Downloaded from gsabulletin.gsapubs.org on August 9, 2013

Geological Society of America Bulletin

"Fracture cleavage" in the Duluth Complex, northeastern Minnesota

M. E. FOSTER and P. J. HUDLESTON

Geological Society of America Bulletin 1986;97, no. 1;85-96
doi: 10.1130/0016-7606(1986)97<85:FCITDC>2.0.CO;2

Email alerting services click www.gsapubs.org/cgi/alerts to receive free e-mail alerts when new articles cite this article

Subscribe click www.gsapubs.org/subscriptions/ to subscribe to Geological Society of America Bulletin

Permission request click http://www.geosociety.org/pubs/copyrt.htm#gsa to contact GSA

Copyright not claimed on content prepared wholly by U.S. government employees within scope of their employment. Individual scientists are hereby granted permission, without fees or further requests to GSA, to use a single figure, a single table, and/or a brief paragraph of text in subsequent works and to make unlimited copies of items in GSA's journals for noncommercial use in classrooms to further education and science. This file may not be posted to any Web site, but authors may post the abstracts only of their articles on their own or their organization's Web site providing the posting includes a reference to the article's full citation. GSA provides this and other forums for the presentation of diverse opinions and positions by scientists worldwide, regardless of their race, citizenship, gender, religion, or political viewpoint. Opinions presented in this publication do not reflect official positions of the Society.

Notes

Geological Society of America

THE
GEOLOGICAL
SOCIETY
OF AMERICA

"Fracture cleavage" in the Duluth Complex, northeastern Minnesota

M. E. FOSTER*
P. J. HUDLESTON
Department of Geology and Geophysics, University of Minnesota, Minneapolis, Minnesota 55455

ABSTRACT

An unusual cleavage is developed in troctolitic rocks of the late Precambrian Duluth Complex in northeastern Minnesota. It is localized in a zone ~250 m wide, and it is parallel to the axes of gentle folds in rhythmic banding. The rocks show no textural signs of ductile deformation, and the cleavage is defined by fractures that are fairly straight and evenly spaced, but on a fine scale, somewhat irregular, multiple, anastomosing, and locally crosscutting. Textures indicate a purely extensional mode of development. Cleavage refraction across interfaces of bands of different composition is common. Serpentine coats the fracture surfaces and suggests that fluids were present to produce this alteration during formation of cleavage.

The geological relationships and serpentinization suggest conditions of cleavage formation of 1.5–1.8 kb and 350 to 500 °C. The presence of extensional fractures indicates that cleavage formation involved crack propagation at high fluid pressures. The absence of shear fractures or ductile deformation features, and the nature of the cleavage refraction indicate that differential stresses were low.

Localization of the cleavage seems to be due to the presence of a structure, probably a dike, below the layered intrusive rocks. This could have served to localize stresses and reactive fluids at high pressure during a late stage of continental rifting.

INTRODUCTION

It is usually an easy matter to distinguish between the planar fabric elements that constitute cleavage and joints in rocks. Cleavage is penetrative or closely spaced and usually associated with ductile deformation, recrystallization, or solution transfer; whereas joints are more widely spaced and usually associated with brittle deformation. Most cleavages appear to be "compressive" features; that is, they form at high angles to the direction of maximum compression in rocks and are commonly subparallel to the XY plane of the strain ellipsoid (with axes $X \geqslant Y \geqslant Z$) (Siddans, 1972). Cleavages can less commonly be identified as "shear" features; they develop in planes of high shear strain (Hobbs and others, 1976, p. 245; Platt and Vissers, 1980). Joints are structures that develop as brittle fractures, usually forming in "tensile" mode perpendicular to the least compressive/most tensile principal stress (Segall and Pollard, 1983; Segall, 1984).

Despite these genetic and morphological differences, the distinction between the two structures is not always straightforward. A smooth transition from jointing to spaced cleavage may occur across a region of increasing intensity of deformation (Siddans, 1977), and planar discontinuities in one locality have been called joints where widely spaced and fracture (or spaced) cleavage where closely spaced (Price and Hancock, 1972). Such situations lead to problems in nomenclature and raise questions about the origin of the structures. Many different processes can contribute to cleavage development, and so it seems wise to use morphological terms to describe cleavage (Powell, 1979; Borradaile and others, 1982) and to avoid terms with genetic connotation such as "fracture cleavage" and "strain slip cleavage."

The purpose of this paper is to describe an unusual structure that has many of the characteristics of a spaced cleavage and yet is clearly the result of brittle fracturing; it is in fact a true "fracture cleavage." The structure is developed in intrusive mafic igneous rocks that otherwise show no sign of penetrative tectonic deformation. We have found no detailed description of this type of structure in the literature, although similar structures in similar rocks have been observed elsewhere (Jackson, 1961, Fig. 20; Wager and Brown, 1967, Fig. 34; Page, 1977; P. Weiblen, 1982, personal commun.). Irvine (1967, Figs. 4.4, 4.5) refers to such a structure in the Duke Island Complex, Alaska, as "closely spaced joints."

Our aim is to describe the morphological and spatial characteristics of this structure and its relation to other structural features and to come to some conclusions about its origin and the conditions of formation. We will take up the question of nomenclature in a later section but will anticipate the results of this discussion and refer to the structure as a "cleavage" throughout the paper, although it should be stressed that the structure satisfies the usual definitions of both cleavage and joints.

GEOLOGICAL SETTING

The Duluth Complex is a massive body of intrusive rock, forming part of the late Precambrian rift structure which cuts across the midcontinent region of North America (Weiblen and Morey, 1980). The rocks of the complex are mainly gabbroic and anorthositic in composition, and they intruded rocks of the North Shore Volcanic Group (Green, 1982), a series of mostly basaltic lava flows with intercalated sediments. All of these rocks are fairly well exposed in the northeast corner of Minnesota (Fig. 1). The area in which the cleavage occurs lies just west of Duluth at Bardon Peak (Figs. 1, 2). This is the southwest terminus of the presently exposed rocks of the complex, which may have been truncated by a fault to juxtapose the intrusive rocks of the complex with the somewhat younger Fond du Lac sedimentary rocks found just a few kilometres to the south (Schwartz, 1949; Foster, 1981).

The rocks at Bardon Peak are part of the layered troctolitic series of Taylor (1964) and Weiblen and Morey (1980). They form an intrusion, ~4,500 m thick, between the older and structurally overlying anorthositic rocks of the complex to the east and the underlying Ely's Peak basalts of the North Shore Volcanics to the west. They form a band about 10 km wide trending north-south. The area has been mapped at scales of 1:24,000 by Taylor (1964) and about 1:3,000 by Foster (1981).

---

*Present address: Exxon Company, U.S.A., P.O. Box 3116, Midland, Texas 79702.

Geological Society of America Bulletin, v. 97, p. 85–96, 15 figs., 1 table, January 1986.

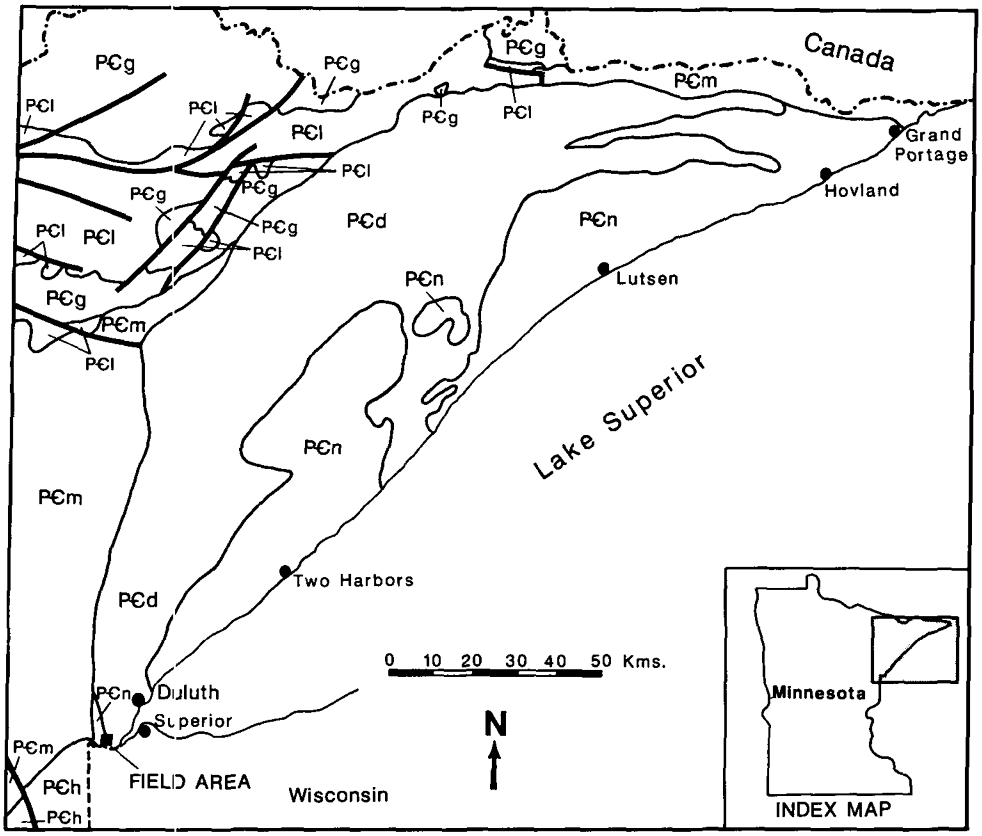


Figure 1. Generalized geological map of the Duluth Complex and surrounding rocks (after Sims, 1970), showing the location of the field area. PЄl, lower Precambrian metasedimentary and metavolcanic rocks; PЄm, middle Precambrian sedimentary rocks; PЄh, upper Precambrian Hinkley and Fond du lac Formations; PЄg, granitic rocks, mostly lower Precambrian; PЄd, Duluth complex; PЄn, North Shore Volcanic Group.

