4. Strain distribution and geometry of folds

By

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Abstract.—The pattern of strain of folds produced by buckling is charcteristically different from that of folds produced by bending. It is therefore possible to distinguish between these two types of folds in the field, a circumstance which is significant for tectonic analysis since buckling folds and bending folds are results of quite unlike systems of stresses.

It is well known that parallel folds can only occur through a limited thickness of stratified complexes. Experiments show that, if the thickness of a stratified complex is greater than this limit, the style of folding varies continuously from concentric in the outer layers to similar or chevron-type in the central part of the complex.

The detailed pattern of finite strain within the layers of buckled stratified bodies may be demonstrated experimentally by markers drawn on the surface of bodies consisting of alternating layers with unlike competency (i.e. rigidity or viscosity). The geometry of deformation of such markers during buckling of the multilayer may then be compared with the attitude of various structural features in similarly buckled rocks. Such comparison shows, for example, that schistosity in general is normal to the direction of maximum finite compressive strain rather than parallel to the directions of maximum shear within each individual layer.

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All types of folding of layered rocks are characterized by periodically varying motion across originally more or less straight planar or linear structures. Within this element of common character it is, however, room for a number of properties by which unlike classes of folds may be distinguished.

For the purpose of discussion in this paper folds in layered rocks may be classified in two mechanically distinct groups, viz.:

(*a*) Folds whose periodically varying transversal displacement is a secondary effect of compression parallel to layering. In order that layer-parallel compression shall result in folding it is necessary that the layers display unlike mechanical properties, or that the rocks are mechanically anisotropic in the sense that they shear more easily along directions quasi-parallel to compression than in other directions.

(*b*) Folds whose periodic or heterogeneous transversal motion represents a heterogeneous strain not generated by compression along layers with contrasted mechanical properties. Anisotropic or contrasted mechanical properties of the rocks are not necessary, and originally planar or linear markers are important only in order to show the sinuous strain.

Folds of the first kind form for similar reasons that slender columns and thin plates in strength-of-material theory collapse when the longitudinal compressive force reaches a critical value. An account of this phenomenon may be found in most texts on strength-of-material theory, see for example TIMOSHENKO (1960). The process is often called *buckling*, and the terms *buckling folds* or *buckle folds* have been used (HILLS, 1953, p. 89) for this kind of folds in rocks.

When the primary force which gives rise to curvature in plates and beams has a significant component normal to the planar or linear body, strength-of-material scientists classify the deformation as *bending*. The second class of folds above is accordingly termed *bending folds*. It is of interest for tectonic analysis that one be able to distinguish between these two classes because such distinction in the field yields valuable information about the direction, and to some extent about the relative magnitude, of the forces that have produced the deformation.

We shall discuss some of the characteristic features of these two classes of folds in layered rocks.

Buckling folds

A sheet of rock or other material that is thin relative to the length will usually buckle under a compressive deviatoric stress that acts parallel to the sheet. Let us imagine that such a sheet is floating about in a region in outer space where the pull of gravity is equal in all directions and thus neutralized. Under such conditions a compressive force acting along the sheet will make it buckle to one single half-wave, independent of the length of the sheet.

The situation is generally different if the sheet is in contact with, or inclosed in, another material and/or acted upon by gravity. The strength and viscosity of adjacent materials tend to prevent the sheet to buckle to one single half-wave. For such a long fold too much energy would be dissipated by the contact strain in the surroundings. Less energy would be dissipated if a given shortening of the system is accommodated by a number of short buckles. This is shown by KIENOW (1942), BIOT (1957 and 1961) and RAMBERG (1959, 1962 and 1963*b*). A certain characteristic initial wavelength or length of arc tends to establish itself, the ratio between length of arc and thickness of layer depending upon the mechanical properties of layer and adjacent rocks.

If the layer rests on the earth's surface, or at a boundary between two deep-seated rocks with unlike density, then the acceleration due to gravity also executes a control over the wavelength as discussed by SMOLUCHOWSKI (1909 and 1910), BIOT, op. cit., and RAMBERG (1963*a*).

Buckling folds may be distinguished from bending folds by the following criteria:

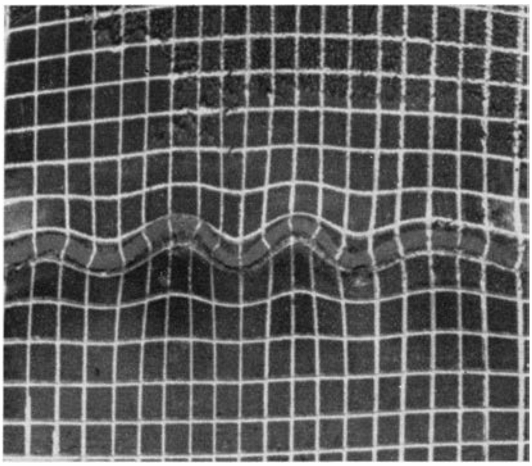

Fig. 1. Buckling folds of 1.7 mm thick stiff rubber sheet sandwiched between slabs of soft rubber. Note distortion of original orthogonal marker lines in contact region adjacent to buckled sheet.

(*a*) Shortening in a direction parallel to the originally more or less straight layer and normal to the fold axis is associated with buckling folds (unless the layer has been rotated from the segment of compression into the segment of extension in the strain ellipsoid, see RAMBERG, 1959).

The shortening may be recognized by structural features independent of the buckles. Such features as strained fossils or boulders, schistosity, rotated or otherwise strained mineral grains, occurrence of boudins etc. in the surrounding rocks may indicate whether or not shortening has occurred parallel to the buckled layer.

(*b*) The pattern of contact strain in rocks adjacent to a buckled layer is quite characteristic and unlike the strain pattern adjacent to a similar layer folded by bending. The convolved contact strain on either side of a buckled layer decreases in magnitude rather rapidly with distance from the buckled sheet. In case the adjacent rocks are the same on either side of a buckled layer, the contact strain is symmetrically distributed on either side of the layer as shown in Fig. 1. It is noteworthy that the contact strain is negligible and hardly visible beyond one wavelength from the buckled layer.

Bending folds formed as defined above do not show a symmetric distribution of a convolved contact strain that diminishes with distance on either side of a given layer. To the contrary, if the amplitude of the sinuous strain decreases with distance on one side of the folded layer, the strain must increase with distance on the opposite side of the given layer such as indicated in Fig. 2.

