Journal of Structural Geology, Vol. 4, No. 2, pp. 177 to 195, 1982
Printed in Great Britain

0191-8141/82/020177-18 \$03.00/0

Repeated parallel deformation in part of the eastern Sierra Nevada, California and its implications for dating structural events

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(*Received* 26 *June* 1981; *accepted in revised form* 18 *January* 1982)

Abstract—Study of a thick section of late Paleozoic to mid-Cretaceous sedimentary and volcanogenic rocks in east-central Sierra Nevada has revealed an involved structural succession not readily apparent when analysed under the traditional assumptions of structural analysis (e.g. parallel structures are of the same age).

Earliest structures in the area occur as sparse folds in late Paleozoic rocks, whereas in Triassic to mid-Cretaceous rocks earliest structures occur as penecontemporaneous slumps. Upon these earliest structures are superimposed slaty cleavage with associated lineations and subsequent crenulations. The slaty cleavage across the area is statistically parallel, as are the axial planes of crenulations which fold the slaty cleavage. Such a succession would traditionally be interpreted as representing two periods of deformation, the first forming the slaty cleavage and the second the crenulation of the slaty cleavage. There is evidence, however, to indicate that the slaty cleavage itself was formed during more than one period of deformation and the same may be true for the crenulations. Dykes emplaced in Jurassic rocks have been dated (U/Pb) as mid-Cretaceous and lie parallel to what is probably an early slaty cleavage direction. The dykes, however, also bear a slaty cleavage, albeit weaker than in the host rock. In addition, quantitative strain determinations of rocks in the area show that the older units are more strongly deformed than the younger units. These and other data suggest that the statistically parallel slaty cleavage and related structures (folds, lineations, etc.) found in the Jurassic and older rocks have formed during at least two, and possibly three, increments of strain, each increment separated by a lengthy period of geologic time, possibly as much as 45 Ma or more. Crenulations of the slaty cleavage at any point (subsequently formed after each period of slaty cleavage formation) may even predate slaty cleavage formed later at another nearby point.

While it is possible to set up a chronology between earlier (tectonic and/or penecontemporaneous slumps) and later structures (slaty cleavage, folds, lineations, etc.), it is not valid to designate for the entire area a relative time sequence of formation of slaty cleavage and crenulations in the Jurassic or older rocks by the usual methods (e.g. S_2, S_3, F_2, F_3, etc.). These later structures can only be designated as Mesozoic. Only in the youngest stratigraphic unit in the area, which has been subjected to one deformation (mid-Cretaceous), can a valid structural succession be applied areally.

We suggest that multiphase, parallel structures, comparable to those we have described, may be a relatively common phenomenon in orogenic belts. Until one arrives at a thorough understanding of the detailed stratigraphy and the absolute ages of units in key relationships to the structures, it may only be possible to delineate the broadest of time sequences for the structures concerned.

INTRODUCTION

SINCE 1950 the detailed structural history of fold mountain belts in many parts of the world has been given considerable scrutiny. Successive periods of deformation have been shown to be characterized by structural and metamorphic features which allow a chronological sequence of structural events to be established for a given area. Orientation of structural elements is often used as one of the important criteria in identifying a given period of deformation. In particular, slaty cleavage with a distinct statistical orientation is commonly assumed to have been formed during one period of deformation. Our work on rocks of the eastern Sierra Nevada of California has shown that slaty cleavage accompanied by folds and lineations with essentially the same statistical orientations are the result of at least two successive periods of deformation that were separated by a substantial period of geological time. Correlation of periods of deformation (S_1, S_2, F_1, F_2, etc.) over larger areas by the usual methods of similarities in style, orientation, metamorphic characteristics, etc. becomes untenable.

The purposes of this paper are (1) to present data supporting our interpretation of the deformational history of the rocks in the area under consideration, (2) set forth our evidence for discriminating between one, versus two, deformations that produced the observed structures and (3) discuss some of the problems of determining the structural successions in orogenic belts, using the Sierra Nevada as the primary example.

ROCK TYPES AND AGES OF STRATIGRAPHIC UNITS

The Ritter Range (Fig. 1) is one of the most extensive areas of country rock in the east-central Sierra Nevada, and is made up of a thick sequence of metamorphosed

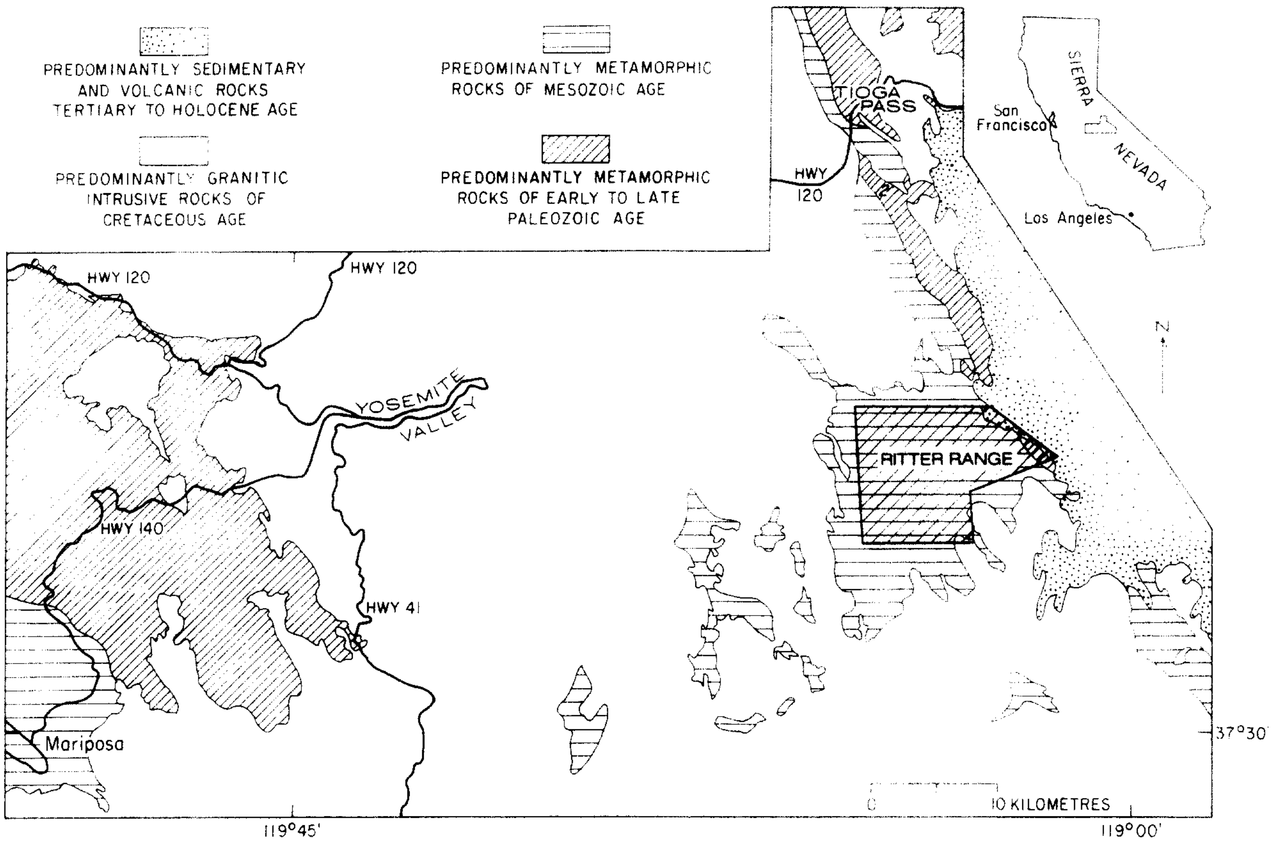

Fig. 1. Geological sketch map showing the locality of area under consideration within the central Sierra Nevada, California.

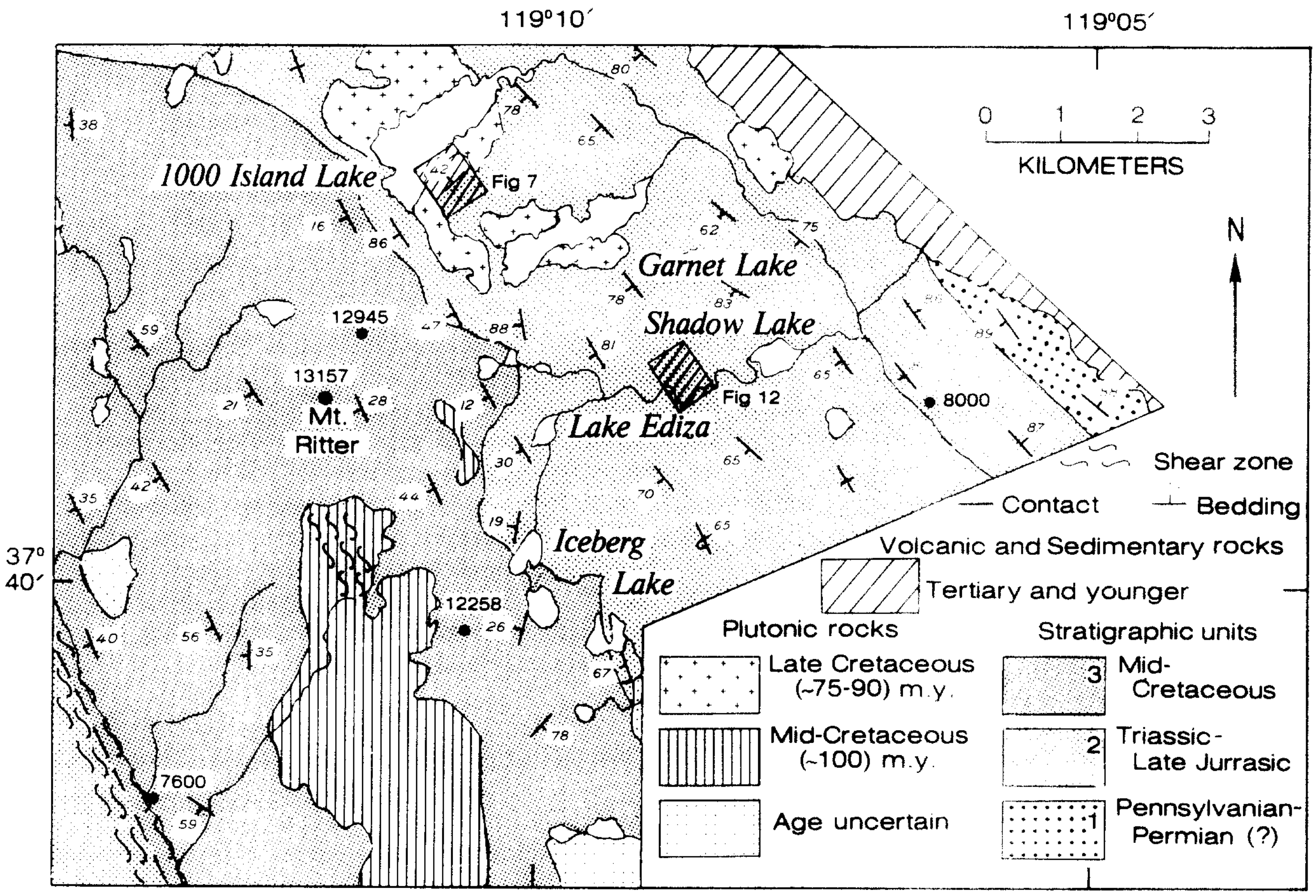

Fig. 2. Larger scale map of the hachured area in Fig. 1. Geology has been greatly simplified into three stratigraphic units, each of a distinct age as shown.

sedimentary and volcanic rock ranging in age from early Palaeozoic to mid-Cretaceous (Rinehart *et al.* 1959, Huber & Rinehart 1965, Fiske & Tobisch 1978). The hachured area in Fig. 1 is part of an area mapped by Huber & Rinehart (1965) on a scale of 1:62,500. Our larger scale mapping (1:6000 in the eastern and 1:12,000 in the western part of the area) supplemented by radiometric age dating of specimens carried out by T. W. Stern (U/Pb) and R. W. Kistler (Rb/Sr) of the U.S. Geological Survey allows us to enlarge on and reinterpret the basic data provided by Huber & Rinehart (1965).