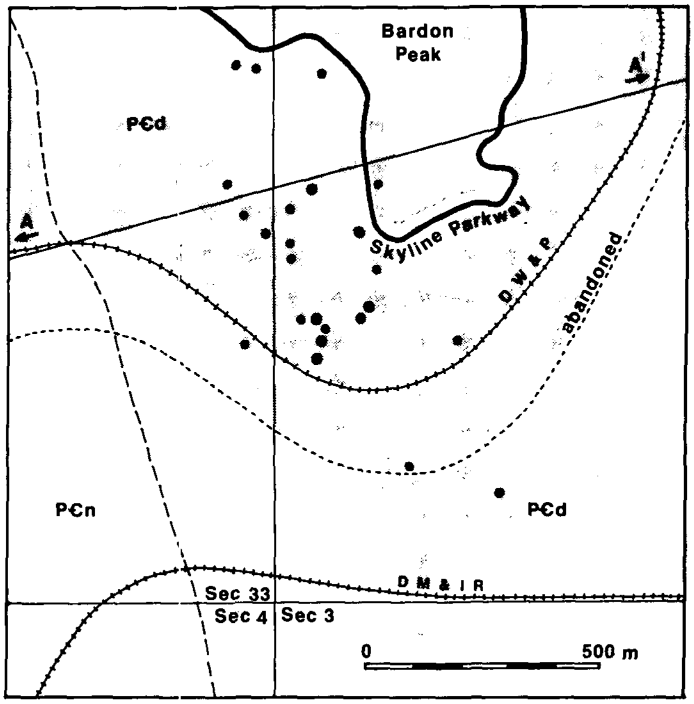


Figure 2. Map of the Bardon Peak field area showing the distribution of outcrops (shaded), with presence and intensity of cleavage development indicated by dots. Dot size is a relative measure of cleavage development (large dots indicate best developed cleavage). PЄn, North Shore Volcanic Group (Ely's Peak basalts); PЄd, Duluth complex.

A well-developed rhythmic-layering in these rocks is defined by varying proportions of olivine and plagioclase. In addition to modal layering, the gabbros exhibit a planar orientation of tabular grains, mainly plagioclase, within and parallel to the banding, that imparts a type of igneous lamination to the rocks (Grout, 1918).

The main constituents of the troctolites at Bardon Peak are plagioclase and olivine. Plagioclase (An$_{65}$) makes up 30%–80% of the rock by volume. The crystals are lath-shaped, 0.5–10 mm long, and generally oriented parallel to the modal layering. Olivine, with an average composition of Fa$_{39}$ (Taylor, 1964), usually makes up 50%–70% of the rock but sometimes much less. It occurs as fairly equant crystals, about 0.5 mm in diameter. Augite forms 2%–15% of the rock, occurring as large poikilitic crystals, or less frequently, interstitially. Magnetite and ilmenite together form $<5\%$ of the rock. The crystals are small and subhedral but occasionally large and poikilitic. Minor amounts of orthopyroxene, biotite, hematite, chlorite, and hornblende also occur, and serpentine is common in samples that show secondary alteration.

Regional Structure

The rocks of the mid-continental rift occupy a linear zone in the crust, as much as 200 km wide and extending from Kansas to Minnesota (King and Zeitz, 1971; Craddock and others, 1970; Chase and Gilmer, 1973; Green, 1983). The structure is complicated by the existence of separate centers of volcanism and intrusion, crosscutting faults, the development of flanking and overlying sedimentary basins, and in the vicinity of Duluth, by the intrusion of the Duluth Complex (Craddock, 1972; Davidson, 1982; Green, 1983). Surface outcrops and geophysical data suggest that the North Shore Volcanics and the base of the troctolitic intrusion dip 10°–20° southeast at Bardon Peak; similar dips in this direction continue along the north shore of Lake Superior to form the northwestern limb of the so-called Lake Superior Syncline (Craddock, 1972). The southeastern limb of this fold, made up of lava flows and sedimentary rocks, is exposed in northern Wisconsin and upper Michigan, and is seen to dip from 35° west to near vertically. The fold probably formed as a result of protracted regional subsidence (Craddock, 1972), although it may be in part related to a late-stage compressional event of the same trend (Green, 1983).

Following the Keweenanan rifting event, no significant tectonism has affected the mid-continent region, although modest fault reactivation has occurred, at least into the Paleozoic (Austin, 1972; Bagdadi, 1981).

Primary Structures

A brief description of the primary layered structures is necessary, because the layering to some extent controls the development of the later cleavage. Layering in intrusive igneous rocks is common, easily seen, and has received a lot of attention (Balk, 1937; Wager and Brown, 1967; Irvine, 1980). Two types of primary layering may be identified at Bardon Peak: rhythmic banding and igneous lamination (Taylor, 1964).

The rhythmic banding in the Duluth Complex was originally described by Grout (1918). It is defined by systematic variations in the proportions of olivine and plagioclase, enhanced in outcrop by weathering. The layers are usually a few centimetres thick, although faint layers as much as 15 m thick have been reported near Duluth (Taylor, 1964). The boundary between the layers can be either sharp or gradational. The upward transition from a feldspathic layer to an olivine-rich layer is typically quite abrupt, but that from an olivine-rich layer to a plagioclase-rich layer is gradational. Textural variations, too, are systematic, as the plagioclase is usually coarser grained in the feldspathic layers than in the olivine-rich layers. Coupled with the character of the interlayer contact, this imparts a type of reversed graded-bedding to the layers.

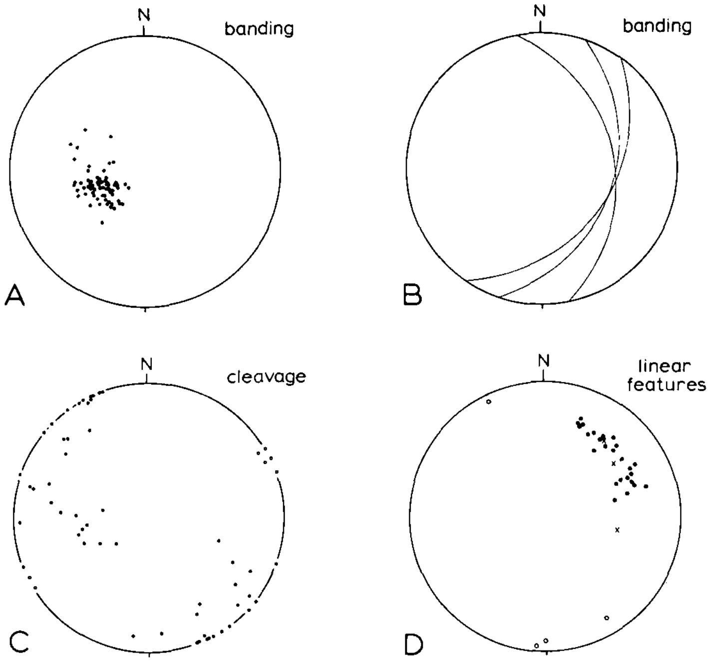


Figure 3. Structural data for rocks at Bardon Peak in equal-area, lower-hemisphere projection. A. Sixty-five poles to rhythmic banding. B. Rhythmic bands from one locality defining a minor warp axis. C. Forty-six poles to cleavage; dots for main set and circles for second set. D. Linear features. Banding/cleavage intersections shown by dots (main cleavage) and circles (secondary cleavage). Local fold or warp hinges shown by crosses.

Intimately related to the rhythmic banding, there is an igneous lamination (Wager and Deer, 1939) or fluxion structure (Grout, 1918; Taylor, 1964) that is defined by the preferred orientation of tabular plagioclase crystals parallel to the modal layering. This lamination is most readily seen in thin section and is best developed in the finer-grained olivine-rich layers.

The rhythmic banding and associated igneous lamination at Bardon Peak dip east-northeast at about 30°, with minor fluctuations (Fig. 3A), suggesting a broadly concordant intrusion (Taylor, 1964), although they do appear to dip more steeply than layering in the underlying and overlying North Shore Volcanics. There are no major folds or faults in the area, but a few local warps exist (Figs. 3B, 3D).

The presence of rhythmic banding and the parallel igneous lamination is strong evidence that the layered troctolites are bottom crystal cumulates that formed under conditions of convective magma flow and the action of density currents (Irvine, 1980).

THE CLEAVAGE

The structure in question at Bardon Peak was first noticed by Schwartz (1949, p. 88), who described:

> banded gabbro which has close-spaced jointing with the attitude at an angle to the bedding such as is usually characteristic of fracture cleavage. This suggests some movement of the gabbro when it had reached at least a partially solid state.

This is the sole reference to the structure in the Duluth Complex.

Distribution

At Bardon Peak, the cleavage is developed exclusively in the rocks of the troctolitic series, but it is not found in all the exposures of troctolite. The cleavage is largely resticted to outcrops situated on the southwestern slope of Bardon Peak (Fig. 2). These outcrops together define a north-northwest–south-southeast–trendng zone about 450 m wide, though the localities exhibiting the strongest cleavage are confined to a zone barely 200 m wide.

Morphology

In outcrop, the cleavage usually appears as a set of closely spaced parallel or subparallel traces, often branching and occasionally crossing at small angles. "Selected" cleavage planes at a greater spacing (5–20 cm) may be more pronounced and appear as typical joints (Figs. 4A, 4B). The cleavage is normally oriented at a high angle to the layering, extending across an individual layer, disappearing or changing form and orientation in the layers above and below (Fig. 4). The structure appears in some instances as a set of closely spaced joints, restricted to a single layer, and in others it appears penetrative to the naked eye.

Differential development of cleavage in adjacent layers is typical, as seems to be the case for the structure illustrated by Irvine (1967, Fig. 4.5). It is very clearly noted in thin section. This phenomenon must be controlled by compositional differences. A characteristic spacing between cleavage traces is observable in outcrop and ranges anywhere from a few millimetres (Fig. 4C) to several centimetres for the coarser structures (Figs. 4A, 4B). The length of cleavage trace usually depends on the thickness of the encompassing layer. From the effects of differential weathering, the cleavage is seen as narrow, rust-colored grooves in the outcrop face.