(*c*) In buckling folds of a single enclosed layer the latter is always more competent than the adjacent rocks otherwise the particular distribution of stresses that generate the secondary sidewise motion in buckling will not develop. Such

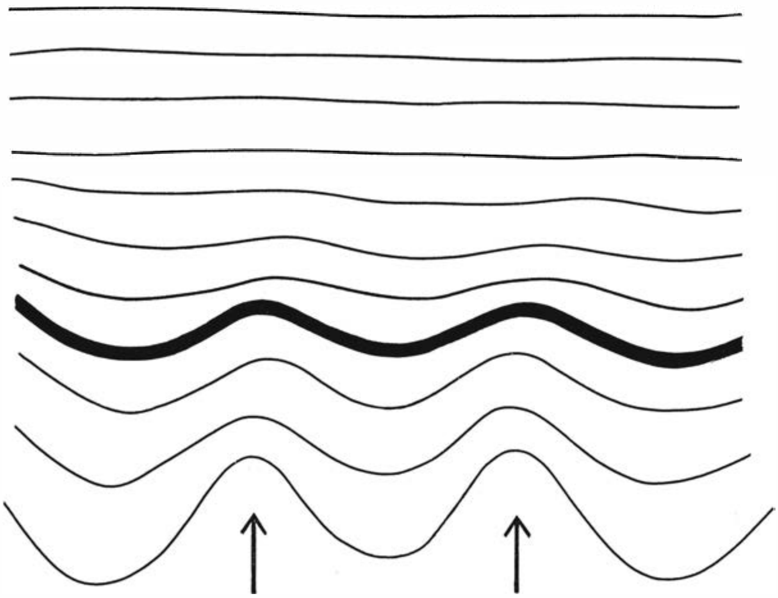

Fig. 2. Layer (black) folded by bending. Arrows indicate localities and direction of maximum compressive stress. Note gradual decrease of amplitude by increasing distance from the stress-generating sites.

a contrast in competency is not necessary and generally not present in bending folds. Relative competency of rock types can be established in the field by means independent of the buckled layers themselves. Distribution of fractures, relative ease of flow of rocks, distribution of boudinage structures etc. indicate relative competency. If the kind of rock which constitutes a given folded layer, by independent observations is found to be less competent than the adjacent rocks, buckling can usually be excluded as the fold-generating process. On the other hand, if a folded layer by independent means is shown to consist of a type of rock that is more competent than the host, the buckling model may well be the correct one.

In rock complexes with strongly anisotropic resistance to shear strain, buckling may occur without contrasted competency of buckled layer and adjacent rocks.

(*d*) For buckling folds the relationship between the ratio: length of arc/thickness and the ratio of the viscosities, μ_1/μ_2, or of the elasticity moduli, G_1/G_2, of layer and enclosing rocks, are

$$\lambda/D = 2\pi\sqrt[3]{\frac{1}{6}\frac{\mu_1}{\mu_2}}, \tag{1}$$

and

$$\lambda/D = 2\pi\sqrt[3]{\frac{1}{6}\frac{G_1}{G_2}} \text{ respectively.} \tag{2}$$

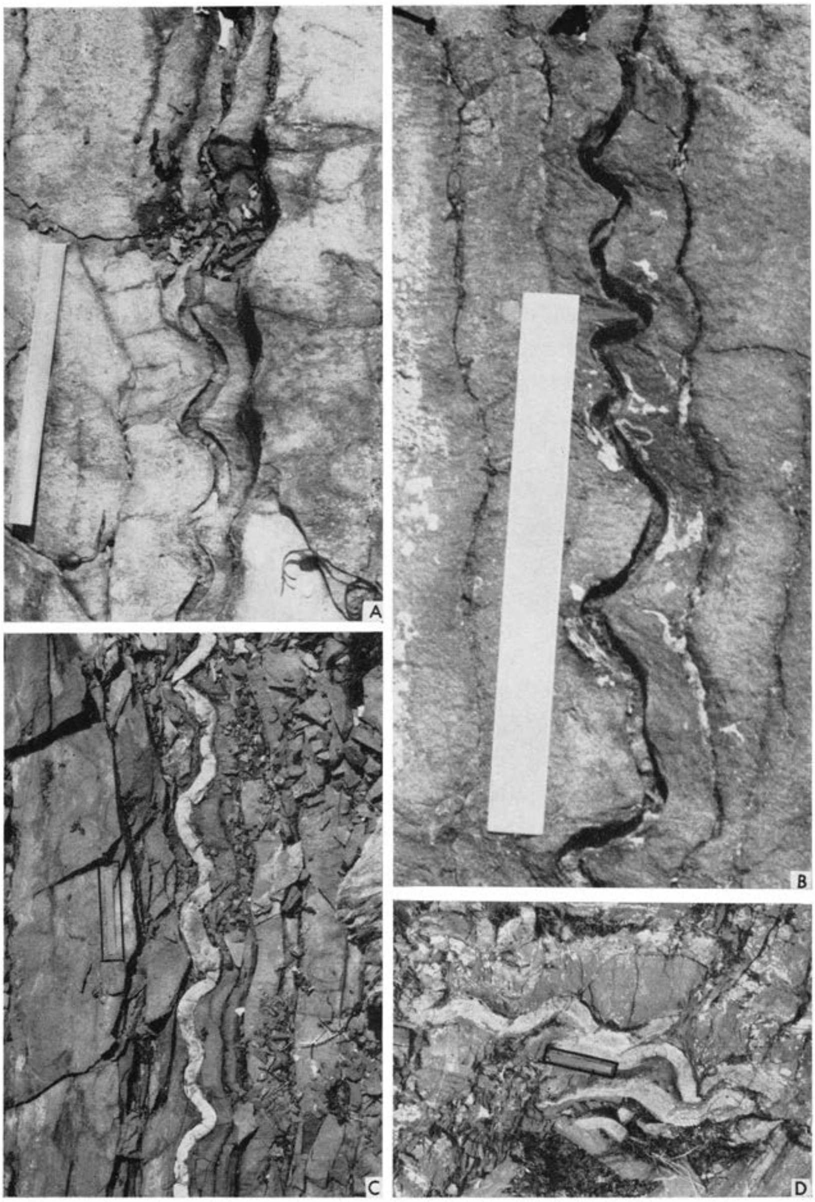

Fig. 3. Buckled veins or layers of chiefly quartz in siltstone. The photographs are of unlike scale as shown by the 6 inches long ruler which is the same on all four pictures. [On *C* and *D* the ruler is framed by black ink.] Since the buckled layers appear approximately equally thick on all four photographs, (note: the prints are adjusted to fit that condition) the apparent lengths of the 6 inch ruler show that the buckled layers in *A*, *C*, and *D* are respectively 1.87, 6.3 and 8.2 times thicker than the vein in *B*. Inasmuch as the apparent length of arc of the folds is approximately the same in the four photographs, the four different folded layers are rather geometrically similar, i.e. the λ/D ratio is approximately the same in spite of the considerable difference in absolute dimensions.

Eq. (1) is only valid for substances which behave as so-called Newtonian bodies, i.e. substances whose coefficient of viscosity is independent of magnitude of applied stress and rate of change of strain. Eq. (2) is valid for elastic materials.

To the extent these equations apply to rocks it follows that for several layers with unlike thickness of a given type of rock (e.g. quartzite) enclosed in a given host (e.g. shale) the λ/D ratio should be constant and independent of the thickness, D, of the quartzite layers. Fig. 3 furnishes a good demonstration of this feature that is characteristic for buckling.

Such a simple relationship between length of arc of folds and thickness of layer does not exist for bending folds. Eqs. (1) and (2) consequently furnish a means of distinguishing between buckling folds and bending folds. For example, if the λ/D ratio of a certain layer is found to be much greater than for other layers of the same kind of rock in the same kind of surroundings, buckling could hardly be the fold-generating process.

