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Fold reorientation and quartz microfabric in the Okanogan dome mylonite zone Washington : kinematic and tectonic implications

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FOLD REORIENTATION AND QUARTZ MICROFABRIC
IN THE OKANOGAN DOME MYLONITE ZONE, WASHINGTON:
KINEMATIC AND TECTONIC IMPLICATIONS

by

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ABSTRACT

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Fold Reorientation and Quartz Microfabric in the Okanogan Dome Mylonite Zone, Washington: Kinematic and Tectonic Implications

Director: Donald W. Hyndman DWH

Mylonitization of crystalline rocks in the Okanogan dome, north-central Washington, reoriented tight folds in high-grade metasedimentary gneiss and formed type I crossed-girdle c-axis fabric patterns (Lister and Willliams, 1979) in granodiorite. Rocks studied include: (1) thinly layered hornblende-biotite-plagioclase paragneiss; (2) homogeneous granodiorite orthogneiss; (3) mylonite derived from the gneisses.

Sillimanite-zone regional metamorphism and deformation of sediments formed a schistosity parallel to compositional layering and a sillimanite-biotite mineral lineation. Syn- to post-metamorphic deformation formed tight (20-30°) similar-style folds with fold axes at a high angle to mineral lineation. Fold axes plunge gently east and axial planes dip steeply south. Several granodiorite to tonalite plutons intruded the high-grade paragneisses after folding, indicated by non-folded, planar late-stage pegmatite and aplite dikes.

Mylonitization of paragneiss formed a foliation which dips moderately southwest and contains a subhorizontal quartz-biotite-plagioclase mineral elongation lineation that plunges both S 50-65° E and N 50-65° W. Steeply southeast-dipping extensional crenulation cleavage in layered paragneiss apparently formed during mylonitization as a non-coaxial strain feature, and indicates bulk dextral shear viewed northeast. Transposition of the regional metamorphic fabric and progressive reorientation of post-metamorphic folds in the mylonite zone during progressive simple shear, formed uniformly coplanar and colinear structures. Combined flattening and rotation of fold elements with respect to mylonitic shear foliation indicates mylonitization occurred during eastward displacement of upper plate rocks over rocks within Okanogan dome.

Mylonitization of granodiorite formed a ductile shear fabric consisting of a subplanar foliation and unidirectional quartz-biotite-plagioclase extension lineation. Quartz c-axis preferred orientations from mylonitic orthogneiss show diffuse patterns toward the interior of Okanogan dome, and more well-developed patterns toward the dome margin, indicating greater degrees of ductile strain at higher structural levels within the mylonite zone. Type I crossed-girdle c-axis fabric patterns from the granodiorite mylonites indicate mylonitization involved non-coaxial progressive simple shear strain, and the fabric asymmetry indicates a dextral or clockwise sense of shear in the mylonites viewed northeastward.

A regional top-to-the-east sense of shear in the Okanogan dome mylonite zone is indicated both by mesoscopic pre-existing fold reorientation and quartz c-axis microfabrics. Mylonitization occurred during ductile crustal shear in a deep-seated intracontinental thrust or decollement zone, which displaced upper-plate allochthonous rocks to the east relative to autochthonous rocks exposed in the Okanogan dome.

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INTRODUCTION

Okanogan dome in north-central Washington is a high-grade crystalline terrain which contains rocks and structures representative of a long period of geologic and tectonic activity. Early episodes during formation of the Okanogan dome reflect middle crustal orogenesis and differ from later shallow crustal structural events, just as in the tectonic evolution of equivalent rocks in the Shuswap metamorphic complex of British Columbia proposed by Brown and Read (1983). The principal tectonic feature relating Okanogan dome rocks to orogenic deformation is a thick ductile shear zone, perhaps associated with a system of deep-seated subhorizontal continental thrust faults. Later shallow crustal deformation was brittle and resulted in the rise, arching and exposure of rocks comprising the Okanogan dome. In this paper I discuss the early deformation and strain history of rocks in the dome before regional uplift and exposure as a crystalline structural culmination. Principal episodes of deformation in the early history of Okanogan dome involve, in order from oldest to youngest: syn-tectonic high-grade regional metamorphism, folding, and ductile crustal shear or mylonitization, of metasedimentary and plutonic rocks associated with a continental-arc tectonic environment. Structural analysis of mesoscopic folds in metasedimentary gneisses, and quartz c-axis preferred orientations in Okanogan dome orthogneisses indicate that regional crustal shear within a broad ductile mylonite zone transported high-grade crystalline rocks southeastward over equivalent-grade rocks presently exposed in Okanogan dome. Simple-shear strains in the mylonite zone were strong enough to reorient pre-existing folds, and the deformation style indicates an environment greatly different from that which formed a brittle zone of brecciation at the margin of the Okanogan dome during a late period of deformation.