We have divided the rocks of the area into three stratigraphic units, from east to west (Fig. 2). Stratigraphic unit 1, which consists of Pennsylvanian and Permian(?) metasedimentary rocks that are moderately deformed. The age of these rocks has been designated by Huber & Rinehart (1965) and is based on lithologic correlation with similar but fossil-bearing strata occurring several miles to the south. Stratigraphic unit 2 consists of Triassic and Jurassic metavolcanic rocks which are weakly to strongly deformed. The age of these rocks is known from fossil occurrences as well as radiometric dating (U/Pb method) and ranges from 214 Ma at the base to 153 Ma at the top of the unit (Huber & Rinehart 1965, Fiske & Tobisch 1978, p. 220). Stratigraphic unit 3 consists of middle Cretaceous metavolcanic rocks which are undeformed to strongly deformed. These rocks have been radiometrically dated by U/Pb and Rb/Sr methods and span the interval 101–98 Ma (Fiske & Tobisch 1978, p. 220).

The rocks of unit 1 consist of black slate (locally pebbly), thin-bedded siltstone, porcellanite(?), bedded chert(?) bearing relict radiolaria(?) and lesser amounts of thin-bedded quartzite. Resting unconformably(?) upon unit 1 is a thick sequence of volcanogenic rocks (unit 2) consisting of ash flows, lava flows, sills, lapilli tuffs, tuff-breccias and fine-grained slaty tuffs, with thin, discontinuous zones of pure and impure limestone (see Fiske & Tobisch 1978). Our field observations and structural data from the present area support the interpretation that bedding in units 1 and 2 is statistically parallel. To the north, however, the contact between these two units has been interpreted as an unconformity (Kistler 1966) while to the south it has been interpreted as a fault (Bateman *et al.* 1963, Morgan & Rankin 1972). Although this contact appears to have undergone a complex and variable history along its length, on a regional basis it is very likely that there has been tectonic movement between the deposition of units 1 and 2.

Upon unit 2 lies a second volcanogenic unit (unit 3) with clear angular unconformity. This unconformity, along with structural data presented in the following pages, indicates that considerable deformation of the rocks in units 1 and 2 took place prior to deposition of unit 3. The principal rocks types of unit 3 consist of ash-flow deposits (in part welded), a massive caldera collapse megabreccia (Fiske & Tobisch 1978, Fiske *et al.* in press), fine-grained tuffs, tuff-breccias and sills. Deformation of the rocks in unit 3 is often domainal. In addition, complete recrystallization or extensive coarsening of grain size in rocks of all the units is not common, except around plutons and in limestone bodies. As a result, much of the primary texture in the metavolcanic rock is preserved, even though regional metamorphism has affected the entire sequence of rocks. The anorthite content of plagioclase in pelitic, quartzofeldspathic, basic and calcareous assemblages exceeds An_{20}, even in rocks a considerable distance from exposed plutons. Most of the rocks appear to be mineralogically stable in the lower amphibolite facies of regional metamorphism (Turner 1968), although the fine grain size of the typical specimen is more akin to slate and greenschist facies rocks. Areas of lower metamorphic grade may be present locally, however, especially in undeformed zones of Cretaceous rocks of unit 3.

STRUCTURES IN THE ROCK

It has long been known that the country rock of the Sierra Nevada contains more than one set of folds (Clark 1960a, Best 1963, Kistler 1966). In east-central Sierra Nevada, generally, three sets of tectonic structures (e.g. cleavage and/or folds, lineations, etc.) have been documented from the Paleozoic rocks, while two sets of structures are reported from the Mesozoic rocks. Our findings generally concur with these observations, although some of our interpretations differ from those of other workers (see Tobisch & Fiske 1976). Recent work in rocks of comparable age in the western Foothill Belt, however, is uncovering areas of previously undetected, increased structural complexity (Bogen 1979, Schweickert 1979, Merguerian & Schweickert 1980).

Earlier deformation

Folds which predate the slaty cleavage are found in both stratigraphic units 1 and 2. The early folds are scarce in unit 1 and only two examples have been found, both in refolding relationships with later folding (Fig. 3a). These early folds are very tight to nearly isoclinal and lack an accompanying axial plane cleavage, so that it appears as if the early folding in unit 1 was not accompanied by recrystallization. Fold axes of later folds in unit 1 have extremely variable orientations over short distances (2 m), which could be interpreted to indicate the presence of more widespread early folds not readily discernible in outcrop.

In unit 2, early folds occur on both a large and small scale, in two zones. In one locality, we observed folds with wavelengths of as much as 100 m and amplitudes approaching 200 m. These folds are within a stratigraphic horizon nearly 700 m thick, which shows extreme lithologic disruption and can be termed a chaotic breccia over much of its outcrop. We have also observed that slaty cleavage cuts across both limbs of the folds with the same sense, indicating that cleavage post-dates the folds. Based on these and other field observations, we have interpreted both large- and small-scale structures as the result of gravity slumping associated with volcanic activity; details are given elsewhere (Fiske & Tobisch 1978, pp. 214–216).

Slump deposits also occur in unit 3, but no large-scale coherent folds of this genesis have been identified.

Later deformation

The later deformations are found in all three stratigraphic units, although they are variably developed in units 2 and 3. These structures are clearly tectonic in origin. Cleavage is the most prominent planar element (Fig. 3b), and varies in intensity and morphology from a continuous, slaty structure in fine-grained rocks, to a spaced or fracture-type cleavage in the most competent or weakly deformed horizons. Minor folds accompanying the slaty cleavage are not common. Their interlimb angle tends to vary from close to tight (Fleuty 1964) or rarely isoclinal in unit 1 and the lower parts of unit 2 (Fig. 3c) where shortening has exceeded 50% (Tobisch *et al.* 1977), but moderately open to tight in the upper part of unit 2 where the average shortening is around 35% (Fig. 3d). Only a few minor folds with axial plane slaty cleavage were observed in unit 3.

Folded slaty cleavage is found in all three units and usually occurs as small-scale folds and crenulations, commonly with quartz–epidote sheets parallel to their axial planes (Fig. 3e). In a few exposures they occur in conjugate pairs (Fig. 3f) (Tobisch & Fiske 1976). Lineations of various kinds occur, and are usually associated with the slaty cleavage. They may be classified into two types (see Tobisch & Glover 1971, p. 2211): Type 1, parallel to fold axes, include bedding/cleavage intersections, boudinage necklines, mullions and striping. Type 2, parallel to the maximum extension direction in the rock, include streaking, elongate minerals (principally biotite) and alignment of the long axes of volcanic particles. The two most common lineations are streaking and bedding/cleavage intersections; fold and crenulation axes associated with both slaty cleavage and crenulations are less common linear features.

ORIENTATION OF STRUCTURAL ELEMENTS

We have divided the mapped area into four subareas so that variations in orientation of the structural elements can be easily grasped (Fig. 4).

Orientations of the earliest structures have not been plotted on the equal-area nets (Fig. 5) because of sparse data. In unit 1, the axial plane of these early folds strikes NNW (160°) and has a steep dip, while their axes have subhorizontal plunges (Fig. 3a). Axial planes of large-scale penecontemporaneous slump folds in unit 2 have strikes varying between NW–SE and NE–SW (see for example, Fig. 7) and axial plunges generally moderate to steeply to the south.

The orientation of the later structural elements is very similar in the late Paleozoic and the Triassic to Jurassic rocks, and hence they are both included in subarea 1 (Fig.

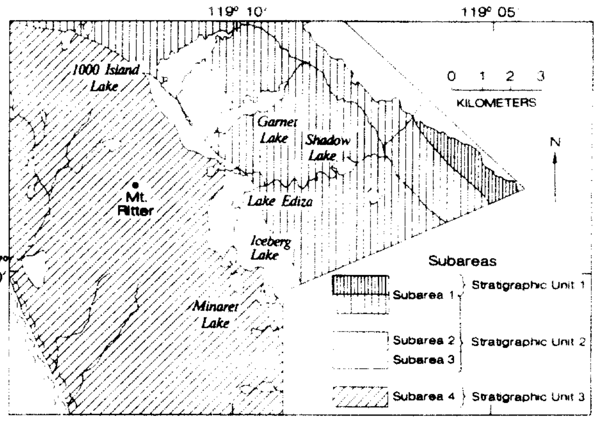

Fig. 4. Map showing the subareas used in analyzing the structural data from the three stratigraphic units.

5). Bedding (Fig. 5a) shows predominantly steep dips with some scatter as a result of mesoscopic folding, especially common in the late Paleozoic sedimentary rocks. A plot of poles to the slaty cleavage shows relatively tight groupings and shows that the cleavage strikes northwest and dips steeply (Fig. 5b). Lineations of type 1 associated with the slaty cleavage, however, show considerable variations (Figs. 5b & d). In unit 1 this variation in orientation is probably a result of superimposition of the cleavage on earlier folds, while in unit 2 field observations in a number of exposures suggest that the variation is more likely to be the result of differential flow in incompetent zones containing slate, marble and fine tuffs. This differential flow is interpreted as having rotated the fold axes in varying degrees towards the direction of maximum extension in regions of higher strain. Lineations of type 2 (maximum extension) fall into a reasonably tight grouping (Fig. 5c).

In subarea 2, the slaty cleavage and associated large-scale folds gradually swing around from their usual northwest orientation to a west–northwest or east–west, orientation (Figs. 6a & b and 7). This swing in cleavage and fold orientation is interpreted as resulting from the presence of large-scale pre-existing slump structures (Fig. 7) which have locally reoriented the strain ellipsoid to accommodate further deformation using a mimimum amount of energy (Flinn 1962, Tobisch 1967). Subsequent emplacement of the plutons in the immediate vicinity may have further modified the orientations of these later structures into a more easterly position than normal (Figs. 2 and 7).

In subarea 3 slaty cleavage is poorly developed or absent. This subarea is made up of rocks that occur along the wall of a mid-Cretaceous caldera (Fiske *et al.* 1977, in press) and we speculate that this large slab of Jurassic country rock has slumped en masse into the caldera. The relatively flat dip of bedding and slaty cleavage (Fig. 6c) is interpreted as having resulted from rotation of originally steeply dipping planar surfaces by slumping to their present gentler dips.