Bending folds

In such folds the more or less periodically varying motion normal to layering is not a direct result of compression parallel to the layers. If anything it is the other way around: the periodic or irregular motion normal to layering may result in shortening in a direction parallel to the original layers.

In bending folds contrasted competency of layer and adjacent rocks is not necessary. Layering may for example well be replaced by imaginary marker surfaces (i.e. surfaces whose sole purpose is to make the strain visible) in an isotropic and mechanically uniform rock, yet such markers would display a wavy pattern if the rock was exposed to a stress system characteristic of bending folds. A stress system that is characteristic of buckling folds, viz. maximum compression making less than 45° with layering, will not result in folding of such marker surfaces in a mechanically uniform rock.

One may imagine a number of situations in which stress patterns capable of creating bending folds may be generated. Bending folds are often formed in regions adjacent to an irregular surface of a competent rock body enclosed in less competent laminated rocks. A row of competent inclusions also commonly gives rise to bending folds in adjacent strata. When a complex of competent and incompetent rocks is exposed to anisotropic stress, the local stress pattern in the neighbourhood of the competent bodies is apt to be of a nature that causes periodic movement transverse to layering in the adjacent incompetent rocks. Figs. 4 and 5 show examples of this kind of development.

The rise of domes also results in what here is classed bending folds in superincumbent strata, see Robinson (1923). This particular kind of bending folds is called supratenuous in the terminology of structural geology.

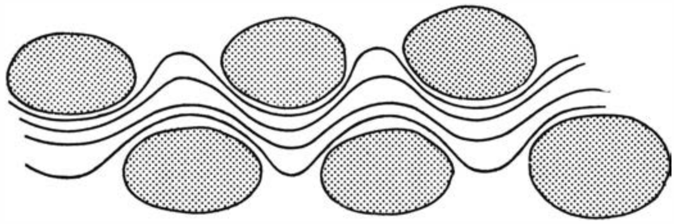

Fig. 4. Bending folds in stratified conglomerat matrix formed by compression more or less normal to original layering.

There are two features of the strain pattern of bending folds that are particularly significant:

(*a*) The contact strain that generates bending folds around competent inclusions or adjacent to an irregular surface of a competent rock, does not penetrate very far into the surroundings. The tendency of incompetent rocks to conform to the contours of competent bodies is only noticeable within a boundary zone that is no thicker than the "wavelength" of the irregularities, or the spacing between inclusions of competent bodies such as, for example, a row of boudins. The "wavelength" of the irregularities on the surface of the competent rock or the spacing between the inclusions that generate the particular heterogeneous strain displayed as folds in adjacent strata, is approximately equal to the wavelength of the folds themselves as shown in Figs. 4 and 5. This is significant because it means that the particular feature (for example some competent bodies, or a rising dome) that generates a given set of bending folds cannot be more than one wavelength away from the folded layer in question. In well exposed regions this condition furnishes an excellent means of determining whether or not given folds are caused by bending. For example, in Fig. 6 there is no indication of anything that could have generated a stress field characteristic of bending, in spite of the fact that an area much wider than the

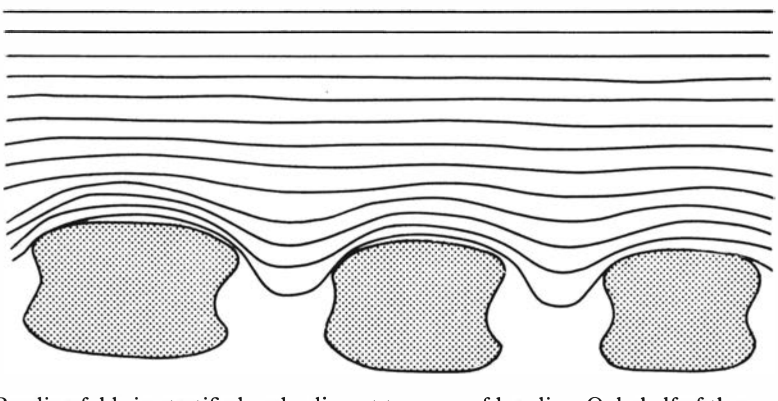

Fig. 5. Bending folds in stratified rock adjacent to a row of boudins. Only half of the symmetric structure shown.

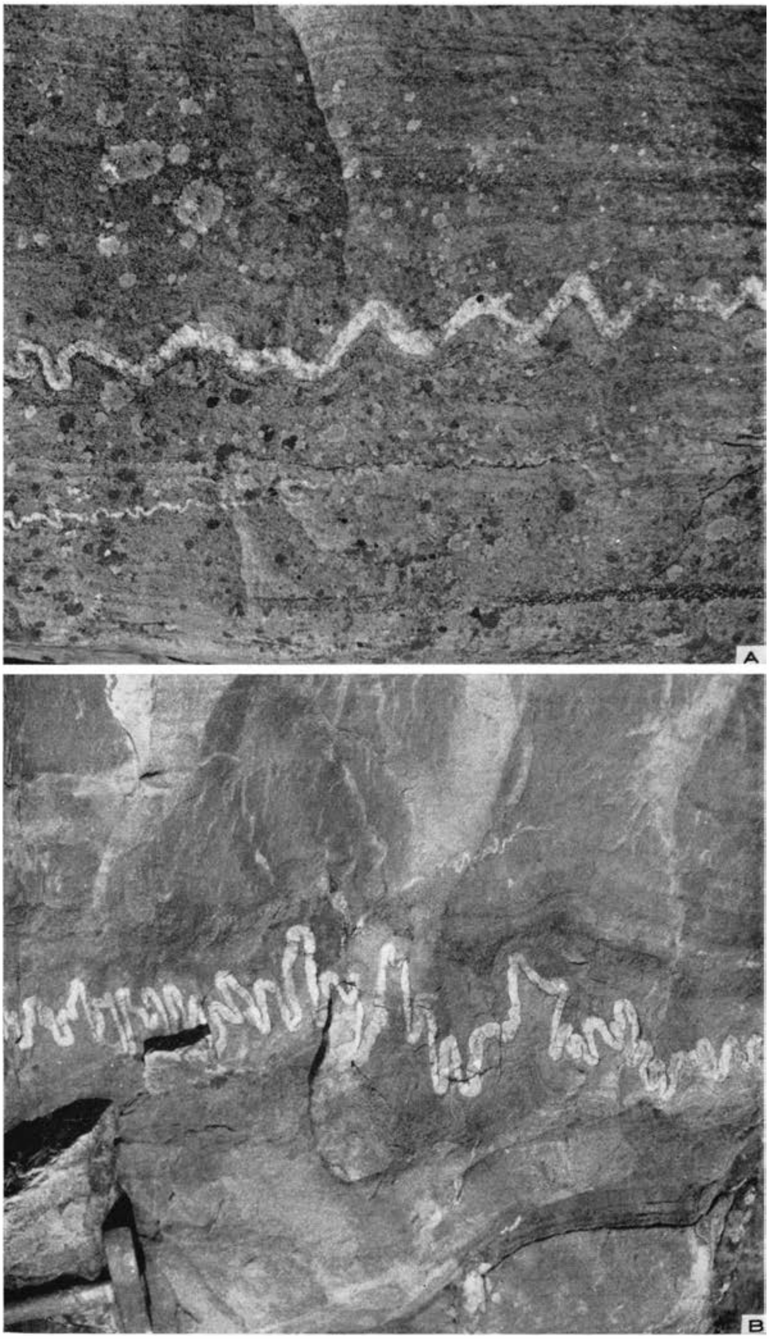

Fig. 6. Ptygmatic quartz veins in micaceous quartzite, Tanvik, Hemne, Norway. The veins have buckled in response to the compressive stress whereas the host rock some distance away from the veins has been compressed parallel to layering in an uniform manner.

