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PRESSURE SOLUTION AND THE DEVELOPMENT OF
CLEAVAGE IN THE BARABOO QUARTZITE

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PRESSURE SOLUTION AND THE DEVELOPMENT OF CLEAVAGE IN THE BARABOO QUARTZITE

By

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ABSTRACT

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OF CLEAVAGE IN THE BARABOO QUARTZITE

By

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Evidence which is consistent with the formation of spaced and slaty cleavages in the quartzite of the Baraboo quartzite by pressure solution of quartz is presented. The amount of spaced and development of slaty cleavage varies with the amount of phyllosilicate in the rock. Petrographic evidence provides support for the pressure solution mechanism through: insolubles in cleavge planes, truncated grains, pressure shadows, interpenetrative grains. Rotation and fracture are found to be unlikely mechanisms for, respectively, slaty and spaced cleavage formation although recrystallization of the phyllosilicates may have obliterated evidence. Other processes, such as recrystallization and intracrystalline slip probably aided pressure solution in cleavage formation.

Crenulation cleavage is not well developed, prohibiting conclusions regarding it.

Cleavage refraction seems to have formed through a combination of local variations in stress and strain due to viscosity differences (Dieterich, 1969) and differential volume loss due to differences in solubility (Steuer and Platt, 1980).

DEDICATION

to my family: "We all need people we can turn to knowing that being with them is coming. . . home."

Anonymous

and my friends: "The touch of your love is a gift I will never forget."

Anonymous

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Thank you,

Mary

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INTRODUCTION

Many ideas have been brought forth on mechanisms for the origin of the different types of cleavage, but the subject is still contentious even after more than a century of work. There are three general types of cleavage noted in the literature: slaty, fracture and crenulation cleavage. Mechanisms used to account for the development of slaty cleavage include grain rotation, recrystallization and new grain growth, crystal slip or diffusion, and grain breakage. Fracture cleavage has generally been attributed to closely spaced jointing or extension fractures which are then cemented. Crenulation cleavage has been thought to have formed by buckling and/or folding of pre-existing cleavages. During recent years, there has been increasing emphasis placed on the importance of pressure solution as a mechanism in the development of slaty and spaced (fracture) cleavage (Alvarez, Engelder and Lowrie, 1976; Beach, 1979; Bell, 1978; Beutner, 1980; Groshong, 1975a,b; Wood, 1974). Its role in the development of crenulation cleavages has also been noted (Gray, 1979; Gray and Durney, 1979; Fletcher, 1977; Marlow and Etheridge, 1977). Pressure solution is defined as an intergranular diffusive mass transfer process in the fluid phase in response to stress gradients around grains (Beach, 1979); in other words, solution which occurs preferentially at the contact surface of grains where the external pressure exceeds the hydraulic pressure of the interstitial fluid (Gary, McAfee, Jr., and Wolf, 1974).

This study is concerned with the role of pressure solution (diffusive mass transfer processes) as a possible mechanism in the origin of the cleavage present in the Baraboo Quartzite. The Baraboo syncline in south-central Wisconsin has been a classic area for the study of rock cleavage since the early 1900's (Riley, 1947). Weidman (1904) provided a comprehensive study of the structural geometry of the area and his work was extended, especially in the area of rock cleavage by Van Hise and C. K. Leith (1911) and by Mead (1940) in the early part of this century (for a full discussion see Dalziel and Dott, 1970). Earlier work in the area by Van Hise (1893) and C. K. Leith played a large role in their classic works on cleavage, Van Hise's (1896) article "Deformation of Rocks" and C. K. Leith's (1905) treatise "Rock Cleavage". More recent published studies on the area include a structural study by Riley (1947), a comprehensive review of the geology of the area, by Dalziel and Dott (1970). Dott and Dalziel's more recent (1972) work on the age and correlation of the Quartzite and Dalziel and Stirewalt's (1973) paper on folding and cleavage development.

Cleavage in the Quartzite includes a penetrative slaty cleavage present in the associated phyllite which extends locally into the quartzite (the slaty cleavage S_{1E} of Dalziel and Dott, 1970), a predominant spaced cleavage (the quartzite cleavage S_1' of Dalziel and Dott, 1970) and in some phyllosilicate rich areas, a slight crenulation of the pre-existing cleavage. Other minor cleavages are also present which are not dealt with in this study. See Figure 3 for Dalziel and Dott's (1970) classification of cleavages.

The association of phyllosilicates with what appears to be, from petrographic evidence, enhancement of pressure solution was first noted by Heald (1956) in connection with his work on stylolites. The mechanism(s) remain(s) unclear, but the presence of clays or phyllosilicates appears to increase the amount of pressure solution which takes place (Bosworth, 1981; Gresens, 1966; Kerrich, Beckinsale and Durham, 1975; Durney, 1972b; Heald, 1956; Weyl, 1959; Thomson, 1959). Therefore, the major portion of this study is twofold: 1) to compare differences in lithology with the amount and orientation of cleavage which is present in these domains, and 2) to look for evidence of pressure solution (i.e., truncated grains along spaced cleavage margins, quartz overgrowths forming slaty cleavage orientations), while not ignoring evidence which might go against this case, (i.e., fractured grains through which spaced cleavage passes).

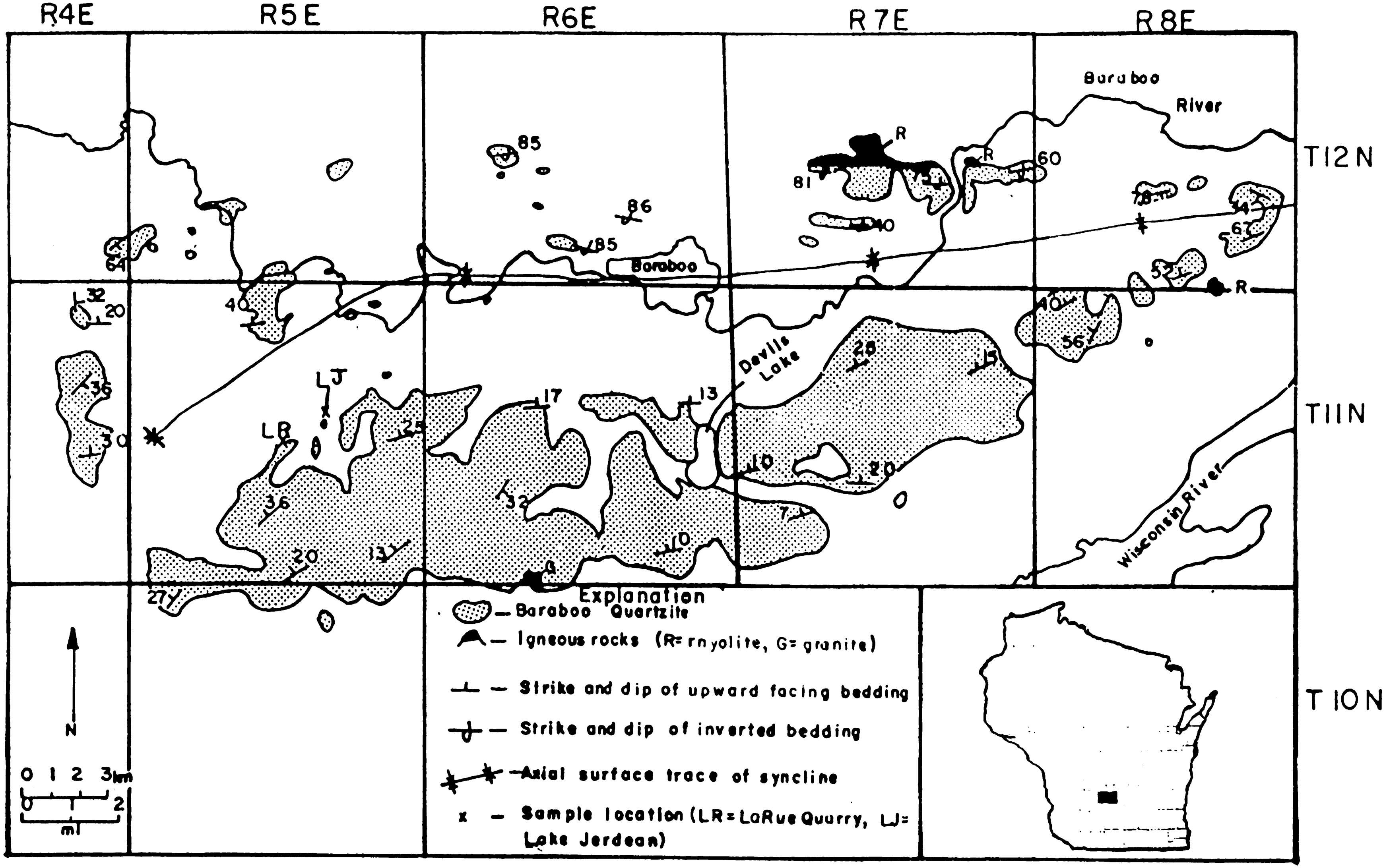
Methods used in this study were petrographic, consisting of point counts, cleavage counts and measurements, length/width and orientation measurements, and descriptive analyses of thin sections of the quartzite. Graphs were plotted comparing the amount of spaced cleavage with the percentage of phyllosilicates, and comparing the orientation and length/width ratios to each other and to the amount of phyllosilicates. Histograms were constructed to provide clearer comparisons of the orientation of the grains with the cleavage and bedding orientations. Statistical methods used included correlation analyses to assess the relationship between phyllosilicate content and the amount and intensity of the cleavages, and analysis of variance to test for reproducibility of data.

GEOLOGY OF THE AREA

In the area around Baraboo, Wisconsin, massive pink, maroon or purple colored exposures of quartzite form an elongate ring of hills known as the Baraboo Ranges (Dalziel and Dott, 1970). These Precambrian basement rocks are locally exposed, in Sauk and Columbia Counties, along the axis of the Wisconsin arch as inliers in the flat lying lower Paleozoic sediments of the midcontinent region (Dalziel and Stirewalt, 1975). See Figure 1 for exact location of study area.

The Baraboo Quartzite is the lowest formation of the Baraboo Group (Baraboo Series of Wiedman, 1904) which is bounded by two major unconformities. The Group nonconformably overlies a poorly exposed rhyolitic basement complex dated at 1.54 ± .04 b.y. (Rb-Sr dating; Dott and Dalziel, 1972). No evidence has been found of an intrusive contact and the rhyolite contains shards and other structures typical of welded tuffs, thereby supporting a nonconformable contact. The Baraboo Group is overlain unconformably by the Dake Quartzite, also of Precambrian age, a coarse grained quartzite containing angular pebbles of quartz and quartzite, and a high proportion of chlorite and sericite (Dalziel and Dott, 1970). No exposures of the Dake have been positively identified, although two outcrops of Baraboo Quartzite have been identified at times as Dake, so that the Dake is known only from drilling records. The Precambrian succession in the area is topped by the Rowley Creek Slate, a gray quartz-chlorite sericite slate, which overlies the Dake. See Figure 2 for stratigraphy and approximate thicknesses of beds (Dalziel and Dott, 1970).

Figure 1. Pre-Paleozoic outcrops in the area of Baraboo, Wisconsin. Inset map - regional setting. Sample locations denoted by X. (From Dalziel and Stirewalt, 1975; Dalziel and Dott, 1970.)


R4E
R5E
R6E
R7E
R8E
T12N
T11N
T10N
Baraboo River
Baraboo
Devils Lake
Wisconsin River
LR
LJ
R
G
Explanation
— Baraboo Quartzite
— Igneous rocks (R= rhyolite, G= granite)
— Strike and dip of upward facing bedding
— Strike and dip of inverted bedding
—Axial surface trace of syncline
x — Sample location (LR= LaRue Quarry, LJ= Lake Jerdean)
N
0 1 2 3 km
0 2
mi

Rowley Creek Slate (maximum known thickness 149 feet)

Dake Quartzite (maximum known thickness 214 feet)

(Unconformity)

Freedom Formation (dolomite and ferruginous slate; minimum thickness 1000 feet)

Seeley Slate (maximum known thickness 370 feet)

Baraboo Quartzite (thickness over 4000 feet)

(Unconformity?)

Rhyolitic "basement" (thickness unknown)

Figure 2. Precambrian stratigraphy of the Baraboo District. (From Dalziel and Dott, 1970).

The Baraboo Group consists of, in ascending order, the Baraboo Quartzite, the Seeley Slate and the Freedom Formation. The Seeley Slate is a grey or green slate, generally uniform in appearance and containing fine stratification as well as a well developed cleavage. The Freedom Formation consists of dolomite, ferrugenous slate and iron ore along with other minor rock types. The Seeley Slate and the Freedom Formation are also known only through subsurface drilling records (Dalziel and Dott, 1970).

The Baraboo Quartzite is comprised of more than 80% quartz which occurs as medium to coarse sand sized grains and sporadic rounded granules and fine pebbles, along with some micas (pyrophyllite, sericite and chlorite), and a small percentage of heavy minerals such as zircon, rutile and iron oxides. Numerous layers or lenses of more argillaceous material appear in the quartzite. They vary in size from only a few inches thick to several feet, and consist chiefly of pyrophyllite and quartz with minor quantities of muscovite and hematite. Dalziel and Dott (1970) refer to them as "phyllite, quartz phyllite, or phyllitic-quartzite...because they exhibit more pronounced recrystallization and less regular foliation than a slate, but are finer grained than a schist". The assemblage (quartz-pyrophyllite-muscovite-hematite) of these layers indicates that the Baraboo Quartzite reached the lower greenschist facies of regional metamorphism, and from experimental pyrophyllite stability data a temperature as high as 410-430^{o}C at 2-4 kilobars of water pressure (Dalziel and Dott, 1970; Kerrich, 1968).

The entire Precambrian succession of the area has been folded into a complex doubly plunging asymmetric syncline, about 25 miles long and with a maximum width of 10 miles. The hills of the Baraboo Range rise to a maximum elevation of 700-800 feet above the Wisconsin River valley and form a structural and topographic basin infilled with Paleozoic and Pleistocene sediments. The axial surface of the syncline strikes approximately east-northeast to west-southwest and dips steeply to the north-northwest. The north limb is nearly vertical and the south limb dips generally about 15^{o} to the north (Dalziel and Dott, 1970). The deformation of the quartzite and surrounding pelitic rocks was produced during several phases as is shown by the sequence of cleavages which has developed. This polyphase deformational history may well have occurred as part of a single progressive deformation event in response to one regional stress system with the various structures merely recording stages within this continuous strain history (Dalziel and Dott, 1970).

Dalziel and Dott (1970), Figure 3, describe the several types of cleavage which occur in the Baraboo Quartzite and associated phyllite: 1) a penetrative cleavage in the phyllite (extending also into the quartzite) which they term slaty cleavage (S_{1E}), 2) a fracture cleavage in the phyllite which they call phyllitic cleavage (S_1) and which is associated with 3) a closely spaced parting in the quartzite which they label quartzite cleavage (S_1'). Other cleavages noted in the quartzite include a locally more highly penetrative foliation, a crenulation of the phyllitic and quartzite cleavage, and some secondary cleavage resulting from closely spaced jointing (Dalziel and Dott, 1970). As previously noted, this study will concern itself with the following: the quartzite cleavage (S_1') which consists of a distinct fracture or spaced cleavage marked by discrete surfaces formed by concentrations of aligned phyllosilicates; the locally highly penetrative cleavage (S_{1E}); and the occasionally visible, fine crenulation cleavage also seen in the quartzite.

AGE	STRUCTURES			
Primary	S_0 Bedding (master bedding) (also current bedding, and perhaps some preferred grain and pebble orientation		Ripple marks ?Some preferred grain and pebble orientation	
Post-sedimentation, pretectonic	Color banding			
Early Main Phase	S_{1E} *Slaty cleavage (phyllite)	S_1' Quartzite cleavage (quartzite)	S_0/S_{1E} intersection (indistinct)	S_0/S_1' intersection
Main Phase	S_1 Phyllitic cleavage (phyllite), deforming forming S_{1E} Axial surfaces of tight asymmetric minor folds (folding S_0)		S_0/S_1 intersection Axes of tight asymmetric minor folds Longrain (mineral alignment) in S_1 (phyllite) Slickensides on S_0 (quartzite) Boudin axes	transitional
Late Main Phase	S_{1L} Faint strain-slip or crenulation cleavage (deforming S_1 in phyllite)		Axes of fine crenulations (deforming S_1 in phyllite)	
Secondary Phase	S_2 Strain-slip or crenulation cleavage (conjugate; deforming S_0 and, mainly, S_1 in phyllite) Axial surfaces of minor chevron folds (deforming S_0, S_1, and S_1'		Axes of crenulations (deforming S_0 and, mainly, S_1 in phyllite) Axes of minor chevron folds deforming S_0, S_1 and S_1'	
Late Phase	Axial surfaces of open minor folds (deforming S_0, S_1, and S_2)		Axes of open minor folds	
Various	Joints, some quartz-filled quartz "veins", including en echelon tension gashes in bands, some deformed gashes (probably post-second phase) Breccia zones and faults (?)		Slickensides on bedding current bedding, fracture cleavage and joint surfaces Boudin Axes	
Doubtful	Local highly penetrative cleavage in quartzite Second cleavage (closely spaced jointing) in quartzite Other cleavages (closely spaced jointing) in quartzite		Intersection of highly penetrative cleavage in quartzite with bedding Intersection of second cleavage in quartzite with bedding Intersection of other cleavages in quartzite with bedding	

Figure 3. Chart showing Dalziel and Dott's (1970) classification of the cleavage and other mesoscopic structures in the Baraboo Quartzite. (From Dalziel and Dott, 1970).

PREVIOUS WORK

There is, at present, much controversy concerning mechanisms for the origin and development of cleavage. Classification of cleavages also presents a problem and this may be due, in part, to their tendency to grade into one another, often within a few feet in the same rock. Cleavages can be classified into three general types based on certain simple criteria. The first of these criteria is whether the cleavage is penetrative or nonpenetrative. Penetrative cleavage is pervasive, uniformly distributed through the rock and without discontinuities. Slaty cleavage is a penetrative cleavage. Nonpenetrative cleavage is the opposite of penetrative cleavage, and generally divides the rock into domains or microlithons which differ in lithology. Nonpenetrative cleavage is further subdivided based on the amount and type of orientation present (another criterion) into spaced and crenulation cleavages. Spaced cleavage does not imply or require a foliation or alignment of grains; while crenulation cleavage does in that it is the crenulation of a previous aligned mineral fabric which gives rise to the cleavage.