Okanogan dome is one of a number of crystalline terrains in the northwest United States and southwest Canada, including the Kettle, Bitterroot, and Spokane (Priest River) domes in the U. S., and the Valhalla, Thor-Odin and Frenchman Cap domes of the Shuswap metamorphic complex in British Columbia. These domes are part of a discontinuous string of Cordilleran metamorphic core complexes extending from southeast British Columbia to northern Mexico (Crittenden and others, 1980; Armstrong, 1982)(Fig. 1). Use of the terms gneiss dome (Fox and others, 1977) and metamorphic core complex (Coney, 1980; Cheney, 1980) to describe the Okanogan dome are compatible; however, the latter is more inclusive of the characteristics shared by many crystalline terrains of the Cordillera. Use of the term "dome" in this paper refers to both geomorphic and structural features. Geomorphically the Okanogan range exhibits a low domal profile between the Okanogan and Sanpoil Rivers, particularly as viewed to the east from the Okanogan River Valley. Gently west-dipping flanks result from preferential erosion along foliation planes in a broad domal mylonite zone. Formation of the dome refers to structural and plutonic events which raised high-grade, deformed crystalline rocks from depth to their present position and which preceded erosional exposure of the domal crystalline terrain. Discussion in this paper is limited to the southwest portion of the Okanogan dome and stems from geologic mapping and structural analysis of rocks in the Omak Lake, Washington 15′ quadrangle (Fig. 3).

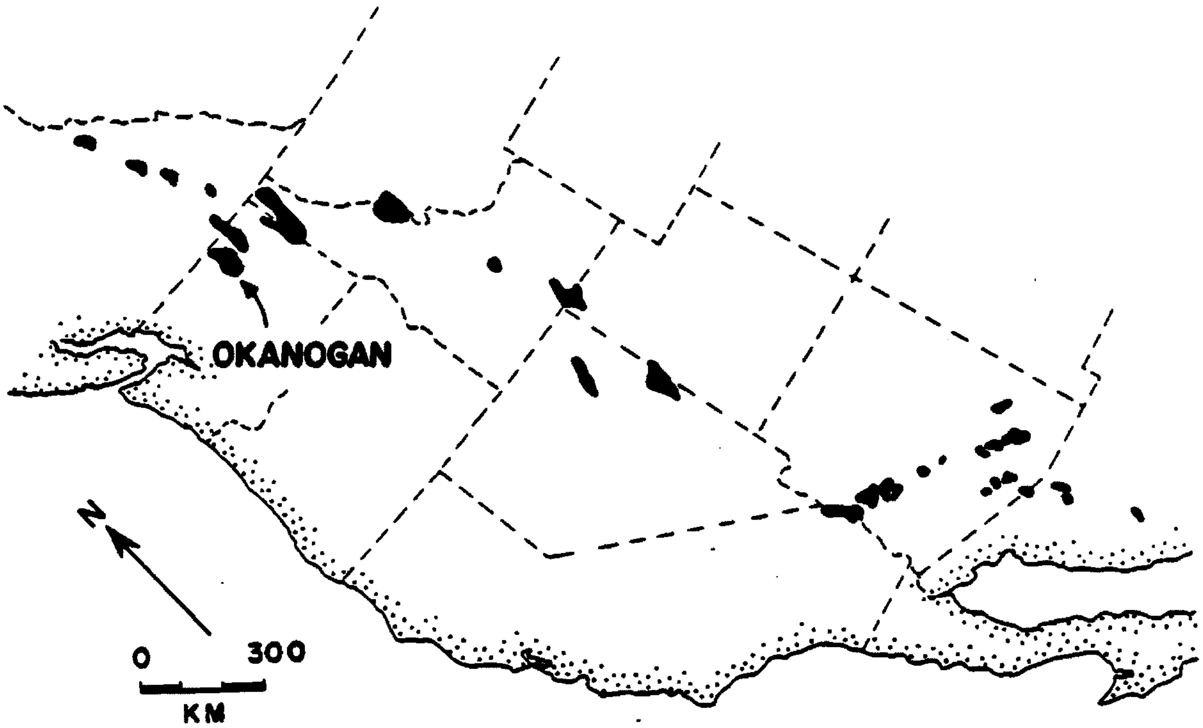


FIG. 1. Location map of the North American Cordillera (after Coney, 1980). Metamorphic core complexes shown in black.

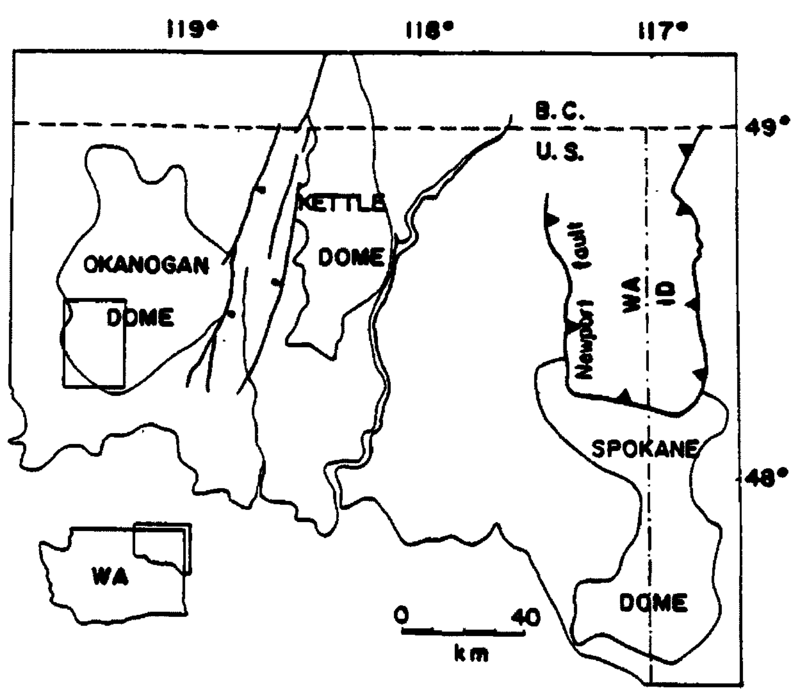


FIG. 2. Location map of northeastern Washington (after Cheney, 1980), showing location of the Okanogan, Kettle, and Spokane domes. Ball and stick symbols show downthrown side along high-angle faults of the Republic Graben. Study area outlined by rectangle, and is area of map in Figure 3.