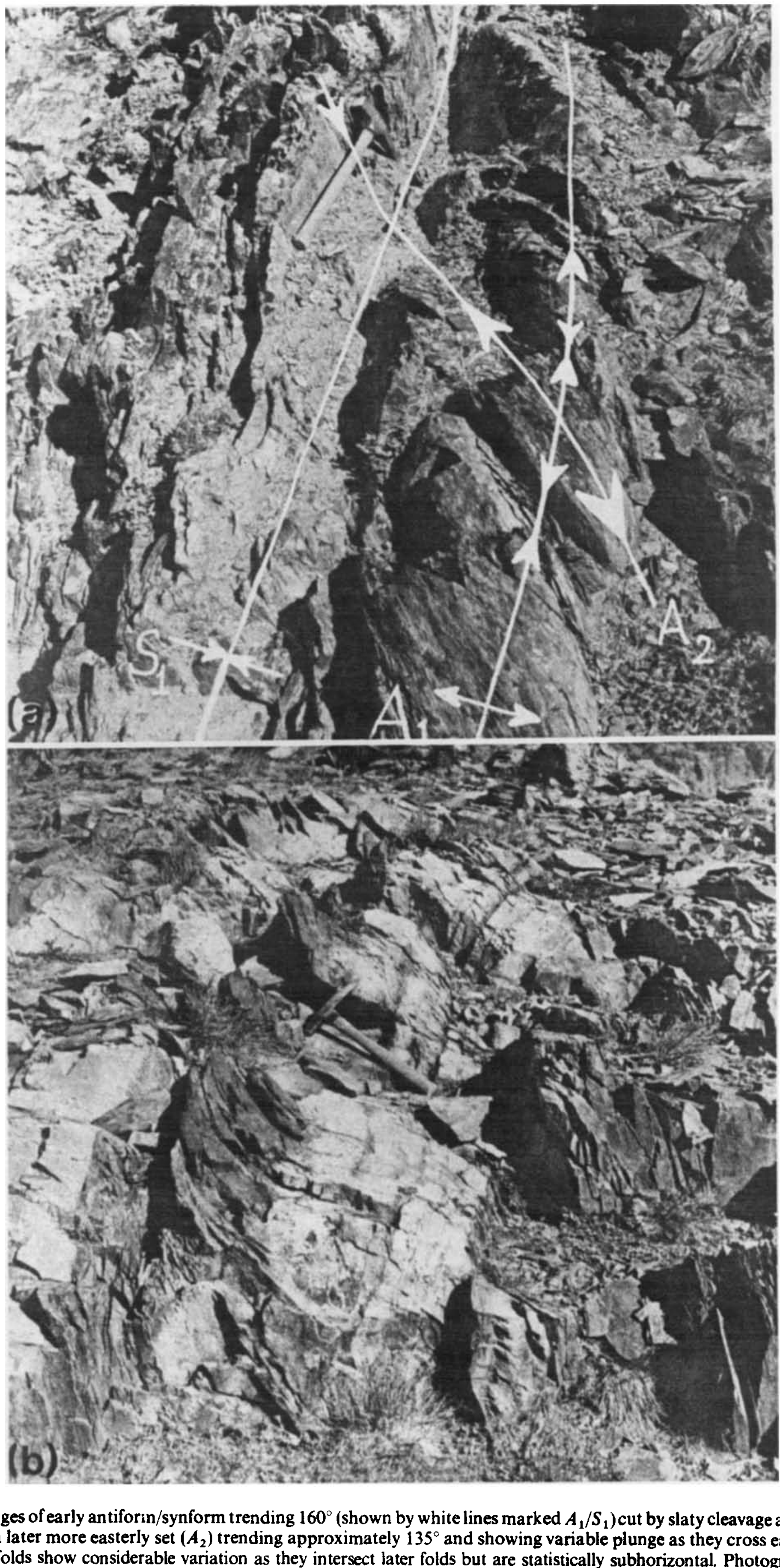

Fig. 3. (a) Hinges of early antiform/synform trending 160° (shown by white lines marked A_1/S_1) cut by slaty cleavage and small-scale folds of a later more easterly set (A_2) trending approximately 135° and showing variable plunge as they cross early folds. Axes of early folds show considerable variation as they intersect later folds but are statistically subhorizontal. Photograph from late Paleozoic rocks of unit 1. Hammer handle in upper part of photograph is approximately 35 cm in length. (b) Slaty cleavage refracting through volcaniclastic rocks of unit 2, consisting of tuffaceous slates and sands. Hammer handle about 30 cm in length.

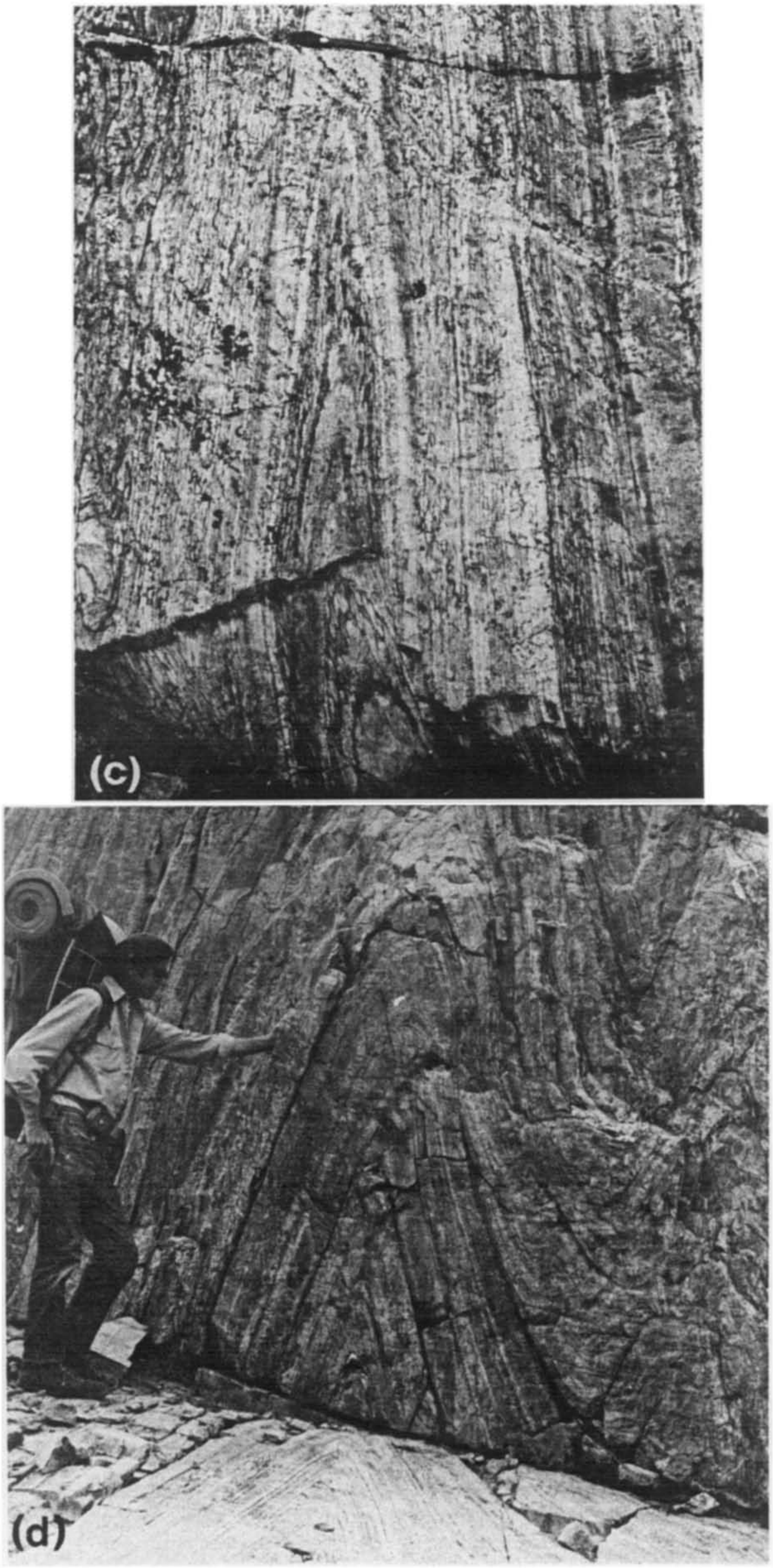

Fig. 3. (c) Very tight, nearly isoclinal minor fold in tuffaceous slate from a zone of high strain in the lower part of unit 2. Pencil near top of photograph is about 8 cm in length. (d) Tight minor fold in bedded tuffaceous rocks in the upper part of unit 2 where shortening is about half that shown in Fig. 3(c).

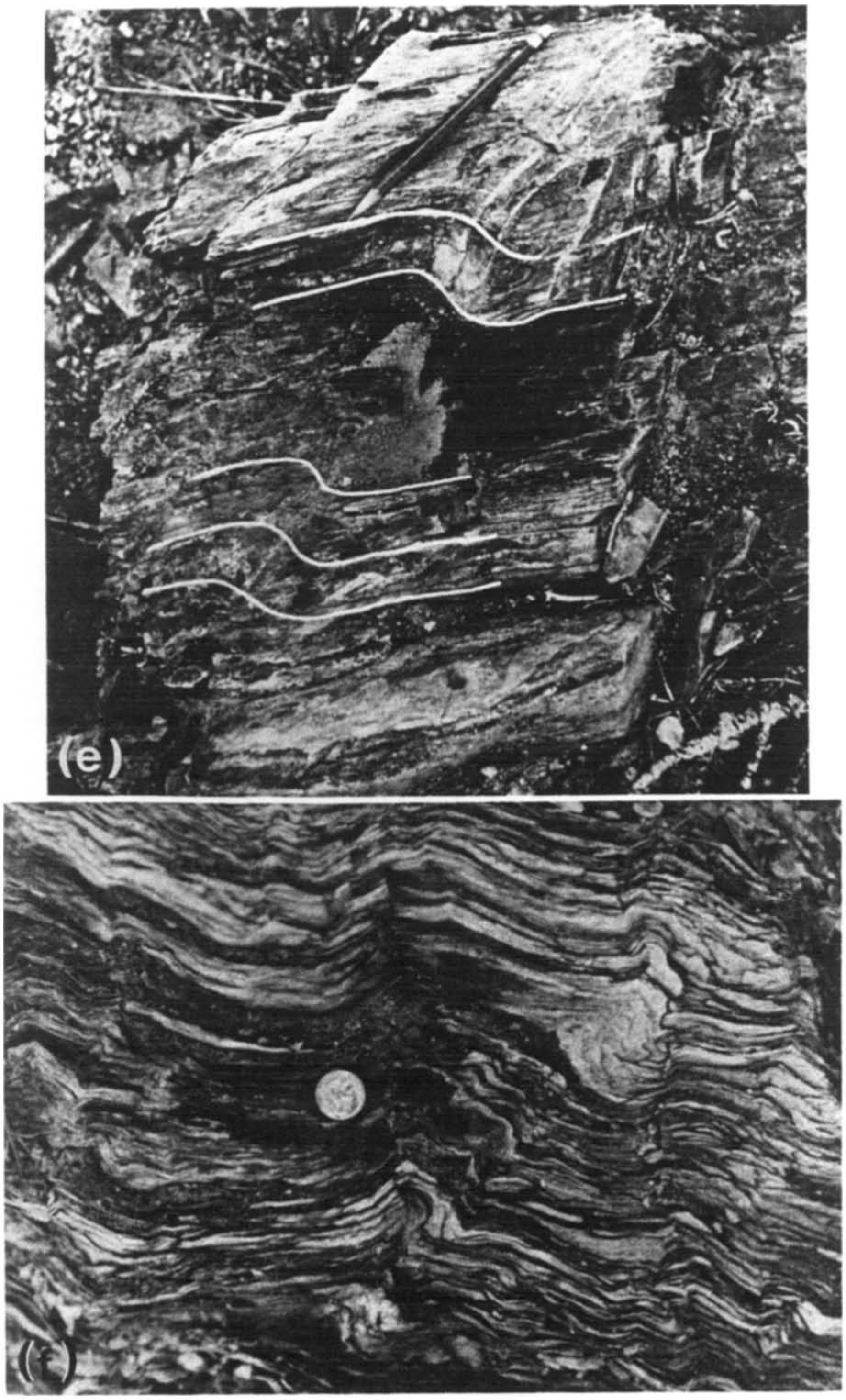

Fig. 3. (e) Small-scale fold and crenulations (Z-symmetry) folding the slaty cleavage in black slate of unit 1. Some form surfaces outlined in white. Note quartz-epidote layers lying parallel to the axial plane of the folds, a common feature in parts of the area. Pencil is about 25 cm in length. (f) Conjugate crenulations developed in a zone of intense slaty cleavage development in unit 2 near its base. Coin is about 2.5 cm in diameter.