There are departures from the typical patterns of cleavage. The near vertical "normal" cleavage is paired locally with another, oriented approx-

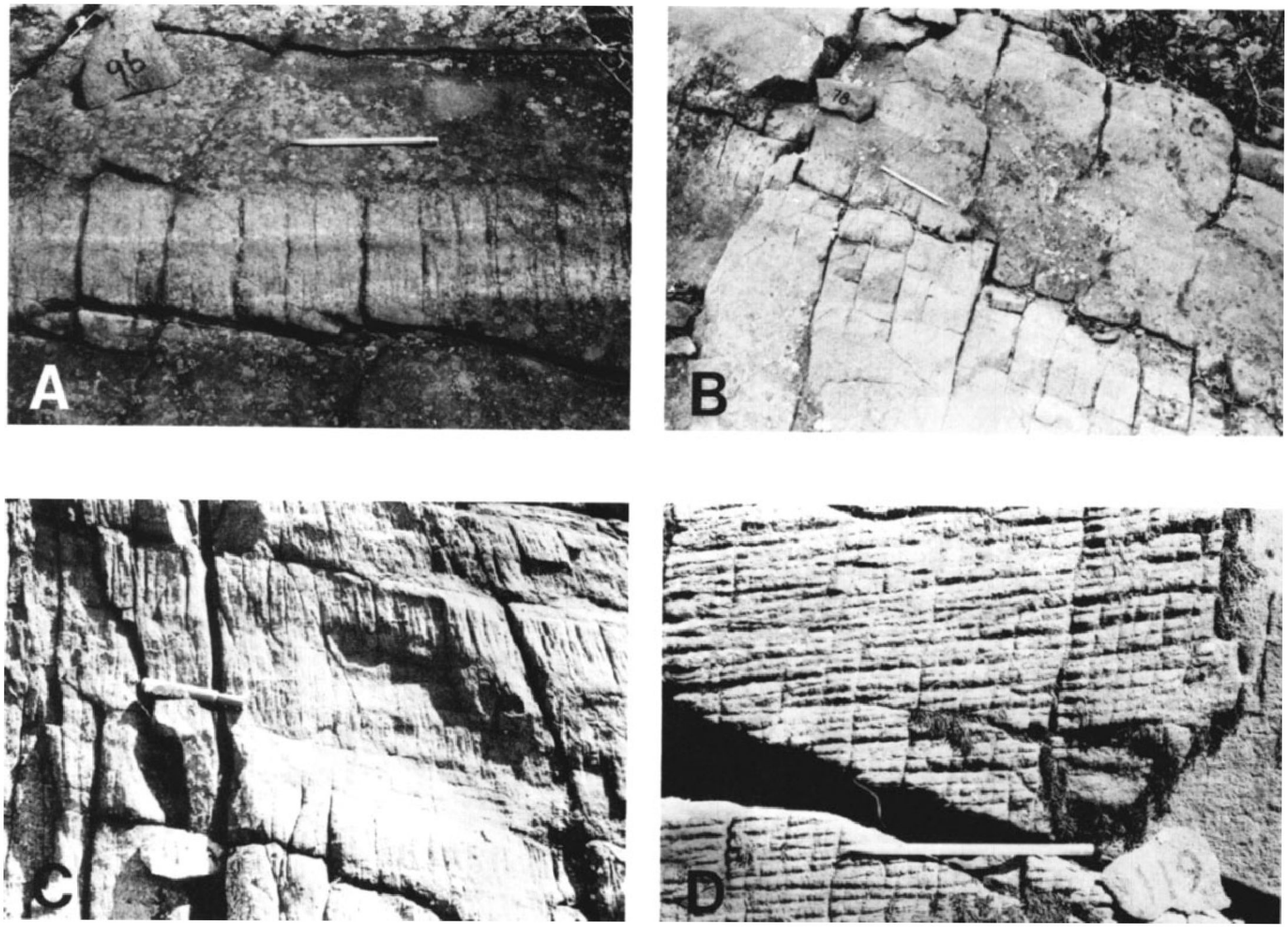


Figure 4. Appearance of cleavage and rhythmic banding in outcrop (with a pencil or marker parallel to the banding). A. Preferential development of cleavage in one band; B. cleavage defined in part by closely spaced joints; C. well-developed cleavage and banding; D. unusual checkered pattern resulting from two nearly orthogonal cleavages, viewed on surface parallel to layering.

imately perpendicular to it and the layering; as the spacing as well as the weathering characteristics are similar for the two sets of structures, a distinctive checkered pattern is produced (Fig. 4D), reminiscent of pencil cleavage. Cleavage refraction is quite common, and will be discussed later.

Several patterns of cleavage in thin section can be distinguished. In the simplest, the structure consists of discrete subparallel fractures, with major traces several centimetres in length and spaced from 1 mm to about a centimetre apart (Fig. 5A). Minor fractures are much more numerous and less continuous along their length; they are spaced somewhat irregularly down to a fraction of a millimetre, and generally less than 5 mm in length (Fig. 5A). Branching of these smaller cracks is very common, and an anastomosing structure is produced (Fig. 5B). At the smallest scale, the pattern is confused by an abundance of microcracks in all orientations. The larger fractures are fairly straight overall, cutting through mineral grains regardless of size, shape, orientation, or composition; they do, however, consist in detail of smaller straight-line segments arranged in a weak, irregular, zigzag pattern, with occasional offsets and branches. The largest fractures are only ~1/10 mm wide and are characteristically filled with alteration products described below. The smaller fractures have lesser degrees of alteration associated with them.

Rose diagrams to represent the degree of preferred orientation of the cleavage traces in each thin section are also shown in Figure 5. These diagrams were constructed by determining the orientation of all cleavage segments 1 mm long, thus eliminating the smallest fractures. The total number of segments measured per specimen is labeled "N" in the rose diagrams. The layers displaying the straight, evenly spaced cleavage in thin section (Fig. 5A) are those in outcrop display the most distinct cleavage (Fig. 4).

The cleavage is often more obvious in thin section than in outcrop. An example is shown in Figure 5B (from the outcrop shown in Fig. 4B), in which the major cleavage traces anastomose and are themselves composed of a series of smaller-scale anastomosing fractures. Where a zone of smaller-scale fractures crosses a plagioclase grain, it consists of narrower and more numerous fractures than where such a zone crosses an olivine grain. The fractures either cut through grains or weave around them. The igneous lamination is evident in the photograph and is aligned at a high angle to the cleavage. The two structures are apparently unrelated. Although the "braided" cleavage of Figure 5B does show a distinct preferred orientation, it is not as well defined as that shown in Figure 5A.

An even weaker preferred orientation of fracture traces is apparent in

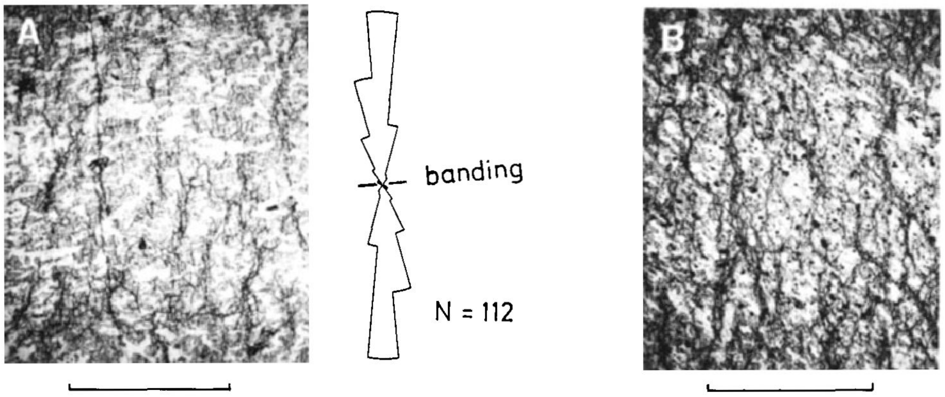


Figure 5. Cleavage in thin section and ordinary light, showing varying degrees of intensity and preferred orientation, with corresponding rose diagrams (for A and C only). Scale bars 1 cm long. A. In plagioclase-rich layer. B, C. In more olivine-rich layers, with greater intensity and less preferred orientation.

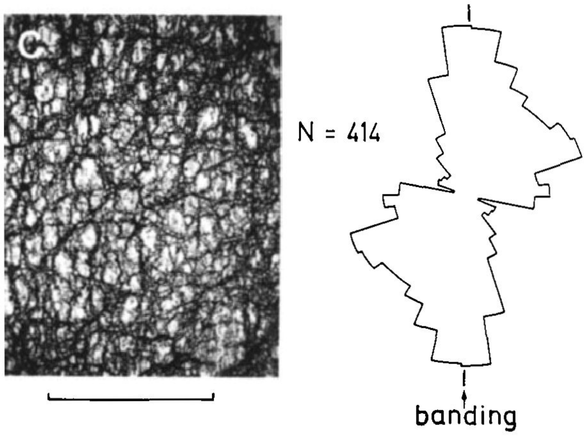


Figure 5C. The high-density, almost reticular pattern in this example represents the opposite extreme to the straight, parallel, and evenly spaced fractures shown in Figure 5A.

Observations in thin section and outcrop suggest that a relationship exists between (a) cleavage frequency and (b) intensity and composition of the host rock. Cleavage frequency was determined by drawing equally spaced parallel lines perpendicular to the mean cleavage trace on a photomicrograph and counting the number of intersections of cleavage traces along that line per unit length. At lower percentages of olivine ($<40\%$) there is no obvious relationship between cleavage frequency and amount of olivine, but for olivine $>40\%$ a correlation exists (Fig. 6). At very high concentrations of olivine, cleavage frequency is particularly high.

Refraction

The orientation of the cleavage is often distinctly different in layers of different modal proportions. Cleavage traces are oriented at a higher angle to the banding in the plagioclase-rich layers than in the olivine-rich layers. Good examples of cleavage refraction are common (Fig. 7), although unrefracted cleavage approximately at right angles to the layering also occurs, and refracted cleavage in the olivine-rich layers is often cryptic on the outcrop scale. Such refraction becomes obvious in thin section, however (as shown in Fig. 7D), and in the bimodal distribution of fracture traces shown in the rose diagram derived from it. In Figure 7D, an olivine-rich layer about 0.5 cm thick displays the characteristic close spaced anastomosing cleavage, in contrast to layers above and below that exhibit wider spaced, more discrete, and more linear cleavage traces. Similar relationships are shown in Figure 7B, with one unusual feature: the fractures in the olivine-rich layer are concentrated adjacent to the interlayer contact, forming an en echelon pattern.

Because banding is fairly constant in orientation, one might expect a systematic relationship between the angle of intersection of banding and cleavage and modal mineralogy. This does appear to be the case (Fig. 8), although the scatter of data is large. Where measurements were made in single samples, the angle is always greater in the plagioclase-rich layers than in the olivine-rich layers. An alternative way of representing the data on cleavage refraction is to plot the angle that cleavage makes with the layering on one side of a contact, against the angle on the other side (Fig. 9). Some constraints on the nature of refraction are imposed by requirements of stress and strain compatibility across the contact, as is discussed below.

Orientation and Relation to Rhythmic Layering and Folds

There is a wide scatter of poles to cleavage on a stereoplot, with a preference for northeast strikes and steep dips (Fig. 3C). There is also a tendency for cleavage to make high angles with bedding (Fig. 4). Most poles lie in a broad girdle that has its pole plunging at about 30° to 055°. Most of the scatter along the girdle is due to cleavage refraction; the lowest dips are associated with the olivine-rich layers. The poles in this girdle can be attributed to one "group" of cleavages. A second more or less orthogonal "group" strikes north-northwest and has near-vertical dips. This cleavage is local and usually found when the first group is weak or absent.