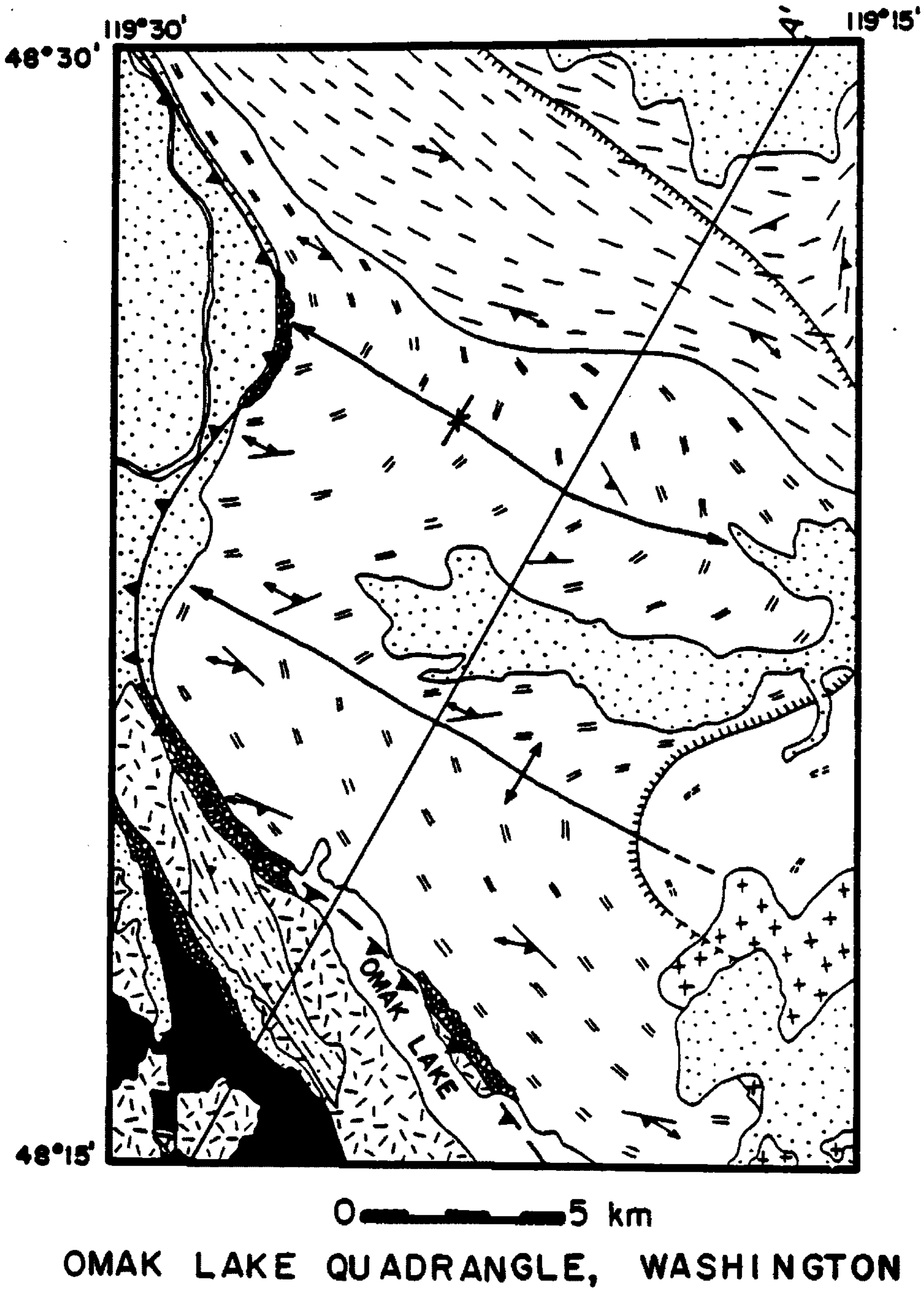
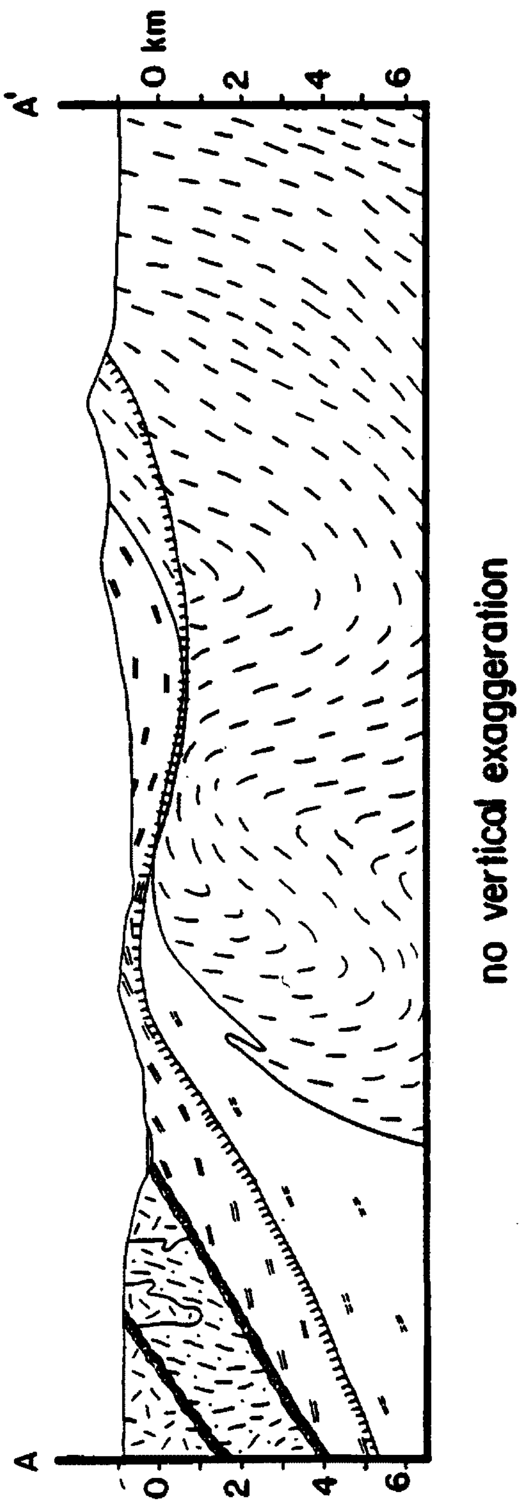