Slaty Cleavage

The role of pressure solution in the development of slaty cleavage was noted early in cleavage studies. Slaty cleavage is defined as a penetrative foliation or alignment of inequant mineral grains (Gary, McAfee and Wolf, 1974). Sorby in 1879 recognized the importance of pressure solution transfer in the deformation of individual grains (Siddans, 1972). Rotation and deformation of constituent grains and recrystallization were also recognized by various authors

(Sharpe, 1849; Sorby, 1853, 1856, 1857; Lyell, 1835; Sedgwick, 1835; Darwin, 1846; Harker, 1886) as important mechanisms for slaty cleavage formation at about this time (Siddans, 1972). Sorby (1853, 1856) and other early writers believed rotation was the primary process in cleavage development (Siddans, 1972). Van Hise (1896) felt that recrystallization and new grain growth were the most important mechanisms but that flattening and the rotation of old and new mineral grains were important also. He discounted the effects of the rotation into parallelism of random original particles. Leith (1905) emphasized the interaction of various processes to form slaty cleavage. The processes he felt were of primary importance were the flattening of old mineral particles and the growth of new grains by the processes of crystallization and recrystallization. He also listed gliding along definite planes in minerals, rotation of the particles toward a parallel arrangement and granulation without rotation, but did not feel these processes played as large a role in cleavage development. Discussions on the origins of slaty cleavage, then, have focused from the beginning on two main mechanisms: rotation and recrystallization. Rotation of pre-existing inequant grains as the primary orienting mechanisms has received support in the literature from Alterman (1973), Braddock (1970), Carson (1968), Davies and Cave (1976), Clark (1970), Geiser (1974) Lebedeva (1978), Maxwell (1962), Moore and Geigle (1972), Oertel (1970), Powell (1972a,b, 1973), Roy (1978), Tullis and Wood (1972), and Williams (1972). Maxwell (1962) added an interesting hypothesis to the concept of grain rotation as an orienting mechanism. Working in the Delaware water gap area of New Jersey and Pennsylvania, he theorized that slaty cleavage was a prelithification phenomenon resulting from high pore water pressure in unconsolidated sediments. Support for Maxwell's hypothesis is found in several other studies, noteably Moench (1966), Carson (1968), Braddock (1970), Clark (1970), Powell (1972a,b, 1973), Roy (1978). His theory has, however, been widely questioned (Beutner, 1978, 1980; Beutner, Janan and Simon, 1977; Epstein, 1974; Geiser, 1975; Gregg, 1979) and is not currently thought to be able to account for the genesis of most slaty cleavage (Wood, 1974). Recrystallization as the predominant orienting mechanism in slaty cleavage formation has received support from DeVore (1966, 1969a,b), Burger (1974), Durney (1972, 1976), Holeywell and Tullis (1975), Knipe and White (1977), Manktelow (1979), Wintsch (1978), Bell (1978), Epstein (1974), Kamb (1959, 1961). Studies (Oertel, 1970; Ramsay, 1976) also suggest that a combination of rotation and recrystallization processes could be responsible for reorientation of the minerals.

Pressure solution which, it can be argued, is a specific recrystallization mechanism (Wood, 1974) has been re-emphasized in recent years. Epstein (1974) accounted for the development of cleavage in the Delaware water gap area of Pennsylvania and New Jersey (the same area as Maxwell's, 1962, study) through the dissolution of quartz, and thus the flattening of grains, and the deposition of new overgrowths on the original quartz grains leading to an oriented mineral fabric. Other recent studies which have concluded that pressure solution was primarily responsible for slaty cleavage development are Burger (1974), Wintsch (1978), Beutner (1978, 1980), Bell (1978), Groshong (1976), Beutner, Janan and Simon (1977), Geiser (1975), Durney (1972, 1976), Holeywell and Tullis (1975) and Wood (1974). A 1976 Penrose Conference on cleavage took note of the re-emergence of pressure soution as an orienting mechanism for slaty cleavage development (Platt, 1976) and four years later at another Penrose Conference on the role of pressure solution, this was again emphasized (Engelder, Geiser and Alvarez, 1981). Experimental studies on salt mica specimens by Means and Williams (1972, 1974) and Means (1977) have also noted solution effects associated with the reorientation of micas. For excellent summaries of research on slaty cleavage, see Siddans (1970) and Wood (1974).

Spaced or "Fracture" Cleavage

Spaced or "fracture" cleavage, then, is defined as a nonpenetrative cleavage consisting of closely spaced planes or surfaces that divide the rock into a series of tabular bodies or microlithons (Hobbs, Means and Williams, 1976). No orientation of grains within the microlithons is implied or required although some orientation may be present. Sedgwick in 1835 theorized that cleavage planes were discrete planes of fracture due to mechanical tension and was careful to distinguish between this "fracture" cleavage and jointing (Siddans, 1972). Later studies by Van Hise (1896) and Leith (1905) attributed "fracture" cleavage to closely spaced jointing. Becker (1896) believed cleavage planes (fracture cleavage) were a product of shear strain with accompanying brittle failure. Most studies of "fracture" or spaced cleavage since that time have concluded that the mechanism responsible for this type of cleavage was some form of jointing or fracture; in other words, brittle failure. For instance, Mead (1940) divided spaced cleavages into two types: fracture cleavages which were due to rock fracture and shear cleavages which were closely spaced surfaces of shear displacement on which platy minerals may have developed. Later Knill (1960) defined fracture cleavage as a series of closely spaced microfaults or fractures. Still later, Price and Hancock (1972) in a study of fracture cleavage and like structures, attributed fracture cleavage to hydraulic fracturing during phases of high pore fluid pressure and extension of the rock. The recent structural study of the Baraboo quartzite by Dalziel and Stirewalt (1975) suggests that Price and Hancock's (1972) mechanism may be used to account for the quartzite cleavage (S_1') in the Baraboo.

More recent work on "fracture" cleavage (Alvarez, Engelder and Lowrie, 1976; Alvarez, Engelder and Geiser, 1978; Beach, 1974, 1979; Borradaile, 1977; Durney, 1972a, 1976; Engelder, Geiser and Alvarez, 1981; Geiser, 1974;

Groshong, 1975a,b, 1976; McClay, 1977; Means, 1975; Platt, 1976; Roben, 1979; Stringer and Treagus, 1980; White and Johnston, 1981; Woodland, 1982) suggests that the term "fracture" cleavage is inappropriate and that the more appropriate term is spaced cleavage which carries no genetic implications. This recent work has shown some spaced cleavages to be a result of dissolution (pressure solution) and the concentration of insoluble residues into cleavage planes in a process comparable to the formation of stylolites. The first of these studies dealt with carbonates where pressure solution processes are generally more easily identified. Plessman (1964), using partially dissolved fossils as evidence, described "cleavage planes" formed by the concentration of micas in carbonate rich rocks as analogs of stylolites in limestones (Williams, 1972). Choukroune (1969) noted two types of stylolites in deformed limestones in the Alps -bedding plane parallel or sedimentary stylolites and transverse or tectonic stylolites (spaced cleavage?) which are perpendicular to tension gashes and perpendicular to the maximum compression direction. Working with limestones Alvarez, Engelder and Lowrie (1976), Alvarez, Engelder and Geiser (1978), and Groshong (1975a,b) have shown that pressure solution acted to form spaced cleavages. They used evidence such as fossils that truncate against cleavage traces, bedding offsets, and the concentration of quartz and other minerals less soluble than limestone along the cleavage surfaces. Spaced cleavages in pelitic rocks have also quite recently been attributed to pressure solution. Durney (1972) noted that "fracture" cleavage can be produced by pressure solution "as the source region is enriched in insoluble matter and depleted in soluble matter". His work was quickly supplemented by other authors (Geiser, 1974; Groshong, 1976; McClay, 1977; Robin, 1979; Stringer and Treagus, 1980; White and Johnston, 1981; and Woodland, 1982), all of whom noted the relationship of pressure solution or diffusion processes to the formation of spaced cleavages.

Beach (1979) went so far as to say, "The most widespread record that terrigenous sedimentary rocks have deformed by a pressure solution mechanism is seen in the development of spaced cleavages..." Experimental work by Means (1975) on micaceous sandstones also showed evidence of possible solution effects leading to the presence of mica films along cleavage planes. A 1980 Penrose Conference report, by Engelder, Geiser and Alvarez (1981), again noted the importance of pressure solution in the development of "fracture" cleavage.

Crenulation Cleavage

Crenulation cleavage is a nonpenetrative cleavage which is characterized by a finite spacing of cleavage planes between which there occurs thin, tabular bodies of rock displaying a crenulated cross lamination (Gary, McAfee and Wolf, 1974). It has also been known as slip or strain slip cleavage, but as these terms carry genetic connotations they will not be used herein. Early studies such as that of Leith (1905) attributed crenulation cleavage to fractures and microfaults along the limbs of developing microfolds. Recent studies by Cosgrove (1976), Gray and Durney (1979), Gray (1976, 1977), Fletcher (1977) and others have related crenulation cleavage not only to buckling (microfolding), but also to pressure solution which aids in mineral migration from the limbs to the crests of these microfolds. Evidence cited is the presence of the more soluble minerals along the crests of these microfolds and the insoluble or less soluble material along the limbs which has led to a mineralogical differentiation and a definite fabric.

A short lucid explanation of past and current thoughts on cleavage development is found in Hobbs, Means and Williams (1976, Chapter 5).

Pressure Solution

In 1862, Thompson formulated principles dealing with the solution of crystal aggregates under differential stress (Kerrich, 1978). These principles form the basis for discussion of the phenomenon of pressure solution. Pressure solution is defined as solution which occurs preferentially at the contact surfaces of grains where the external pressure exceeds the hydraulic pressure of the interstitial fluid (Gary, McAfee and Wolf, 1974). Thompson's principles were first applied to the study of rock deformation by Sorby (1879). Sorby ascribed many deformation features such as slickensides, slaty cleavage, schistosity, veining, stylolites and impressed pebbles with associated insoluble residues to the effects of pressure solution (Kerrich, 1978). Other authors such as Van Hise (1896) and Harker (1885) also noted the role of pressure solution in deformation at about this time. Little progress was made in research on the relationship of pressure solution to deformation structures or on any detailed theory of pressure solution in the early part of this century (Kerrich, 1978). Russell (1935) in work on crystal growth and solution with alum crystals and salt found that crystals would dissolve under a local stress and be redeposited away from that stress. This offered further support to the theory of pressure solution. He questioned, however, whether this process could play a very large role in deformation. Some interest in pressure solution effects was reinstituted with work by Thomson (1959), Heald (1955, 1956) and others on stylolites and indented pebbles toward the midpoint of the century.

Interest in pressure solution as a deformation mechanism has been revived in recent years by new work on the origins of cleavage and schistosity (see above). Problems with the theoretical base of pressure solution remain (Robin, 1978) and this has led at least one worker to question the very existence of the process (Deelman, 1975, 1977). Experimental work by Sprunt and

Nur (1976, 1977) with the compaction of quartz sands has supported the existence of pressure solution as a deformation mechanism, although these studies have been questioned (Deelman, 1977). In general, however, "Experimental as well as petrographic evidence for pressure solution have generally been accepted as convincing" (Robin, 1978). Theoretical modeling of the pressure solution process still presents many problems despite the recent surge of interest in the subject (Paterson, 1973; Stephansson, 1974; Weyl, 1959; DeBoer, 1975; Durney, 1976; Bathurst, 1958; Kamb, 1961; Engelder, 1981; Elliott, 1970, 1973). This is perhaps due to the complexity of the subject which must include mineral reactions, solution and redeposition, diffusion and flow interactions, all within a nonhydrostatic stress system, open to external factors (Beach, 1979). Progress has been made. Kamb (1961) related nonhydrostatic thermodynamic theory to the dissolution of solids, while Paterson (1973) has related this same theory to pressure solution. Elliott (1970, 1973) has found that pressure solution obeys a diffusion flow law with a linear stress to strain ratio. It does seem, however, as though it will be some time before a unified body of theory is developed.

The use of the term pressure solution has been questioned by some authors (Bosworth, 1981; Engelder, Geiser and Alvarez, 1981). They believe that because it is difficult to separate the effects of stress and strain in metamorphic rocks and because the solution effects seen may be due to a combination of both stress and strain; it might be better to use a term such as diffusive mass transfer for the process (Engelder, Geiser and Alvarez, 1981). This is a valid argument, but because the use of the term pressure solution for these solution-reprecipitation processes is widely accepted in the literature, this term will continue to be used in this study.

Some studies have recognized that the presence of clay minerals (phyllosilicates) in the rock seemed to increase the amount of pressure solution that occurred (Bosworth, 1981; Gresens, 1966; Kerrich, Beckinsale and Durham, 1975; Durney, 1972b; Heald, 1956; Weyl, 1959; Thomson, 1959). This was mainly on the basis of petrographic observations, and not quantified. There have been some attempts to explain these observations. Weyl (1959) thought that clay films might provide pathways for more rapid diffusion, with diffusion being the slowest or rate controlling step and thus increase the rate of pressure solution. Heald (1956), one of the first to note the role of clays, felt that in some way the clays acted as catalysts to the pressure solution process. Thomson (1959) stated that the clays promote pressure solution by providing a microenvironment of high pH at grain contacts. The hydrolytic weakening of quartz, the promotion of recrystallization of quartz by the presence of water (Jones, 1975; Hobbs, 1981) and the adsorbed water layers on clays (Weyl, 1959) are all fairly well known phenomena which also might provide explanations for this effect. There is still, however, no firm theoretical basis for this widespread observation of the enhancement of the pressure solution process by clays.

For a more complete review of past and present research on pressure solution, see Kerrich (1978).

SAMPLE LOCATION AND THIN SECTION PREPARATION

Samples were taken from two locations: the LaRue Quarry (SW¼, NW¼, sec. 22, T11N R5E, about one mile south of LaRue from Sauk County Highway PF, North Freedom Quadrangle), and an outcrop next to Lake Jerdean (SW¼, SE¼, SE¼, sec. 15, T11N R5E, about three quarters of a mile southeast of LaRue from Sauk County Highway PF, North Freedom Quadrangle). Both locations are in the southwest portion of the syncline, fairly close to one another (see Figure 1). These areas were chosen for their excellent exposures of cleavage along with the availibility of a wide range in lithologic composition from phyllite-rich to phyllite-poor within a small area. Two locations were chosen in order to test that results obtained were reproducible from more than a single location. Samples were bedding plane oriented, where possible, as they were collected. Only one sample was obtained from the Lake Jerdean area. It is, however, a large oriented specimen showing a wide range in lithology.

Sections were cut perpendicular to bedding cleavage intersections so as to provide a standard orientation. This plane was chosen since it would be, presumably, closest to the XZ plane of the finite strain ellipsoid and would therefore show the greatest strain ratio (the maximum difference in strain). Several sections were cut that were not oriented in this manner so that a comparison might be made. Bedding orientations were carefully preserved and indicated on the sections made from oriented samples. Sections were made so as to obtain a wide variation in lithology. Several unusually large (3 x 5 or larger) sections were made in order to facilitate textural comparisons.

ANALYTICAL METHODS

Point counts of 1000 points per section were done on all slides to obtain a percent phyllosilicate versus whole rock ratio. This should give a maximum confidence interval of ± 3% according to Pettijohn, Potter and Siever (1973, Figure A-3). Counts were made of the number of spaced cleavage planes on each slide. Areal measurements were also made of the amount of spaced cleavage per centimeter (measured normal to the cleavage). Using only the oriented sections, length to width ratios of individual quartz grains were measured coupled with measurements of orientations of the long axis of each grain. Before doing this, it was thought desirable to attempt to remove phyllosilicates from the slides in order to obtain more accurate measurements. This was done by etching with hydrofluoric acid and when this did not work, by a sodium pyrosulfate fusion technique developed to remove clay minerals from soil samples (Blatt, Jones and Charles, 1982). Four slides were prepared using this technique and comparisons were made to slides which were not prepared in this manner. As there was no discernible difference in the measurements from the phyllosilicate-removed slides and the others, this part of the procedure was discarded. Orientations of bedding and spaced cleavage planes from these sections were at the same time measured. Due to the anastomosing nature of the cleavage, the orientations of the cleavage planes varied on each slide, so up to ten measurements were made and the average taken as the cleavage orientation. The range of the differences in orientation was from 15° to 37° with the average difference being about 25°. One-hundred-fifty quartz grains were measured per oriented section, although it was found that the results were largely reproducible after 50 such measurements. Point counts were also made to compare the amount of polygonized quartz with the amount of detrital quartz. All of the sections were then examined and any salient features noted.

Graphs were then prepared. The percentage of phyllosilicates for each slide was plotted against the number of spaced cleavage planes per centimeter. Orientation measurements were graphed against length/width (axial) ratios for each of the oriented sections. The orientation of both spaced cleavage planes and bedding were also noted on these graphs. Previous to this, to facilitate comparisons between graphs, transformation of the data was done so that all bedding planes were arbitrarily oriented at 90°. Histograms were also prepared showing both the frequency, in 10° class intervals, of long axis orientations of quartz grains and the orientation of both spaced cleavage and bedding for each slide. The average axial ratio for each section was also plotted against the percentage of phyllosilicates in these slides.

Statistical testing of the data was done using both parametric and where appropriate nonparametric tests. To test for differences not due to chance between the axial ratio measurements from the sections, an analysis of variance (ANOVA) was done. Tests were performed to see if there was a correlation between the amount of phyllosilicate and the axial ratio averages. Due to the possibility that the data on the orientation of the quartz grains is bimodal, a nonparametric test, the Kruskal-Wallis one way analysis of variance was used to test for differences not due to chance between these measurements. Correlation and linear regression of the spaced cleavage versus percent phyllosilicate data were used to delineate their relationship. Correlation and linear regression were performed also on the polygonized quartz versus percent phyllosilicate data.

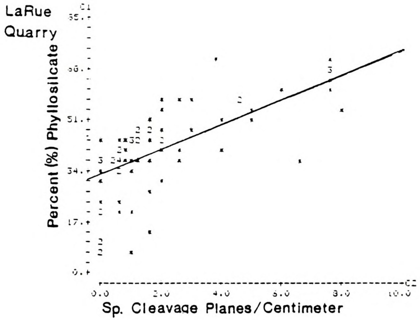
As a further check on the reproducibility of the axial ratio and orientation data, four replicate data sets were prepared and comparisons between these and the original data were made. Students t test was used to check the ratios to see if the original data was essentially reproduced in the duplicate data set for the axial ratio measurements. A nonparametric test, the Mann Whitney sign rank test, performed the same task on the orientation data. Also, a χ^2 test was done on histogram data from these duplicate and original data sets to provide another check.

RESULTS

Data Summary

The results of the point counts and spaced cleavage measurements are found in Figure 4. Length to width ratios of sections from the LaRue samples are graphed against the long axis orientation measurements for those sections in Figure 5 and Figure 6 contain the histograms which correspond to these graphs. Figures 7 and 8 contain the comparable graphs and histograms for the Lake Jerdean sections. Note that bedding and cleavage plane orientations for each section are marked on the graphs and histograms along with the percentage of phyllosilicates in that section. Figure 9 shows the average axial ratio for each section plotted against the percentage of phyllosilicates in that section. Figure 10 shows the percent phyllosilicate plotted against the percentage of polygonized quartz in each section.