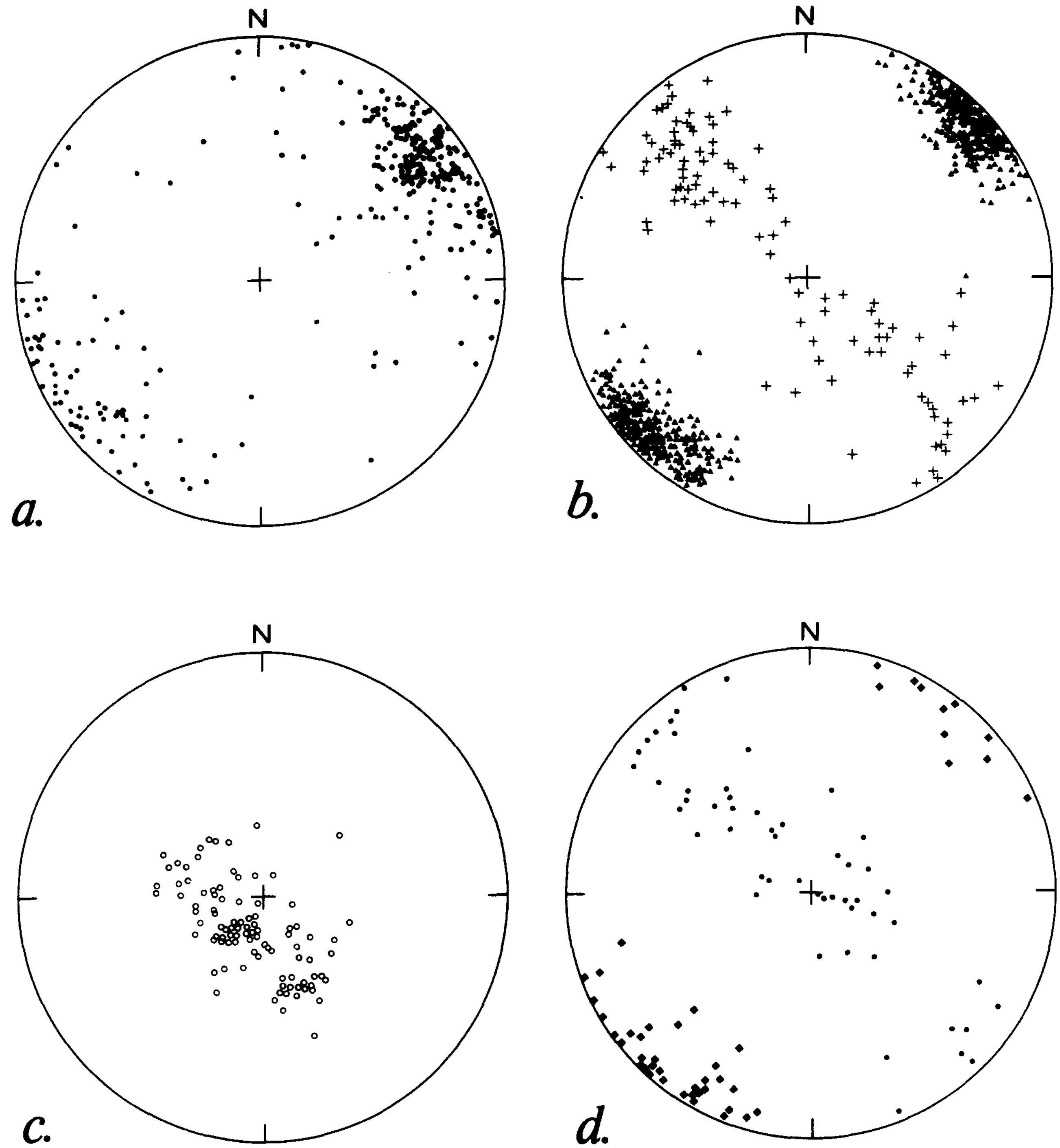

Fig. 5. Equal area projections from subarea 1. (a) poles to bedding; (b) poles to slaty cleavage (triangles) and lineations of type 1 shown by crosses (see text); (c) lineations of type 2 (see text); (d) poles to axial planes of minor folds associated with slaty cleavage (diamonds) and their axes (filled dots).

Most of the tectonic structures described from subareas 1–3 also occur in subarea 4 with two notable differences: (a) folds associated with the slaty cleavage are uncommon as are crenulations of the slaty cleavage. The latter occur only in a zone of intense slaty cleavage development along the western boundary of subarea 4 (Fig. 2); (b) while subarea 1 and to a lesser degree parts of subarea 2 show a pervasive development of slaty cleavage, subarea 4 contains large areas where slaty cleavage is either spottily developed or entirely absent. This domainal character of the slaty cleavage in subarea 4 is not dependent on lithology and appears to be a function of the mechanics of deformation.

The orientation of structural elements in subarea 4 (Figs. 8a & b) is very similar to that of subarea 1 (Fig. 5), with the exception that the dip of planar elements and plunge of lineations are not as steep as those in subarea 1. The scatter of bedding orientation (Fig. 8a) is the result of mesoscopic folding as well as gravity slumping during formation of the megabreccia that makes up a large part of unit 3 (Fiske & Tobisch 1978, Fiske *et al.* in press).

Crenulations and small-scale folds that fold the slaty

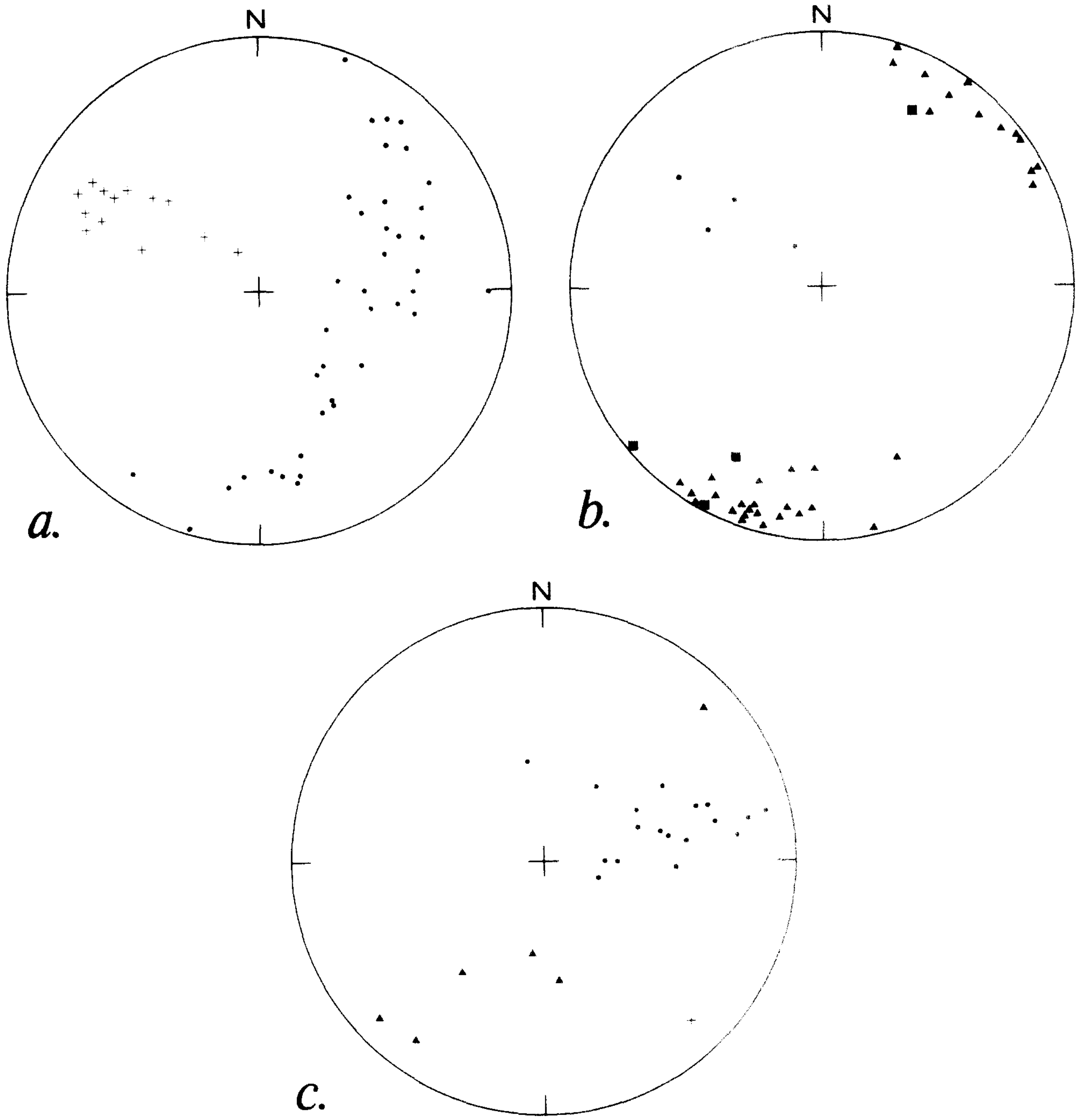

Fig. 6. Equal area projections from subareas 2 and 3. (a) poles to bedding (dots) and lineations of type 1 (crosses); (b) poles to slaty cleavage (triangles), poles to axial planes of minor folds associated with slaty cleavage (squares) and their axes (dots); (c) poles to bedding (dots), poles to slaty cleavage (triangles) and lineation of type 1 (cross). See text for explanation.

cleavage are present in subareas 1 and 4, and appear to be best developed in zones of intense slaty cleavage development. As reported elsewhere, we have recorded two sets of crenulations which appear to have formed contemporaneously, since they occur in conjugate relation in a number of exposures, and they have never been observed to mutually overprint each other (compare Tobisch & Fiske 1976). The average strike of the axial planes of the two sets have westerly and northerly orientations (Figs. 8c & d), and are associated with *Z*- and *S*-sense of movements, respectively. The scarcity of crenulation data from sub-area 4 reflects the fact that the intense slaty cleavage necessary for the formation of crenulations is present only over a relatively small area compared to subarea 1.

QUANTITATIVE STRAIN DATA AND THEIR INTERPRETATION

In an earlier study, we presented a detailed analysis of the quantitative strain the rocks have undergone in part of subarea 1 (Tobisch *et al.* 1977). We have extended this study with twelve more determinations, six from mid-Cretaceous rocks (unit 3) and the remaining six from parts of the Triassic to Jurassic section (unit 2). The measured and calculated data are given in Fig. 9 and Table 1. The extreme scatter in strain magnitude and the lesser scatter in the symmetry component is a function of both the rock type (finer-grained horizons taking up more strain than breccias, sills, etc.) and a true variation in total strain

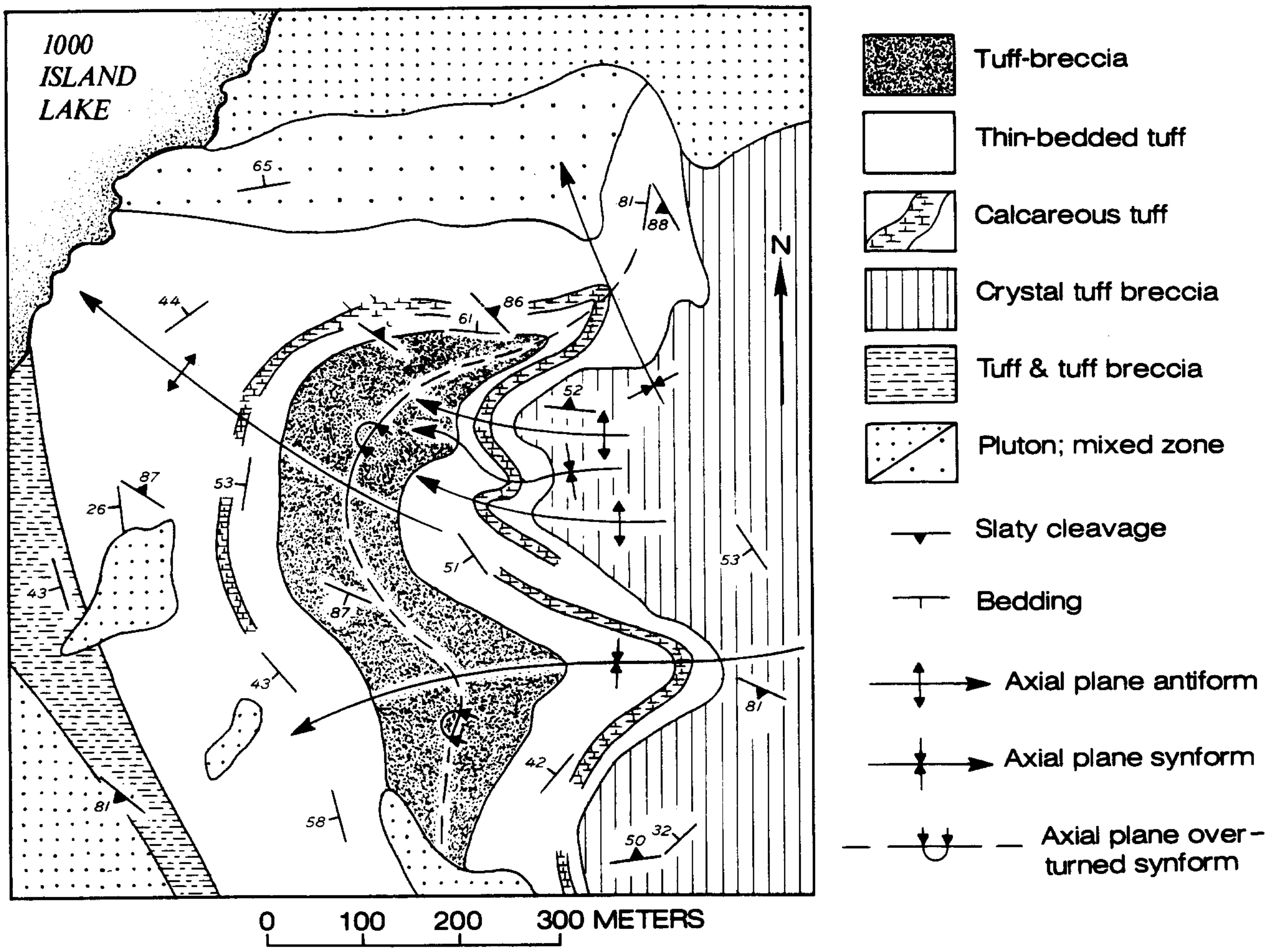

Fig. 7. Simplified geological map of area from unit 2 between Garnet and Thousand Island Lakes (see Fig. 2) showing change in orientation of axial planes of later folds as a result of various factors including refolding of a large-scale slump fold (see text).

across the area. The intense strain (84% shortening) recorded in specimens 8 and 9, for example, was determined from accretionary lapilli in tuffaceous slate and it is noteworthy that, even under such high strains, the shape of the accretionary lapilli is retained with enough clarity to be measured accurately. In a generalized way, the symmetry component of the mid-Cretaceous rocks (samples 1–6) tends to lie very close to the line of plane strain, while in the Triassic and Jurassic rocks (samples 7–12), the symmetry of deformation has a greater component of flattening as demonstrated in our earlier study (Tobisch *et al.* 1977, fig. 19a). There are not enough samples to define a possible deformation path more specifically.