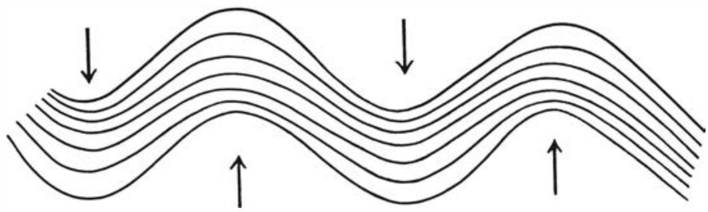

Fig. 7. Change of thickness due to bending of originally even-thick strata. Arrows indicate distribution of bending forces.

length of arc of one wave is exposed. The folds shown in Fig. 6 are therefore not of the type here defined as bending folds (or shear folds, see below), but belong to the buckling type.

(*b*) Since the primary forces that generate bending in general are distributed as indicated in Fig. 7 the change of thickness of originally even-thick layers tends to display the pattern shown in the figure. This is exactly opposite to the distribution of layer-thickness in folds formed by buckling as indicated in Fig. 8.

It should be mentioned at this point, though, that buckled multilayers with a great many individual sheets display very little change in thickness of the central layers (RAMBERG, 1963 b).

According to HILLS, op. cit., p. 89, so-called shear folds and flow folds are two categories of folds that should be distinguished from the buckling- and the bending categories.

In case shear folds actually are formed as generally assumed, viz. by differential slip parallel to the axial plane, it seems, however, that the stress system which generates shear folds is principally the same as that which generates bending folds, compare Figs. 9 and 7. The difference between folds which display discrete planes of shearing parallel to the axial plane, and other bending folds that have been strained in a continuous, plastic manner is found in unlike mechanisms of deformation rather than in unlike stress systems.

Flow folds probably do not either represent a category of folds different from the bending- and buckling types. Flow folds may be of either class, the sole peculiarity about flow folds being that the rocks constituting such folds

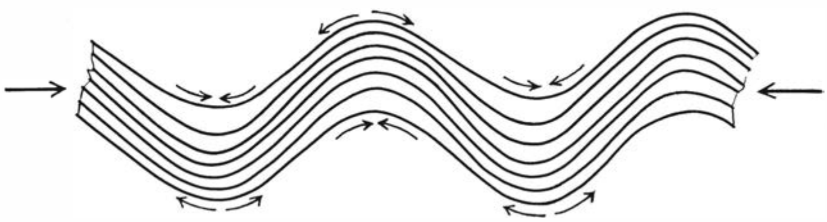

Fig. 8. Change of thickness due to buckling of originally even-thick strata.

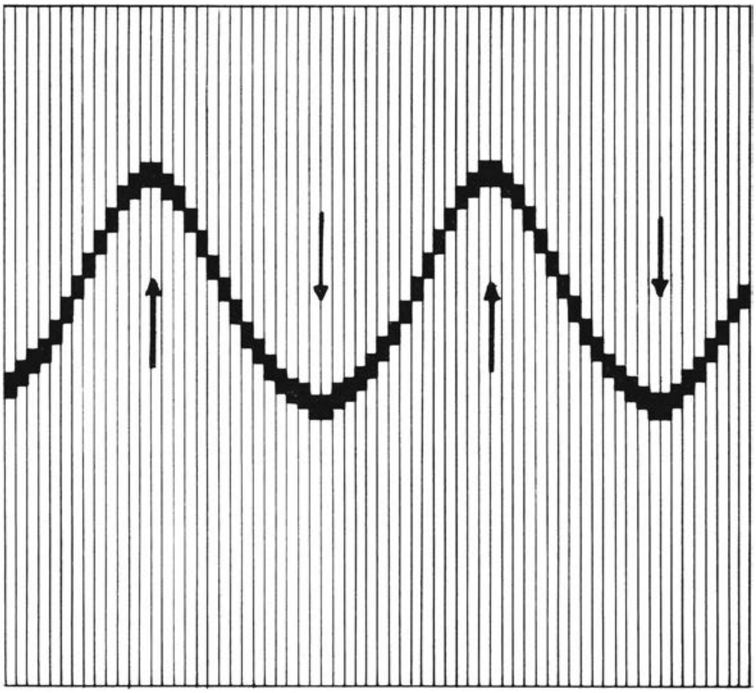

Fig. 9. Pure shear folds in layer (black), according to convention. Arrows indicate forces needed. Probably not a likely mechanism of fold formation.

either were more plastic, or that the process of deformation was more slow than in the cases of more "ordinary" bending- and buckling folds.

The possibility that flow folds represent turbulent flow in deep-seated crystalline rocks seems untenable because the Reynold's number, which determines the conditions at which laminar flow changes to turbulent flow, is generally much too small for crystalline rocks to permit turbulence.

Concentric folds and similar folds

It is a matter of simple geometry that concentric or parallel folds cannot extend throughout the whole thickness of a stratified complex that is thicker than a certain limit as determined by the wavelength and amplitude of the folds (see calculation in the text of Fig. 10). This fact is mentioned in most texts on structural geology such as for example HILLS, op. cit., p. 82, and BILLINGS (1954), p. 58.

It can also be argued, on basis of the distribution of stresses, that similar folds cannot generally prevail across the entire thickness of a stratified rock complex. During folding, layers close to the boundaries of the complex are subject to stresses sensibly different from those existing in the central layers.

Inasmuch as unlike stress distribution generally gives unlike finite strain in plastic or viscous deformation the style of folding must change as one goes from the boundaries to the interior in a layered complex. Thus the folds cannot be exactly similar.

As thus neither of the two chief idealized geometric categories of folds, viz. concentric and similar, can in general persist throughout the entire thickness of multilayered rocks, it becomes a problem of some significance to determine how the style of folding may change across layered complexes.

Some light was thrown on that problem during the author's experiments with buckling of multilayered test pieces of rubber of unlike rigidity (RAMBERG, 1962 and 1963*b*).

Multilayered rectangular blocks consisting of two sets of alternating rubber sheets with unlike rigidity were compressed parallel to layering. If the whole multilayer was thin relative to the wavelength quasi-concentric folds formed, but if the multilayer was thick relative to the wavelength the style of folding changed from quasi-concentric in the outermost few layers to similar (or chevron type) in the central portion as indicated on Fig. 11.