Statistical analyses, at the 95% confidence level, showed definite differences not due to chance between the sections. Also at the 95% confidence level, the four duplicate data sets replicated the original data. The correlation coefficient (r) for the graph of spaced cleavage versus percent (%) phyllosilicate was .894 ($p < .01$, $df = 7$) for the sections from the Lake Jerdean sample and .709 ($p < .01$, $df = 74$) for the sections from the LaRue samples. The correlation coefficient for the graphs of average axial ratios versus percent (%) phyllosilicate was .773 ($p < .05$, $df = 7$) for the Lake Jerdean sections and .973 ($p < .01$, $df = 5$) for the LaRue sections. The correlation coefficients for the graphs of percent phyllosilicate versus polygonized quartz were .845 ($p < .01$, $df = 5$) for the LaRue graph and .659 ($p < .05$, $df = 7$) for the Lake Jerdean graph.



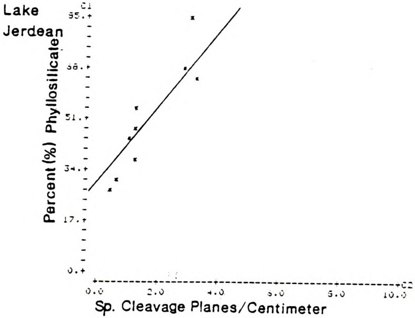


Figure 4. Graphs: the number of spaced cleavage planes per centimeter compared to the amount of phyllosilicate for each section. Regression lines are indicated. Top: LaRue Quarry. Below: Lake Jerdean.

Figure 5. LaRue Quarry Graphs: the orientation of the long axis plotted against the axial ratio (long axis/short axis) of each quartz grain. Bedding and average spaced cleavage orientations, as well as the range of spaced cleavage are indicated on each graph. The graphs are in order of increasing phyllosilicate content. *Point concentration from histograms, Figure 6.

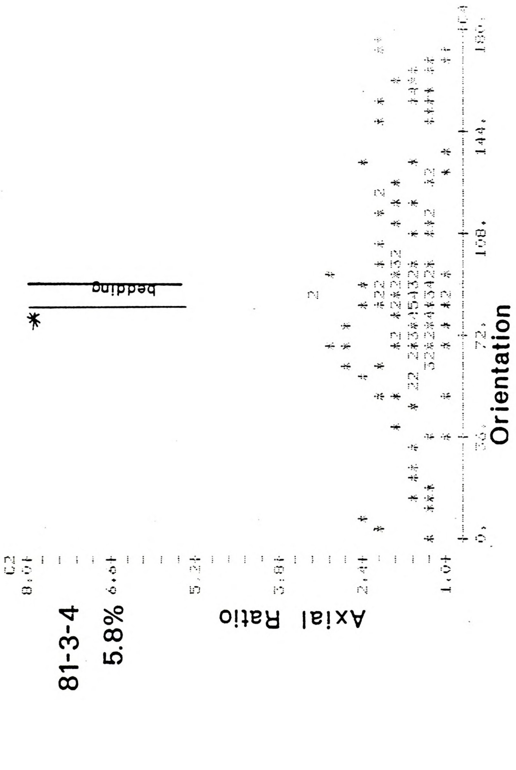


Figure 5a.

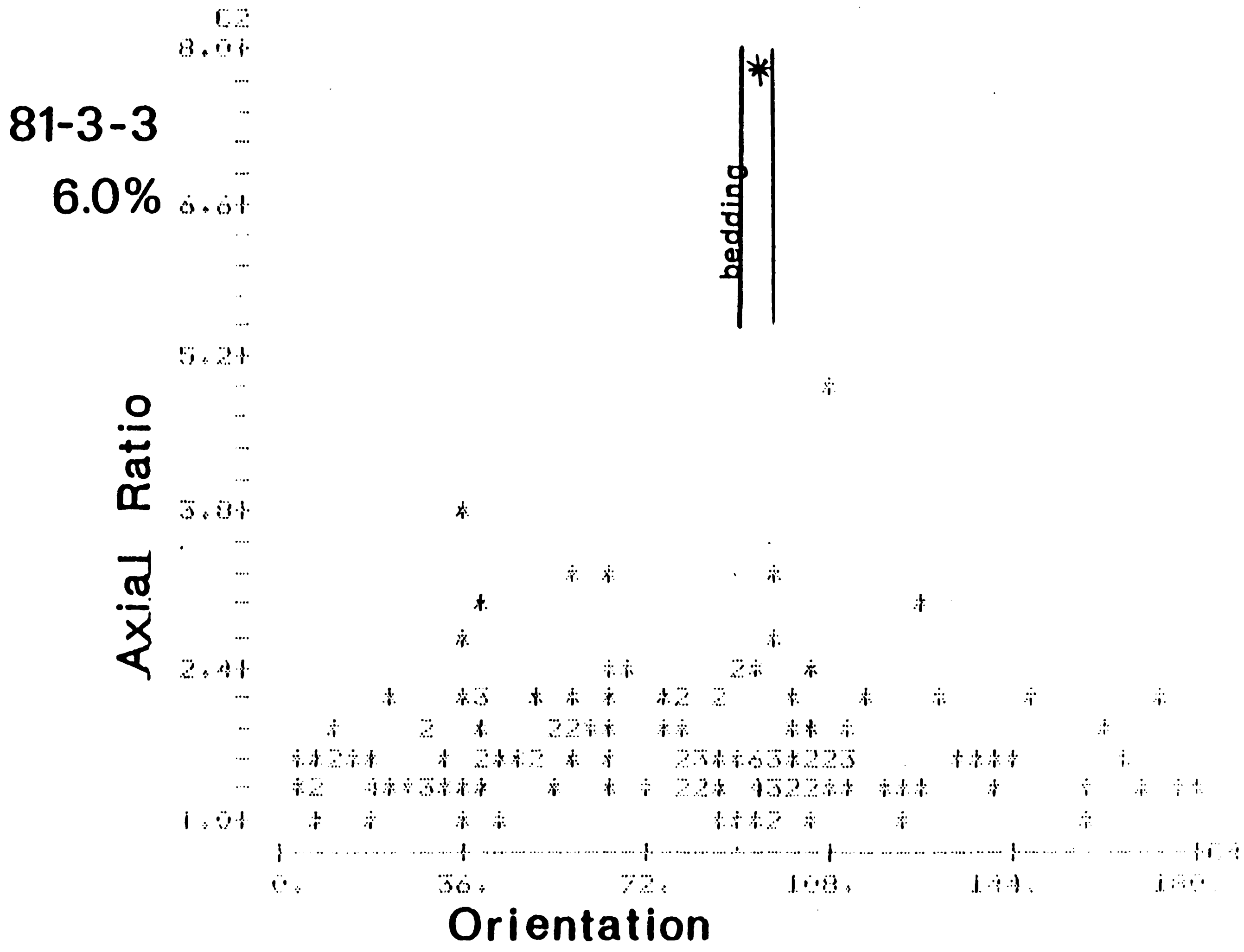

81-3-3
6.0%
Axial Ratio
8.0
6.6
5.2
3.8
2.4
1.0
bedding
0.
36.
72.
108.
144.
180.
Orientation

Figure 5b.

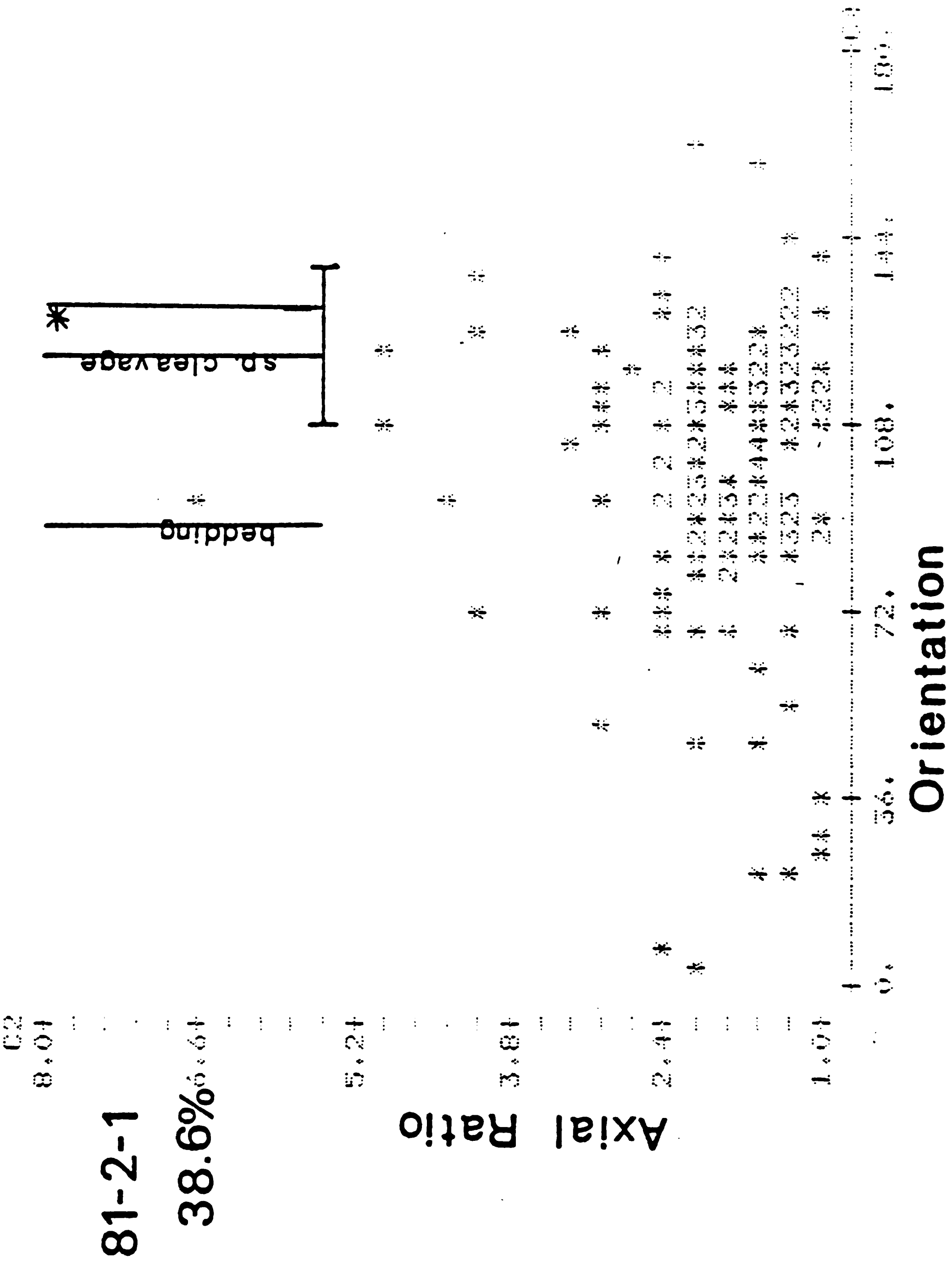


Figure 5c.

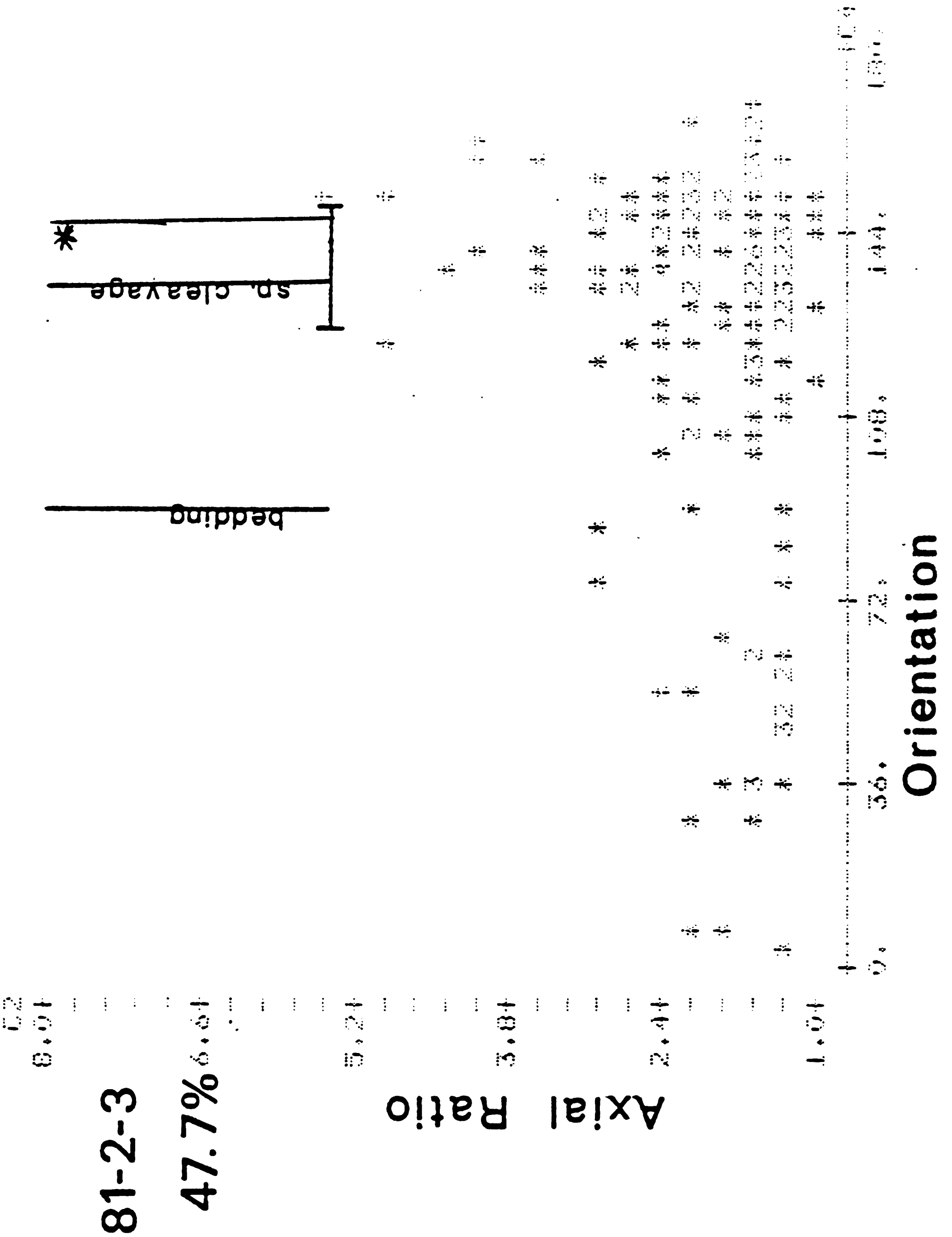

81-2-3
47.7%
Axial Ratio
1.04
2.44
3.84
5.24
8.04
Orientation
0.
36.
72.
108.
144.
180.
bedding
sp. cleavage

Figure 5d.

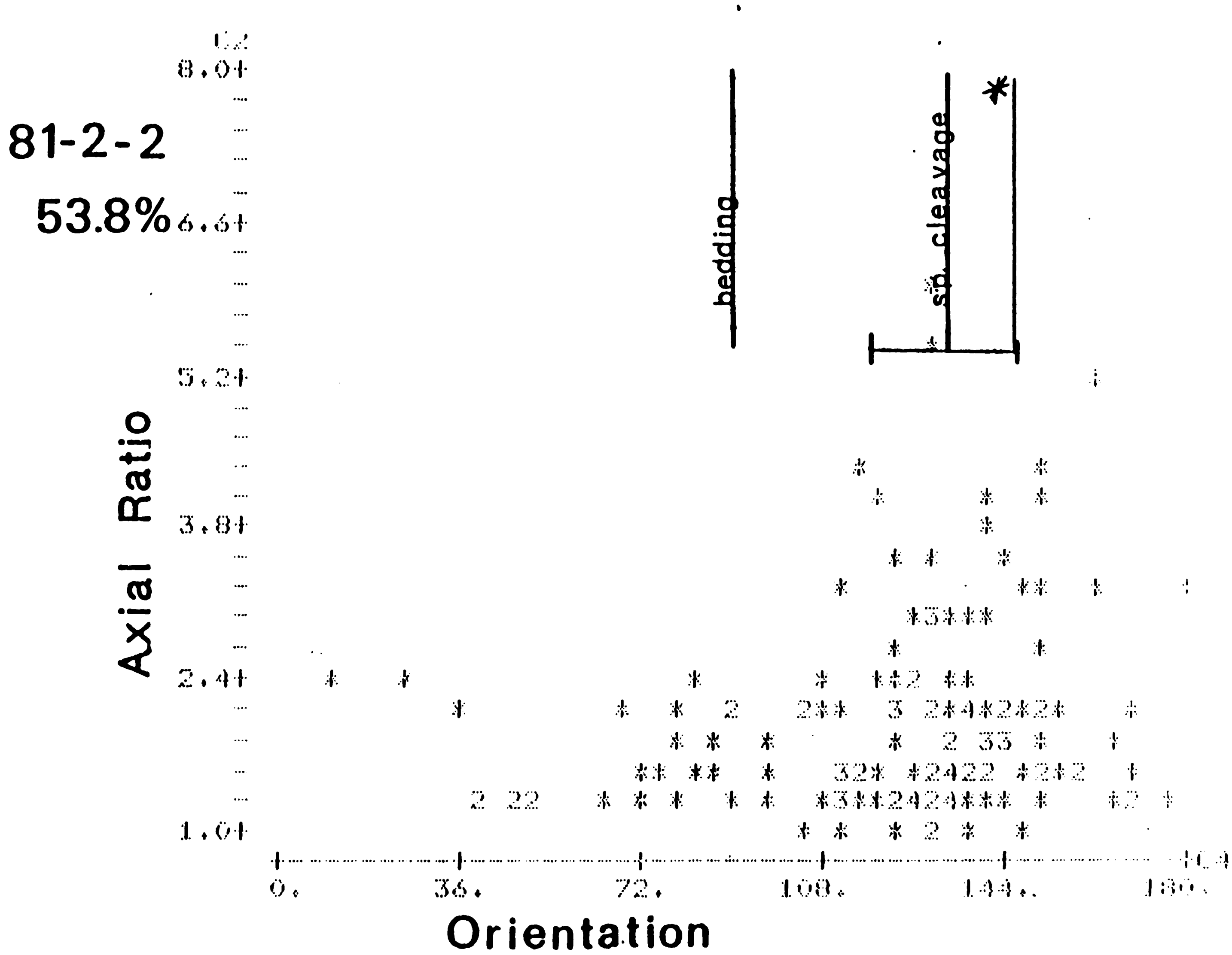

81-2-2
53.8%
Axial Ratio
Orientation
bedding
sp. cleavage
8.0
6.6
5.2
3.8
2.4
1.0
0.
36.
72.
108.
144.
180.

Figure 5e.