In Fig. 10, we have produced a map which shows the variations in flattening strain (shortening normal to the slaty cleavage) that takes place across units 1, 2 and 3 (identified in Fig. 2). Given the sparse data in certain areas, it seems most meaningful to represent the variations in strain by three zones of deformation, each zone covering a range in values of shortening. With the data from our previous work (Tobisch *et al.* 1977) and the data presented here (Table 1, Fig. 9), we have extended the deformation zones along strike into areas where there are no 'hard data' and have based the delineation of these extended zones on the qualitative observation of innumerable exposures in the field. These observations included visual estimations of (1) axial ratios of strain markers, (2) degree of alignment of long axes of strain markers parallel to cleavage and (3) degree of development of slaty cleavage. Close observations using these criteria throughout the area and comparison with many specimens of known strain determined in our previous study, give us confidence that the extrapolation of the deformation zones across the area is reliable as a first-order approximation.

It is clear that the rocks of units 1 and 2 (Pennsylvanian and Permian(?) to Late Jurassic) have undergone an overall higher degree of penetrative strain than the rocks of unit 3 (mid-Cretaceous). With the data presented in our earlier study (Tobisch *et al.* 1977) as well as that presented here (Table 1, specimens 7–12), we estimate that the present 6–7 km combined thickness of units 1 and 2 was 12–14 km in thickness prior to the onset of slaty cleavage formation (for details of the method, see Tobisch *et al.* 1977, p. 36). Strain determinations in unit 3 are few and, therefore, reconstruction of the amount that deformation has thinned the mid-Cretaceous section can only be approximated. Based on our 6 strain determinations and innumerable qualitative observations in the rocks of unit 3, we estimate that the present 4-km thick section was approximately 4.5 km thick prior to deformation, with

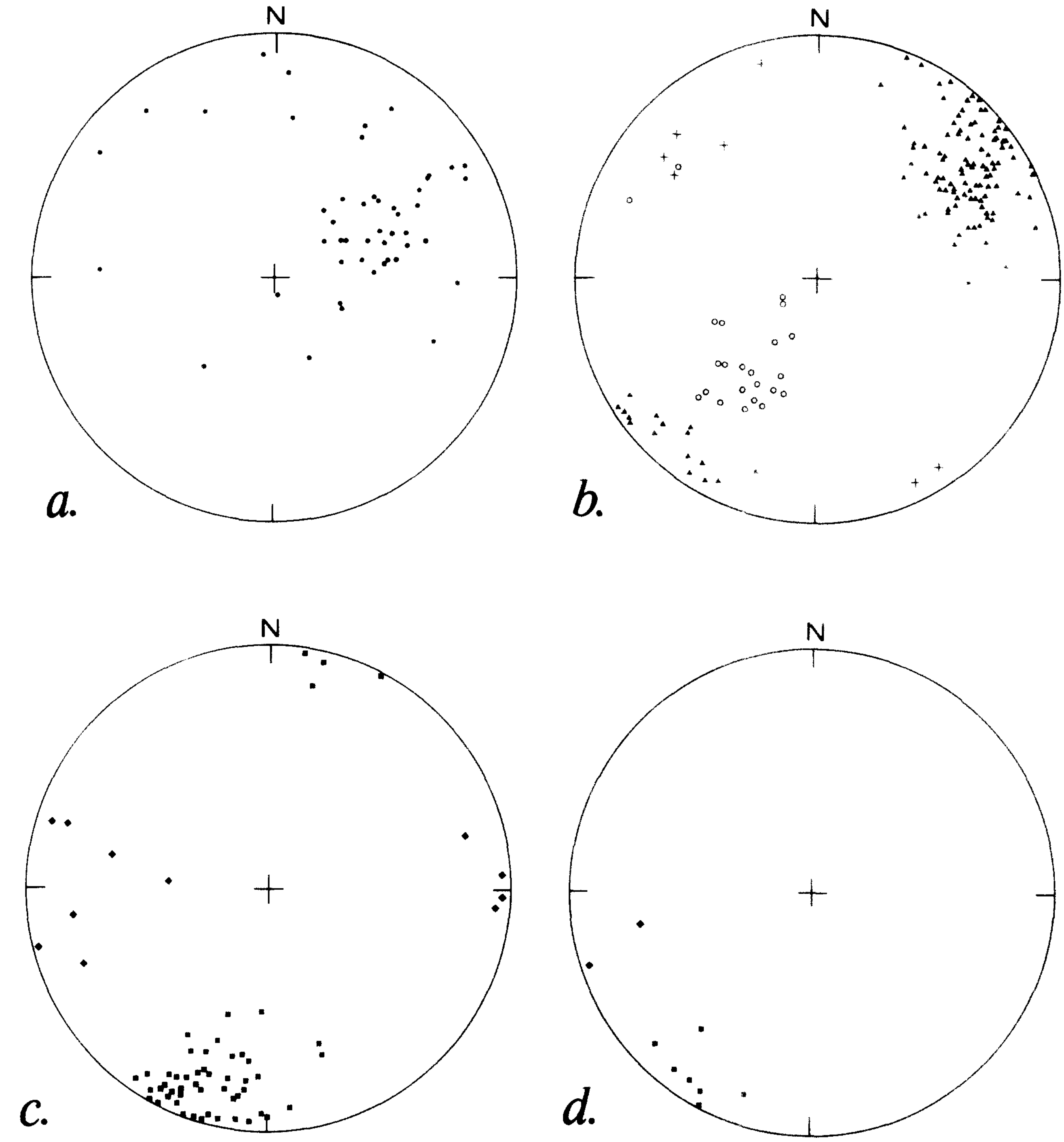

Fig. 8. Equal area projections from subareas 1–4. Subarea 4, (a) poles to bedding (dots); (b) poles to slaty cleavage (triangles), lineations of type 1 (crosses) and type 2 (open circles); (c) subareas 1–3, poles to axial planes of crenulations showing *Z*-sense of movement (squares) and *S*-sense of movement (diamonds); (d) subarea 4, designation same as (c).

most of the thinning taking place in zones where shortening exceeds 20% (Fig. 10). In summary, the present 10–11 km thickness of the entire stratigraphic section (units 1–3) was approximately 17–19 km prior to the deformations which affected these rocks.

STRUCTURAL CHRONOLOGY

In the preceding pages, we have attempted to describe the various structures with a minimum of labelling as regards their relative or 'absolute' ages. We have adopted this approach for specific reasons. In the stratigraphic units shown in Fig. 2, slaty cleavage and associated folds have a similar statistical orientation across the entire area; orientations of axial planes of the two sets of crenulations are also comparable between subareas 1 and 4 (Figs. 5, 6 and 8). Fossil and radiometric ages of the rocks of all three units make it certain, however, that these essentially parallel structures are not everywhere of the same age.

The mid-Cretaceous age determined from rocks of unit

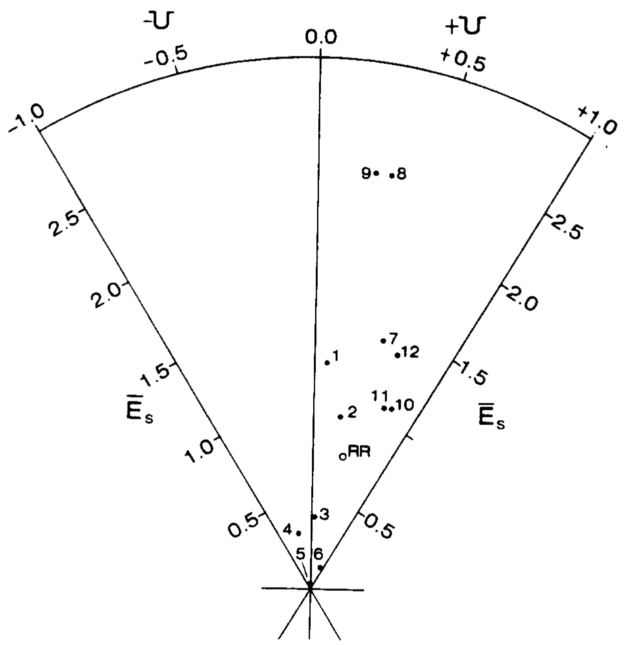

Fig. 9. Hsü diagram showing the magnitude and symmetry of strain of 12 specimens from units 2 and 3, (see Table 1 and Fig. 10). RR represents average strain ellipsoid in Ritter Range rocks calculated from data given in Tobisch *et al.* (1977).

3 requires that the deformation there must be no older than mid-Cretaceous. On the other hand, the late Paleozoic to Late Jurassic rocks of units 1 and 2 leave a greater time span for their deformation. Indeed, a Late Jurassic (Nevadan) deformation has long been held to be the major event responsible for structures observed in the central Sierra Nevada (Knopf 1929, Taliaferro 1942, Bateman *et al.* 1963, Best 1963, Bateman & Clark 1974, Schweickert & Cowan 1975). Conclusions as to the age of the Nevadan deformation by these and other workers have been based upon stratigraphic and structural observations, while more recent work also integrates radiometric age dating of the rocks. These types of data clearly indicate that on a regional scale much of the Sierra Nevada had been subject to at least one, and in various places two or more, periods of deformation prior to Cretaceous time, and that these deformations were accompanied by cleavage formation (e.g. Sharp & Saleeby 1979, Merguerian & Schweickert 1980, Nokleberg & Kistler 1980). In the immediate area of the Ritter Range, the unconformity between units 2 and 3 indicates that there had been deformation before mid-Cretaceous. While we cannot unequivocally demonstrate from the present data that slaty cleavage accompanied deformation, we think it is highly likely, judging from the regional history of the central Sierra Nevada and other indirect evidence from the Ritter Range area. Slaty cleavage in units 1 and 2, therefore, probably formed during both Jurassic and mid-Cretaceous times. We will now consider the evidence.