It is noteworthy that the different style of folding formed simultaneously over the entire test piece in response to a given stress distribution along the boundaries of the composite block. Since the mechanical properties and the dimensions of the two sets of layers were the same throughout the body it follows that the unlike style was solely due to a change in stress distribution across the composite body. A detailed discussion of this stress distribution is however outside the scope of the present paper.

It is of interest to note that the typical zig-zag shape of the internal layers is not caused by breaks at original weak points in the layers because the layers consisted of very flexible rubber. Release of the compressive force made the layers straighten out without leaving sign of the hinge points. It is the particular stress distribution in the central portion of the composite block which gives rise to the zig-zag shape.

The tests were performed on elastic materials which unquestionably are quite unlike rocks in rheological behavior. But theoretical considerations of viscous multilayers discussed by the author (op. cit.) show that a very similar change of fold-style would occur during buckling of plastic multilayers whose rheological properties are analogous to rocks. Experimental tests on such materials (modelling clay, putty and other materials) by BAYLY (1962) showed this to be true.

The change of fold-style from quasi-concentric in the marginal layers to similar (chevron) in the central layers is characteristic for multilayers whose layering is uniform throughout. This ideal fold-style will be modified in various manner if thickness and/or rheological properties of one or both sets of layers vary in the complex.

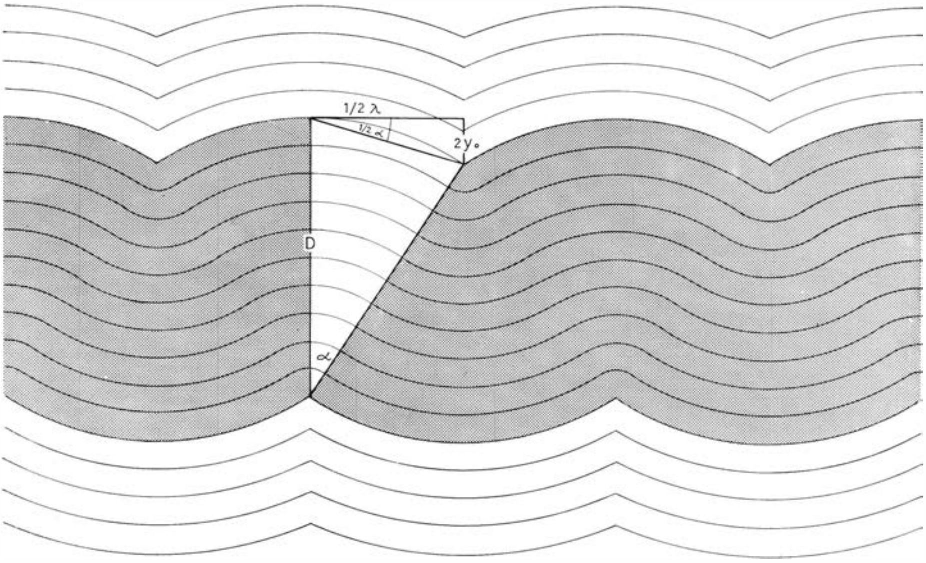

Fig. 10*a*.

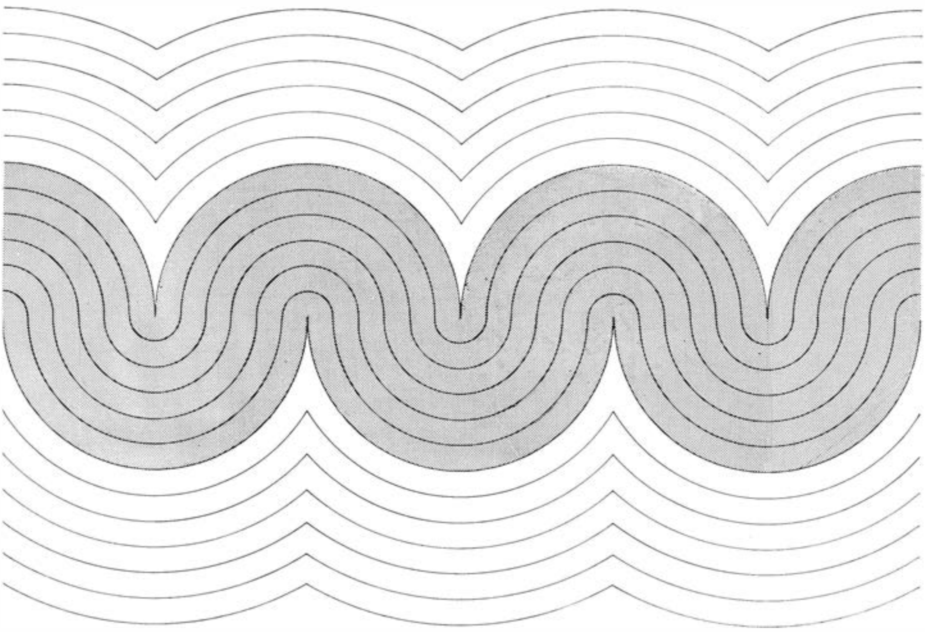

Fig. 10*b*.

Fig. 10. Ideal concentric folds consisting of circular arcs. *D* is maximum thickness of the portion of the complex which consists of ideal concentric folds not effected by shortening and thickening of the layers. All layers within the region *D* have identical length of arc along a complete run from crest to crest. Maximum thickness, *D*, increases with increasing wavelength if the amplitude, y_0, remains constant, but decreases with increasing y_0 if λ remains constant, compare Figs. *a* and *b*.

Folds and schistosity

There are a number of contradicting viewpoints as to the relationship between the attitude of schistosity in tectonites and the orientation of the strain- and/or stress ellipsoids. (The discussion below is limited to the kind of schistosity that is represented by parallel arrangement of platy and rod-shaped minerals such as micas, chlorite, hornblende and sillimanite, or parallel arrangement of deformed ellipsoidal grains of such minerals as quartz and feldspars.)

In many situations it may be difficult to decide among suggested relationships such as the following: (*a*) schistosity is parallel to the two long axes in a triaxial strain ellipsoid; (*b*) it is parallel to one of the directions of maximum shear movement in the strain ellipsoid; (*c*) it is normal to maximum compressive stress in the stress ellipsoid, or (*d*) the schistosity is parallel to one of the directions of maximum shear stress.

It obviously would be desirable to know in individual cases with certainty the relationship, if any, between schistosity and the geometry of strain and/or stress which presumably have generated the schistosity. Such knowledge would be very helpful for tectonic studies in gneissic areas, for example, where a monotonous schistose structure is the only thing which may inform about the geometry and magnitude of deformation which the region has experienced.

In order to study how schistosity is related to the geometry of the systems of strain and stress within a given limited region, one may for example hunt for structural features which are not directly related to the schistosity, but which are related in a known manner to the same strain and stress systems that (presumably) generated the schistosity. By comparing such structural features with the orientation of the schistosity it is in principle possible to decide how the latter is related to the pattern of the systems of stress and strain.