Banding/"main" cleavage intersections all cluster in the northeast quadrant of the stereonet, falling in the same region as the axes of minor warps in the rhythmic banding and the general fold axis defined by the distribution of poles to banding (Fig. 3D). The geometry is that typically seen in low-grade folded and cleaved rocks (Wilson, 1982, Fig. 6.6).

Banding/"other" cleavage intersections trend subhorizontally in a south to south-southeast direction and do not coincide with any fold axes.

Alteration

Many of the fractures defining the cleavage are coated or filled with green or brown serpentine, identified by X-ray powder diffraction as antigorite, chrysotile, and lizardite, although these were not all found in the same sample. The serpentine probably developed by hydrothermal alteration of olivine at a late stage in the magmatic process, and if this is so, the cleavage formation also occurred at this time.

The degree of alteration depends on the host mineralogy and fracture intensity. The finer fractures are only slightly affected by alteration, whereas the largest fractures commonly have up to 1 mm of serpentine filling; in

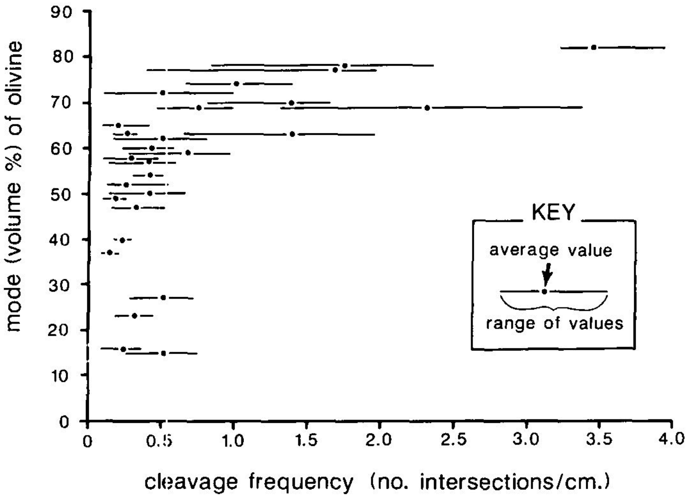


Figure 6. Relationship between cleavage density or frequency and amount of olivine in the rock. Each dot represents average values for one thin section (or part of a thin section if more than one layer is present), and the bars give the range of densities for a series of traverses across the slide.

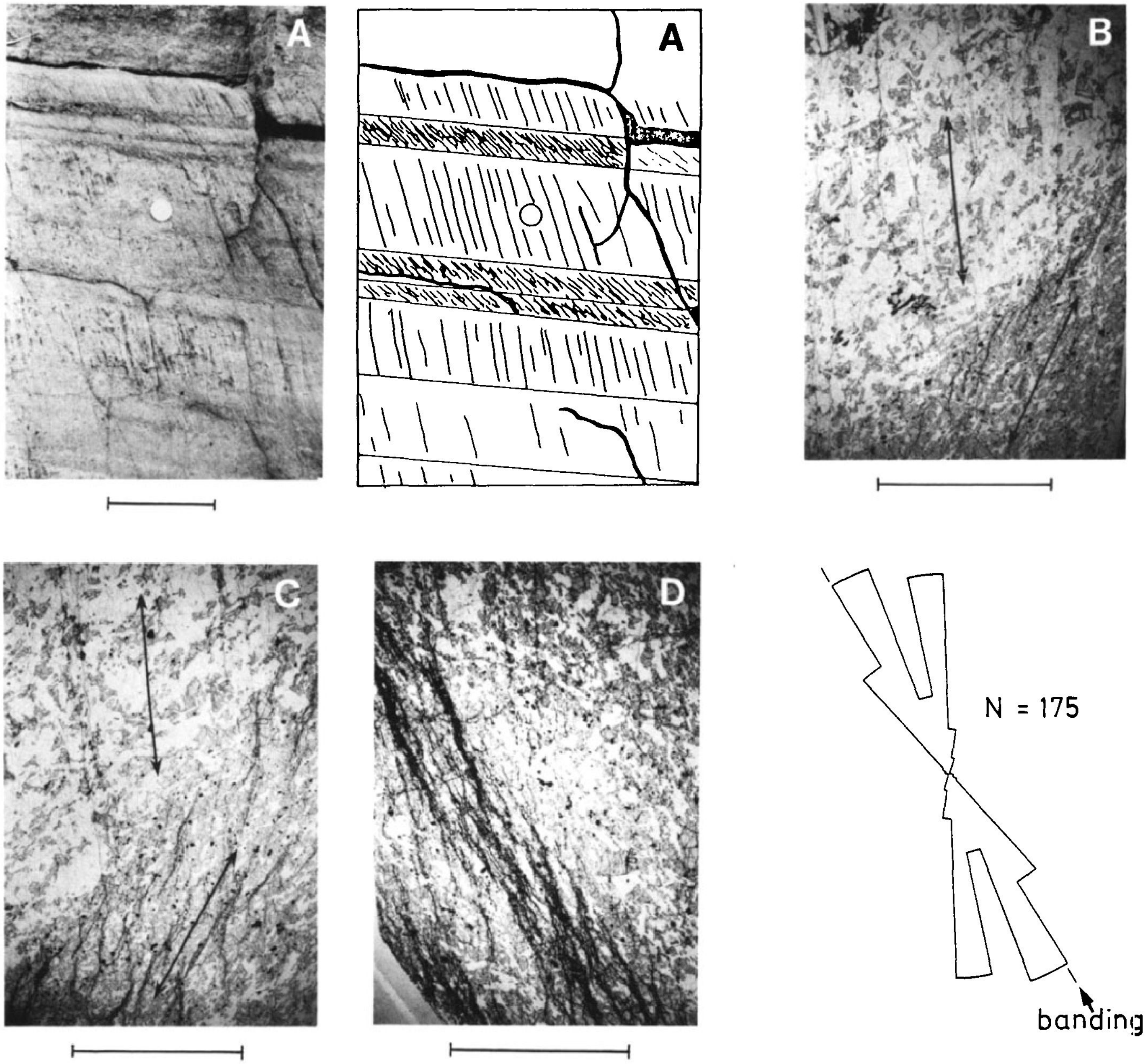


Figure 7. Cleavage refraction. A. In outcrop: cleavage at higher angle to banding in plagioclase-rich layer than in olivine-rich layer. Scale bar 10 cm. B, C, D. Similar relationship in thin section, ordinary light. Scale bars 1 cm. D. With corresponding bimodal rose diagram. Thin, olivine-rich band shows cleavage refracted and of greater intensity than in the adjacent rock.

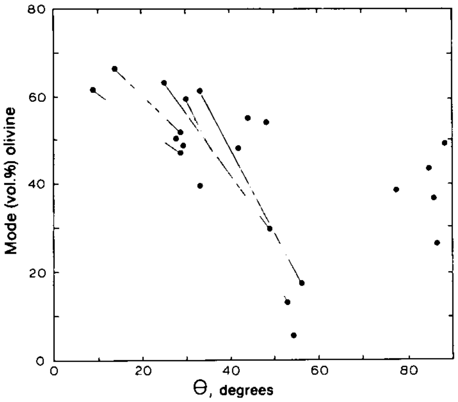
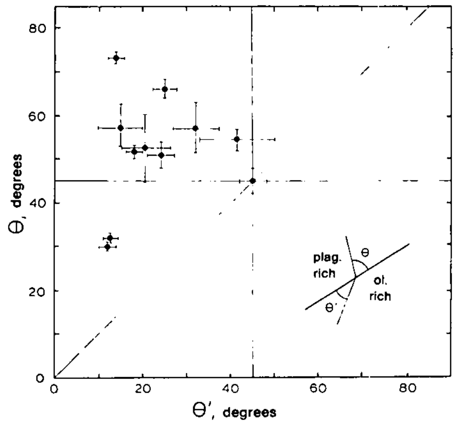


Figure 8. Relationship between the angle (θ) that the cleavage makes with the interlayer contact and amount of olivine in the rock. Connected points are from a single thin section on either side of an interface; otherwise, each point represents one thin section.

Figure 9. Cleavage refraction data. θ and θ′ are angles in plagioclase-rich layers and olivine-rich layers, respectively, that cleavage makes with the interface between the layers.



fact, it is the weathering of the alteration products that makes the structure so distinct in outcrop. Alteration is best developed when the cleavage crosses olivine grains or olivine-rich layers, but a coating occurs in plagioclase grains too, indicating transport of material along the fractures. Some olivine-rich layers which have the most intense and irregular type of fracturing are almost completely altered to serpentine. Magnetite is commonly found in association with serpentine in the samples showing greater degrees of alteration.

Apart from alteration along cleavage fractures, the rock is fresh and displays clear primary igneous textures.

ORIGIN OF CLEAVAGE

There is little doubt that the cleavage in the rocks at Bardon Peak resulted from brittle fracturing, in the sense that its formation was not preceded by any appreciable permanent deformation (Paterson, 1978). A number of pieces of textural evidence can be adduced to support this contention.

The rocks display exclusively primary igneous textures, with no sign of recrystallization or new grain growth parallel to the cleavage, as might be expected under metamorphic conditions. In addition, on a finer scale, there is no evidence of any intracrystalline deformation such as mechanical twinning, undulatory extinction, kinking, or deformation bands. There is also no evidence for cataclasis, either by fracturing along grain boundaries or along cleavage surfaces. The lack of such deformation-induced microstructures, plus the morphology of the cleavage fractures themselves, clearly indicate a brittle mode of deformation. The alteration found along the fracture planes is contemporaneous with the fracturing or postdates it.

Under very low confining pressures in experiments (<0.1 kb), most rocks tend to behave brittlely, and the development of extension fractures is favored. In isotropic rocks, extension fractures are usually oriented perpendicular to the least principal stress direction (compression positive), and the separation is normal to the fracture surface. In tensile tests, specimens tend to rupture along single planes. In compressional experiments, extension fractures are manifest as "axial splitting," and occasionally a series of close-spaced, parallel fractures is produced (Gramberg, 1965, Fig. 8). More or less periodic fractures, of the kind seen in the Bardon Peak rocks, have not been produced experimentally.