Poles to slaty cleavage in the late Palaeozoic to Late Jurassic rocks in units 1 and 2 (subarea 1) have a statistical orientation very close to that of poles to slaty cleavage in the mid-Cretaceous rocks of unit 3 (subarea 4; Figs. 11a & b). If we take the poles to slaty cleavage as the Z-axis and type 2 lineations as approximating the X-axis of the strain ellipsoid ($X > Y > Z$), the axes of the bulk strain ellipsoid determined from mid-Cretaceous rocks is essentially parallel to that determined in the older rocks of units 1 and 2 (Fig. 11c). From this relationship, we might assume that the cleavage in all units is the same age, but radiometric age determination (as well as regional considerations) indicate otherwise. Midway in unit 2, two felsic dykes have been emplaced into the section in an orientation that parallels the mean slaty cleavage direction which cuts across the bedding at a small angle in that area (Figs. 2 & 12). Further to the north the larger of the two dykes becomes parallel to bedding, while the smaller dyke continues parallel to cleavage. It seems improbable that these dykes would be emplaced at a small angle to bedding if there were no preferred fabric in that direction. Indeed, such igneous bodies will seek out the avenues of least resistance as they are emplaced, most likely along

Table 1. Strain data and their interpretation on 12 specimens, six from the mid-Cretaceous rocks and the remaining six from the Triassic to late Jurassic rocks. Samples were analysed as outlined in Tobisch *et al.* (1977). Location of the specimens is shown on Fig. 10

Spec. No.	Strain (ε) on plane XZ	XY	YZ	Internal consistency	$\bar{\varepsilon}_s$	ν	% Change in axes of strain ellipsoid X	Y	Z	
1	0.91	0.41	—	—	1.3	0.10	140	6	−61	
2	0.59	0.24	0.45	0.10	0.99	0.30	87	15	−53	
3	0.29	0.14	—	—	0.41	0.04	37	1	−26	Strat. unit 3
4	0.23	0.15	—	—	0.33	−0.30	30	−4	−18	mid-Cretaceous
5*	~0	~0	—	—	0	0	0	0	0	
6	0.09	0	—	—	0.14	1.0	7	7	−11	
7	0.84	0.25	0.70	0.11	1.4	0.48	123	35	−67	
†8	1.67	—	1.11	—	2.4	0.33	350	47	−84	
†9	1.68	—	1.06	—	2.4	0.26	375	37	−84	Strat. unit 2
10	0.70	0.09	—	—	1.1	0.73	72	41	−58	Triassic to
11	0.73	0.11	—	—	1.1	0.70	77	41	−59	Late Jurassic
12 EII	0.86	0.22	0.70	0.06	1.4	0.52	115	38	−66	

* Strain measured on both XZ and XY planes was less than the average "apparent strain" for undeformed lithic lapilli tuff; see Tobisch *et al.* 1977, Table 1, p. 28.

† Accretionary lapilli; EII, ash flow ellipsoid, type II (see Tobisch *et al.* 1977, pp. 31–32); strain measured from lithic lapilli in all other specimens.

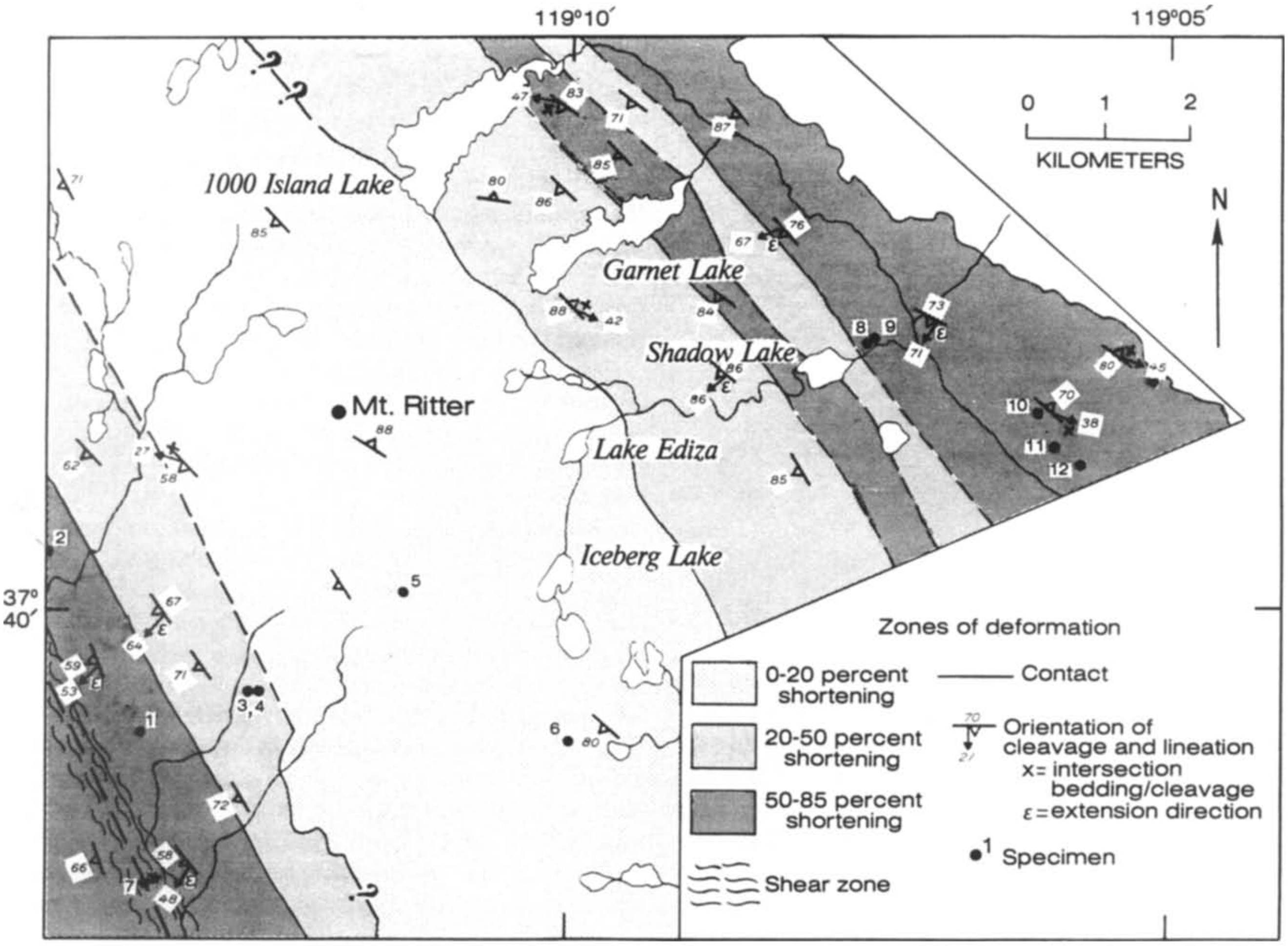

Fig. 10. Map of area (plutons removed) showing variation in shortening normal to cleavage that occurs (see text). Zones may contain some localized areas where shortening is either greater or less than that indicated on the figure.

bedding, joint or fault surfaces. No obvious joint surfaces are oriented at a small angle to bedding in the area, but slaty cleavage, and locally faults which parallel the cleavage, do have such an orientation. From the above discussion, therefore, we think that the most plausible interpretation of the dyke orientation is that they were emplaced parallel to a pre-existing fabric in the rock, in this case slaty cleavage. The age of the pre-existing fabric (slaty cleavage), however, could be either Jurassic or possibly mid-Cretaceous, having formed in large part just before dyke emplacement. The intensity of cleavage development, however, may help resolve this problem of age. Slaty cleavage is less well-developed in the dykes than in the surrounding host rock. While this may be accounted for in part by competency differences or timing of emplacement of the dykes relative to cleavage formation, strain measurements indicate that the Cretaceous rocks (unit 3) are overall much less deformed than the Jurassic and older rocks (units 1 and 2). Mean shortening in unit 2, for example, approaches 45% (Tobisch *et al.* 1977), with extreme shortening (84%) and elongation (375%) occurring in some zones (Figs. 9 & 10; Table 1). Cretaceous deformation in unit 3, although determined from considerably less samples, produced a mean shortening which is estimated at less than 15% when the weakly to undeformed parts of the section are considered.

Our interpretation is that the pre-existing fabric (slaty cleavage) in the host rock is most likely to be of Jurassic age, which is also in keeping with the regional structural history established in adjacent areas. A U/Pb age of the larger dyke which intrudes along this earlier cleavage has been dated as mid-Cretaceous (101 Ma, believed to be good to $\pm 5\%$, T. W. Stern, written communication, 1976). From these data, we interpret the relatively weak slaty cleavage in the dykes to have formed during the mid-Cretaceous deformation, while the relatively strong cleavage in the host rock reflects the cumulative strain recorded during statistically parallel Jurassic and mid-Cretaceous periods of deformation (Fig. 11). Extending this interpretation to the area as a whole, we think the evidence suggests that units 1 and 2 have been subjected to two periods of parallel slaty cleavage formation, separated by a considerable period of geologic time. The strain in unit 3, however, represents only the mid-Cretaceous deformation.

Radiometric ages (U/Pb) of rocks in unit 3 cluster around 100 Ma (Fiske & Tobisch 1978), whereas tectonically undeformed plutons of the Tuolomne Intrusive

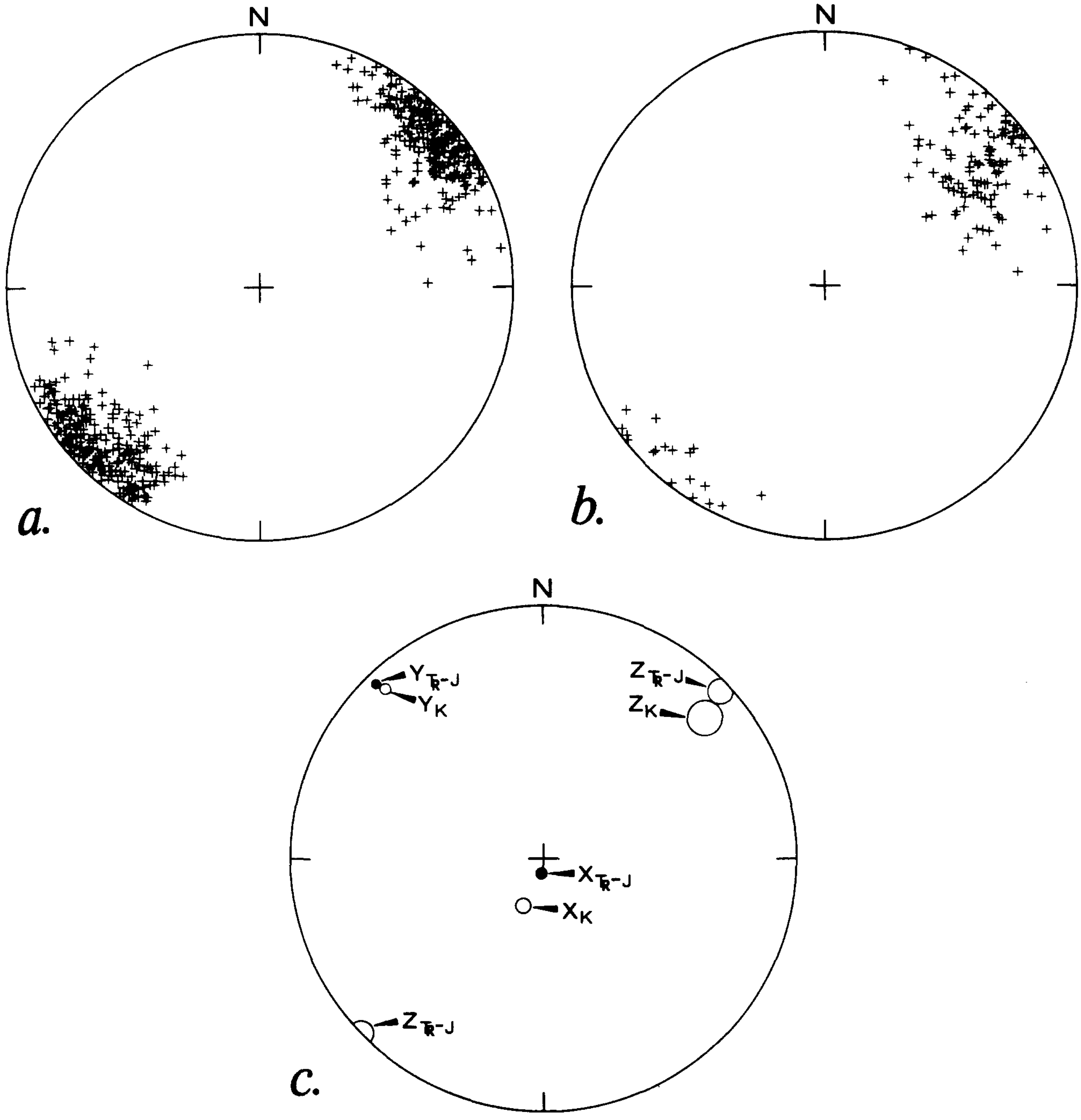

Fig. 11. Equal area projections showing: (a) and (b) Poles to slaty cleavage from subarea 1 (a) and subarea 4 (b); (c) Axes of the strain ellipsoid as determined from data in Figs. 5, 8 and 11, showing the nearly co-axial nature of the strain ellipsoid associated with deformations in Late Jurassic and mid-Cretaceous. Size of circles of the Z-axis indicates the 95% confidence level obtained by statistical analysis of the population of points shown in Figs. 11a & b.