There are unfortunately few structural features whose relationship to the

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This sort of fold style is physically unrealistic and must be considered a purely geometric demonstration used to show that the thickness over which concentric folding may persist is limited.

The relationship between maximum thickness, amplitude and wavelength follows from simple trigonometry:

$$2y_0 = \tfrac{1}{2}\lambda \ \mathrm{tg}\ \tfrac{1}{2}\alpha = \tfrac{1}{2}\lambda \frac{\sin\alpha}{1+\cos\alpha}.$$

$$\sin\alpha = \tfrac{1}{2}\ \lambda/D, \text{ and}$$

$$\cos\alpha = \frac{D-2y_0}{D}, \text{ which gives}$$

$$D \leqslant \frac{1}{16}\frac{\lambda^2}{y_0} + y_0.$$

Pinching in the core of the folds will occur if $y_0 > 1/4\ \lambda$, this destroys the ideal concentric pattern. Thus the limitation $y_0 \leqslant 1/4\ \lambda$ is also imposed upon geometrically ideal (but physically unrealistic) concentric folds.

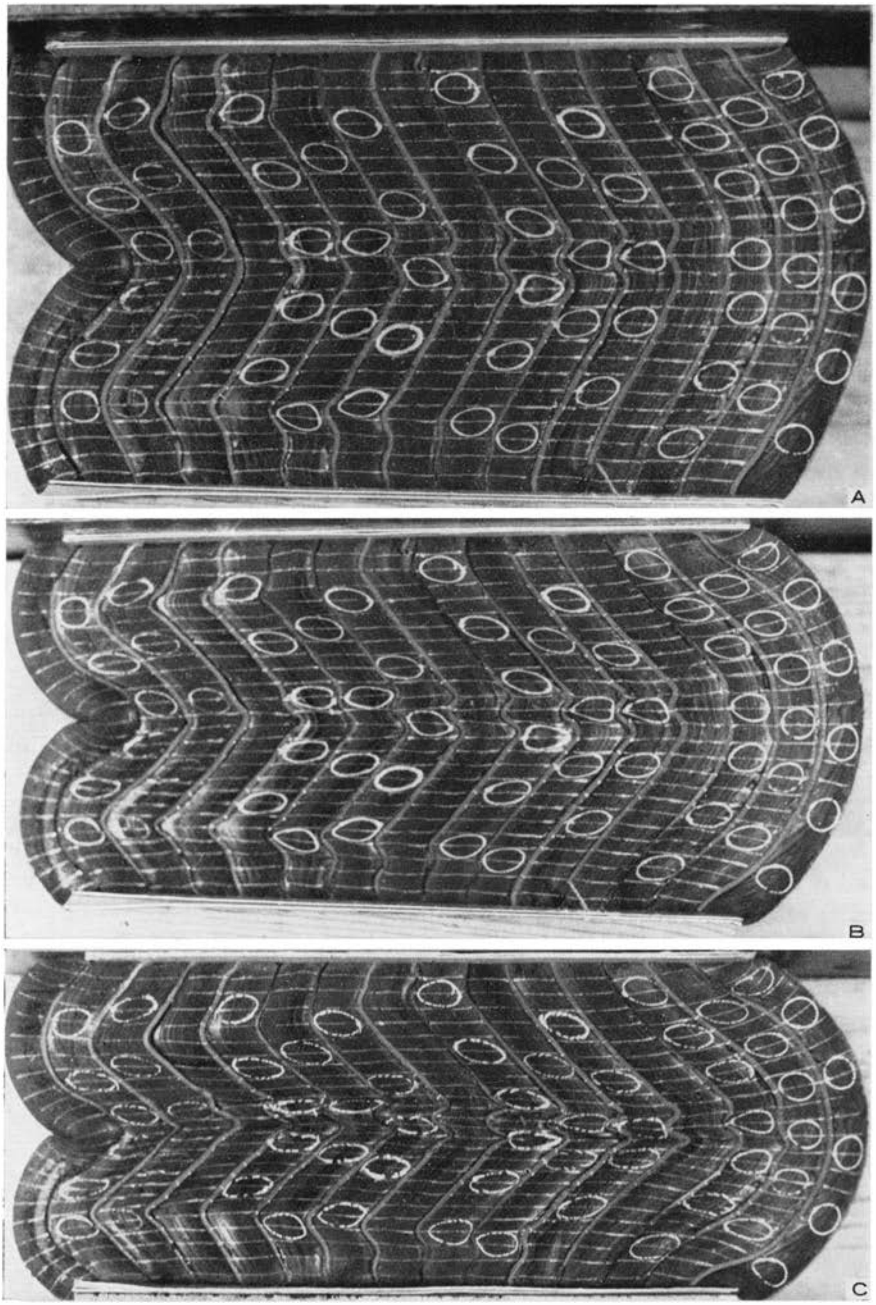

Fig. 11. Test slab consisting of 1.7 mm thick sheets of rather stiff rubber interlayered with 12 mm thick sheets of soft rubber. *A*, *B* and *C* display increasing amount of compression. Note tendency to develop concentric fold pattern in the outermost 2–3 pairs of layers whereas almost ideal chevron-type folds have formed in the central portion of the multilayer. Details of the strain are shown by the strain ellipses as formed from original circles.

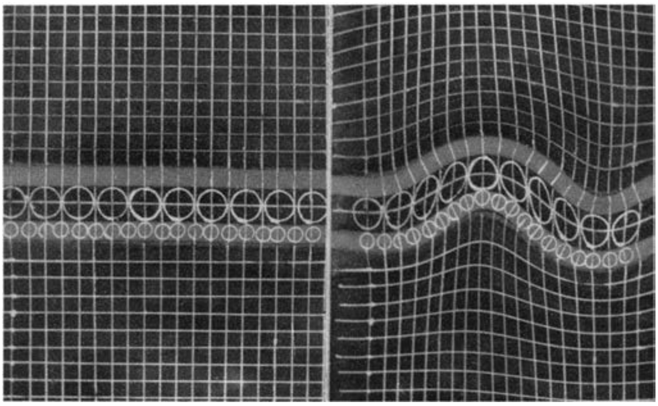

Fig. 12. Two 6 mm thick stiff rubber sheets (light grey) interlayered with softer rubber (black). *A* unstrained state, *B* buckled state. Note distribution of strain as shown by distorted markers.

strain or stress systems is agreed upon among geologists. The disagreement probably arises from the difficulty of determining the relationship between deformation structures in rocks and the generating stresses. The relationship between certain structures and the pattern of finite strain is better understood. Some of these relationships may be used as guides toward establishing possible correlations between schistosity and the geometry of finite strain in tectonites.

Structural features which may be used as guides of the above-mentioned type are for example: deformed boulders in conglomerates with schistose groundmass (see Flinn, 1956 and Billings, op. cit., p. 342), strained oolites in limestone (Cloos, 1947), occurrence of boudins in schists and gneisses (Ramberg, 1955), and the occurrence of ptygmatic veins in schists and gneisses (Kuenen, 1938; Ramberg, 1959). The concept of *tectonic regimes* (Harland and Bayly, 1958) is of interest in this connection.