At slightly higher pressures, most rocks are still brittle or semi-brittle, but the dominant mode of failure in test specimens is by shear fracture. The throughgoing fractures are formed by a complex process involving the coalescence of small-scale extensional fractures, and perhaps shear fractures and grain boundary cracks, and crushing of grains (Dunn and others, 1973; Wong, 1982). They are generally oriented at an angle of less than 45° to the maximum compressive principal stress, and displacement is parallel to the fracture surface. Under low confining pressure, only one shear fracture usually forms; at higher pressures, a conjugate pair often appears; and sometimes failure is distributed through a closely spaced zone of fractures rather than a single discrete break (Wawersik and Brace, 1971, Pl. 4B).

Apart from brittle fracturing, there is one other process that may cause a cleavage to develop in rocks in which the mineral grains of the microlithons remain undisturbed. This is pressure solution, a process now recognized as playing a major role in the development of cleavage in low-grade metamorphic rocks. Morphologically, structures resulting from pressure solution may resemble brittle fractures very closely, and in fact much of what has been termed "fracture cleavage" and was thought by many to be the result of shear failure (Billings, 1972, p. 398) is now recognized as being the result of pressure solution (Nickelsen, 1972; Groshong, 1976). The anastomosing nature of many cleavages produced by pressure solution (Nickelsen, 1972; Means, 1982) is very similar to the cleavage pattern commonly seen in the olivine-rich layers at Bardon Peak. Evidence for pressure solution includes (1) fabric elements truncated by removal, not offset by shear, and (2) fabric elements truncated against narrow zones of the relatively insoluble components of the rock (Groshong, 1976). Neither is evident at Bardon Peak.

Figure 10 shows the expected sense of offset of twinned plagioclase laths along cleavages formed under conditions of contraction (pressure solution), shear, and extension. For grains oriented perpendicular to the fractures, the offset is negligible (an argument against shear), but for those oriented at low angles to the cleavage, some degree of offset is normally seen. A single offset crystal (Fig. 11A) or group of similarly aligned crystals can be interpreted in several ways. Because of igneous lamination, neighboring plagioclase laths inclined in opposite senses to the cleavage trace are uncommon, but where found, the offset always indicates an extensional origin (Figs. 10, 11B). This is the strongest textural evidence for an extensional mode of failure.

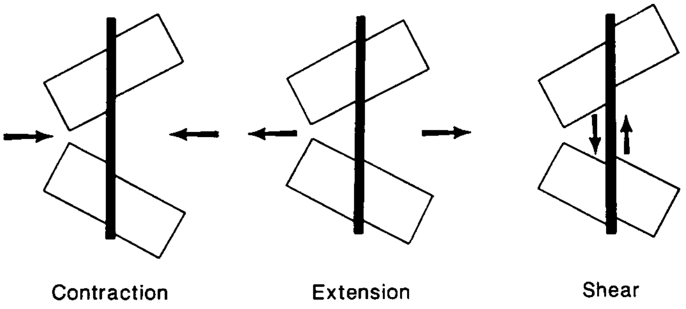


Figure 10. Sketches to show the expected sense of offset along the traces of cleavage formed by contraction (pressure solution), extensional failure, and shear.

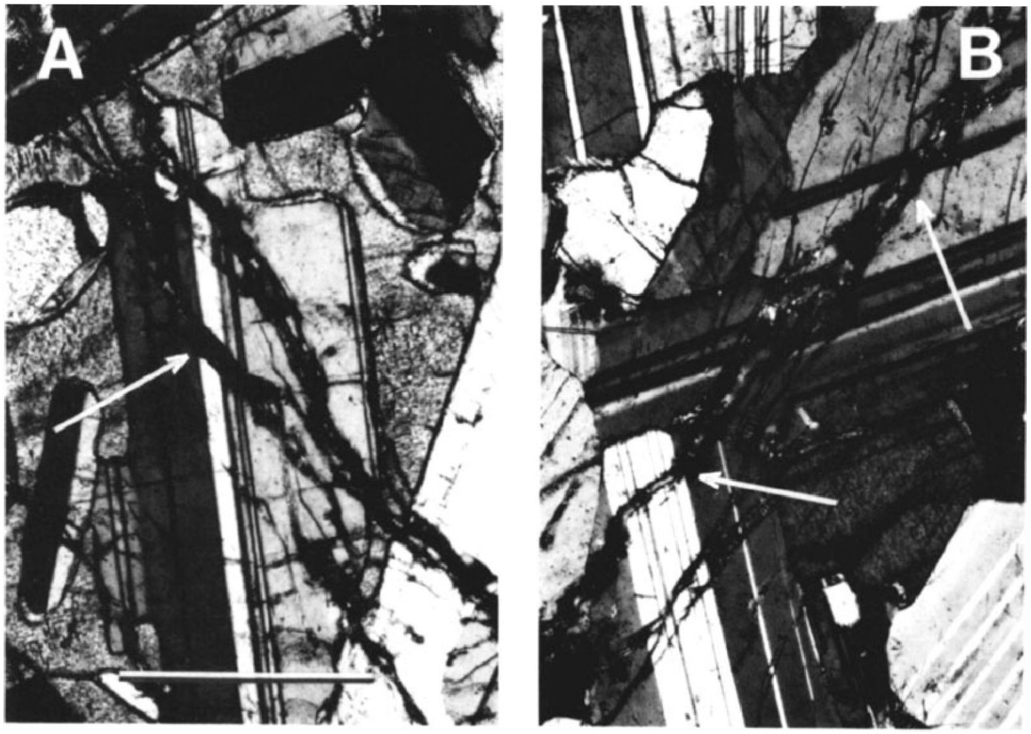


Figure 11. Photomicrographs under crossed nicols to show offset along cleavage of twins in plagioclase. A. Ambiguous offset, where displacement can be interpreted as due to either extension or left-lateral shear. Scale bar 1 mm. B. Unambiguous offset (as in Fig. 7), indicating an extensional displacement associated with cleavage formation. Scale as in A.

Temperature and Pressure Conditions of Cleavage Formation

Minimum temperature and pressure can be estimated approximately if it is assumed that during cleavage formation the rocks were buried beneath 1–2 km of North Shore Volcanics (J. C. Green, 1985, personal commun.) plus another 4 km, at least, of overlying gabbro of the anorthositic series. With a pressure gradient of 0.29 kb/km, the pressure would be 1.45–1.75 kb. The serpentine-forming reaction of forsterite + water has been determined experimentally to occur between 430°–470 °C at pressures of between 1 to 3 kb (Scarfe and Wyllie, 1967), and the reaction of forsterite + water to give chrysotile + brucite has been found to occur between 385°–400 °C in the same range of pressures (Helgeson and others, 1978). Similar reactions of iron-bearing olivine and water will occur at somewhat lower temperatures. It thus seems that the temperature at the time of fracturing may have been in the range of 350°–500 °C.

Under high confining pressures and in the absence of fluids in rocks, extensional fractures will not occur. The presence of serpentine suggests that fluids occupied the fractures soon after they developed; in fact, it seems likely that such fluids were present at the time of fracture formation in the form of late-stage magmatic volatiles. Fluids would raise crack pressure in the rock and facilitate brittle failure, perhaps by hydraulic fracturing if the fluid pressure was high enough to make the effective stress equal to the tensile strength of the rock (Secor, 1969; Bredehoeft and others, 1976; Pollard and others, 1983). Hydraulic fracturing has been invoked to account for fracture cleavage (Price and Hancock, 1972) and jointing in sedimentary rocks (Secor, 1969). It also seems highly likely that, in the presence of hot, reactive fluids, processes of stress corrosion aided propagation of the fractures, and that propagation rates were subcritical (Atkinson, 1982). Branching of the kind normally associated with dynamic crack propagation (Lawn and Wilshaw, 1975) is not observed.

Thermal stresses, induced during cooling from the crystallization temperature (~1000 °C) to the temperature of cleavage formation (≤500 °C), are also likely to have played a role in fracture development. The coefficient of thermal expansion of olivine is about twice that of plagioclase, ignoring effects of crystal anisotropy, and a cooling olivine-rich layer will attempt to contract along its length more than a plagioclase-rich layer. This would give rise to an effective component of layer parallel tension in the olivine layers and compression in the plagioclase layers. This may account for the greater intensity and frequency of fractures in the olivine-rich layers than in the plagioclase-rich layers (Fig. 6). The extensional fractures themselves would have provided the increase in volume necessitated by the alteration of olivine to serpentine.

Fine fractures radiating away from the contacts of individual grains of olivine and plagioclase probably are due to local thermal stresses induced during cooling.

Significance of Cleavage Refraction

If the cleavage is accepted, as argued above, as being due to extensional fracturing, it should develop perpendicular to the least compressive stress, σ_3. Cleavage refraction then represents stress refraction at the time of cleavage formation. It can be readily shown that in continuous layers of different material properties, stress refraction is required to maintain stress and strain continuity at the contacts between layers (Treagus, 1973, 1981). The situation is illustrated in Figure 12. Primed quantities refer to one side of the interface, unprimed to the other. Normal stress $\sigma_y = \sigma'_y$, and shear stress $\tau_{xy} = \tau'_{xy}$ at the interface must be equal. Assuming homogeneous stresses within each layer and making use of standard equations for stress (Jaeger, 1969, equations 3.13, 3.14) lead to a relationship between the orientation of σ_1 and σ'_1 in adjacent layers:

$$\frac{\tan 2\theta}{\tan 2\theta'} = \frac{(\sigma'_x - \sigma'_y)}{(\sigma_x - \sigma_y)} \qquad (1)$$

When constitutive relations are introduced, equation 1 can be expressed in terms of material properties. We will consider only the case of plane strain here, but the basic conclusions still hold true for a general strain. For

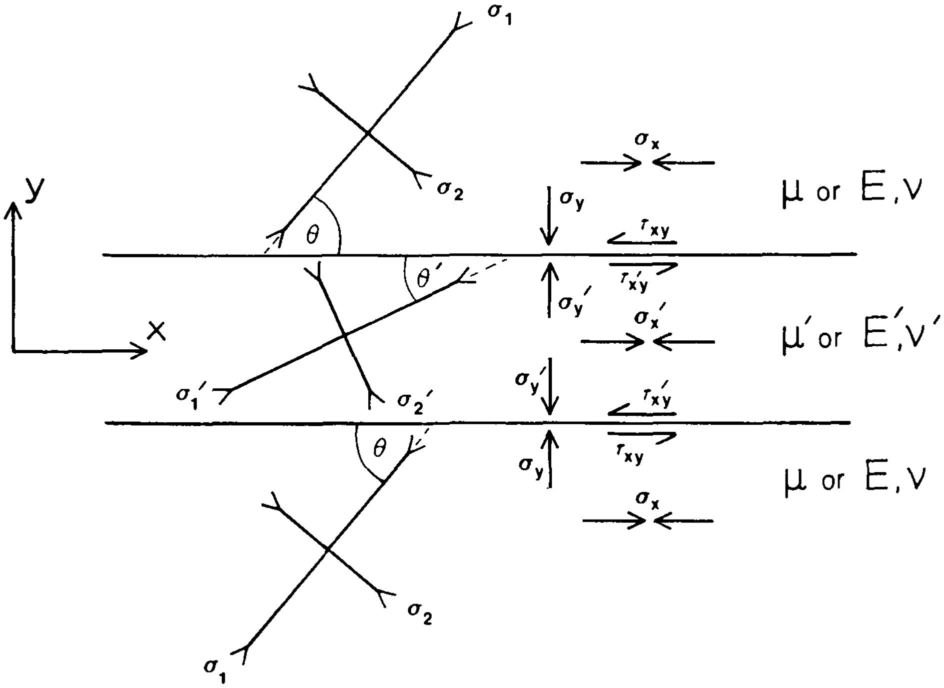


Figure 12. Refraction of principal stresses across an infinitely long layer with physical properties different from those of the embedding medium. Homogeneous isotropic materials and plane strain.