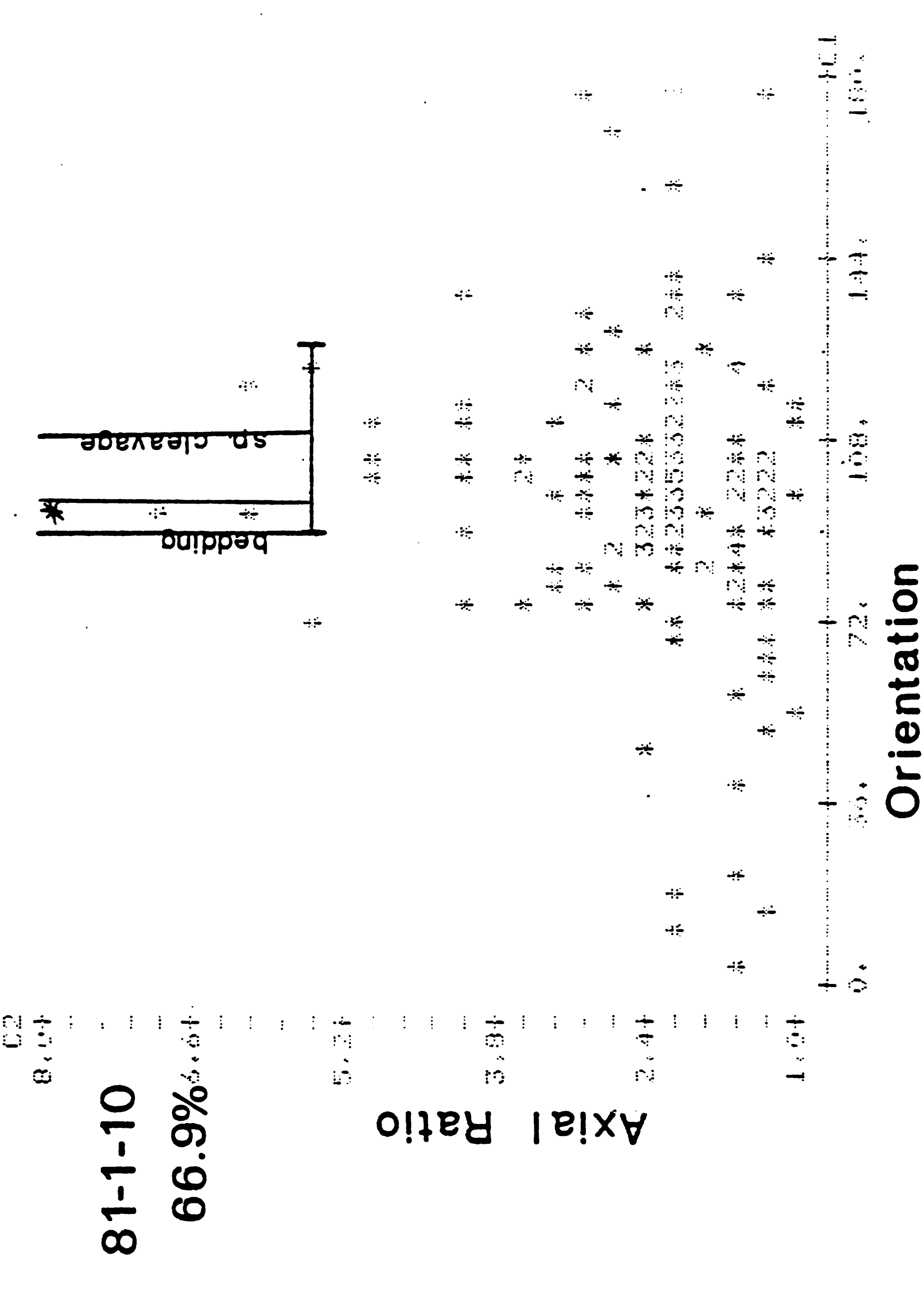


Figure 5f.

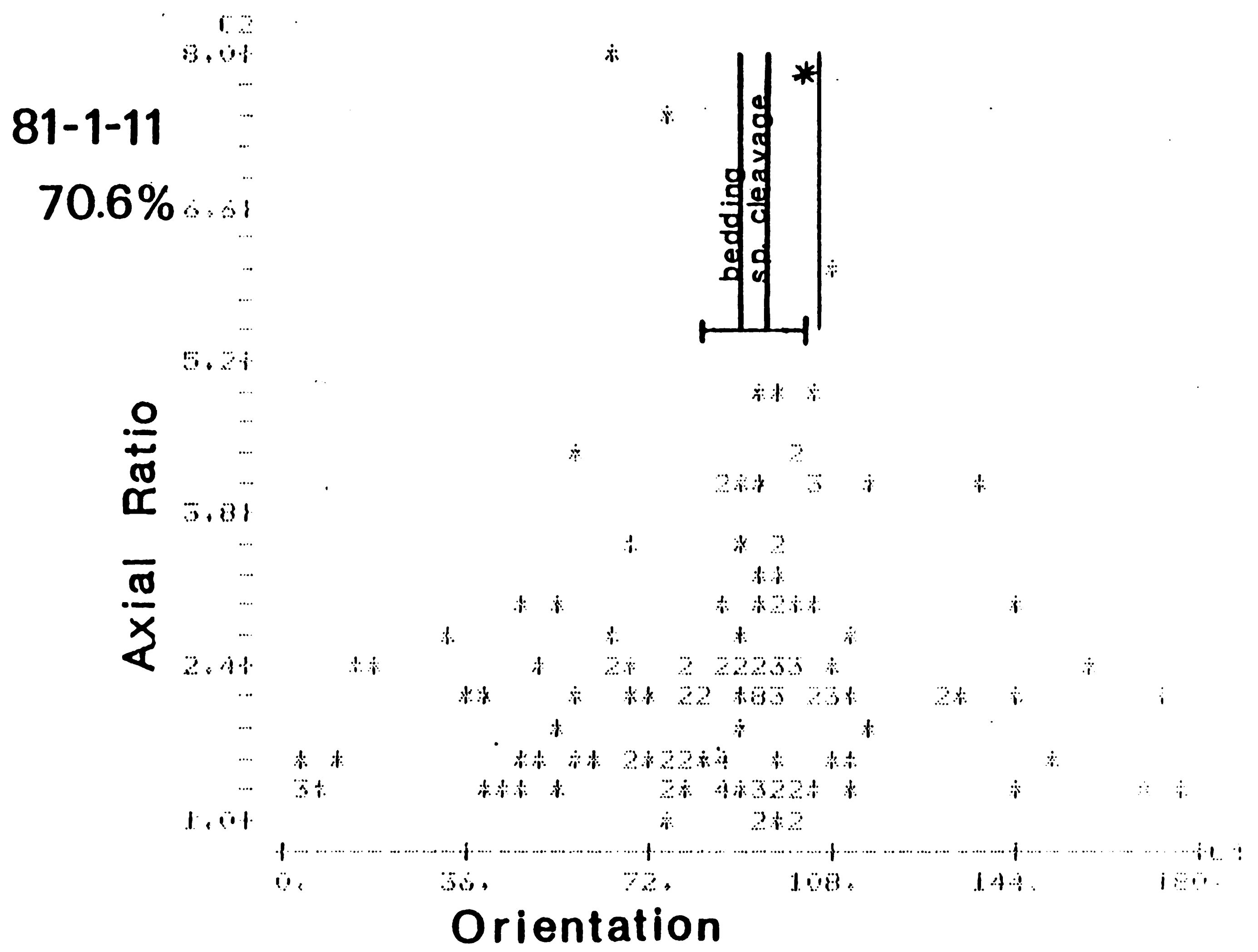

81-1-11
70.6%
Axial Ratio
Orientation
bedding
sp. cleavage
8.04
6.61
5.24
3.81
2.44
1.04
0.
36.
72.
108.
144.
180.

Figure 5g.

Figure 6. LaRue Quarry histograms indicating the frequency in 10° class intervals, of long axis orientations of quartz grains. Bedding, and spaced cleavage range and average orientation are indicated on each histogram. The histograms are in order of increasing phyllosilicate content and correspond directly with the preceding graphs.

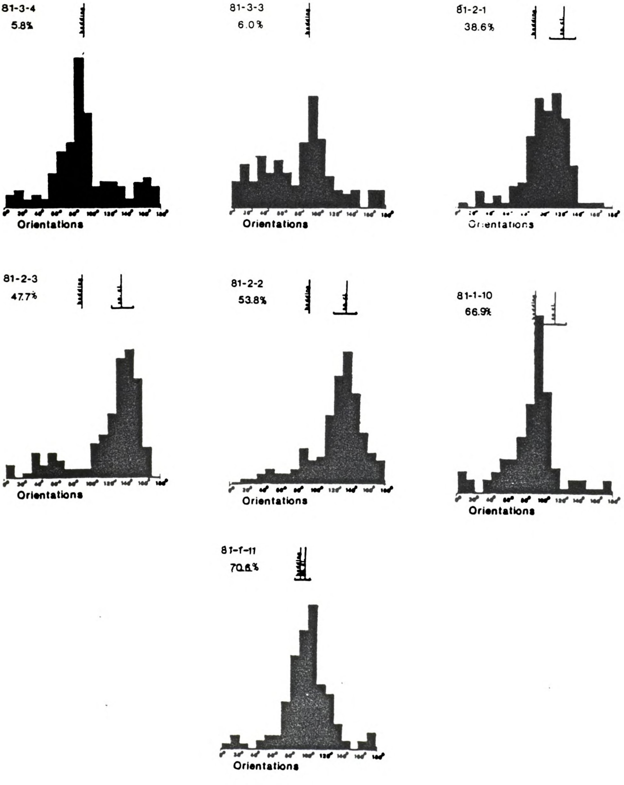

81-3-4
5.8%
Orientations
81-3-3
6.0%
Orientations
81-2-1
38.6%
Orientations
81-2-3
47.7%
Orientations
81-2-2
53.8%
Orientations
81-1-10
66.9%
Orientations
81-1-11
70.6%
Orientations

Figure 6.

Figure 7. Lake Jerdean Graphs: the orientation of the long axis plotted against the axial ratio (long axis/short axis) of each quartz grain. Bedding and average spaced cleavage orientations, as well as the range of spaced cleavage are indicated on each graph. The graphs are in order of increasing phyllosilicate content. *Point concentrations from histograms, Figure 8.

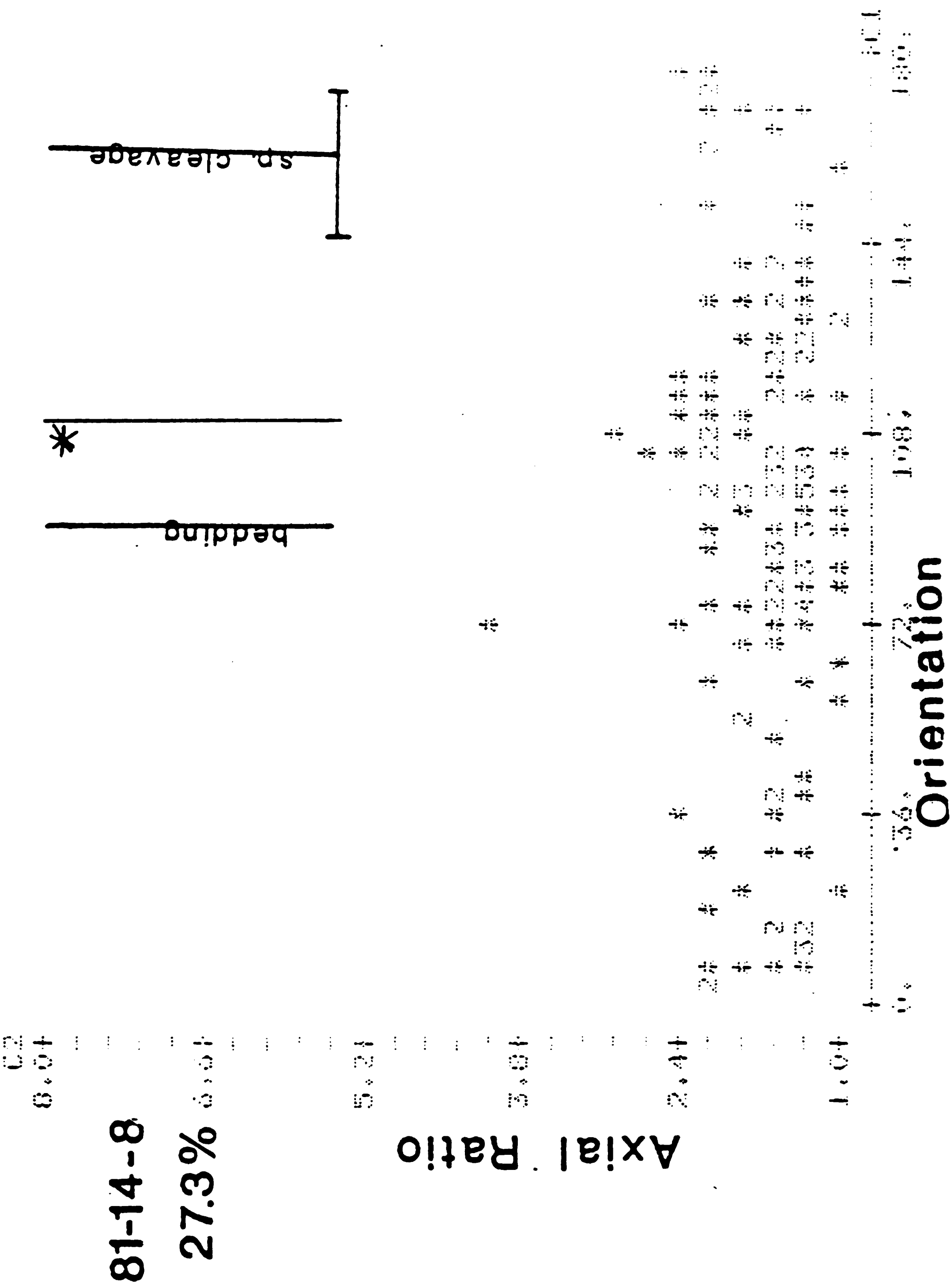


Figure 7a.

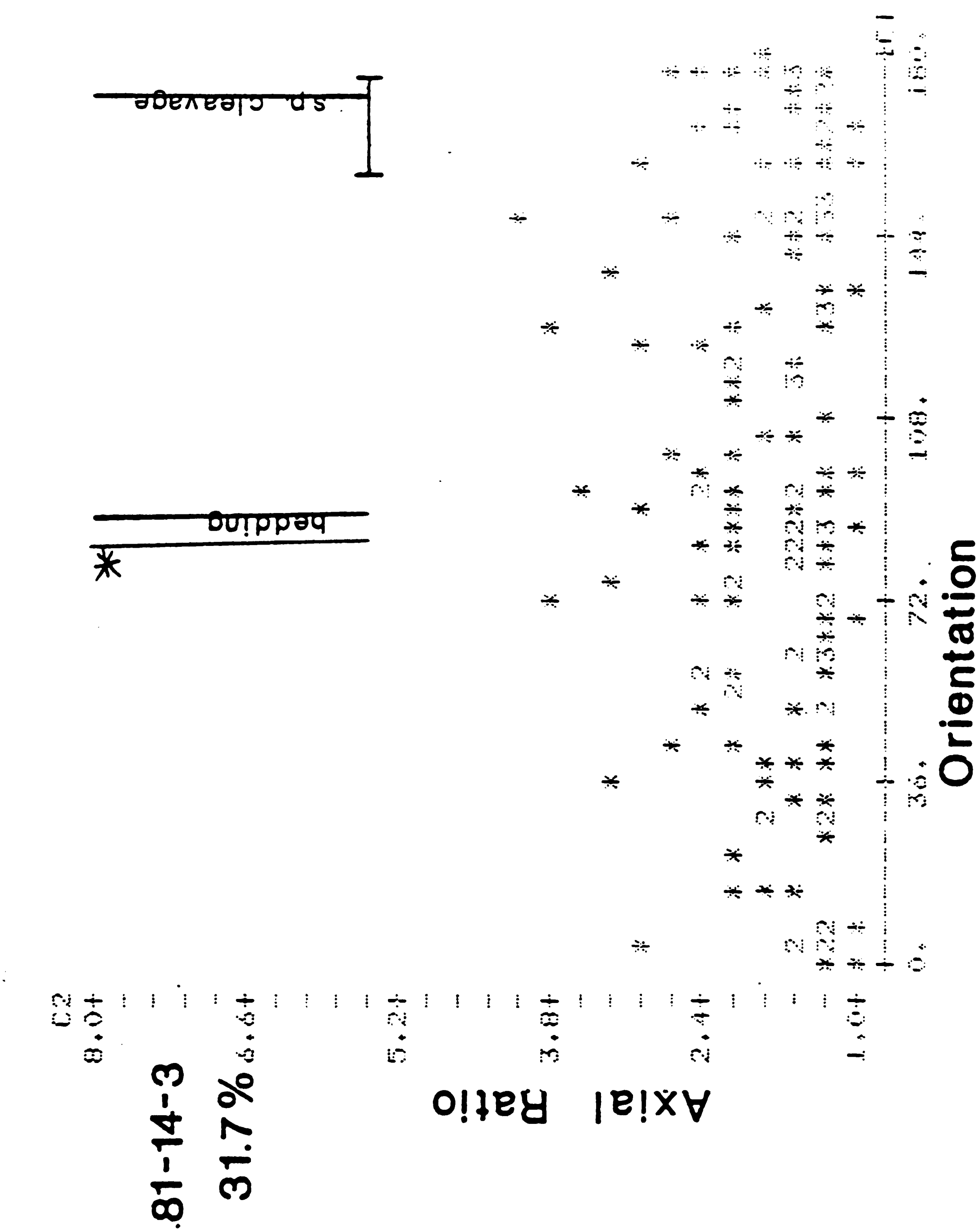


Figure 7b.

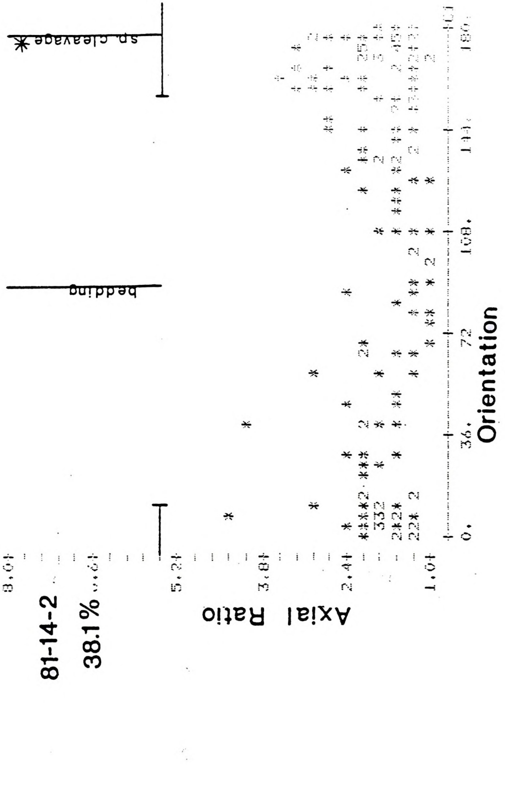

81-14-2
38.1 %
Axial Ratio
1.04
2.44
3.84
5.24
bedding
sp. cleavage
0.
36.
72
108.
180.
Orientation

Figure 7c.