FIG. 3. Geologic map and cross section of study area in Okanogan dome, Omak Lake 15′ quadrangle, Washington; geology by J. W. Goodge and V. L. Hansen in 1982. Course of the Okanogan River shown in northwest corner of map area by long double line. Conventional structure symbols used, as explained in legend.

FIG. 3. con't.



MAP UNITS

Quaternary deposits; undivided

Miocene basalt of the Columbia River; unconformably overlies undeformed rocks of the upper plate.

Granitic rocks; undeformed; ages unknown. Lined: upper plate rocks, including equigranular biotite granodiorite and K-feldspar-megacryst biotite granodiorite. Crosses: lower plate rocks, leucocratic porphyritic granite, intrudes lower plate orthogneisses.

Orthogneisses of Okanogan dome lower plate. Includes K-feldspar-megacryst biotite granodiorite, equigranular biotite granodiorite, and hornblende-biotite tonalite, all containing numerous pegmatite and aplite dikes. All major phases contain a mylonitic fabric consisting of foliation and mineral lineation, increasing in intensity toward the dome margin. Intrudes paragneisses of the lower plate.

Layered paragneisses, undivided assemblage in upper plate (lines and dots) and lower plate (lines). Upper plate rocks include hornblende-biotite gneiss, augen gneiss, sillimanite-muscovite-biotite schist, amphibolite, quartzite, marble, pyroxene calc-silicate gneiss and metamorphic dunite. Lower plate rocks (Tonasket gneiss) include hornblende-biotite gneiss, augen gneiss, sillimanite-garnet-muscovite-biotite schist, biotite quartzite, amphibolite, and pyroxene calc-silicate gneiss. Both sequences are compositionally layered, contain a well-formed schistosity and mineral lineation, and are folded locally into tight similar-style folds. Lower plate rocks contain a superimposed mylonitic fabric near the margin of Okanogan dome.

MAP SYMBOLS

lithologic contact

chloritic breccia zone

low-angle fault; inferred between areas of chloritic breccia

limit of mylonitic deformation; gradational; mylonitized rock on hatchured side

trace and plunge of antiform

trace and plunge of synform

strike and dip of mylonitic foliation

trend and plunge of mylonitic lineation

Authors of previous studies provide several interpretations of structures in the Okanogan dome, and formulate quite different models of the origin of this crystalline complex. Initial investigations by Pardee (1918) and Waters and Krauskopf (1941) led these workers to interpret the crystalline-cored dome as an extensive plutonic sequence which developed mylonitic and cataclastic textures along its border during emplacement. Fox and others (1976, 1977) concur with the thermal mechanism involved in this hypothesis; however, they propose that high-grade metamorphism and plutonism related to "mobilization and diapiric intrusion" of crystalline rocks into lower-grade metamorphic rocks was accompanied by brittle deformation along the perimeter of the thermally uplifted gneiss dome. Snook (1965) disagrees with these interpretations. Based on a study of gneissic rocks along the western side of the Okanogan dome, he suggests that metamorphism of rocks in the dome significantly pre-dated mylonitization along the dome border, and that dome formation is a consequence of regional gentle folding and high-angle block faulting. Cheney (1980) proposes a fourth hypothesis which attempts to unify formation of the Kettle and Okanogan domes; he contends that early Cenozoic thrusting and cataclasis separated metamorphic and plutonic basement rocks from upper-plate Precambrian and Tertiary layered rocks, followed by structural doming during post-Eocene gentle folding.