Suite, dated at 90–85 Ma, have been observed to cut deformed rocks of mid-Cretaceous age immediately west of the area shown in Fig. 2 (D. L. Peck, personal communication 1979, Stern *et al.* 1981). These data suggest that the mid-Cretaceous deformation in the area took place about 92 $\pm$ 7 Ma. The absolute age of the late Jurassic (Nevadan) deformation in part of the west-central Sierra Nevada has been recently refined through extensive field work and isotopic dating by a number of workers. Their data suggests a relatively short but wide-spread deformational period between 150–140 Ma (Stern *et al.* 1981, P. C. Bateman, personal communication 1980, Warren Sharp, personal communication 1981). This interpretation if it can be extended to the eastern part of the central Sierra Nevada (Ritter Range), suggests that there was a hiatus of 40–60 Ma between the two parallel deformations that affected the rocks of units 1 and 2.

DISCUSSION

Over the last three decades, extensive structural work in orogenic belts around the world has revealed a distinct pattern: early deformations, usually accompanied by

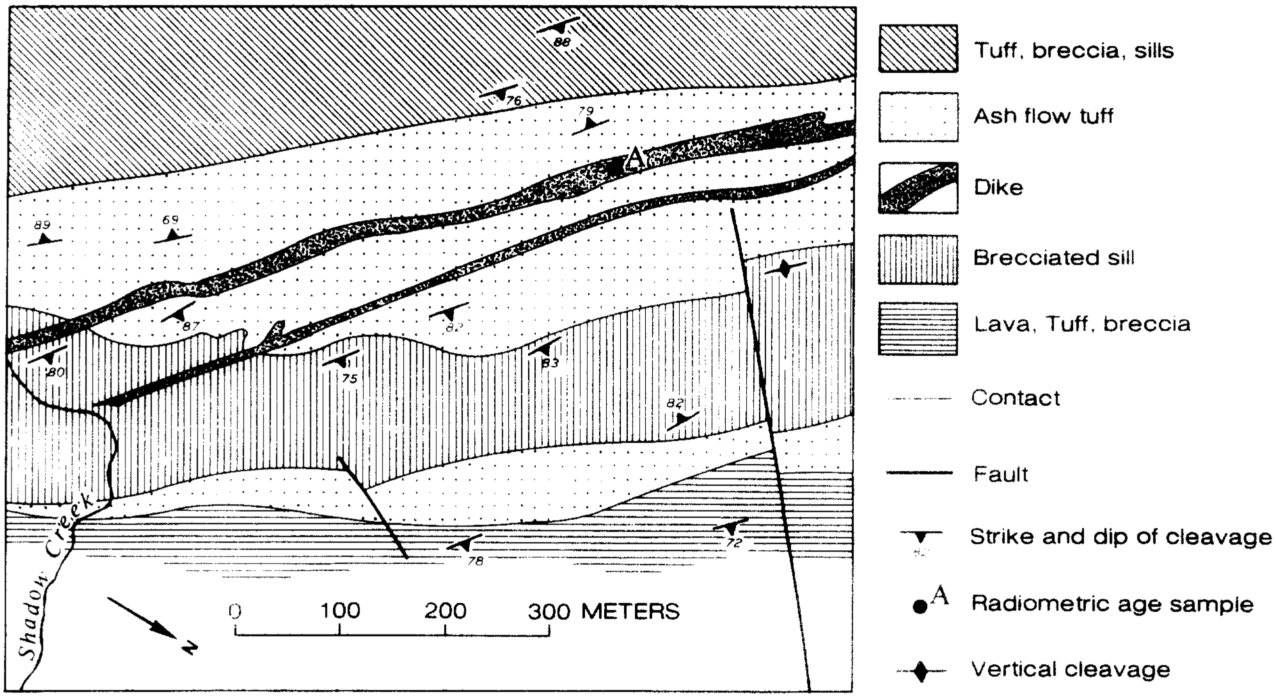

Fig. 12. Greatly simplified geological map from the middle part of unit 2 (see Fig. 2), showing the cross-cutting nature of the dykes and their essentially parallel orientation to slaty cleavage in the area.

extensive moderate, to sometimes higher-temperature metamorphism, form penetrative slaty cleavages or schistosities, while later deformations, accompanied by less extensive, often lower-temperature metamorphism, usually form crenulations of early formed planar structures. Cleavages and axial planes of folds with parallel orientations are usually considered to be contemporaneous. While there may be various geological cases which do not fit such a simple model, this sequence is well-established in the geological literature and is commonly used to distinguish successive periods of deformation in multideformed orogenic belts. The more exact knowledge provided by radiometric age dating, however, is allowing the field geologist to see more clearly the spatial and chronological relationships of structures mapped in the field. Detailed mapping and radiometric ages determined in rocks of the present area (Fiske & Tobisch 1978, this paper) have provided us with a chronological framework which suggests that the well-established model for setting up the structural succession is not suitable for accurately determining the structural evolution of this area.

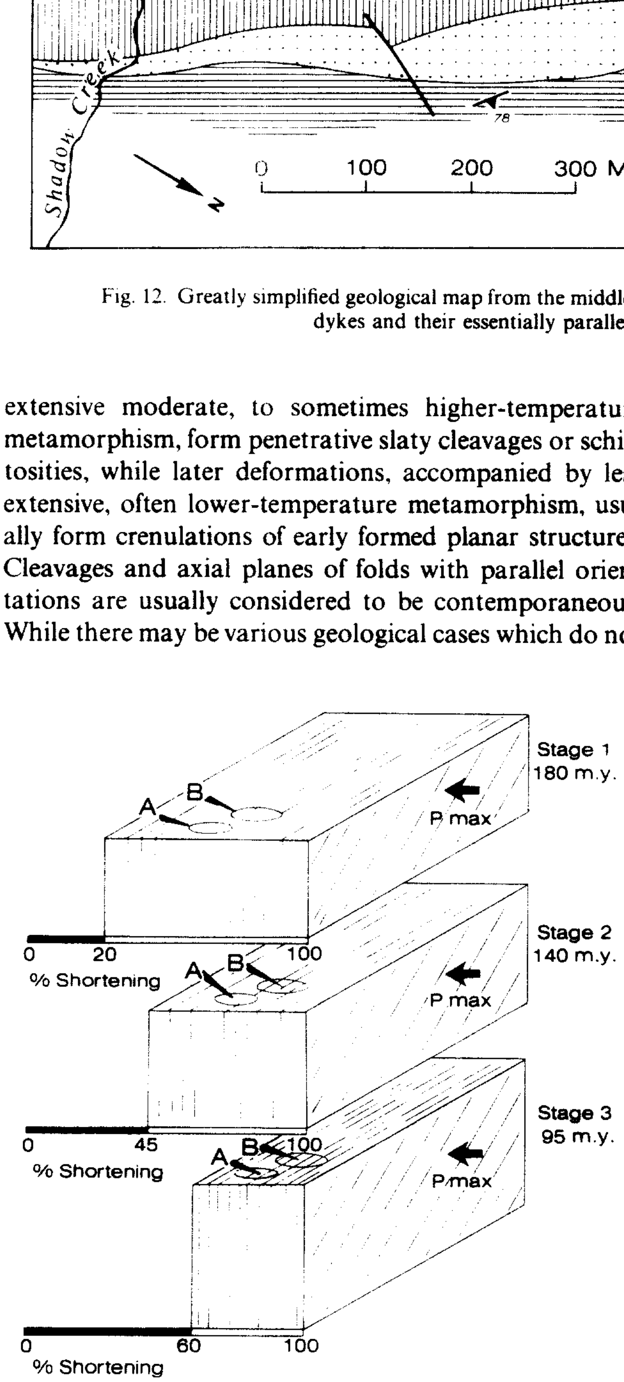

Fig. 13. Diagrammatic illustration showing the progressive development of slaty cleavage by repeated parallel deformation in three stages (see text).

To illustrate this point, let us consider a model showing a sequence of events (Fig. 13) in which a block of rock undergoes three successive periods of deformation where the orientation of the strain ellipsoid remains essentially constant. Each period of deformation adds about 15–25% shortening to the rock body. Possible times of deformation are shown on the right. In stage I, the rock has been shortened about 20% resulting in domainal development of slaty cleavage. Area B shows some cleavage formation and the development of open folds, while area A, possibly only tens of metres away, still appears undeformed. Such a scenario, of domainal cleavage development on a large scale, is probably a common mechanism by which cleavage develops over larger areas and is well shown in parts of the mid-Cretaceous rocks, within the zone of 0–20% shortening illustrated in Fig. 10. In stage II, 40 Ma later, the rocks are again deformed (co-axially) adding 25% shortening to the bulk strain. Existing cleavage domains are expanded and intensified, while new domains are formed. Area B has shortened further, with more cleavage development and tightening of existing folds, while cleavage in area A, for various reasons, remains poorly developed. In stage III, 45 Ma later, a further 15% shortening is added co-axially. Expanding cleavage domains have now mostly coalesced so that the entire rock body shows cleavage and/or folding at various degrees of intensity. Area B shows cleavage which represents in excess of 50% shortening, while area A, having absorbed less total strain, shows cleavage and fold development commensurate with 25–30% shortening. Orientation of slaty cleavage in areas A and B is statistically parallel, as it is throughout the whole rock body. Yet the time of formation of the cleavage and folds in the two adjacent areas (possibly only metres apart) may be vastly different, locally as much as 85 Ma. Labelling the cleavage/folding in the area as S_1/F_1 for example, would be very misleading, considering that the bulk strain is the product of three distinct periods of deformation separated by marked periods of time.

We suggest that the sequence of events outlined above and in Fig. 13 may be applicable to unit 1 and the lower part of unit 2 (Fig. 2). Two such stages for these rocks have now been documented, although the suggested amount of bulk shortening during each stage is only an estimate. Recent work in the west-central Sierra Nevada indicates still earlier stages of deformation prior to and during the Early and Middle Jurassic (Sharp & Saleeby 1979, Merguerian & Schweickert 1980). It is entirely possible that prior to deposition of the Mid- to Late Jurassic rocks, a pre-Middle Jurassic deformation affected unit 1 and the lower part of unit 2 as suggested in Fig. 13. Indeed, results of our earlier study (Tobisch *et al.* 1977, Fig. 2b) indicate that the bulk shortening in the Triassic and Early Jurassic rocks of zone 1 is approximately 60%, while shortening in the Mid- to Late Jurassic rocks of zones 4–6 is approximately 35%. As mentioned earlier, our estimation for bulk shortening in the mid-Cretaceous rocks of unit 3 is about 15%. From these data, it would appear reasonable to suggest that the total strain shown in the Triassic and Early Jurassic rocks (eastern part of unit 2) may be the cumulative result of three successive nearly parallel deformations, possibly taking place in increments of about 15–25% shortening in each deformation pulse (compare Fig. 13). Rocks of unit 1 very likely underwent a similar strain history, although the total strain includes that derived from deformation which preceded the deposition of the Triassic to Early Jurassic volcanic rocks which overlie it.

In addition to the problems of identifying the age of slaty cleavage, the presence of conjugate as well as single sets of crenulations and small-scale folds of the slaty cleavage needs to be considered. It is widely accepted that slaty cleavage in most instances forms essentially normal to the maximum compression direction (P_{max}). Crenulations showing layer-parallel shortening, on the other hand, whether occurring in conjugate or single sets, are formed with P_{max} lying parallel or subparallel to the cleavage (or surface being folded). This has been amply documented in numerous experimental studies as well as field examples (e.g. Paterson & Weiss 1966, Means & Williams 1972, Tobisch & Fiske 1976). Crenulation and folding of slaty cleavage is a common phenomenon in orogenic belts around the world. In an area where repeated co-axial deformations has formed parallel slaty cleavages such as shown in the central Sierra Nevada, when did the crenulations form? Are we to believe that after a slaty cleavage has formed by a succession of two or three co-axial compressional phases where P_{max} is oriented normal to the length of the mountain belt, that P_{max} suddenly rotates 90° into a position parallel to the belt (and to the slaty cleavage) to form the crenulations? We prefer to view the formation of crenulations and accompanying folding of the cleavage as a late but integral part of each pulse of slaty cleavage development, as suggested in our earlier study (Tobisch & Fiske 1976). Whether or not the mechanism of "elastic recovery" we suggested is correct (Tobisch & Fiske 1976, p. 1419), the concept of a close temporal relation between slaty cleavage and subsequent crenulation or folding of the cleavage appears to us to be the most plausible model. We have observed the association of crenulations with fault zones in various parts of the central Sierra Nevada (cf. Clark 1960a, b). While this association is not necessarily ubiquitous in all parts of the Sierra Nevada or other orogens, it does suggest that the rocks may have been in a semi-brittle state in those zones during crenulation and fold formation, and that this environment may be appropriate to the "elastic recovery" model we have proposed.