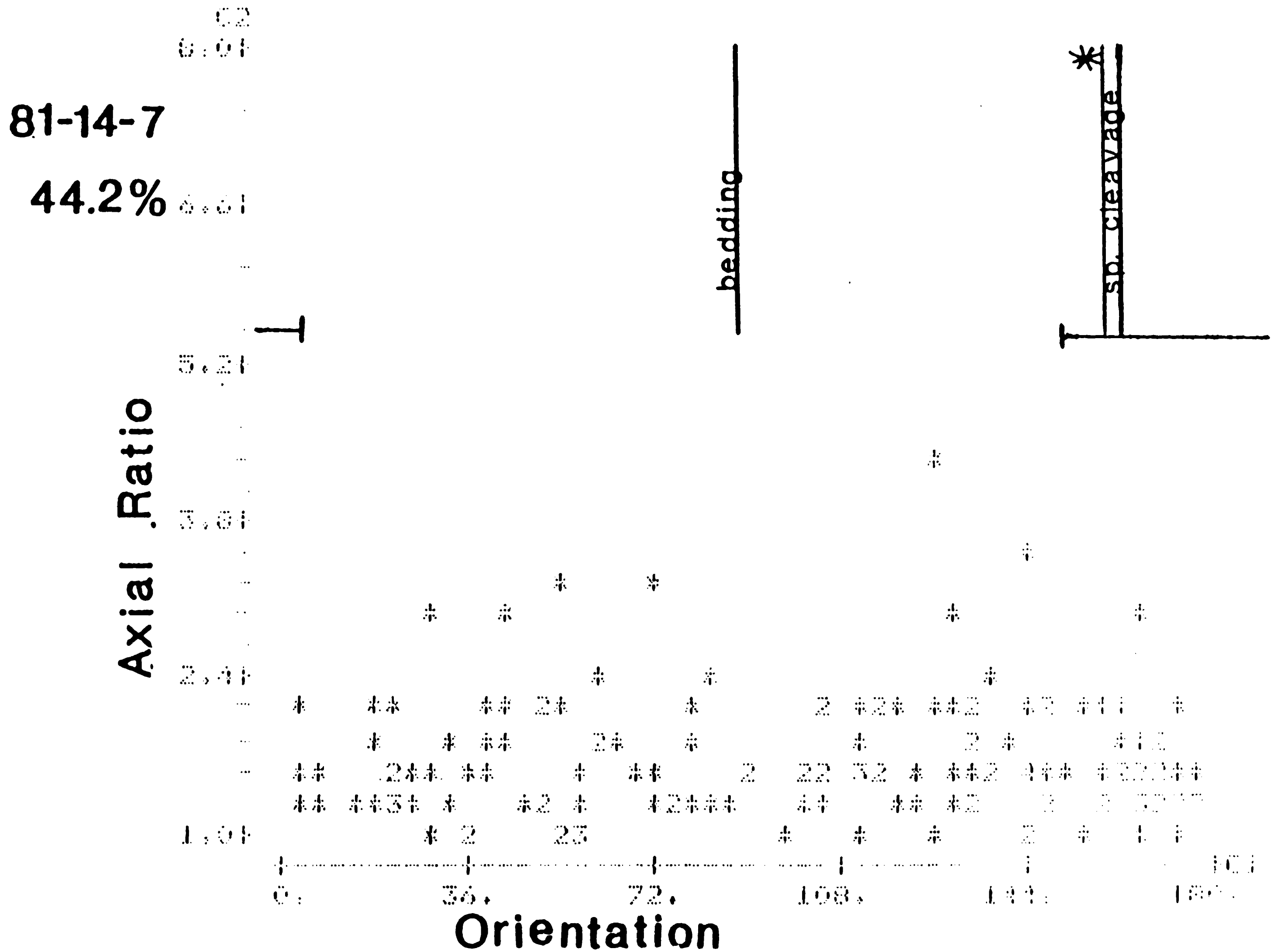


Figure 7d.

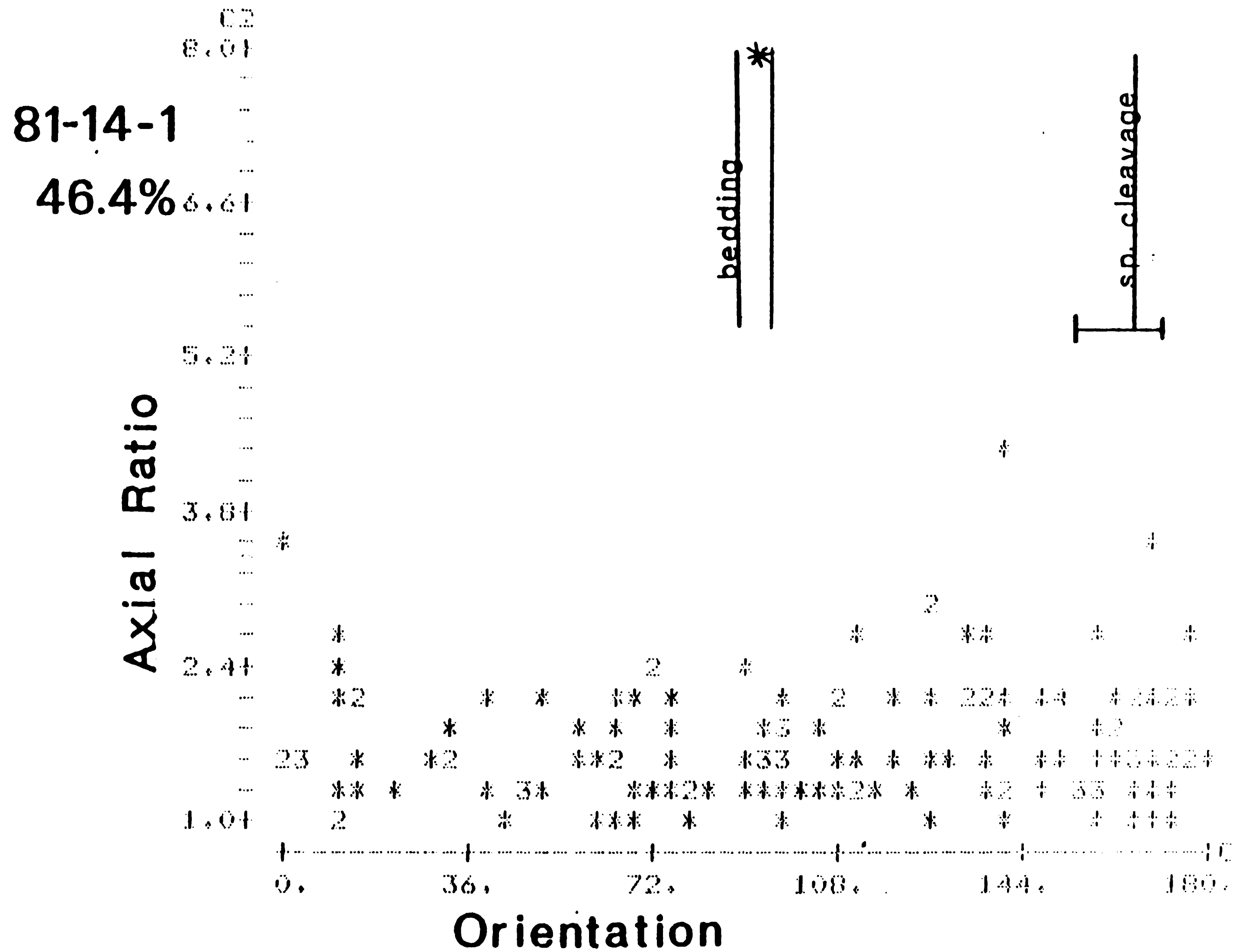

81-14-1
46.4%
Axial Ratio
Orientation
bedding
sp. cleavage
8.01
6.61
5.21
3.81
2.41
1.01
0.
36.
72.
108.
144.
180.

Figure 7e.

81-14-6

56.1%

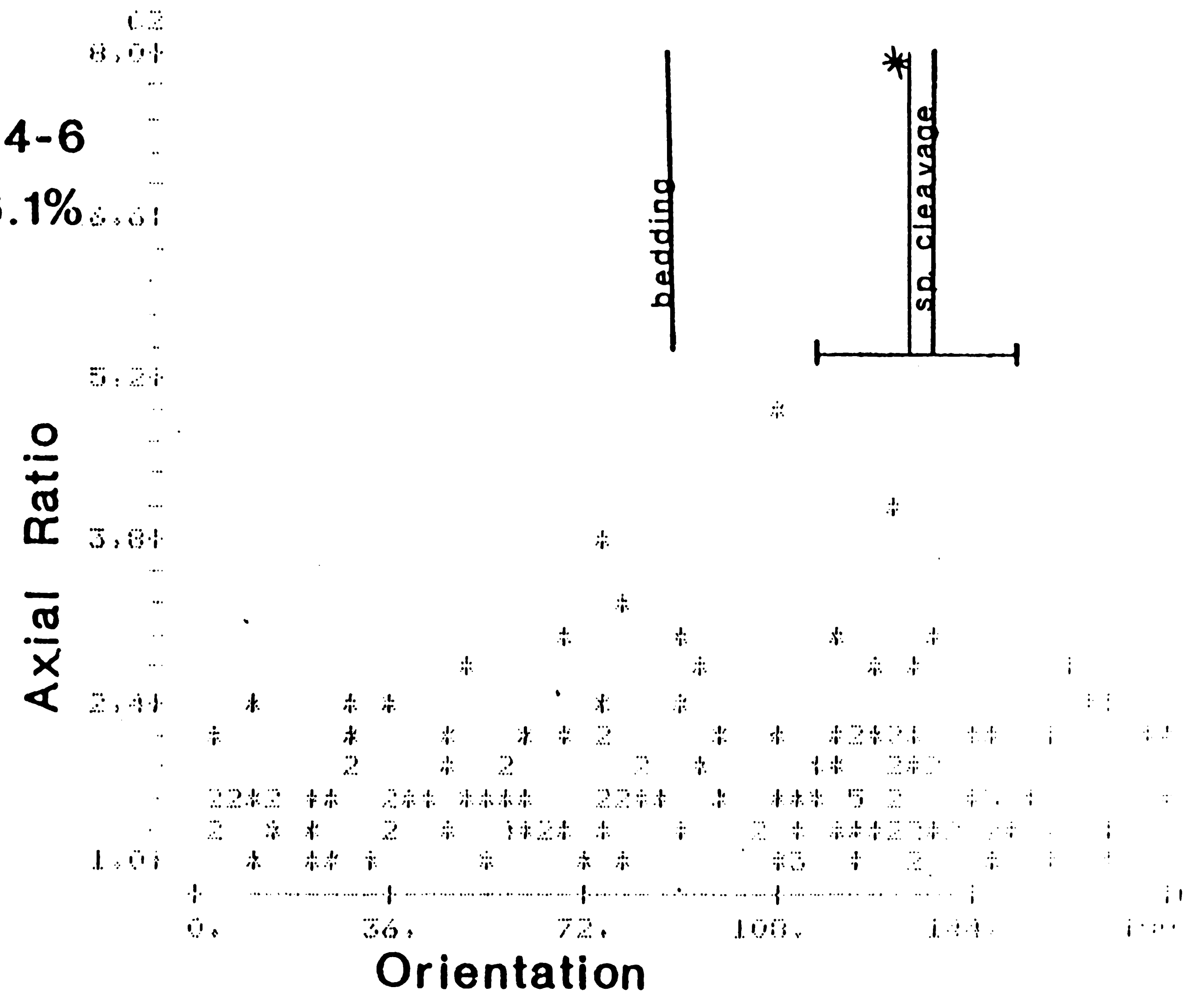


Figure 7f.

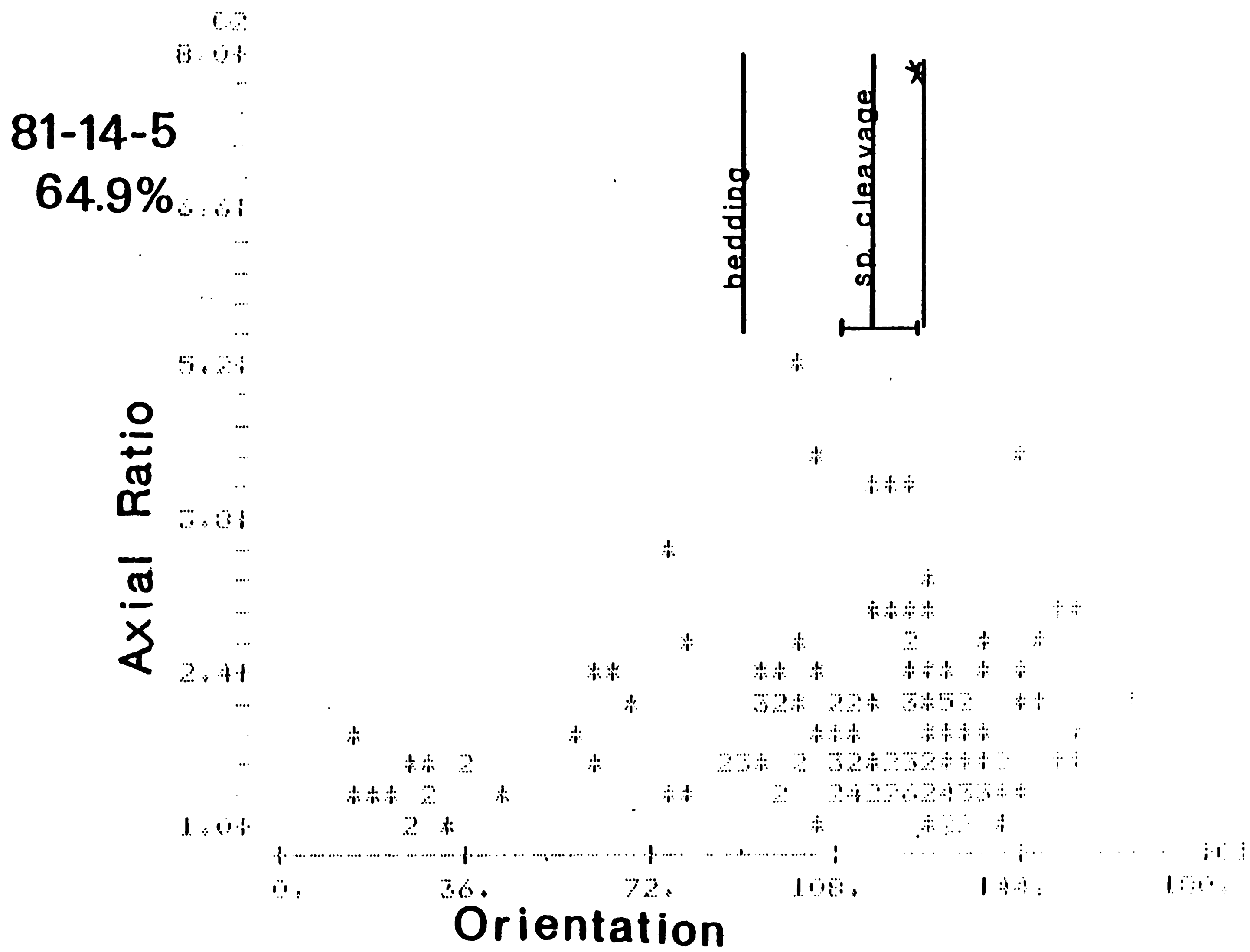
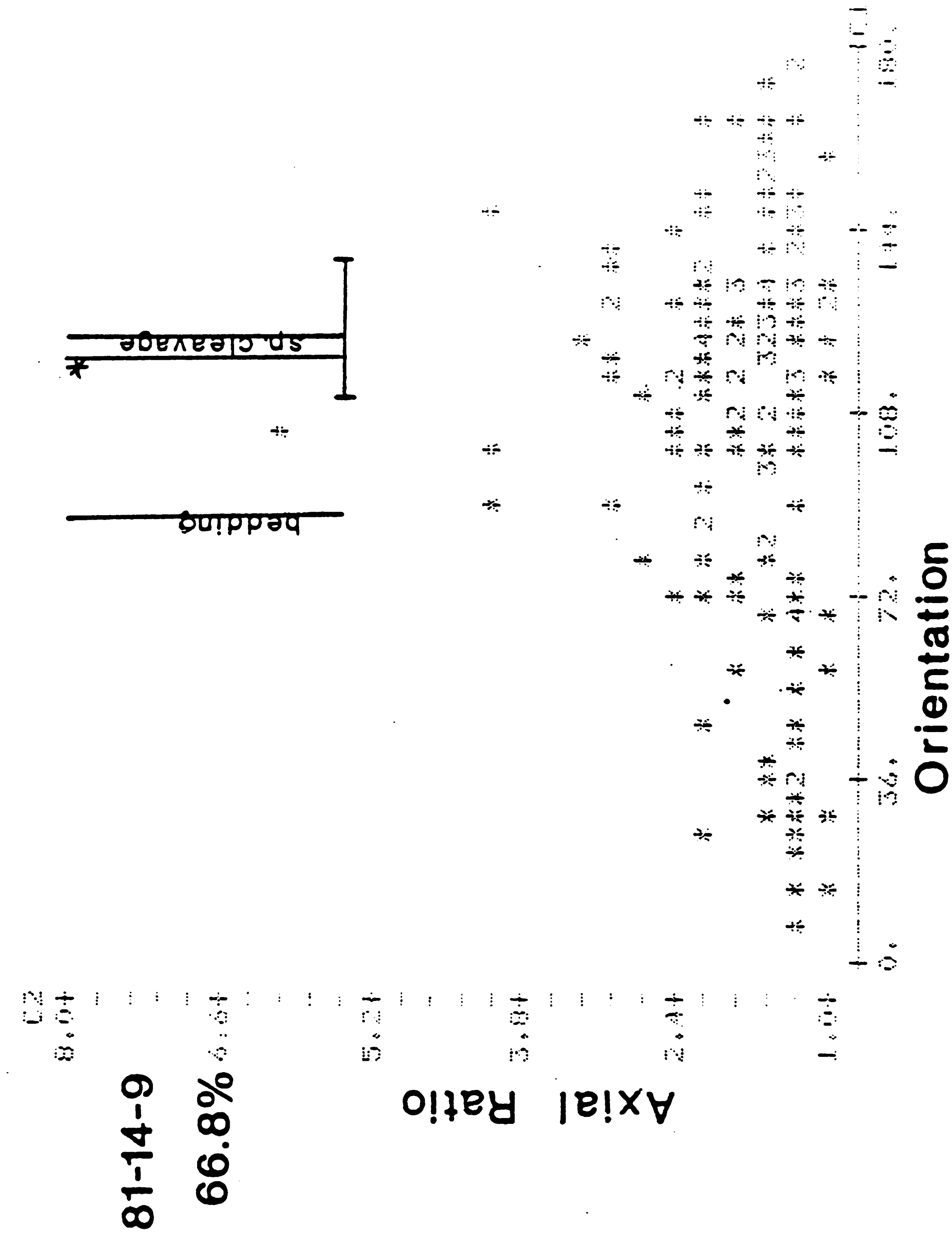

81-14-5
64.9%
Axial Ratio
Orientation
bedding
sp. cleavage
8.04
5.21
3.81
2.44
1.04
0.
36.
72.
108.
144.
180.