Newtonian viscous materials, the layer-parallel strain rate $\dot{e}_x = \dot{e}'_x$, and it can be shown (Treagus, 1973) that:

$$\frac{\tan 2\theta}{\tan 2\theta'} = \frac{\mu'}{\mu} \qquad (2)$$

where μ and μ' are viscosities. There is no refraction at $\theta = 0$ and 90°, and also at $\theta = 45°$ (it will be seen that $\dot{e}_x = \dot{e}'_x = 0$ at this position and that the stress state is identical in both layers). Thus θ and θ' must always be both $\geqslant 45°$ or both $\leqslant 45°$. If we look at the data for the Bardon Peak cleavage (Fig. 9), we find that in all but two cases, this condition is *not* satisfied. For instance, typical values of θ and θ' are 55° and 20°.

Viscous rheology is clearly inappropriate for the present case, and it is instructive to find the equivalent to equation 2 for elastic materials. The expression is:

$$\frac{\tan 2\theta}{\tan 2\theta'} = \frac{G'(e'_x - e'_y)}{G(e_x - e_y)} = \frac{E'(1+\nu)\,(e'_x - e'_y)}{E(1+\nu')\,(e_x - e_y)} \qquad (3)$$

where G is the shear modulus, E Young's modulus, and ν Poisson's ratio. The expression cannot be given purely in terms of elastic constants because of compressibility. For incompressible elastic materials, $\nu = 0.5$, $e_x = -e_y$ and the expression is just as for Newtonian materials (equation 2), so that θ and θ' must both be $\leqslant 45°$ or $\geqslant 45°$.

With $\nu < 0.5$ the situation is not so simple, and we examine it in the following way. We make use of the standard relationships (Jaeger, 1969, equations 14.8, 3.11, 3.19):

$$Ee_x = (1 - \nu^2)\,\sigma_x - \nu(1+\nu)\,\sigma_y \qquad (4)$$

$$\tan 2\theta = 2\,\tau_{xy}/(\sigma_x - \sigma_y) \qquad (5)$$

$$(\sigma_1 - \sigma_2)^2 = 4\,\tau_{xy}{}^2 + (\sigma_x - \sigma_y)^2 \qquad (6)$$

with similar equations for primed quantities, and note again that $\sigma_y = \sigma'_y$, $\tau_{xy} = \tau'_{xy}$, and $e_x = e'_x$. It is useful to find the conditions for no refraction when $\nu = \nu' \neq 0.5$, and when θ and θ' are not 0° or 90°. To make the problem dimensionless, all stress components are expressed in terms of stress difference over mean stress (in the plane), $s = 2(\sigma_1 - \sigma_2)/(\sigma_x + \sigma_y)$. Note that, for no refraction, $\sigma_x = \sigma'_x$ (from equation 5), so that $\sigma_1 = \sigma'_1$, $\sigma_2 = \sigma'_2$ and $s = s'$. Equations 4 to 6 were solved for different values of ν and s, and the results are shown in Figure 13. For any value of ν, two angles of no refraction are possible besides $\theta = 0°$ and 90°, corresponding to the two principal stresses. For σ_1 alone, the angle of no refraction is $\geqslant 45°$. We note that the smaller the value of ν or the smaller the stress difference, the more the angle of no refraction departs from 45°. We also note that the angle of no refraction is independent of E/E'. If, for example, $\nu = 0.2$ and $s = 2$, then (for σ_1) $\theta = \theta' = 63°$. This means that for stresses of this magnitude, θ and θ' must both be $<63°$ or $>63°$ for stress refraction to occur.

Are there stress conditions which allow $\theta = 55°$ and $\theta' = 20°$, values typical for Bardon Peak? Values of $\nu = \nu' = 0.3$ are reasonable for gabbroic rocks, as derived for example from shear and compressional wave velocities in Carmichael (1982). E and E' will almost certainly not differ by more than an order of magnitude and probably by much less (Carmichael,

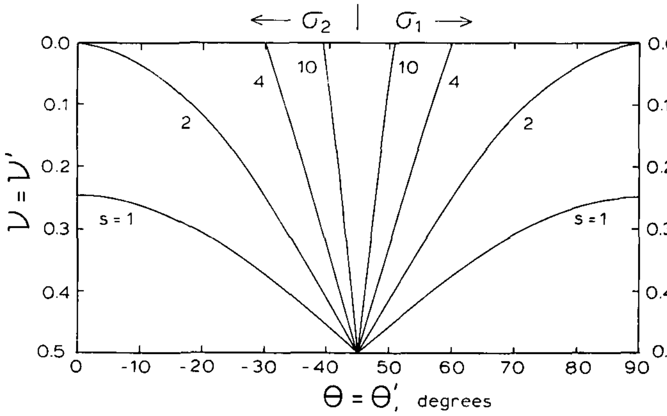


Figure 13. Conditions for no refraction of stresses (for $\theta = \theta' \neq 0°$ or 90°) across plane interface between elastic media where $\nu = \nu'$ at various relative stress levels, $s = 2\,(\sigma_1 - \sigma_2)/)\sigma_x + \sigma_y)$. To the right of the line $\theta = \theta' = 45°$ the angle refers to σ_1, to the left σ_2.

TABLE 1. RELATIVE STRESS DIFFERENCES

ν	ν'	E/E′	S	S′
0.3	0.3	10	0.41	0.87
		0.1	1.40	1.01
		1.2	0.07	0.10
		0.83	0.08	0.11
0.35	0.3	10	0.42	0.90
		1	0.09	0.07
		0.1	1.03	0.86

Note: $s = 2(\sigma_1 - \sigma_2)/(\sigma_x + \sigma_y)$ and $s' = (\sigma'_1 - \sigma_2)/(\sigma'_x + \sigma'_y)$, for refraction given by Bardon Peak data, $\theta = 55°$, $\theta' = 20°$, for various combinations of E/E′ and ν and ν'. The procedure used to find these was, for given ν, ν' and E/E′, first to arbitrarily set a value for $\sigma_y = \sigma'_y$ and hence use equation 5 to find σ_x and σ'_x in terms of τ_{xy}. Expressions for σ_x and σ'_x were then substituted into equation 4 (two equations) which can be solved simultaneously for τ_{xy}. Equation 6 can then be used to find $(\sigma_1 - \sigma_2)$, and hence s.

1982). If equations 4–6 are solved for E/E′ = 10, 0.1, 1.2, and 0.83, one finds solutions exist in all cases (Table 1). Either layer may be the 'competent' member of the pair, and if E/E′ is close to unity, very small stress differences may give rise to the observed refractions.

It seems quite likely that $\nu \neq \nu'$, and data in Carmichael (1982) suggest that the values of Poisson's ratio in gabbroic rocks may differ by up to about 0.1. A possible pair of values are 0.35 and 0.3. The case for no refraction at various E/E′ and s is given in Figure 14. At low values of relative stress difference, the angles of no refraction depart significantly from 45°. For the values of ν selected, there is a singularity at E/E′ = 1.04, causing great divergence from 45° of the angle $\theta = \theta'$ at any stress level for E/E′ in the vicinity of 1.04.

For $\theta = 55°$, $\theta' = 20°$, $\nu = 0.35$, and $\nu' = 0.30$ (a possible set of data for the Bardon Peak rocks), resulting relative stress differences for various E/E′ are listed in Table 1. Note again that either layer may be "competent" and that very low stress differences can account for the observed refraction if E/E′ = 1.

The above considerations indicate that very low relative stress differences can account for the observed refraction of the cleavage if E/E′ is close to unity (varying ν and ν' within reasonable limits in the above calculations does not alter this conclusion). Larger relative stress differences are possible if E/E′ is an order of magnitude greater or less than 1, but because the pressure under which fracturing occurred was high, it seems likely that shear failure under high pore pressure or ductile deformation would have occurred if stress difference was equal to mean stress (s = 1). The scatter of data (Fig. 9) can be accounted for by variation in composition (and hence elastic constants) and by variation in local stress orientation with respect to the layering. The existence of pairs of refraction angles as extreme as $\theta = 70°$, $\theta' = 15°$ is further suggestive of low relative stress differences.

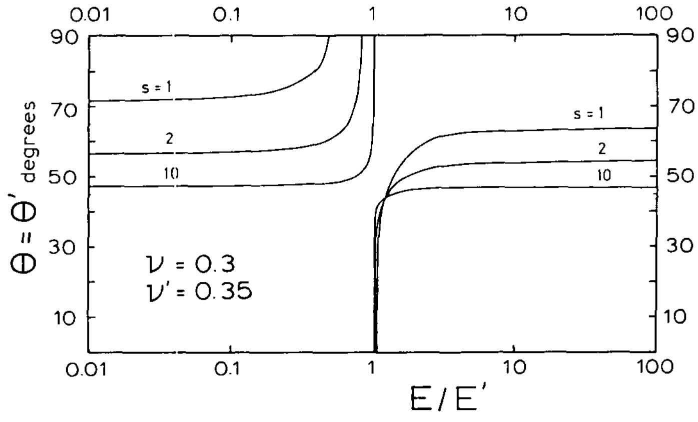


Figure 14. Conditions for no refraction of stresses (for $\theta = \theta' \neq 0°$ or 90°, for σ_1 only) across plane interface between elastic media for which $\nu = 0.3$ and $\nu' = 0.35$, for various values of E/E′ and relative stress difference, s.

