleavage duplex eventually cuts the base of this precocious zone of cleavage, imparting simple shear and dragging the cleavage against the thrust, as is illustrated in Figs. 2, 3 and 8. In Figs. 13(b) and (c) note the propagating tip line (T) advancing left as the platy cleavage extends from 1 to 2.

(3) One and 2 may be repeated several times before imbricate thrust faults ramp through the cleavage zone from the floor thrust to the contact with the overlying uncleaved shale (Figs. 13c & d). The thrust at 180 m in Fig. 3, Section 1 is an example of weak development of the roof thrust at this stage. Tip lines now exist for both the propagating floor and roof thrust (Fig. 13d).

(4) As 1, 2 and 3 repeat, the cleavage duplex is formed with floor thrust and roof thrust bounding a zone of sigmoidal cleavage (Figs. 13e & f). The duplex has become a zone of progressive transfer of slip from floor to roof through highly strained rocks that serve as a shear boundary between large, more internally passive tectonic units, as best shown by the thin cleavage duplex in Figs. 4 and 5.

The strain concentrations that initiate cleavage duplexes may form either in front of the tip line of a thrust fault entering the section from below or in front of stiff rock types enclosed in ductile rocks. Cleavage that forms in front of the tip line of a propagating cleavage duplex is later cut by floor thrust, ramps and roof thrust (Figs. 13c–e). Formation of cleavage locally converts shale from a ductile to a stiff member of the sedimentary sequence. These zones of cleaved shale then behave like the stiff members (carbonate or sandstone) in a fault duplex. Strain concentrations related to rock-type in a horizontal sedimentary sequence undergoing uniform layer-parallel shortening may initiate thrust faulting where stiff, unyielding strata abut ductile strata. Small scale examples are cleavage halos around concretions, but regional facies changes from sandstone to shale may provide large scale examples. Both the Mahantango Formation and the Marcellus Shale have sandstone facies to the southeast that may promote stress concentrations in the Selinsgrove Junction area, leading to zones of rock cleavage at the tips of sandstone facies and eventual thrusting toward the nothwest.

Acknowledgements—Thanks to Jake Hossack for field observations and discussions that lead to the concept of cleavage duplexes and to David Egan and Ronald Miller for their careful student papers which contributed to my knowledge of the rocks at Selinsgrove Junction. The comments of an anonymous reviewer were most helpful in preparing the final manuscript. Funding for this study was provided by the donors of the Petroleum Research Fund, administered by the American Chemical Society.

REFERENCES

Bosworth, W. 1983. Foreland deformation in the Appalachian Plateau, central New York: the role of small-scale detachment structures in regional overthrusting. *J. Struct. Geol.* **6**, 73–81.

Boyer, S. E. & Elliott, D. 1982. Thrust systems. *Bull. Am. Ass. Petrol. Geol.* **66**, 1196–1230.

Cooper, M. A., Garton, M. R. & Hossack, J. R. 1983. The origin of the Basse Normandie duplex, Boulonnais, France. *J. Struct. Geol.* **5**, 139–152.

Dahlstrom, C. D. A. 1970. Structural geology in the eastern margin of the Canadian Rocky Mountains. *Bull. Can. Petrol. Geol.* **18**, 332–406.

Damberger, H. H. 1974. Coalification patterns of Pennsylvanian coal basins of the eastern United States. *Spec. Pap. geol. Soc. Am.* **153**, 53–74.

Epstein, A., Epstein, J. B. & Harris, L. D. 1977. Conodont color alteration—an index to organic metamorphism. *Prof. Pap. U.S. geol. Surv.* **995**.

Epstein, J. B. 1974. Metamorphic origin of slaty cleavage in eastern Pennsylvania. *Geol. Soc. Am. Abs. with Progr.* **6**, 724.

Faill, R. T. & Wells, R. G. 1974. Geology and mineral resources of the Millerstown quadrangle, Perry, Juniata and Snyder Counties, Pennsylvania. *Penn. geol. Surv.* Atlas 136 (4th Ser).

Frey, M., Teichmuller, M., Teichmuller, R., Mullis, J., Kunzi, B., Breitschmid, H., Gruner, U. & Schwizer, B. 1980. Very low-grade metamorphism in external parts of the Central Alps: illite crystallinity, coal rank and fluid inclusion data. *Eclog. geol. Helv.* **73**, 173–203.

Harris, A. G., Harris, L. D. & Epstein, J. B. 1978. Oil and gas data from Paleozoic rocks in the Appalachian Basin: maps for assessing hydrocarbon potential and thermal maturity (conodont color alteration isograds and overburden isopachs). *U.S. geol. Surv.*, Misc. Geol. Inv., Map I-917-E.

Kepferle, R. D., Potter, P. E. & Pryor, W. A. 1981. Stratigraphy of the Chattanooga shale (U. Dev. & L. Miss.) in vicinity of Big Stone Gap, Wise Co., VA. *U.S. geol. Surv. Bull.* **1499**, 1–20.

Levine, J. R. 1983. Tectonic history of coal-bearing sediments in eastern Pennsylvania using coal anisotropy. Ph.D. thesis, The Pennsylvania State University.

Mullis, J. 1979. The system methane–water as a geologic thermometer and barometer from the external part of the Central Alps. *Bull. Mineral.* **102**, 526–536.

Nickelsen, R. P. 1981. Rock cleavage, strain discontinuities and thrust faulting. *Geol. Soc. Am. Abs. with Progr.* **13**, 520.

Nickelsen, R. P. & Cotter, E. 1983. Silurian depositional history and Alleghanian Deformation in the Pennsylvania Valley and Ridge. *Guidebook for the 48th Annual Field Conference of Pennsylvania Geologists, Penn. geol. Surv.*, Harrisburg, PA.

Perry, W. J., Jr. 1978. Sequential deformation in the central Appalachians. *Am. J. Sci.* **278**, 518–542.

Pierce, K. L. 1966. Bedrock and surficial geology of the McConnelsburg Quadrangle, Pennsylvania. *Penn. geol. Surv.* Atlas 109a.

Pohn, H. A. & Purdy, T. L. 1982. Disturbed zones: indicators of deep-seated subsurface faults in the Valley and Ridge and Appalachian structural front of Pennsylvania. *U.S. geol. Surv.*, Open File Report 82-967.

Reks, I. & Gray, D. R. 1982. Pencil structure and strain in weakly deformed mudstone and siltstone. *J. Struct. Geol.* **4**, 161–176.

Wheeler, R. L. 1978. Slip planes from Devonian Millboro shale, Appalachian plateau province: statistical extensions of discfold analysis. *Am. J. Sci.* **278**, 495–517.

White, D. 1925. Progressive regional carbonization of coals. *Trans. Am. Inst. Min. Met. Engng* **71**, 253–281.

Wood, G. H. Jr., Trexler, J. P. & Kehn, T. M. 1969. Geology of the west—central part of the Southern Anthracite Field and adjoining areas, Pennsylvania. *Prof. Pap. U.S. geol. Surv.* **602**.

Wood, G. H. & Bergin, M. J. 1970. Structural controls of the Anthracite Region, Pennsylvania. In: *Studies of Appalachian Geology Central and Southern* (edited by Fisher, G. W., Pettijohn, F. J., Reed, J. C., Jr. & Weaver, K. N.). Interscience, New York, 147–160.