This paper summarizes structural relationships of deformed rocks in the southwest portion of the Okanogan dome, outlines an early deformation history prior to structural doming, presents evidence for regional top-to-the-east ductile crustal shear from structural analysis of folds and quartz c-axis preferred orientations, and suggests kinematic and tectonic interpretations for early deformation of rocks in the Okanogan dome. Results of this study expand on more general geologic and structural relationships in the southwest portion of the Okanogan dome previously discussed by Goodge (1983) and Goodge and Hansen (in press).

GEOLOGIC SETTING

Metamorphic core complexes, or gneiss domes, of the northern Cordillera region in Canada and the northern United States lie within the Omineca crystalline belt, a region comprised largely of plutonic and metamorphic rocks (Monger and others, 1972). In southern British Columbia these domes comprise high-grade metamorphic and plutonic rocks of the Monashee Complex, and they are bounded along the Okanogan and Columbia Rivers by superimposed mylonite zones and low-angle faults (Read and Brown, 1981; Ross, 1981; Brown and Murphy, 1982; Brown and Read, 1983). Such superposition of structures continues southward along the Columbia River in the east flank of the Kettle dome and along the Okanogan River in the western flank of the Okanogan dome (Snook, 1965; Cheney, 1980; Rhodes and Cheney, 1981; Hansen, 1983a; Goodge and Hansen, in press). In addition to structural similarities, the Okanogan dome is continuous in lithology and metamorphic grade to the north with high-grade crystalline rocks of the Shuswap metamorphic complex. Okanogan dome is separated from the Kettle dome to the east by Tertiary volcanics and clastic sediments of the Republic graben, formed along steep normal faults (Fig. 2). Miocene basalts of the Columbia Plateau lie southwest of the Okanogan dome and end abruptly at the southwest dome boundary marked by the Okanogan River Valley and Omak Lake.

Okanogan dome is flanked to the southwest by plutonic and high-grade crystalline rocks across the Omak Lake fault, and to the west by Triassic sedimentary rocks across the Okanogan fault (Snook, 1965).

Rocks in the southwest portion of the Okanogan dome include, in decreasing order of abundance, homogeneous coarse-grained orthogneiss, high-grade layered paragneiss, leucocratic porphyritic granite and microdiorite dikes (Goodge and Hansen, in press)(Fig. 3). Of primary importance to this study are the paragneisses and orthogneisses. Compositionally varied paragneiss (tgn) contains folds which serve as passive strain markers in the mylonite zone, and the dominant orthogneiss rock type, K-feldspar-megacryst biotite granodiorite (olgd), displays a well-developed mylonitic quartz c-axis fabric. Although contact relationships between orthogneiss and paragneiss are obscured by deformation, orthogneiss is the younger of these major units. This relationship is shown by planar, late-stage pegmatite and aplite dikes in orthogneiss, indicating intrusion after folding of layered paragneisses.

The layered assemblage of high-grade metasedimentary gneisses includes hornblende-biotite-plagioclase gneiss, biotite gneiss, K-feldspar augen gneiss, sillimanite-garnet-orthoclase-biotite pelitic schist, biotite quartzite, amphibolite, diopside calc-silicate gneiss, and garnet-bearing alaskite gneiss. Layered compositional variation and the presence of calcareous, pelitic and quartzitic units indicates these rocks are of sedimentary origin. Hornblende-biotite gneiss and biotite gneiss are arealy dominant. Metamorphism reached sillimanite zone of the amphibolite facies but probably did not exceed this level;

sillimanite-orthoclase disequilibrium indicates less than sillimanite-orthoclase zone metamorphism. Throughout the paragneisses, several of the broad lithologic units are thinly layered on a scale ranging from millimeters to several centimeters. Broad compositional or lithologic differences probably reflect primary sedimentary variation, whereas the small-scale strongly layered appearance may be the result of metamorphic differentiation. Most rocks, particularly fine-grained mafic-rich units, contain a prominent mineral lineation and schistosity formed by biotite, hornblende and sillimanite. Metamorphic schistosity is everywhere subparallel to layering, and combined they form a general metamorphic foliation.

Megacrystic granodiorite is a light- to medium-gray (CI = 5-7), homogeneous-textured plutonic rock containing conspicuous pink to gray K-feldspar megacrysts (2-8 cm long) comprising 1 to 5 percent total rock volume. The granodiorite matrix is medium-grained (0.5-5.0 mm) and equigranular, and it contains quartz, plagioclase, biotite, and minor orthoclase, sphene, allanite and opaque oxide. Megacrystic granodiorite and two subordinate orthogneiss rock types, equigranular biotite granodiorite and hornblende-biotite tonalite, are probably contemporaneous phases of a single pluton as indicated by textural and contact relationships (Goodge and Hansen, in press). Snook (1965) proposed a metasomatic origin for the orthogneiss units in the Okanogan dome, but I agree with the conclusions of Waters and Krauskopf (1941) and Fox and others (1976) that these units are of igneous parentage for the following reasons: (1) compositional homogeneity, (2) primary igneous textures in the granodiorites, including abundant euhedral sphene, well-twinned plagioclase, and perthitic orthoclase, (3) inclusions of K-feldspar-megacryst granodiorite in equigranular granodiorite, indicative of magmatic processes, and (4) invariant major-element geochemistry of K-feldspar-megacryst granodiorite (Hansen, 1983b).