CLEAVAGE CLASSIFICATION

The structure at Bardon Peak has an origin more like that of joints than cleavage, yet an appearance more like cleavage than joints. It also clearly satisfies the usual definitions of both structures (Borradaile and others, 1982, and Dennis and Weber, 1979, for definitions and discussions of cleavage, and Dennis, 1967, for a discussion of joint terminology). Features favoring the use of the term "cleavage" are the following: Cohesion is maintained across the planar discontinuities at Bardon Peak, essential for cleavage and usually not the case for joints, although sealed joints are common (Segall and Pollard, 1983). None of the common surface features of joints, such as plumose structure (Price, 1966) is found at Bardon Peak, although they may simply be sealed in and hidden. Very rarely are the fractures as planar and continuous as most ordinary joints (Price, 1966, Pl. 3; Segall and Pollard, 1983, Figs. 2, 3). The anastomosing pattern of the structure is very similar to that of pressure solution cleavage, with the serpentine alteration along fractures forming zones very similar to those produced by insoluble residue from pressure solution. In both structures, the rock between the zones is undeformed. The structure is continuous on the scale of the outcrop (Fig. 4) and is not confined to zones, as is commonly the case for closely spaced joints, although it is controlled by layering.

There is not a great deal written concerning the "gray" area in nomenclature between closely spaced joints and cleavage (Powell, 1979; Dennis, 1967; Hancock, 1982a, 1982b), but the notion that an origin by fracture precludes the use of the term cleavage seems to be implicit in some modern treatments of cleavage. Given the history of the terms and the difficulty of establishing the origin of many cleavage-like structures, this seems unwise. We prefer to follow the increasingly popular practice of basing the terminology of cleavage-like structures on morphology alone (Powell, 1979; Borradaile and others, 1982). Using Powell's scheme, the Bardon Peak structure would be called a "disjunctive spaced cleavage." In addition, the cleavage domains vary from *smooth* to *rough,* and *planar* to *anastomosing* (Borradaile and others, 1982).

With the origin established, qualifying terms can be proposed on the basis of genesis. The Bardon Peak structure is in fact a *fracture cleavage* in the broad sense of the term (Turner and Weiss, 1963) but not in the narrow sense, in which the term is restricted to shear fractures (Billings, 1972, p. 398). The term "fracture cleavage" is not in common usage today and has in fact come into disrepute because many of the structures to which the term has been applied are now recognized as being the result of pressure solution and not fracturing. It is, however, clearly an appropriate term for the structure described in this paper.

SPECULATIONS ON THE TECTONIC SIGNIFICANCE OF THE CLEAVAGE

Two factors must be considered in evaluating the possible tectonic significance of the cleavage: (1) the nature of the stress responsible for cleavage formation and (2) the reason for its localized development.

There are two possible sources for the causative stresses. The first and most likely is the intracontinental rifting event itself. A slight extensional movement during the final stages of cooling could generate effective tensional stresses in rocks with high fluid pressures. The second is a somewhat later tectonic event. Craddock (1972) suggested that during or following the emplacement of the Keweenawan intrusions, the entire Lake Superior

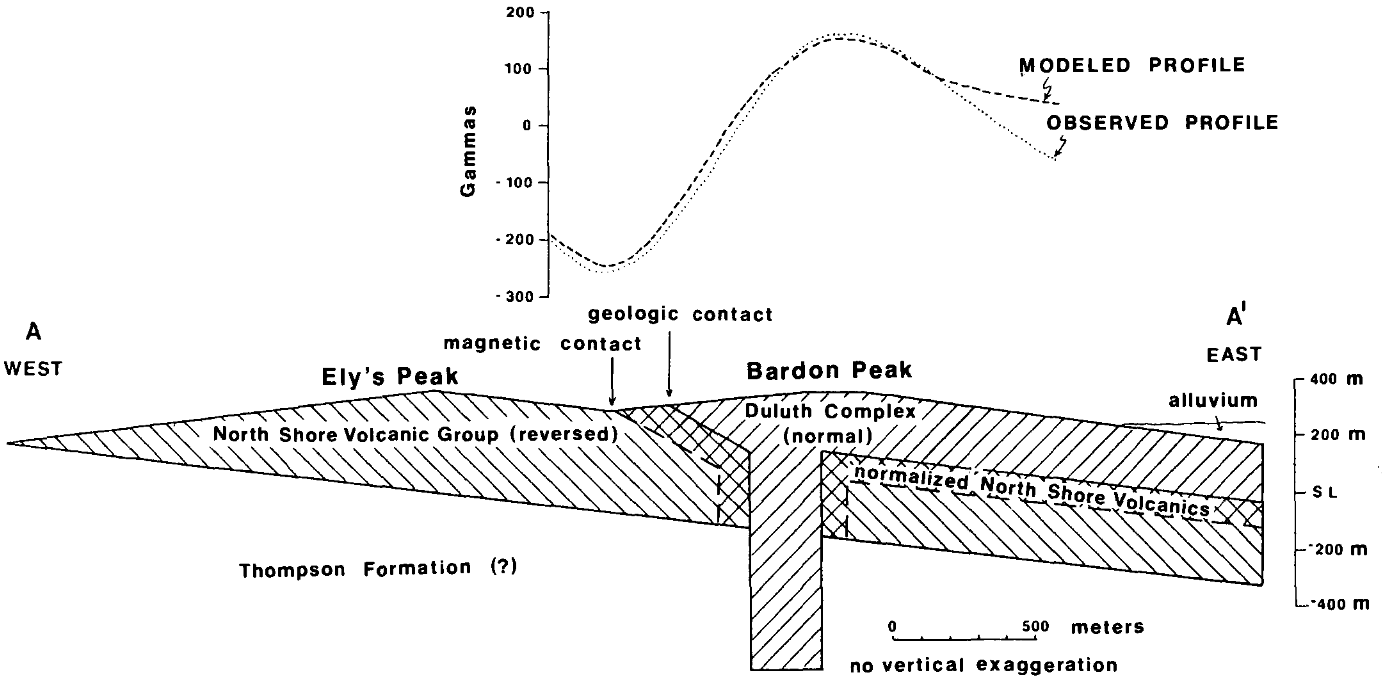


Figure 15. Geologic cross section across the Bardon Peak field area (line of section shown in part on Fig. 2) resulting from a simple two-dimensional model to match high-resolution aeromagnetic data. Parameters used for modeling are: inclination and declination of the Earth's present magnetic field, 76° and 3°, respectively; paleomagnetic inclination and declination for the North Shore Volcanic group (reversed), -55° and 130° (Palmer, 1970) and for the Duluth Complex (normal), 34° and 285° (Beck, 1970); bearing of the profile is 75°.

region was subjected to compressional stresses oriented in a northwest-southeast direction. Because of its generally northeastern strike, the cleavage could not have formed as a direct result of such stresses, but it could conceivably have developed during a later phase of extension or relaxation (Price and Hancock, 1972). Within the field area, the small, northeast-plunging warps in the rhythmic banding (Fig. 3) could be construed as evidence for compression, although direct evidence for extension-relaxation does not exist.

There appear to be no great lateral variations in composition of the troctolites that might have localized cleavage development, although J. C. Green (1985, personal commun.) has stated that the cleavage-bearing rocks are richer in olivine than are the neighboring troctolites. The trend of the zone of localization, generally north-northwest–south-southeast, follows the structural grain of the region, suggesting that some structure at depth may have localized stresses in the overlying troctolites. This idea is supported by aeromagnetic data (Fig. 16; Foster, 1981), which suggests a "root" or "feeder dike" to the Duluth Complex, present below the zone of cleavage localization. The "dike" invoked for the two-dimensional modeling trends at right angles to the line of section and close to the trend of cleavage localization. It is about 250 m wide, and the best developed cleavage is restricted to a zone about 200 m wide.

If this correlation is valid, the implications for cleavage development are clear. First, a gabbroic dike, presumably connected at depth to some magmatic source, could be an inherent zone of weakness compared to the adjacent rocks. The cleavage fractures in the overlying rocks could have resulted from localized stresses due to some differential movement along the dike, whether related to rifting (Pollard and others, 1983, Fig. 21), a pulse in magmatic activity, or a northwest-southeast directed compression (or its related relaxation). The fracturing could not be caused just by extension perpendicular to the dike walls, however, because the individual fractures trend obliquely to the zone containing them. There would need to be some dextral component of shear along the dike to cause this obliquity. The localization of cleavage could also be explained without localization of stress if a zone of high fluid pressure was generated above the dike, which could act as a conduit for the upward migration of hot, late-stage fluids. Extensional fracturing at high pore pressure would thus be restricted, and serpentinization would follow fracturing.

CONCLUSIONS

The following outline of events is proposed to account for the cleavage-like structure in the troctolitic rocks at Bardon Peak.

After crystallization of the intrusion but while temperatures were still high (~350°–500 °C), the rocks were subjected to effective tensional stresses, most likely the result of renewed (or continued) movement along the midcontinental rift, and localized at Bardon Peak in a north-northwest–south-southeast–trending zone of weakness—probably an active or inactive dike. The presence of interstitial fluids at high pressure lead to the propagation of fluid-filled cracks which developed into a set of closely spaced, subparallel extension fractures. The pattern of fracture development was controlled locally by the modal layering of the troctolites. Low stress differences served to inhibit the formation of shear fractures or ductile deformation structures, and they also promoted significant stress (and hence, fracture refraction) between layers of slightly different elastic properties. The fluids which aided crack propagation were also reponsible for the alteration that produced and distributed serpentine along the fractures.

ACKNOWLEDGMENTS

This study was supported by a grant from the University of Minnesota Graduate School. A grant to Foster from the Minnesota Geological

Survey helped to defray the cost of field expenses. We are grateful to these sources of financial support. Advice on the geophysical modeling was provided by Clem Chase and Val Chandler and on the Duluth Complex by Paul Weiblen. The manuscript has been significantly improved by the critical comments of David Pollard, John Green, and an anonymous reviewer.

REFERENCES CITED

Atkinson, B. K., 1982, Subcritical crack propagation in rocks: Theory, experimental results and applications: Journal of Structural Geology, v. 4, p. 41–56.

Austin, G. S., 1972, Paleozoic lithostratigraphy of southeastern Minnesota, *in* Sims, P. K., and Morey, G. B., eds., Geology of Minnesota: A centennial volume: St. Paul, Minnesota, Minnesota Geological Survey, p. 459–473.