Given the model outlined above and in our earlier work (Tobisch & Fiske 1976) for forming slaty cleavage and subsequent crenulations as one continuous process, let us reconsider the process outlined in Fig. 13. Each of the stages represents one distinct deformational pulse producing slaty cleavage, followed immediately by development of domains of crenulation or folding of that cleavage. It is immediately clear therefore that crenulations formed in area B of Fig. 13 during stage II would predate formation of at least part of the slaty cleavage formed in area A during stage III.

If such a sequence of events holds true, one might normally expect to observe overprinting relationships between later slaty cleavage formed in stage III and earlier crenulations formed in stage II. We have not observed such overprinting relationships and would suggest that their absence may be due to two factors: (1) crenulations occur only in very restricted domains, collectively far less than one per cent of the total outcrop area. This fact, in connection with the domainal formation of slaty cleavage as outlined, might render the chance overprinting of later slaty cleavage on earlier crenulations at any one point extremely rare and (2) rocks in which both slaty cleavage and later crenulations have formed will be approaching a state of 'strain hardening' relative to immediately surrounding areas which lack crenulations, simply because the former have absorbed more total strain. It does not seem likely, therefore, that later slaty cleavage would form in a relatively 'strain hardened' rock when it would take less energy to form in an immediately adjacent area where further strain could be accomplished with less resistance. Instead, the crenulated, relatively 'strain hardened' zone may simply rotate towards the existing cleavage direction, with some concomitant tightening of the crenulations in response to further compression normal to slaty cleavage. Such rotation would probably be very difficult to detect except in some unusual circumstances.

Finally, let us return to the model for forming slaty cleavage and subsequent crenulations outlined previously. It is our suggestion that while relative chronology between slaty cleavage and subsequent crenulations can be set up with reasonable assurance at any one point, our data and interpretations indicate that it is not possible to extend that structural succession to any substantial area as a whole. Indeed, an attempt to set up an S_1, S_2, S_3, etc. chronology applicable to the entire area, without extensive radiometric dating, would give an entirely misleading impression of the actual structural history of the area.

CONCLUSIONS

For the rocks of the Ritter Range which we have been considering, we may state with relative certainty that the structural succession in unit 3 (mid-Cretaceous) as a whole is slaty cleavage, followed by crenulations, both within the time period of 92 ± 7 Ma. For units 1 and 2 (late Paleozoic to Jurassic), it is possible to set up a succession from earlier structures to later structures, but it is not possible to extend the succession over a larger area to include the time relationships between subsequent slaty cleavage and crenulation or folding of the cleavage. As illustrated in Fig. 13, the later structures in any one outcrop could have been formed at various periods during successive stages of parallel deformation. It would seem necessary to lump these successive later tectonic structures under the general term 'Mesozoic' until such time that more viable methods are developed to elucidate the structural evolution with greater accuracy.

The questionable validity of correlating structures in time from outcrop to outcrop was treated some time ago by Park (1969) and enlarged upon by Roberts (1977). Our work further underlines the unsuitability of applying such correlations to the present area. In addition, detailed field work and extensive radiometric dating of rocks in the southwestern Sierra Nevada have demonstrated that successive deformations are represented essentially by one statistically parallel foliation (Saleeby 1978, Saleeby & Sharp 1980). Evidence in other orogens (e.g. the Variscan of southwest England (Roberts & Sanderson 1971) and the Blue Ridge of the Virginian Appalachians (L. Glover, personal communication 1981) also indicates that essentially similar structures (i.e. slaty cleavage) have developed in the same or adjacent zones at different periods of time. We suggest that multiphase, parallel structures, comparable to those we have described, may be a relatively common phenomenon in orogenic belts, but as yet remain undetected as a widespread phenomenon. While a suitable model for the valid representation of the structural evolution of multideformed belts has yet to be suggested, it is clear that one cannot necessarily set up a valid structural succession in a larger area based only on the geometrical relationship between structural elements, their respective orientations and the other types of data usually thought to be sufficient to delineate a phase of deformation. To establish a rigorous and detailed structural succession for any given area it may be necessary to gain a thorough understanding of the stratigraphy, the relative and absolute ages of the units, as well as carrying out radiometric dating of rock bodies which have key relationships to the structures involved. Until one arrives at this level of understanding, it will be possible only to delineate the broadest of time sequences for the structures involved.

Acknowledgements—We wish to acknowledge generous support for field work from the U.S. Geological Survey and from the Fluid Research Fund of the Smithsonian Institution. This manuscript was improved by reviews by P. C. Bateman and J. Saleeby.

REFERENCES

Bateman, B. C. & Clark, L. D. 1974. Stratigraphic and structural setting of the Sierra Nevada Batholith, California. *Pacif. Geol.* **8**, 79–89.

Bateman, P. C., Clark, L. D., Huber, N. K., Moore, J. G. & Rinehart, C. D. 1963. The Sierra Nevada Batholith—a synthesis of recent work across the central part. *Prof. Pap. U.S. geol. Surv.* **414D**, 1–46.

Best, M. G. 1963. Petrology and structural analysis of metamorphic rocks in the southwestern Sierra Nevada Foothills, California. *Univ. Calif. Publs geol. Sci.* **42**, 111–158.

Bogen, N. L. 1979. Four sets of structures in the U. Jurassic Mariposa Formation, Tuolomne County, California (abstract). *Geol. Soc. Am.* (Abs. with Programs Cord. Section) **11**, 70.

Clark, L. D. 1960a. Evidence for two stages of deformation in the western Sierra Nevada metamorphic belt, California. *Prof. Pap. U.S. geol. Surv.* **400B**, 316–318.

Clark, L. D. 1960b. Foothills fault system, western Sierra Nevada, California. *Bull. geol. Soc. Am.* **71**, 483–496.

Flinn, D. 1962. On folding during three-dimensional progressive deformation. *Q. Jl geol. Soc. Lond.* **118**, 385–433.

Fiske, R. S., Tobisch, O. T. & Peck, D. L. in press. Ashflow eruption and caldera collapse related to emplacement of the Sierra Nevada Batholith. *Bull. geol. Soc. Am.*

Fiske, R. S., Tobisch, O. T., Kistler, R. W., Stern, T. W. & Tatsumoto, M. 1977. Minarets caldera: a Cretaceous volcanic center in the Ritter Range pendant, central Sierra Nevada, California (abstract). *A. Meet. geol. Soc. Am.* **9**, 975.

Fiske, R. S. & Tobisch, O. T. 1978. Paleogeographic significance of volcanic rocks of the Ritter Range Pendant, central Sierra Nevada, California. In: *Mesozoic Paleogeography of the Western United States* (edited by Howell, D. G. & McDougall, K. A.) Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2.

Fleuty, M. J. 1964. The description of folds. *Proc. geol. Ass.* **75**, 461–492.

Huber N. K. & Rinehart, C. D. 1965. Geologic map of the Devils Postpile Quadrangle, Sierra Nevada, California. *U.S. Geol. Survey Map. CQ* 437.

Kistler, R. W. 1966. Structure and metamorphism in the Mono Craters Quadrangle, Sierra Nevada, California. *Bull. U.S. geol. Surv.* **1221E**, 1–53.

Knopf, A. 1929. The mother lode system of California. *Prof. Pap. U.S. geol. Surv.* **157**, 1–88.

Means, W. D. & Williams, P. F. 1972. Crenulation cleavage and faulting in an artificial salt–mica schist. *J. Geol.* **80**, 569–591.

Merguerian, C. & Schweickert, R. A. 1980. Superposed mylonitic deformation of the Shoo Fly Complex in Tuolomne County, California (abstract). *Geol. Soc. Am.* (Abst. with Programs Cord. Section) **12**, 120.

Morgan, B. A. & Rankin, D. W. 1972. Major structural break between Paleozoic and Mesozoic rocks in the eastern Sierra Nevada, California. *Bull. geol. Soc. Am.* **83**, 3739–3744.

Nolkeberg, W. J. & Kistler, R. W. 1980. Paleozoic and Mesozoic deformations in the central Sierra Nevada, California. *Prof. Pap. U.S. geol. Surv.* **1145**, 1–24.

Park, R. G. 1969. Structural correlation in metamorphic belts. *Tectonophysics* **7,** 323–338.

Paterson, M. S. & Weiss, L. E. 1966. Experimental deformation and folding in phyllite. *Bull. geol. Soc. Am.* **77,** 343–374.

Rinehart, C. D., Ross, D. C. & Huber, N. K. 1959. Paleozoic and Mesozoic fossils in a thick stratigraphic section in the eastern Sierra Nevada, California. *Bull. geol. Soc. Am.* **70,** 941–946.

Roberts, R. L. 1977. The structural analysis of metamorphic rocks in orogenic belts. In: *Energetics of Geological Processes* (edited by Saxena, S. K. & Bhattacharji, S.) Springer, N.Y., 151–168.

Roberts, R. L. & Sanderson, D. J. 1971. Polyphase development of slaty cleavage and the confrontation of facing directions in North Cornwall. *Nature, Lond.* **230,** 87–89.

Saleeby, J. 1978. Kings River ophiolite, southwest Sierra Nevada foothills, California. *Bull. geol. Soc. Am.* **89,** 617–636.

Saleeby, J. & Sharp, W. 1980. Chronology of the structural and petrologic development of the southwest Sierra Nevada foothills, California: Summary. *Bull. geol. Soc. Am.* **91,** 317–320.

Sharp, W. W. & Saleeby, J. 1979. The Calaveras Formation and syntectonic mid-Jurassic plutons between the Stanislaus and Tuolomne Rivers, California (abstract). *Geol. Soc. Am.* (Abst. with Programs Cord. Section) **11,** 127.

Schweickert, R. A. 1979. Structural sequence of the Calaveras Complex between the Stanislaus and Tulomne Rivers (abstract). *Geol. Soc. Am.* (Abs. with Programs Cord. Section) **11,** 127.

Schweickert, R. A. & Cowan, D. S. 1975. Early Mesozoic tectonic evolution of the western Sierra Nevada, California. *Bull. geol. Soc. Am.* **86,** 1329–1336.

Stern, T. W., Bateman, P. C., Morgan, B. A., Newell, M. F. & Peck, D. L. 1981. Isotopic U–Pb ages of zircon from the granitoids of the central Sierra Nevada, California. *Prof. Pap. U.S. geol. Surv.* **1185,** 1–17.

Taliaferro, N. L. 1942. Geologic history and correlation of the Jurassic of southwestern Oregon and California. *Bull. geol. Soc. Am.* **53,** 71–112.

Tobisch, O. T. 1967. The influence of early structures on the orientation of late-phase folds in an area of repeated deformation. *J. Geol.* **75,** 554–564.

Tobisch, O. T. & Glover, L., III. 1971. Nappe formation in part of the southern Appalachian Piedmont. *Bull. geol. Soc. Am.* **82,** 2209–2230.

Tobisch, O. T. & Fiske, R. S. 1976. Significance of conjugate folds and crenulations in the central Sierra Nevada, California. *Bull. geol. Soc. Am.* **87,** 1411–1420.

Tobisch, O. T., Fiske, R. S., Sacks, S. & Taniguchi, D. 1977. Strain in metamorphosed volcaniclastic rocks and its bearing on the evolution of orogenic belts. *Bull. geol. Soc. Am.* **88,** 23–40.

Turner, F. J. 1968. *Metamorphic Petrology.* McGraw-Hill Book Co., Maidenhead, U.K.