The geometric attitude of these structures informs on the geometry and character of the strain in the area such that schistosity may be correlated to the strain.

In addition to the above-mentioned guide structures buckling folds in layered rocks are useful. In such rocks the strain pattern within individual layers is relatively simple as may be demonstrated theoretically and experimentally. If schistosity develops in the layers its attitude may consequently be compared with the known strain pattern.

In order to make the finite strain pattern visible in buckled multilayers consisting of alternating competent and incompetent sheets, markers may be drawn on the unstrained surface normal to the foldaxis of the test pieces and the change of shape of these markers may be studied after buckling, Figs. 12 and 13.

The deformed shapes of original circles show that finite strain is very pro-

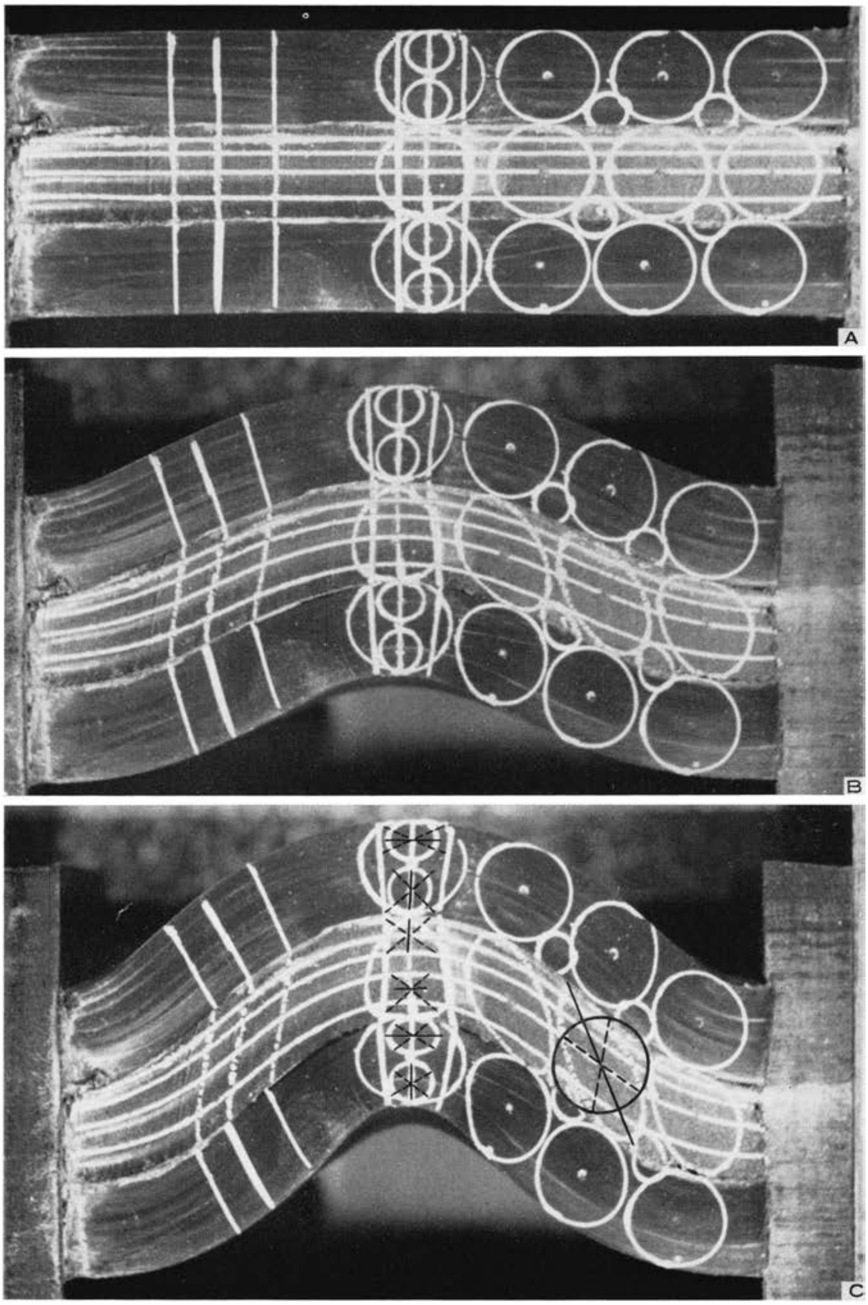

Fig. 13. A layer of soft rubber flanked by two layers of stiffer rubber about 12 mm thick. *A*, *B* and *C* represent increasing amount of buckling. Stippled black lines indicate finite maximum shear directions, full black lines are parallel to finite maximum extensive strain. Black circle on flank of *C* is identical in size to the original unstrained circular markers. Lines between intersections of this circle with the strained ellipse with same centre are almost parallel to the two directions of finite maximum shearing strain. (Although not clearly indicated, the directions of maximum finite shean strain do not coincide with the circular cross sections).

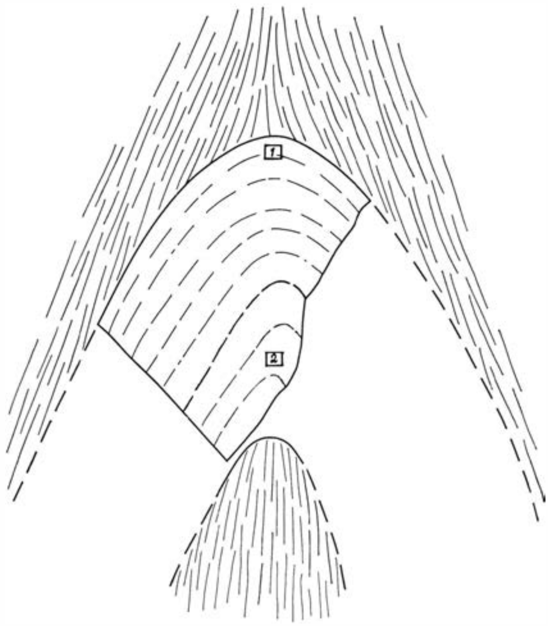

Fig. 14. Part of folded quartzite layer sandwiched between mica schist. Schistosity in mica schist indicated by short lines. The short conformable lines in the quartzite represent original layering as indicated by concentrations of mica. Areas marked 1 and 2 indicate sites where microphotos in Fig. 15 are taken.

nounced in the incompetent layers at the flanks of the folds; considerable strain occurs also in the crest regions in both kinds of layers, but very little strain is displayed in the competent layers at the flanks. This agrees well with observations in the field. In alternating strata of slate and sandstone, for example, or mica schist and quartzite, schistosity or flow cleavage is much more strongly developed in the slate and mica-schist layers at the flanks of folds than in the sandstone and quartzite at these parts of folds. Indeed, in competent sandstone and quartzite layer-parallel lamination of primary sedimentary origin may remain almost unaltered at the flanks. In the crest regions, however, schistosity may be visible in competent as well as incompetent layers, Figs. 14 and 15.