Figure 7g.



81-14-9
66.8%
Axial Ratio
Orientation
bedding
sp. cleavage
8.0
6.6
5.2
3.8
2.4
1.0
0.
36.
72.
108.
144.
180.

Figure 7h.

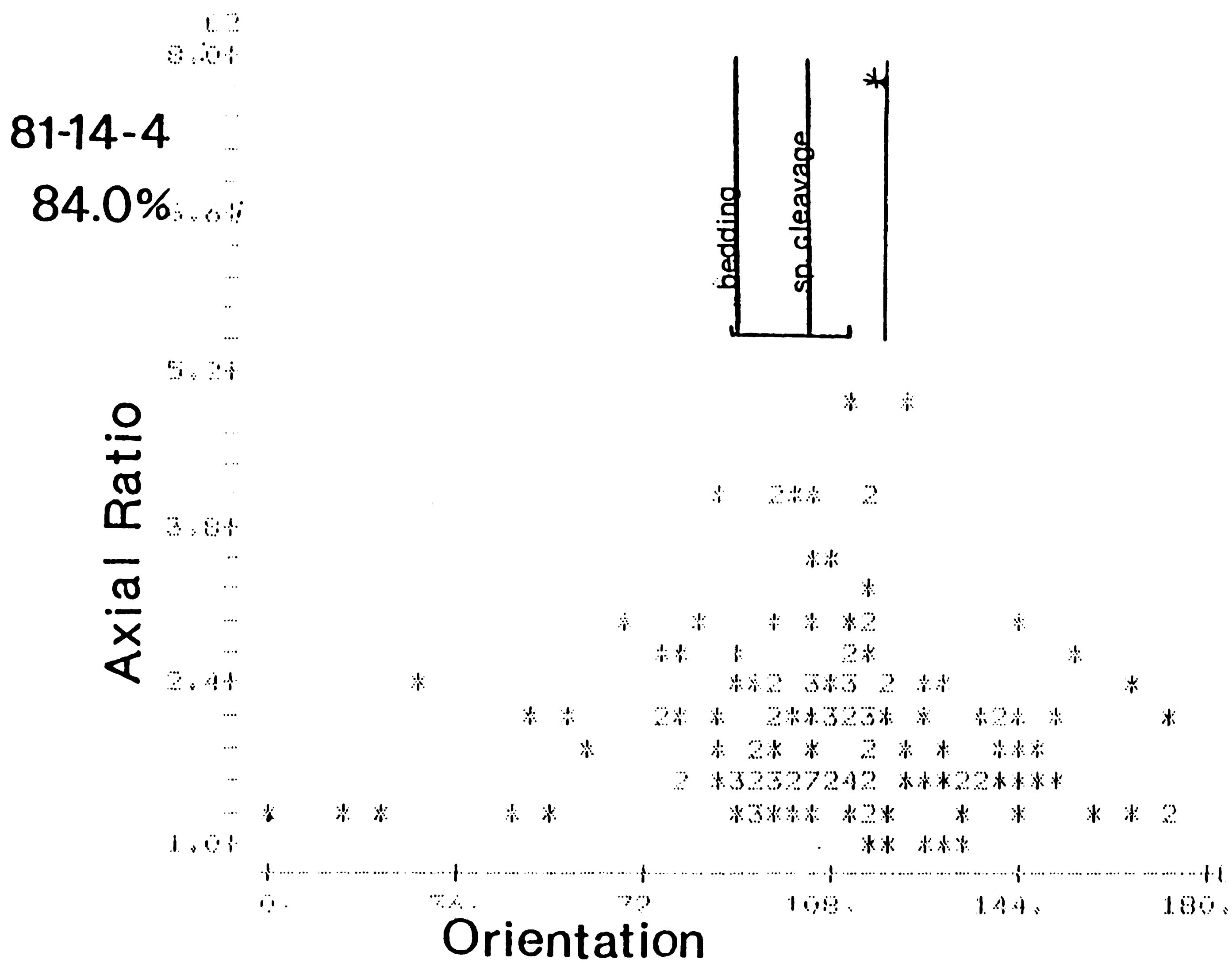
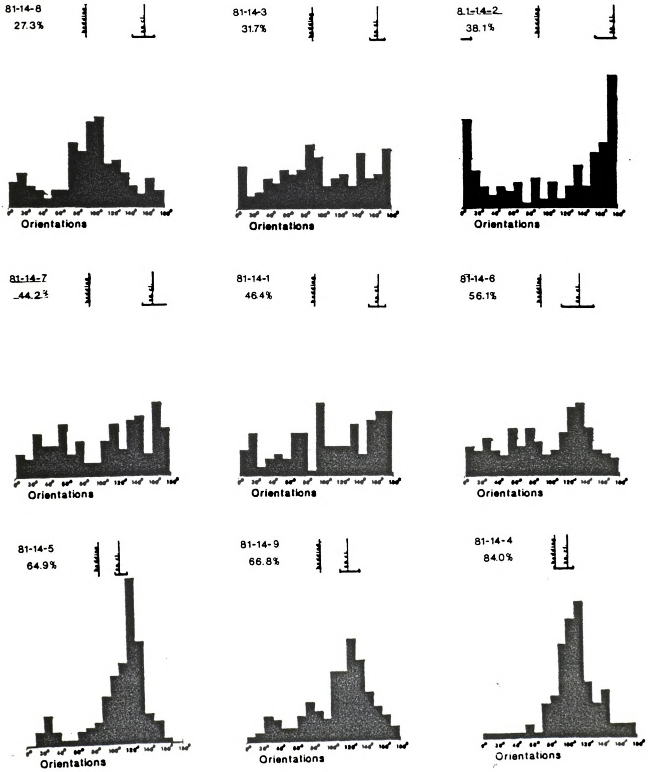

81-14-4
84.0%
Axial Ratio
Orientation
bedding
sp cleavage
8.0
5.2
3.8
2.4
1.0
0.
72
108.
144.
180.

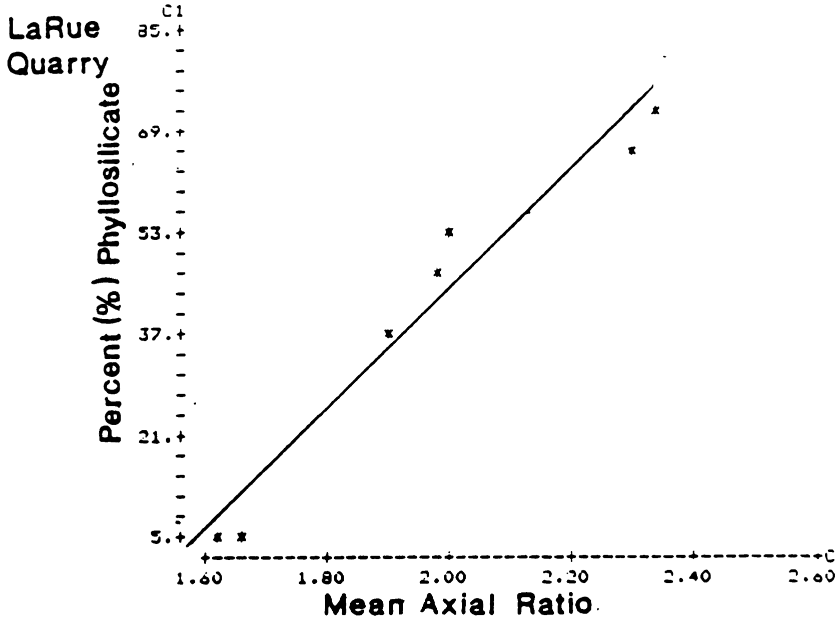
Figure 7i.

Figure 8. Lake Jerdean histograms: indicating the frequency in 10^{o} class intervals, of long axis orientations of quartz grains. Bedding, and spaced cleavage range and average orientation are indicated on each histogram. The histograms are in order of increasing phyllosilicate content and correspond directly with the preceding graphs.



81-14-8
27.3%
Orientations
81-14-3
31.7%
Orientations
81-14-2
38.1%
Orientations
81-14-7
44.2%
Orientations
81-14-1
46.4%
Orientations
81-14-6
56.1%
Orientations
81-14-5
64.9%
Orientations
81-14-9
66.8%
Orientations
81-14-4
84.0%
Orientations

Figure 8.



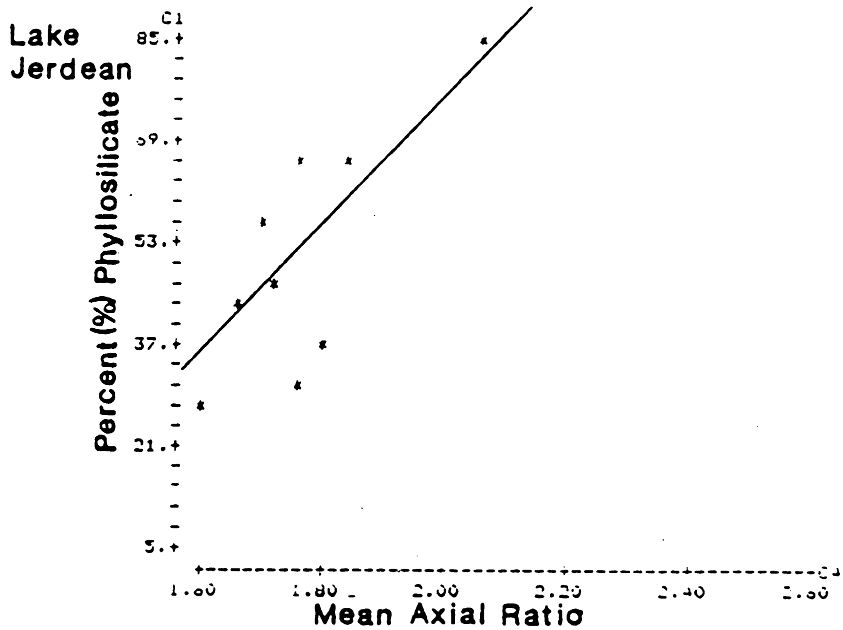


Figure 9. Graphs: the average axial ratio for each section plotted against the phyllosilicate content of the section. Regression lines are indicated. Top: LaRue Quarry. Bottom: Lake Jerdean.

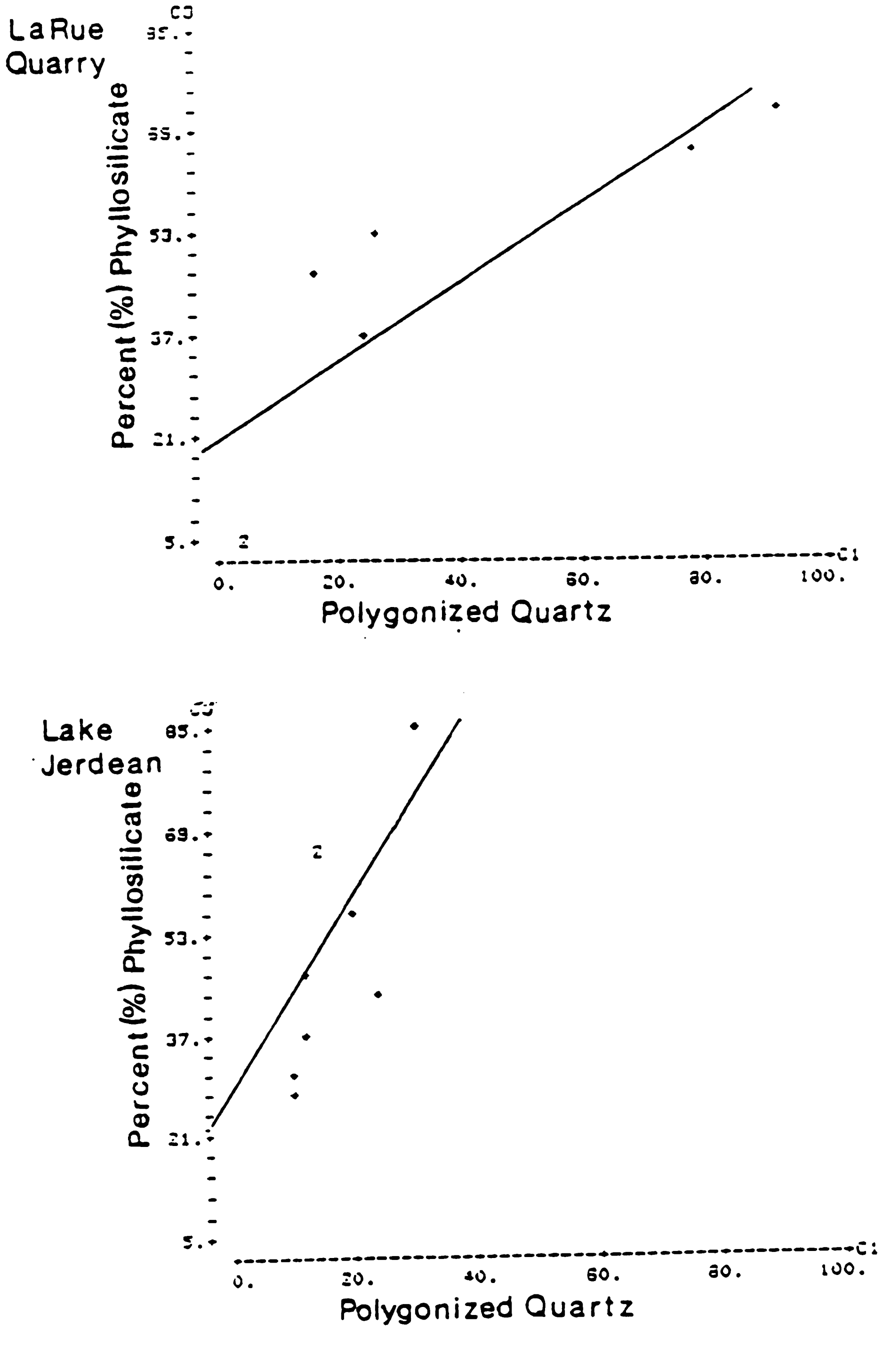


Figure 10. Graphs: percent phyllosilicate versus the percentage of polygonized quartz in thin section. Regression lines are indicated. Top: LaRue Quarry. Bottom: Lake Jerdean.

Linear regression was used to find a best fit line for these graphs and it is marked on the graph.

Description

The spaced cleavage forms anastomosing planes of aligned phyllosilicates. It is generally at right angles to bedding in the pure quartzite. The cleavage is refracted, generally in a gentle curve, as it passes into more phyllitic layers (Figure 11). The width of the spaced cleavage planes ranges from .1 mm to 1 mm. Concentrations of opaques and heavy minerals are often found in the spaced cleavage planes (Figure 12). The number and width of the spaced cleavage planes appears to increase with increasing phyllosilicate richness in the rock.

There seems to have been no major internal recrystallization of the quartz since evidences of strain such as undulatory extinction and deformation lamellae are present. Incipient recrystallization is evident as polygonized quartz occurs at the boundary of many grains. Highly sutured boundaries are common between quartz grains and occasionally one grain is seen to indent another. Dust rings, relict sedimentary grain margins outlined by impurities showing quartz overgrowths, are also common. Pressure shadows and overgrowths are noticeable in all slides (Figures 13 and 15). Quartz grains with microfractures, appearing sliced, are seen fairly often. Also cleavage truncated quartz grains, with flat boundaries adjacent to spaced cleavage planes, appear in most of the sections (Figure 14). Quite commonly, quartz grains are found within the cleavage planes surrounded by phyllosilicates. In all except very quartz rich sections, many of the quartz grains are separated by and surrounded by felted masses of phyllosilicates which appear to have almost intergrown with the quartz grain boundaries. Pressure shadows, fringing stringers of both quartz and phyllosilicate, are often found on the quartz grains. Curved pressure shadows would indicate grain rotation. Two curved pressure shadows were noted, both in phyllosilicate rich sections (Figure 16). Phyllosilicates in all sections are oriented subparallel to the spaced cleavage and parallel to the phyllosilicates within the spaced cleavage. In some of the very phyllosilicate rich sections, areas occur occasionally covering an entire slide which appear to be mylonitized (the grain size is very much smaller than in surrounding areas) (Figure 17). Quartz grain sizes range from .25-2 mm except in these mylonitized areas. The phyllosilicates are very fine grained throughout.

There is a definite contrast between most of the sections cut perpendicular to bedding-cleavage intersections and sections cut at different orientations. In the sections at different orientations, cleavage is wispy and tends to die out more quickly. The phyllosilicate are not as widespread but tend to occur in clusters (blobs). The quartz grains are not generally as elongated and would tend to have a lower axial ratio. The sections cut at other orientations most resemble sections having low phyllosilicate content that are oriented normal to bedding-cleavage intersections. This indicates that, as expected, the strain ratio is not as large in these sections (Figures 18-22).

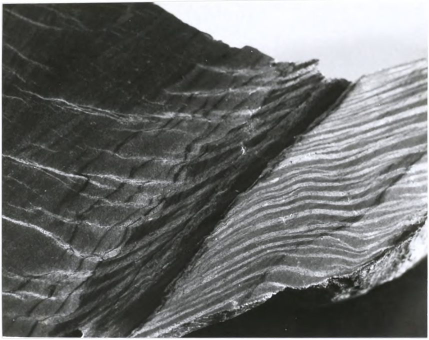


Figure 11. Spaced cleavage and cleavage refraction. Sample from Lake Jerdean. Note the refraction which occurs in passing to more phyllitic beds and the increased amount of spaced cleavage in the more phyllitic beds. Scale: .625" = 1.625".

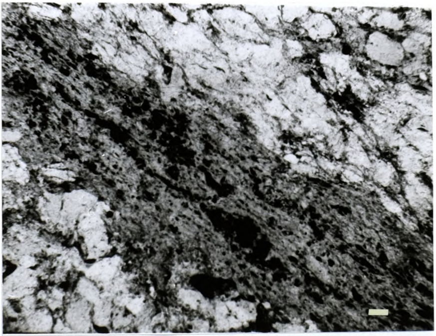


Figure 12. Opaque material in spaced cleavage plane. Plane light. Scale bar = 100µ.