Dominating the form and structural fabric of Okanogan dome rocks is a zone of penetrative ductile mylonite, as defined by Bell and Etheridge (1973), which is not present in rocks southwest of the dome border. This broad mylonite zone is concordant with the dome perimeter, and forms the "flat-iron"-like aprons slanting up eastward from the Okanogan Valley. Thickness of the mylonite zone generally ranges from 1.0 to 1.5 km. The zone is characterized by a gradational change in intensity of deformation such that structurally higher levels within the zone display greater degrees of deformation. Deformation in the mylonite zone is ductile as in most major orogenic shear zones (Mattauer, 1975; Ramsay and Graham, 1970; Ramsay, 1980; Mattauer and others, 1983), and study of feldspar equilibria by Hansen (1983b) indicates that the Okanogan mylonite formed under middle greenschist facies metamorphic conditions. Thus, mylonitization temperatures of less than 400°C (Hansen, 1983b) and sillimanite-zone metamorphism in excess of 500°C (Hyndman, 1972, p. 313) describe not only a thermal gap but a relative age gap between metamorphism and later mylonitization. Along the southwestern boundary of Okanogan dome, wide fault zones marked by extensive brecciation and chloritization indicate that deformation there occurred in a brittle low-temperature hydrothermal environment (Goodge and Hansen, in press). Brecciation within these brittle fault zones completely disrupts highly symmetrical mylonitic fabric elements. Such cross-cutting structural relationships and a significant difference in thermal conditions during deformation indicate that mylonite and breccia represent different stages in the structural evolution of the Okanogan dome. Discussion to follow focuses on deformation related to regional folding and mylonitization only. The significance of the chlorite breccia zone is presented in other earlier studies (Snook, 1965; Fox and others, 1977; Goodge and Hansen, in press).

The mylonite zone as defined by mylonitic foliation is warped into shallowly northwest-plunging gentle folds, axes of which parallel regional trends of a unidirectional mylonitic lineation (Fig. 3). Numerous north-northeast-trending steep joints cut the mylonitic fabric approximately normal to mylonitic lineation and are intruded by undeformed microdiorite dikes.

STRUCTURAL RELATIONSHIPS OF GNEISSIC ROCKS

Orthogneiss.

The texture and fabric of the Okanogan mylonite is well defined in orthogneisses near border areas of the Okanogan dome and becomes increasingly faint with structural depth to the east (see cross-section, Fig. 3). In contrast to several recorded deformation episodes in rocks of the paragneiss assemblage, only mylonitization affected the orthogneiss units. Absence of any evidence for earlier deformation in the orthogneisses indicates these plutons were emplaced during or after the peak of high-grade metamorphism and folding, and makes the orthogneisses ideal for study of quartz preferred orientation related to mylonitization. In this study I consider mylonitic deformation of K-feldspar-megacryst granodiorite (olgd).

The mylonitic fabric consists of several mesoscopic and microscopic features, including a coplanar mineral foliation and lineation. The mylonitic foliation is a broadly planar biotite and quartz fabric which, on a grain scale, appears wavy. The most conspicuous form of the mylonitic foliation occurs in K-feldspar-megacryst granodiorite, in which curved biotite flakes and lenticular quartz wrap around euhedral to subhedral coarse K-feldspar crystals. In the mylonite foliation plane, elongate lenses and streaks of biotite, quartz, and plagioclase define a unidirectional mylonite lineation; the stretched and rod-shaped mineral aggregates which form the lineation represent elongation or extensional structures aligned parallel to the direction of tectonic transport in the mylonite zone (Mattauer and others, 1983; Goodge and Hansen, in press). Mylonite foliation is gently folded about shallow-plunging, macroscopic fold axes which parallel the trend of mylonitic lineation in this area (Fig. 4), and hence may be "a" folds (Mattauer, 1975) if they formed with axes parallel to the direction of tectonic transport during mylonitization.

Prominent in megacrystic granodiorite on both mesoscopic and microscopic scales are sharp planar discontinuities, thin shear surfaces (Mm, after Hansen, 1983b), which cut across mylonitic foliation at acute angles. In their form and relation to foliation planes these shear surfaces resemble S surfaces in other mylonite zones described by Ramsay and Graham (1970), Berthe and others (1979b) and Brown and Murphy (1982). Hansen (1983) discusses their kinematic significance. Shear-surface planes dip uniformly west to southwest parallel to mylonitic lineation plunge and at a steeper inclination than mylonitic foliation. Locally in megacrystic granodiorite mylonitic foliation is