Bagdadi, K., 1981, Tectonic studies of the midcontinent gravity high in east central Minnesota and western Wisconsin [Ph.D. thesis]: Minneapolis, Minnesota, University of Minnesota, 182 p.

Balk, R., 1937, Structural behavior of igneous rocks: Geological Society of America Memoir 5, 177 p.

Beck, M. E., 1970, Paleomagnetism of Keweenawan intrusive rocks, Minnesota: Journal of Geophysical Research, v. 75, p. 4985–4996.

Billings, M. P., 1972, Structural geology (3rd edition): Englewood Cliffs, New Jersey, Prentice-Hall, Inc., 606 p.

Borradaile, G. J., Bayly, M. B., and Powell, C. McA., 1982, Atlas of deformational and metamorphic rock fabrics: New York, Springer-Verlag, 551 p.

Bredehoeft, J. D., Wolff, R. G., Keys, W. S., and Shuter, E., 1976, Hydraulic fracturing to determine the regional in situ stress field, Piceance basin, Colorado: Geological Society of America Bulletin, v. 87, p. 250–258.

Carmichael, R. S., 1982, Handbook of physical properties of rocks, Volume 2: Boca Raton, Florida, CRC Press, 345 p.

Chase, C. G., and Gilmer, T. H., 1973, Precambrian plate tectonics: The Mid-continent gravity high: Earth and Planetary Science Letters, v. 21, p. 70–78.

Craddock, C., 1972, Regional geologic setting, *in* Sims, P. K., and Morey, G. B., eds., Geology of Minnesota: A centennial volume: St. Paul, Minnesota, Minnesota Geological Survey, p. 281–291.

Craddock, C., Mooney, H. M., and Kolehmainen, V., 1970, Simple Bouguer gravity map of Minnesota and northwestern Wisconsin: St. Paul. Minnesota, Minnesota Geological Survey Miscellaneous Map M-10.

Davidson, D. M., Jr., 1982, Geological evidence relating to interpretation of the Lake Superior Basin: Geological Society of America Memoir 156, p. 5–14.

Dennis, J. G., 1967, International tectonic dictionary: American Association of Petroleum Geologists Bulletin, 196 p.

Dennis, J. G., and Weber, K., 1979, Terminology of cleavage and schistosity, *in* Dennis, J. G., Murawski, H., and Weber, K. eds., International tectonic lexicon: A prodrome: Stuttgart, West Germany, E. Schweizerbart'sche Verlagsbuchhandlung (Nägele U. Obermiller), p. 96–122.

Dunn, D. E., LaFountain, L. J., and Jackson, R. E., 1973, Porosity dependence and mechanism of brittle fracture in sandstones: Journal of Geophysical Research, v. 78, p. 2403–2417.

Foster, M. E., 1981, Analysis of small-scale structures in the Duluth Complex at Bardon Peak, St. Louis County, Minnesota [M.S. thesis]: Minneapolis, Minnesota, University of Minnesota, 144 p.

Gramberg, J., 1965, Axial cleavage fracturing, a significant process in mining and geology: Engineering Geology, v. 1, p. 31–72.

Green, J. C., 1982, Geology of Keweenawan extrusive rocks: Geological Society of America Memoir 156, p. 47–55.

——— 1983, Geological and geochemical evidence for the nature and development of the middle Proterozoic (Keweenawan) midcontinent rift of North America: Tectonophysics, v. 94, p. 413–437.

Groshong, R. H., 1976, Strain and pressure solution in the Martinsburg slate, Delaware Water Gap, New Jersey: American Journal of Science, v. 276, p. 1131–1146.

Grout, F. F., 1918, Internal structures of igneous rocks, their significance and origin, with special reference to the Duluth Gabbro: Journal of Geology, v. 26, p. 439–458.

Hancock, P. L., 1982a, Distinction between cleavage and joints using fracture separation, *in* Borradaile, G. J., Bayly, M. B., and Powell, C. McA., eds., Atlas of deformational and metamorphic rock fabrics: Berlin, West Germany, Springer-Verlag, p. 186–187.

——— 1982b, Distinction between cleavage and joints using morphology, *in* Borradaile, G. J., Bayly, M. B., and Powell, C. McA., eds., Atlas of deformational and metamorphic rock fabrics: Berlin, West Germany, Springer-Verlag, p. 188–189.

Helgeson, H. C., Delany, J. M., Nesbitt, H. W., and Bird, D. K., 1978, Summary and critique of the thermodynamic properties of rock forming minerals: American Journal of Science, v. 278A, p. 1–220.

Hobbs, B. E., Means, W. D., and Williams, P. F., 1976, An outline of structural geology: New York, John Wiley & Sons, 571 p.

Irvine, T. N., 1967, The Duke Island ultramafic complex, southeastern Alaska, *in* Wyllie, P. J., ed., Ultramafic and related rocks: New York, John Wiley & Sons, p. 84–97.

Irvine, T. N., 1980, Magmatic density currents and cumulus processes: American Journal of Science, v. 280-A, p. 1–58.

Jackson, E. D., 1961, Primary textures and mineral associations in the ultramafic zone of the Stillwater Complex, Montana: U.S. Geological Survey Professional Paper 358, 106 p.

Jaeger, J. C., 1969, Elasticity, fracture and flow: London, Methuen and Co., 268 p.

King, E. R., and Zietz, I., 1971, Aeromagnetic study of the midcontinent gravity high of central United States: Geological Society of America Bulletin, v. 82, p. 2187–2208.

Lawn, B. R., and Wilshaw, T. R., 1975, Fracture of brittle solids: Cambridge, England, Cambridge University Press, 204 p.

Means, W. D., 1982, Incipient cleavage in fine sandstone, center White Creek, New York, U.S.A., *in* Borradaile, G. J., Bayly, M. B., and Powell, C. McA., eds., Atlas of deformational and metamorphic rock fabrics: Berlin, West Germany, Springer-Verlag, p. 204–205.

Nickelsen, R. P., 1972, Attributes of rock cleavage in some mudstones and limestones of the Valley and Ridge province, Pennsylvania: Pennsylvania Academy of Science, v. 46, p. 107–112.

Page, N. J., 1977, Stillwater Complex, Montana: Rock succession, metamorphism and structure of the complex and adjacent rocks: U.S. Geological Survey Professional Paper 999, 79 p.

Palmer, H. C., 1970, Paleomagnetism and correlation of some middle Keweenawan rocks, Lake Superior: Canadian Journal of Earth Science, v. 7, p. 1410–1436.

Paterson, M. S., 1978, Experimental rock deformation—The Brittle Field: New York, Springer-Verlag, 254 p.

Platt, J. P., and Vissers, R.L.M., 1980, Extensional structures in anisotropic rocks: Journal of Structural Geology, v. 2, p. 397–410.

Pollard, D. D., Delaney, P. T., Duffield, W. A., Endo, E. T., and Okamura, A. T., 1983, Surface deformation in volcanic rift zones: Tectonophysics, v. 94, p. 541–584.

Powell, C. McA., 1979, A morphological classification of rock cleavage: Tectonophysics, v. 58, p. 21–34.

Price, N. J., 1966, Fault and joint development in brittle and semi-brittle rocks: London, England, Pergamon, 176 p.

Price, N. J., and Hancock, P. L., 1972, Development of fracture cleavage and kindred structures: International Geological Congress, 24th, Montreal, Section 3, p. 584–592.

Scarfe, C. M., and Wyllie, P. J., 1967, Experimental redetermination of the upper stability limit of serpentine up to 3-kb pressure [abs.]: American Geophysical Union Transactions, v. 48, p. 225.

Schwartz, G. M., 1949, The geology of the Duluth metropolitan area: Minnesota Geological Survey Bulletin 33, 136 p.

Secor, D. T., Jr., 1969, Mechanics of natural extension fracturing at depth in the earth's crust: Canada Geological Survey Paper 68-52, p. 3–47.

Segall, P., 1984, Formation and growth of extensional fracture sets: Geological Society of America Bulletin, v. 95, p. 454–462.

Segall, P., and Pollard, D. D., 1983, Joint formation in granitic rocks of the Sierra Nevada: Geological Society of America Bulletin, v. 94, p. 563–575.

Siddans, A.W.B., 1972, Slaty cleavage—A review of research since 1815: Earth Science Reviews, v. 8, p. 205–232.

——— 1977, The development of slaty cleavage in part of the French Alps: Tectonophysics, v. 39, p. 533–557.

Sims, P. K., 1970, Geologic map of Minnesota: Minnesota Geological Survey Miscellaneous Map M-14.

Taylor, R. B., 1964, Geology of the Duluth Gabbro Complex near Duluth, Minnesota: Minnesota Geological Survey Bulletin 44, 63 p.

Treagus, S. H., 1973, Buckling instability of a viscous single-layer system, oblique to the principal compression: Tectonophysics, v. 19, p. 271–289.

——— 1981, A theory of stress and strain variations in viscous layers, and its geological implications: Tectonophysics, v. 72, p. 75–103.

Turner, F. J., and Weiss, L. E., 1963, Structural analysis of metamorphic tectonites: New York, McGraw-Hill, 545 p.

Wager, L. R., and Brown, G. M., 1967, Layered igneous rocks: San Francisco, California, W. H. Freeman and Co., 588 p.

Wager, L. R., and Deer, W. A., 1939, Geological investigations of East Greenland, part III: The petrology of the Skaergaard intrusion, Kangerdlugssuag, East Greenland: Meddelefser om Grönland, v. 105, no. 4, 352 p.

Wawersik, W. R., and Brace, W. F., 1971, Post-failure behavior of a granite and diabase: Rock Mechanics, v. 3, p. 61–85.

Weiblen, P. W., and Morey, G. B., 1980, A summary of the stratigraphy, petrology, and structure of the Duluth Complex: American Journal of Science, v. 280-A, p. 88–133.

Wilson, G., 1982, Introduction to small-scale geological structures: London, England, George Allen and Unwin, 128 p.

Wong, T. -F., 1982, Micromechanics of faulting in Westerley granite: International Journal of Rock Mechanics and Mining Science and Geomechanics Abstracts, v. 19, p. 49–64.

MANUSCRIPT RECEIVED BY THE SOCIETY DECEMBER 13, 1984
REVISED MANUSCRIPT RECEIVED AUGUST 23, 1985
MANUSCRIPT ACCEPTED AUGUST 30, 1985
PUBLICATION NO. 1061, SCHOOL OF EARTH SCIENCES, UNIVERSITY OF MINNESOTA

Printed in U.S.A.