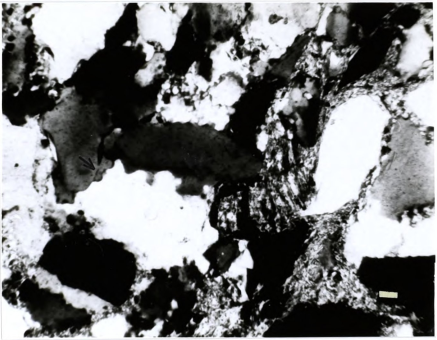


Figure 13. Dust rings and quartz overgrowths (indicated by arrows). Crossed polars. Scale bar = 100μ.

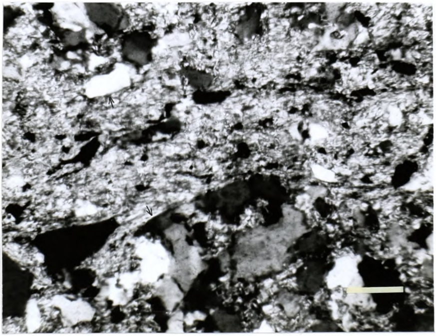


Figure 14. Cleavage truncated quartz grains (indicated by arrows). Crossed polars. Scale bar = 100μ.

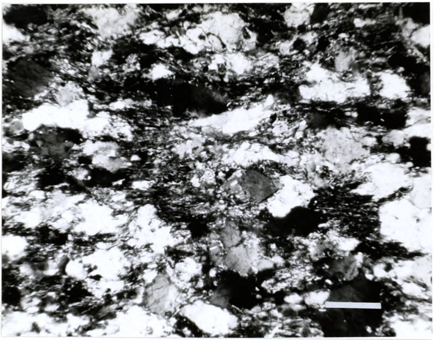


Figure 15. Quartz and phyllosilicate overgrowths. Note the fringing stringers of quartz and phyllosilicate oriented in the cleavage direction. Crossed polars. Scale bar = 100µ.

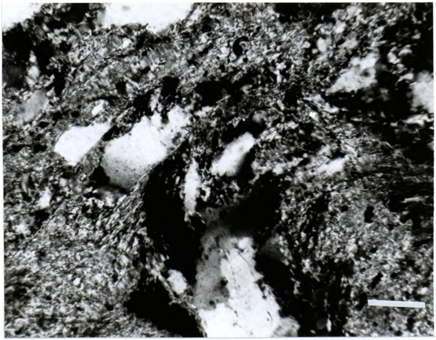


Figure 16. Rotated quartz grain with curved pressure shadow. Crossed polars. Scale bar = 100µ.

Figure 17. Mylonitized quartz in phyllosilicate rich area. Crossed polars. Scale bar = 100µ.

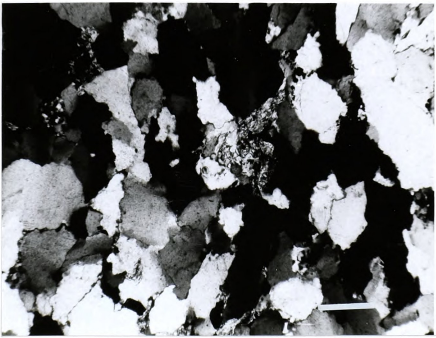


Figure 18. Low phyllosilicate area. 81-3-3. 6% phyllosilicate. Crossed polars. Scale bar = 100µ.

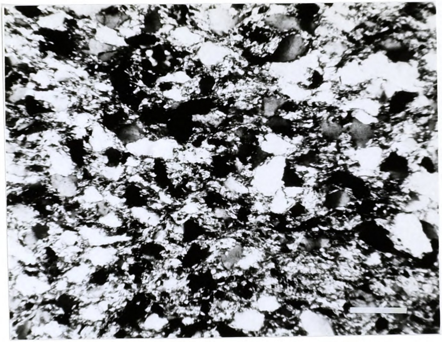


Figure 19. Low-medium phyllosilicate. 81-2-1. 38.8% phyllosilicate. Crossed polars. Scale bar = 100μ.

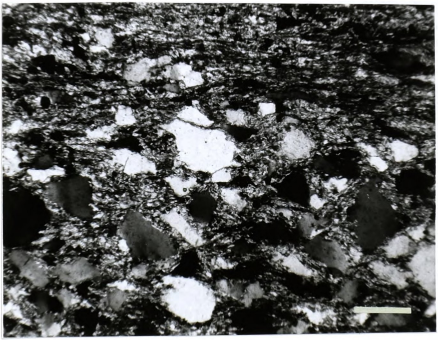


Figure 20. Medium-high phyllosilicate. 81-2-2. 53.8% phyllosilicate. Note the spaced cleavage plane at the top. Crossed polars. Scale bar = 100μ.

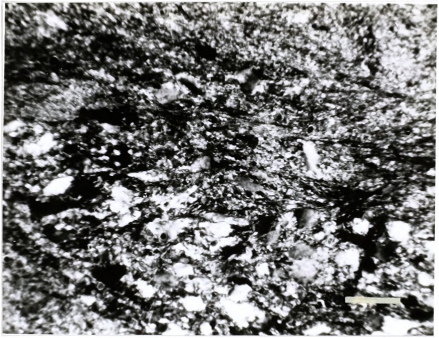


Figure 21. High phyllosilicate. 81-1-10. 66.9% phyllosilicate. Crossed polars. Scale bar = 100μ.

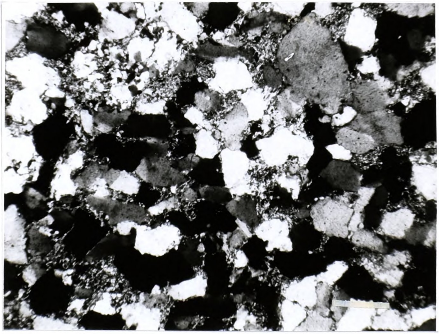


Figure 22. Section cut not oriented normal to bedding cleavage intersection. Note clumped phyllosilicates. More rounded grains. Crossed polars. Scale bar = 100μ.

DISCUSSION

The preceding graphs (Figures 4-8) show a very definite trend; as phyllosilicate content in a rock domain increases, the amount of spaced and the intensity of slaty cleavage also increases. This is what would be expected if pressure solution had played a major role in the formation of the cleavage in the Baraboo Quartzite, because clay (phyllosilicate) minerals are associated with increases in pressure solution (Bosworth, 1981; Gresens, 1966; Kerrich, Beckinsale and Durham, 1975; Durney, 1972b; Heald, 1956; Weyl, 1959; Thomson, 1959).

The first graph (Figure 4) shows the number of spaced cleavage planes per centimeter in each sample plotted against the phyllosilicate content of that particular sample. The number of spaced cleavage planes per centimeter rises sharply as the phyllosilicate content increases; once the percentage of phyllosilicates rises above 20-30%. The 20% phyllosilicate level appears to be a critical point below which little or no spaced cleavage is formed. There exists, then, a strong correlation between the number of spaced cleavage planes per centimeter and the percent phyllosilicate content in each sample. Phyllosilicate content, and thus by inference, pressure solution, does seem to have affected the development of the spaced cleavage.

The next graphs (Figures 5 and 7) and their complementary histograms (Figures 6 and 8) show the effects of increasing phyllosilicate content on slaty cleavage development. Length/width (axial) ratios are plotted against long axis orientation measurements in each graph, and the histograms which correspond to each graph record the frequency, in 10° class intervals, with which each orientation appears. On both the graphs and the histograms, the spaced cleavage and bedding trace orientations are shown. Also, point concentrations, taken from the accompanying histograms, are indicated. These graphs show the reorientation and lengthening of the quartz grains into the spaced cleavage direction as the phyllosilicate content increases. The initial shape of quartz grains should have a low axial ratio (more closely approach sphericity) due to the nature of the grains coming, as they do, from sedimentary sands. Measurements have been made on quartz grains in thin section from a variety of sedimentary rocks (Sneed and Folk, 1958; Moss, 1963; Griffiths, 1967). These have found that most real sedimentary particles (excluding those such as certain crystals and some crinoid ossicles whose axial ratios might have a restricted range) have axial (length/width) ratios between 1.3 and 2 (Sneed and Folk, 1958; Moss, 1963; Griffiths, 1967). The initial long axis orientation of the quartz grains should be parallel to the bedding orientation as clasts will generally lie with their shortest ellipsoidal axes perpendicular to the bedding plane (Elliott, 1970). This means that sections which are cut perpendicular to bedding should show a preferred orientation of elliptical particles parallel to the bedding trace. Initial long axes orientation distributions in sections cut perpendicular to bedding should therefore tend to be either bimodal or unimodal, with unimodal distribution more common and most important (Elliott, 1970; Holst, 1982). In the graphs in Figures 5 and 7, the quartz grains are oriented predominantly in the bedding direction and generally have low axial ratios. The dominant orientation may perhaps be seen more easily in the histograms.

As phyllosilicate content increases, the grains no longer show this strong bedding orientation, and the orientation changes gradually to a somewhat bimodal distribution with a concentration of orientations parallel to bedding and a concentration parallel to the spaced cleavage direction. At the highest phyllosilicate percentages (%), the grains are reoriented to a dominant cleavage orientation. Note also, that the quartz grains in the cleavage orientation have generally higher length/width ratios than those in the bedding orientation indicating a lengthening of the quartz grains in the cleavage orientation.

The first sections on the LaRue graphs (Figures 5 and 6), sections 81-3-4 and 81-3-3, with 5.8 and 6% phyllosilicate, respectively, and no spaced cleavage, show a higher concentration of grains in and around the bedding plane orientation. The axial ratios of the grains are low; not above 3.8 in 81-3-4 nor above 5 in 81-3-3. As the phyllosilicate content increases, the dominant orientation of the grains begins to change. This is seen in the graphs for sections 81-2-1, 81-2-3 and 81-2-2 where the highest concentration of grains begins to fall in and toward the spaced cleavage rather than the bedding orientation. Note also, that the axial ratios of the grains are increasing, with some of the grains reaching a length to width ratio of 6 or more. The spaced cleavage, at the highest phyllosilicate levels, in sections 81-1-10 and 81-1-11, has become very close to the bedding orientation so that it is difficult to separate the effects of each in the orientation of the quartz grains. This is due to cleavage refraction; a fairly well known but not well understood phenomenon where the orientation of cleavage changes as it passes into an area of differing lithology. Although this does make it difficult to separate the cleavage orientation from the bedding orientation, one can see the continued increase in the axial ratios of the grains which in some cases now reaches 7 or 8.

The Lake Jerdean sections, it should be noted, are all from the same sample while the LaRue sections are from three different samples taken from within a few feet in the same quarry. There should, therefore, be a more continuous gradation in lithology and cleavage development from within the Lake Jerdean sections. The Lake Jerdean slides show the same concentration of bedding plane orientations in sections with lower phyllosilicate content (section 81-4-8) as was seen in the LaRue sections, and the same low axial ratio also (Figures 7 and 8). In section 81-14-3, as phyllosilicate content increases, the picture becomes somewhat confused; although, there does seem to be both a small concentration in the bedding plane orientation and also in the cleavage orientation. Section 81-14-2 is anomalous. Its phyllosilicate content is fairly low, yet it shows a strong concentration of grains in the cleavage orientation. Perhaps this is due to some original sedimentary fabric, or to its proximity to the edge of the sample (adjacent to a more phyllitic bed). Again, however, one can see the low axial ratio which has not gone above four for any of the measured grains in the above slides. As phyllosilicate content continues to increase, in sections 81-14-7, 81-14-1 and 81-14-6, one can see the concentration of quartz grain orientations gradually shift to the cleavage orientation. The length/width ratio also increases until it may reach as high as 6. The same problem is encountered in the higher phyllosilicate levels here as was encountered in the LaRue graphs. Cleavage refraction causes the cleavage orientation to become closer to the bedding orientation making it difficult to separate out a dominant orientation of the quartz grains. However, one can again see the continued lengthening of the quartz grains evidenced by the increased axial ratios. This is also shown in Figure 9, where average axial ratios for each section are plotted against increasing phyllosilicate content. A high linear correlation is seen to exist indicating that the length of the quartz grains increases linearly with increasing phyllosilicate content.

If the assumption is made that a high concentration of quartz grains were originally bedding plane oriented and had low axial ratios, and that slaty cleavage has developed due to the gradual reorientation and lengthening of the quartz grains, then it seems that an increase in phyllosilicates (and, by inference, an increase in pressure solution) correlates directly with an increase in slaty cleavage development.

Cleavage refraction occurs where the attitude of the cleavage is changed in passing into domains of differing lithologies and, thus, different physical properties (Steuer and Platt, 1980). Four models have been proposed at various times for the development of cleavage refraction. They are: 1) refraction forms from different processes in rocks of unlike competence, 2) it forms by rotation of shear planes in a competent layer during folding while cleavage in less viscous layers maintains its original attitude, 3) it results from stress-strain distribution changes between layers of varying viscosity, and 4) refraction occurs via differential volume loss between layers, with the passive rotation of a relatively insoluble bed in a soluble matrix (Steuer and Platt, 1980).

The first of these models could not be a cause of refraction for the Baraboo cleavage as it requires a sharp break in lithology and change in competence of the rock. At Baraboo, small changes in lithology bring about a curving cleavage as refraction occurs, and in rocks of differing lithology, the cleavages do not differ in an abrupt fashion but rather grade into one another. Since cleavage development is not thought to occur as a result of shear but normal to the maximimum finite shortening direction, and since the continuous nature of cleavage refraction at Baraboo indicates that it must have occurred somewhat simultaneously with cleavage development, the second model, rotation of shear planes is most probably not a cause for cleavage refraction at Baraboo.

However, in an inhomogenous strain, straight lines may become curved as deformation occurs due to local variations in shear strain. Thus, the third model has some validity. Finite element stress/strain computer modeling of the folding of layers of various viscosities by Dieterich (1969) has produced just such variations and, therefore, presents a strong case for refraction due to viscocity contrasts leading to variations in stress and strain. See Figure 11 for an example of Dieterich's modeling and a comparison with the cleavage refraction in the quartzite.

Steuer and Platt (1980) have only recently proposed the fourth model for cleavage refraction. Their model is based on the fact that often early cleavage lamellae pass undeflected from layer to layer even in rocks of differing lithology, and that there are differences in solubility between areas with only slight changes in lithology, thus explaining cleavage which is refracted (curved) due to small changes in lithology. Their model is a plausible one for the cleavage at Baraboo since the cleavage at Baraboo undergoes a gently curving refraction with a gradual increase in the phyllosilicate content of the rock. The thickness of the spaced cleavage and the number of spaced cleavage planes also both increase with the phyllosilicate content of the rock which is what might be expected if solubility differences (and thereby, pressure solution) were responsible for the formation and refraction of the cleavage at Baraboo (Figure 11).

These last two models are not mutually exclusive. Changes in stress or strain could be caused or increased by volume loss, and volume loss might be increased by increased stress or strain in a domain. There therefore seem to be three possible explanations for the refraction of spaced cleavage in the Baraboo quartzite: 1) that it developed as a result of local variations in stress and strain due to differences in viscosity, 2) that it developed as a result of differential volume loss due to differences in solubility, or 3) that some combination of the above two processes occurred. The third explanation seems the most probable.

Some other features of the quartzite and the cleavage at Baraboo should be noted also. Pressure solution often results in the accumulation of insoluble residues (in quartzite rocks, phyllosilicates, opaque, and heavy minerals); these

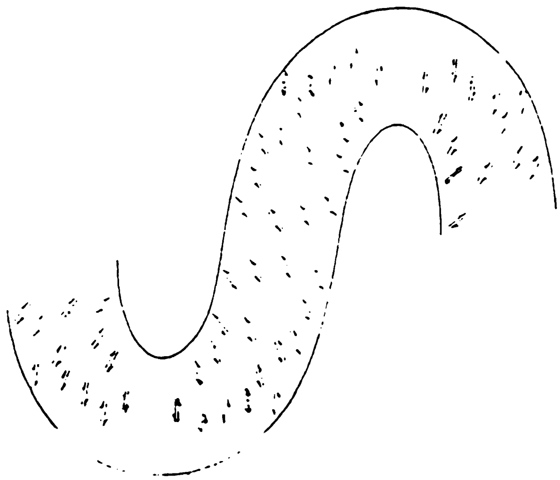


Figure 23a. Dieterich's model. Arrows indicate the sense of shear stress. This is a computer model of a folded layer in a less viscous matrix, 100± average compressive strain, principal axes of stress and strain inclined at large angles. (From Dieterich, 1969.)

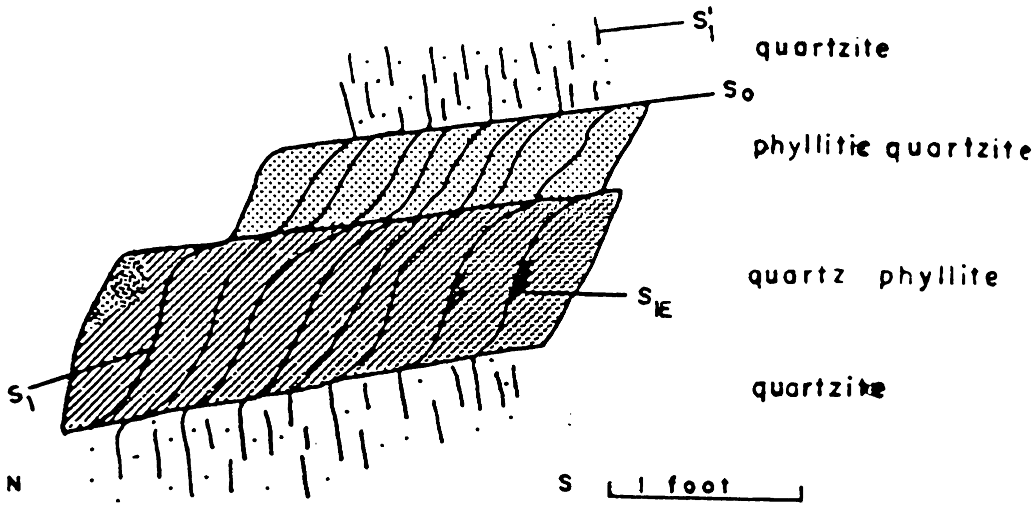


Figure 23b. Refraction of spaced cleavage in the Baraboo Quartzite. (From Dalziel and Dott, 1970.)