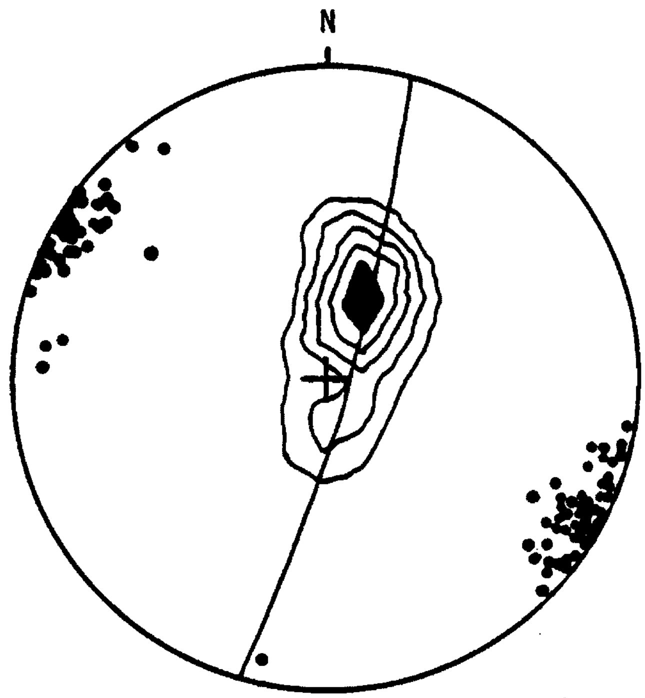


FIG. 4. Stereonet of mylonitic fabric elements in orthogneiss, showing contours of 350 poles to mylonitic foliation, 117 mylonitic lineation poles (dots), and great-circle trace of plane perpendicular to foliation fold axis. Contours 5.00, 8.75, 12.50, 16.25, and 20.00 percent, per 1% area.

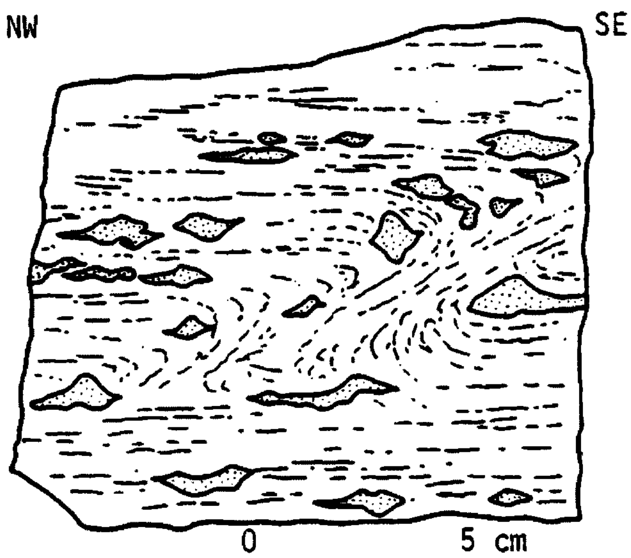


FIG. 5. Layer parallel fold of mylonitic fabric in orthogneiss. View perpendicular to mylonitic foliation and parallel to mylonitic lineation, east to the right. K-feldspar dotted, lines show trace of biotite, and white areas contain quartz, plagioclase, and biotite.

folded around axes lying in the plane of foliation but perpendicular to mylonitic lineation (Fig. 5). Mylonitic foliation is concordant to the limbs of these intrafolial folds, yet also forms a strong axial-plane fabric through their hinges. Initial buckling of the foliation surface in the direction of shear and subsequent layer parallel shear of the fold hinges at increased levels of strain may account for formation of these folds (Rhodes and Gayer, 1977).

On a smaller scale, planar fractures oriented perpendicular to mylonitic lineation cut euhedral megacrysts in the granodiorite. The fractures end abruptly at crystal boundaries, do not cut into mylonitic foliation, appear unrelated to crystal orientation or cleavage, and commonly contain a quartz-biotite-muscovite-epidote mineral assemblage. Formation of the fractures must be a response to mylonitization, and not a post-mylonitic period of brittle deformation, because they cut only megacrysts and not the mylonitic fabric. Such pull-apart fractures indicate that mylonitization involved local extensional strain. As coarse feldspar megacrysts respond with brittle behavior to mylonitization, finer-grained matrix minerals deform as a ductile aggregate around the larger crystals. Quartz deforms under ductile conditions by a combination of grain-size reduction and recrystallization, both measures of strain which becomes more pronounced at higher structural levels in Okanogan dome. Rocks of greatest relative strain contain ribbon quartz grains marked by extreme elongation in the foliation plane parallel to lineation. As with quartz, biotite grain-size is progressively reduced with increased relative strain.

In summary, K-feldspar-megacryst granodiorite exhibits progressive overall shear strain upward within the mylonite zone from an undeformed state exposed at deep structural levels in the Okanogan dome. A mylonitic fabric marked by foliation and a unidirectional east-southeast-trending lineation formed in a direction parallel to tectonic transport within the shear zone. "A"-type folds, axes of which parallel regional mylonitic lineation, substantiate a shear direction parallel to lineation if they formed contemporaneously with mylonitization (Mattauer and others, 1983). A dextral sense of shear as viewed northward within the Okanogan mylonite is given by shear folds in mylonitic foliation. Local extension of brittle feldspar megacrysts occurred during deformation of a quartz- and biotite-rich ductile matrix.

Paragneiss.

Rocks of the metasedimentary gneiss sequence contain structures representative of several periods of deformation and metamorphism, which appear correlative to similar events interpreted in high-grade rocks of the Shuswap metamorphic complex (Wheeler, 1965; Hyndman, 1968; Fyles, 1970; Reesor, 1965; Reesor and Moore, 1971; McMillen, 1973; Brown and Tippett, 1978; Read, 1980; Read and Brown, 1981; Read and Klapacki, 1981; Brown and Read, 1983). Sequentially recognized in the Okanogan dome are deformation structures associated with metamorphism, folding, extensional crenulation, and mylonitization.