The tests shown in Figs. 12 and 13 were made on elastic rubber materials. In such material there is little overall compressive strain normal to the axial plane. But in plastic or viscous substances, such as natural rocks during the process of buckling, a considerable amount of longitudinal compressive strain parallel to layering occurs together with the buckling process as discussed by Biot (1961) and Ramberg (1962). The relative significance of shortening by buckling and by layer-parallel compressive strain respectively, depends upon a number of conditions, but there will always be a significant amount of layer-parallel strain unless the layers are very competent relative to the adjacent rocks or unless a considerable curvature existed in the layers prior to buckling.

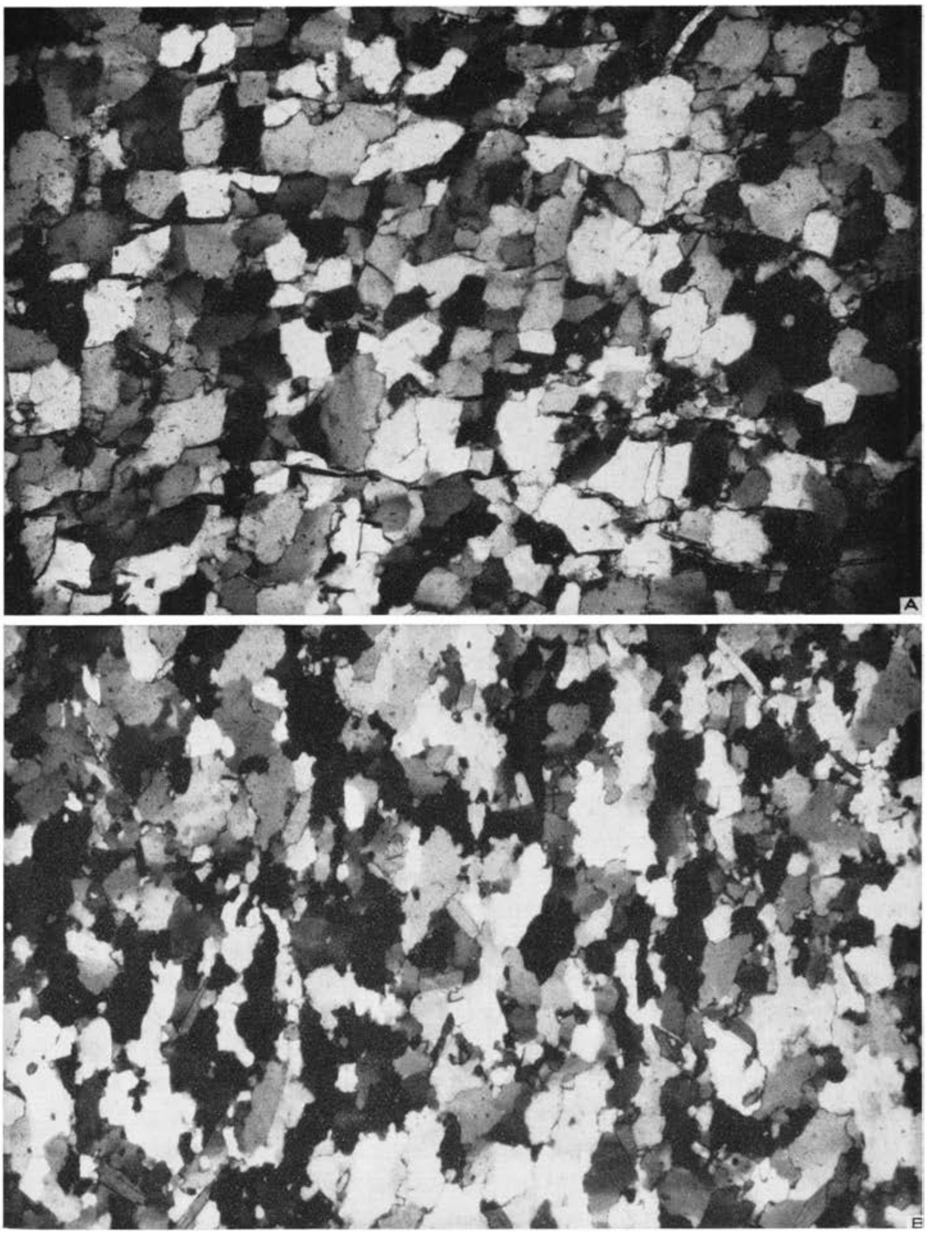

Fig. 15. Microphotos of the folded quartzite layer shown in Fig. 14. *A* is from point 1 in Fig. 14, *B* from point 2. The structure shown in *A* is probably not much altered during the fold movement whereas *B* shows a strongly developed schistosity parallel to the long axis of the strain ellipsoid in the area.

Such curvature increases the bending moment of the compressive stress so that buckling is apt to shorten the multilayer faster than the layer-parallel strain is capable of doing.

The compressive stress that acts on the system tends to rotate the strain ellipsoid at any point along the flanks of the folds in such a manner that the long axis becomes more closely parallel to the axial plane of the folds. Moreover, in the apices of the folds the compression will lengthen one or both of the axes of the strain ellipsoid that are parallel to the axial plane and shorten the axis that is normal to that plane. Hence it is not unlikely that the strain ellipsoids at the convex, stretched part of the crests of the competent layers in the rubber models (Fig. 13), in plastic or viscous materials would display their plane of maximum extension parallel to the axial plane of the fold. When this is the case the schistosity at the apices should be parallel to the axial plane through the entire thickness of the competent layers (assuming that schistosity actually develops normal to the short axis of the strain ellipsoid).

The directions of maximum finite shear strain in the buckled rubber test piece are indicated approximately in Fig. 13. Since the directions of maximum finite shear strain, and of maximum finite longitudinal strain respectively, follow different patterns in buckle-folded layers, it is usually not difficult to decide which pattern the schistosity conforms to in natural rocks. According to most studies flow cleavage or schistosity appears to be parallel to the plane of maximum extension in the strain ellipsoid.

One of the well known difficulties with the hypothesis that schistosity be parallel to the directions of maximum shear strain (or stress) is that there are either two planes of maximum shear strain (in the kinds of strain called simple and pure shear; JAEGER, 1956, p. 24), or that maximum shear strain occurs across a double-cone shaped surface (in three dimensional strain, most common in rocks, RAMBERG, 1959, p. 117 ff.). But the kind of schistosity discussed here (see definition above) is, within any small volume of a rock, parallel to a single plane which occasionally may degenerate to a line, in which case the rock displays lineation without showing schistosity. Such geometric attitude of schistosity is in accord with the assumption that schistosity is parallel to the plane of maximum extensive strain in the strain ellipsoid because there is only one such plane in the strain ellipsoid, and in uniaxial extensive strain this plane degenerates to a single line.

However, the whole question of the attitude of schistosity in relation to the geometry of strain is by no means settled, and it is useful to have opportunities to attack the problem from several sides. Model studies of the distribution of strain in folded layered test-bodies of unlike materials represent a simple but fruitful method of attack.

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