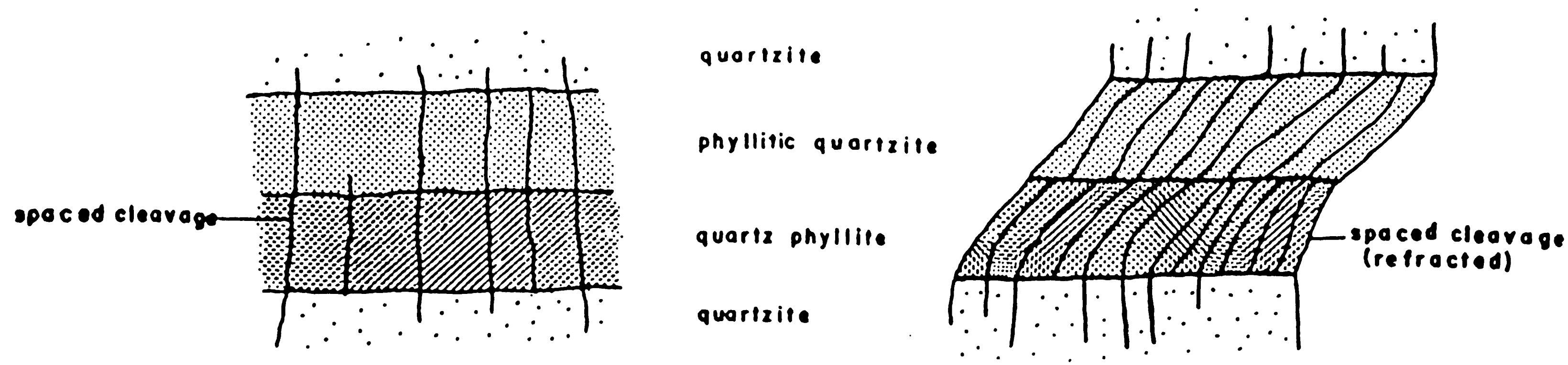


Figure 23c. Steuer and Platt's model. Diagrammatic representation of its application to the Baraboo Quartzite. Cleavage forms originally normal to bedding. As more material is dissolved in the more phyllosilicate rich areas, the existing cleavage is rotated and more cleavage forms in these areas.

form stylolites. If the spaced cleavage planes in the quartzite are formed by a process analogous to that which produces stylolites, one would expect to find the same concentration of opaques, heavy minerals, and phyllosilicates in the cleavage planes. This is exactly what is seen (Figure 12). In this figure, one can see that the zircons and opaques are concentrated in the cleavage planes while outside of the cleavage there is no such concentration. Dust rings are seen on some quartz grains (Figure 13); these indicate that some dissolution and reprecipitation of quartz has occurred although the dust rings quite probably were formed during diagenesis. Pressure shadow overgrowths appear in all sections. These indicate that quartz dissolution and reprecipitation occurred in response to a stress gradient (Figures 15, 26 and 27). Interpenetrative grain boundaries are somewhat common. Quartz grains are also found that are truncated, with flat edges against the spaced cleavage planes as if they had been preferentially dissolved on that boundary (Figure 14).

Other explanations can be found for some of these results. The increase in cleavage development with an increase in clay minerals could be due simply to adsorbed water in the clays enhancing the process of recrystallization. The hydrolitic weakening of quartz (increased recrystallization due to the presence of water) is a well known phenomenon (Blacic, 1975; Griggs, 1967; Jones, 1975) and the correlation found between the amount of phyllosilicate present and the percentage of polygonized grains in the sections indicates that the phyllosilicates do seem to have affected the recrystallization of the quartz. Recrystallization is often thought to be the major cleavage forming mechanism. Recrystallization is the nucleation and growth of new grains, either within grains of like or different phases or primary recrystallization which is the growth of entirely new grains (Hobbs, Means and Williams, 1976; Etheridge and Hobbs, 1974). Pressure solution involves recrystallization by the processes of dissolution, diffusion and reprecipitation in response to nonhydrostatic stress (Beach, 1979; Wood, 1974). Recrystallization can be controlled by local stress and strain, by defects in crystalline structure, by temperature, by pressure, and by chemical activities (Etheridge and Hobbs, 1974).

Stress and strain controlled recrystallization can cause reorientation. DeVore (1969) has stated that the recrystallization of a mineralogically heterogeneous rock under nonhydrostatic stress conditions should result in preferred mineral orientations that are related to the stress direction. This is most easily seen in experimental studies on micas where mica grains have been crystallized and oriented in response to a directed stress (Etheridge, Paterson and Hobbs, 1974; Etheridge, 1971; Means and Rogers, 1964). Defects in crystalline structure can also control reorientation due to recrystallization by affecting the location of sites of nucleation and influencing the process of new grain growth.

Temperature can control whether recrystallization or slip will be more important in the reorientation of grains. Slip is a process which occurs in crystals at pressures high enough to prevent brittle fracture and low enough so that solid state diffusion is unimportant. Slip begins as a rearrangement of atoms in the crystalline structure in a region of high stress concentration and propagates outward. The line separating the slipped region in a crystal from the unslipped region is called a line defect or dislocation. Slip enables a grain to change shape and, therefore, orientation by shear of one portion of a crystal with regard to another (Hobbs, Means and Williams, 1976).

When deformation takes place at higher temperatures or lower strain rates, solid state diffusion becomes important. Diffusion and slip are interrelated as diffusion aids in slip by allowing dislocations to climb from one slip plane to another. When diffusion is important on the scale of grains, grain boundaries may move or newly recrystallized areas may grow. Slip and recrystallization (new grain growth) are difficult to separate, and the patterns of orientation which develop are, in some situations, identical (Hobbs, Means and Williams, 1976). Pressure controls whether recrystallization or brittle fracture will occur; that is whether the rock will deform in a ductile or brittle manner. Chemical activities are a control on new grain growth; they affect if and when it will occur.

Pressure solution differs from other recrystallization mechanisms in that it is an intercrystalline rather than intracrystalline mechanism. It is a process of dissolution, rather than solid state diffusion; diffusion, by intercrystalline rather than intracrystalline processes; and reprecipitation, new grain growth.

Microstructures which develop in deforming rocks are, therefore, the result of mechanical factors, competing controls and the competing and interrelated processes of recrystallization, pressure solution and slip whose effects are quite difficult to separate.

In the Baraboo Quartzite, recrystallization of the phyllosilicates has occurred and the quartz grains have begun to recrystallize; the rocks having undergone greenschist facies metamorphism. Phyllosilicates are entirely recrystallized in a preferred orientation and some nucleation and new grain growth (polygonization) has occurred in the quartz grains also. However, undulatory extinction and deformation lamellae are common features of the quartz grains, so recrystallization does not seem to have occurred on a large scale. Or, the quartz has been recrystallized and subjected to continued deformation. Large scale recrystallization of the quartz has probably not taken place since original quartz boundaries and diagenetic structures such as dust rings are preserved. The seemingly mylonitized areas seen at higher phyllosilicate levels could be considered larger scale recrystallization as they are a result of polygonization of the quartz which could be a recrystallization process. Slip was probably not a major factor in the Baraboo cleavage development since temperatures would have been high enough to make diffusion the more important process. The recrystallization of the phyllosilicates in a definite orientation could have affected the orientation of the quartz grains by a preferred inhibition of normal grain growth and boundary migration of the quartz grains by the phyllosilicates. This alternative does not seem feasible as the phyllosilicates, while exhibiting a preferred orientation, as a whole form a felted mass which completely surrounds many of the quartz grains. This would have inhibited grain growth in all directions, including the cleavage direction. The recrystallization which did possibly occur (resulting in the polygonization of the quartz) could not have resulted in cleavage formation. Polygonization results in a large number of much smaller grains with more equant boundaries forming from the original quartz grains. The polygonitization would not result in a preferred orientation or a larger axial ratio in the quartz grains although it may have been formed by the same process.

There is some evidence indicating that shearing has occurred on the cleavage planes. Slickensides are seen on some cleavage surfaces; and mylonitization of both the quartz and the micas seems to have occurred in some areas. Mylonitization is the formation of fine grained equivalents of adjacent rocks in narrow domains where deformation has been more intense (Bell and Etheridge, 1973; Hobbs, Means and Williams, 1976). In the Baraboo quartzite, recrystallization of the quartz and phyllosilicates seems to have occurred resulting in the breakdown of large grains and the nucleation and growth of smaller strain free grains. No evidence of brittle deformation is seen. Mylonites do not necessarily originate by shearing. Johnson (1967) has suggested that in many mylonite zones, the strain is normal to the fine grained bands and mylonite layering has an origin similar to that of slaty cleavage. Cleavage has been shown by many studies (Harker, 1886; Siddans, 1972; Wood, 1974) to form normal to the short axis of the finite strain ellipsoid making it doubtful that the cleavage formed as a result of shear strain. It is possible that the cleavage, once formed, acted as areas of weakness along which shearing took place; but, according to the above studies, this would have to involve a change in the strain orientation. The results of the previously mentioned studies have also been questioned by some workers (Becker, 1896; Williams, 1976). Dieterich (1969) has shown that small variations in stress and strain may occur along a cleavage plane and this might explain the mylonitized areas or any small areas of shear along the cleavage. The sliced broken quartz grains seen in some of the slides might also be attributed to shearing. Neither the broken grains nor the mylonitization are common features; they are found more often in phyllosilicate rich sections. It seems probable then that any shear which took place was either a later phenomenon, or due to small changes in stress and strain along the cleavage attributable to viscosity contrasts (Dieterich's model, 1969).

Little evidence of grain rotation has been found, although any such evidence (i.e., curved pressure shadows) might have been obliterated by recrystallization. Rotation could result in a bimodal pattern and is also more likely in phyllosilicate rich rocks. But, rotation of grains would not give the gradual change in orientation that is seen on the graphs; nor would it explain the lengthening of the quartz grains and increase in axial ratio in the cleavage orientation. Therefore, although grain rotation may have occurred through the dissolution of nearby grains, it does not appear to have been a major mechanism in the reorientation of the quartz grains and the development of slaty cleavage.

The formation of the spaced cleavage in the quartzite by fracturing, as suggested by Dalziel and Stirewalt (1975) is not supported by petrographic evidence. The spaced cleavage does not pass through grains, splitting them, as one might expect if fracture had formed the cleavage. The phyllosilicates inside and outside of the cleavage planes show a definite preferred orientation subparallel to the cleavage planes which would be unlikely if the cleavage was due to fracturing and subsequent welding by the phyllosilicates. The preferred orientation could, however, have developed during recrystallization. The curving refraction of the spaced cleavage would be difficult to account for by fracturing. Fracture and later phyllosilicate infilling would not appear to be likely mechanisms for the development of the spaced cleavage at Baraboo.

From the graphs and thin section studies, it can be seen that the slaty cleavage and spaced cleavage in the quartzite are related, and grade into one another. Their progression and development also seem to be related. Therefore, it would be more acceptable if any mechanism proposed for the origin and development of one of these cleavages would account for or have an effect on the origin and development of the other. Fracture and grain rotation cannot account for this interrelationship while pressure solution could.

The crenulation cleavage in the quartzite is not well developed. A slight buckling of the phyllosilicates in the spaced cleavage planes is often the only sign of it (Figure 24). This buckling is possibly due to a nearby anisotropy in the cleavage or in the rock (Hammer, 1979). There is no petrographic evidence either for or against pressure solution in these faint buckles but for a slight, not often visible, lineation of the opaques present in the spaced cleavage planes (Figure 25). If pressure solution had occurred, one would expect to find concentrations of insoluble material along the limbs of these microfolds. These concentrations are not seen, but this may be due to the fact that the crenulation cleavage development has not progressed to this point with the above mentioned lineation being the precursor.

Figure 24. Crenulation cleavage (indicated by arrows). Partially crossed polars. Scale bar = 100µ.

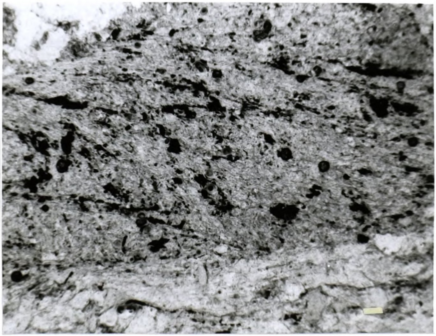


Figure 25. Lineation of opaques. Perhaps the precursor of crenulations? Plane Light. Scale bar = 100µ.

This study, then, proposes a model for the development of the spaced and slaty cleavages in the Baraboo Quartzite by pressure solution. In the case of the spaced cleavage, small concentrations of clay minerals probably occurred as the sediments were deposited. These anisotropies provided loci for the beginning of pressure solution. Where the clay minerals were concentrated, dissolution occurred at a faster rate, and this changed the stress field locally allowing the beginning cleavage plane to grow laterally. This lateral growth continued as quartz grains were dissolved and phyllosilicates, opaques and heavy minerals were concentrated as insoluble residues in the cleavage planes in a process analogous to stylolitization. The quartz dissolved from these planes migrated through pore fluids and was deposited as pressure shadows aiding in the elongation and reorientation of the quartz grains which forms the slaty cleavage. The quartz grains in the clay rich rocks underwent dissolution and overgrowth deposition in pressure shadows on the ends of grains to lengthen and realign them into the same orientation as the spaced cleavage. Pressure shadows form by the growth of minerals of like or different type in zones of low pressure (stress) adjacent to grains. These regions of low stress are a portion of a stress gradient about grains from areas of high stress which are normal to the maximum principal stress to the areas of low stress which are normal to the minimal stress (Kerrich, 1977; Figures 26 and 27). This stress gradient acts as a driving force for the dissolution of the grain in the high stress region and the deposition of material (in this case, quartz) in the low stress region as a pressure shadow. Pressure shadow and overgrowth deposition in the quartzite thus resulted from the dissolution of the grains themselves and the dissolution of grains to form the spaced cleavage planes. Cleavage refraction occurred as a result of a combination of differential volume loss (Steuer and Platt, 1980), and changes in

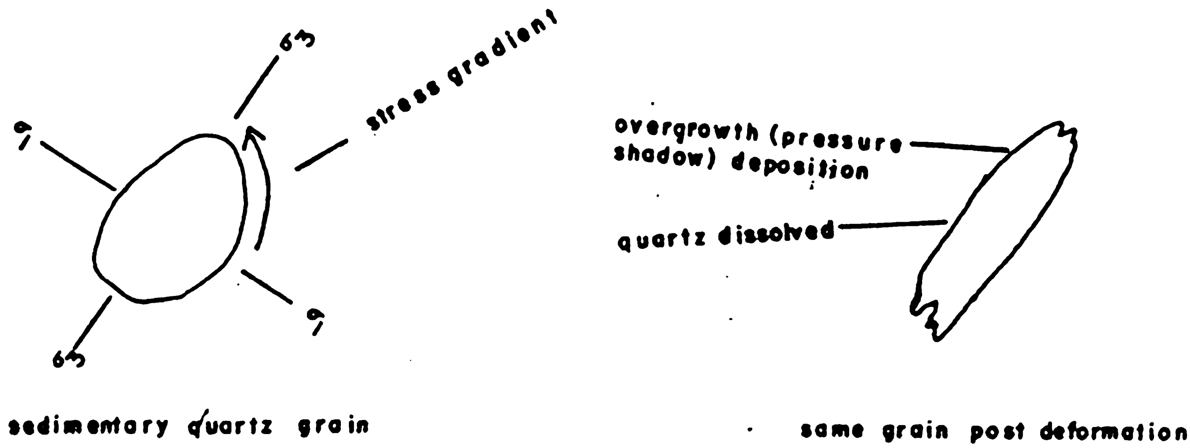


Figure 26. This demonstrates diagrammatically how overgrowth formation would occur. In a stress field, the quartz grain dissolves where the stress is greatest and this material is deposited as an overgrowth where the stress is least.

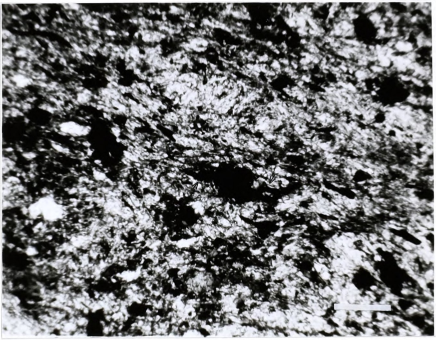


Figure 27. Quartz overgrowth on quartz grain (indicated by arrows). Overgrowths generally form in the direction of least stress. Crossed polars. Scale bar = 100µ.

stress and strain orientation due to viscosity contrast (Dieterich, 1967). Some grain rotation as a result of dissolution probably also occurred.

CONCLUSION

This study has centered on an assessment of the role of pressure solution in the formation of the spaced, slaty and crenulation cleavages present in the quartzite of the Baraboo quartzite. The approach to this was twofold. First, measurement of the amount of phyllosilicate and its relationship to the presence of spaced and slaty cleavage in a domain, and second, petrographic determination of evidence existent which would support or disallow pressure solution as a mechanism for cleavage formation in the quartzite.

It was found that increased phyllosilicate content in the rock resulted in an increase in the amount of both spaced and slaty cleavage developed. There is petrographic evidence for pressure solution in the accumulation of insoluble material in the spaced cleavage planes, and in the occurrence of truncated grains, pressure shadows and overgrowths.

Other mechanisms for cleavage formation, such as recrystallization, grain rotation and fracture are not supported by the evidence. There is very little petrographic evidence of cleavage formation by grain rotation or fracture and neither of these could account for the interrelationship of the spaced and slaty cleavage seen in the quartzite. Recrystallization of the phyllosilicates has occurred and this may have destroyed evidence of rotation and fracture. Little recrystallization of the quartz has occurred. The recrystallization of the quartz which has occurred appears to have been enhanced by the presence of phyllosilicates, perhaps due to hydrolitic weakening of the quartz by water present in the phyllosilicates. The processes of slip, recrystallization and pressure solution are difficult to separate. Pressure solution is a mechanism for recrystallization and aids in the process of intracrystalline slip so that one cannot say with certainty that slip and recrystallization were not part of the cleavage forming process. However, since little recrystallization of the quartz has occurred, except perhaps by the polygonization of the grains, it does not seem likely that recrystallization alone (without pressure solution) was a major cause of cleavage formation.

Theories on cleavage refraction when applied to the refraction of the spaced cleavage at Baraboo indicate that the refraction probably occurred as a result of some combination of differential volume loss due to differences in solubility (Steuer and Platt's model, 1980), presumably through pressure solution, and local variations in stress and strain due to viscosity contrasts (Dieterichs model, 1969).

There is some evidence of shear along portions of the cleavage planes. This may be due to small changes in stress and strain along the cleavages, or to later slip along the spaced cleavage planes.

Crenulation cleavage is not well developed making it difficult to determine what role, if any, pressure solution may have played in its formation. A slight lineation of opaques in some of the spaced cleavage planes may form the beginning of crenulations.

Further study in this area, especially for example, studies of the phyllite using TEM and SEM, more work on cleavage refraction, studies of the crenulation cleavage and chevron folds which are well developed in the phyllite, would certainly be of interest.

The results of this study lead to the conclusion that pressure solution was a definite factor in, if not the mechanism for, the development of cleavage in the Baraboo quartzite. Other mechanisms, such as recrystallization and rotation, cannot be completely excluded. Factors, such as metamorphism and the lithology of the rock, may obscure evidence. Nevertheless, it does seem clear that pressure solution played an important role in the formation of the cleavage in the quartzite of the Baraboo Quartzite.

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