The oldest recognizable deformation fabric is related to high-grade regional metamorphism, and is defined by a well-developed foliation and lineation. Foliation consists of compositional layering and schistosity. Map-scale layering includes the broad lithologic variation described above, and these lithologically distinct units are thinly layered on a scale ranging from millimeters to several centimeters. All rocks, with the exception of calc-silicate and alaskite gneiss which contain negligible micas and amphiboles, contain a planar biotite-hornblende schistosity parallel to thin compositional layers. Most rocks, particularly fine-grained mafic-rich units, contain a mineral lineation formed by recrystallized biotite, hornblende, and, where present, sillimanite. This lineation is distinguished from a younger mylonitic lineation because it is evenly distributed on schistosity surfaces, is formed by the preferred orientation of undeformed biotite and hornblende, and does not include stretched quartz rods or plagioclase augen. Microscopically, the paragneiss units exhibit blastic metamorphic textures, including well-foliated but undeformed biotite, hornblende, and sillimanite (0.5-3.0 mm long), garnet porphyroblasts (2-3 mm) with no detectable rotation, unstrained plagioclase, and weakly undulose to unstrained, serrate quartz (0.25-3.0 mm). These textures indicate that high-grade metamorphic recrystallization and concomitant stress formed a strong preferred mineral orientation without significant shear or translational strains.

The pervasive metamorphic fabric is folded locally throughout the study area (Fig. 3), and textural evidence suggests this deformation occurred during or after the latest stages of regional metamorphism. These folds generally have amplitudes ranging from several centimeters to one meter, have interlimb angles ranging from 30 to 60 degrees, and are uniformly of similar style (Fig. 6)(Ramsay, 1967). A stereographic plot of metamorphic foliation reveals a regional megascopic fold or set of large folds in the northeast part of the study area (Fig. 7a). A continuous distribution of poles to foliation along the connecting great-circle girdle indicates no unique sense of vergence, but two concentrations of poles suggests northeast- and south-dipping limbs. Figure 8 shows common mesoscopic fold profiles, which indicate a northeastward direction of fold vergence not discernible from macroscopic structures. Folding of metamorphic foliation and biotite-hornblende schistosity (Fig. 9), as well as the formation of fold axes perpendicular to mineral lineation, demonstrate that folding occurred as a post-metamorphic phase of deformation. However, some rocks contain microscopic recrystallized polygonal arcs of biotite around fold hinges and a weak axial-plane schistosity, indicating that folding possibly occurred during the latest stages of regional metamorphism. Folding, then, occurred locally throughout the region, during or after high-grade metamorphism, and is of a style indicating deformation by ductile strain with a significant shear component (Ramsay, 1967; Hobbs and others, 1976; Hudleston, 1977). Southeast-directed tectonic compressive stresses are compatible with the regional vergence expressed by mesoscopic folds.

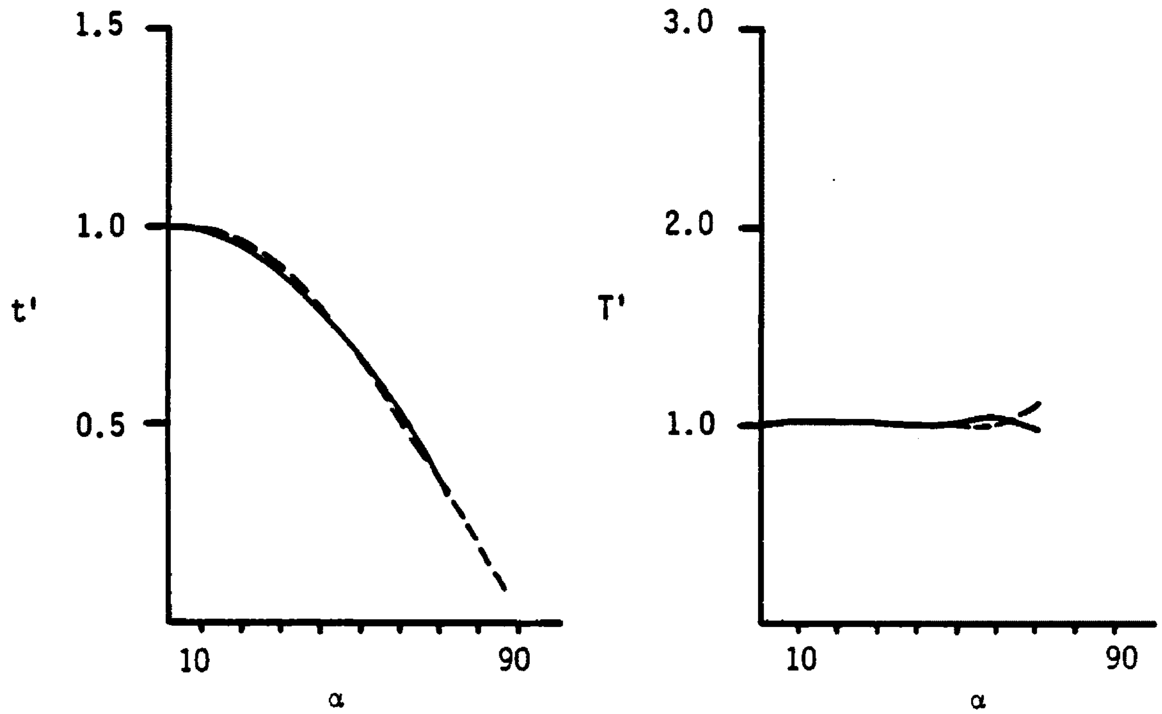
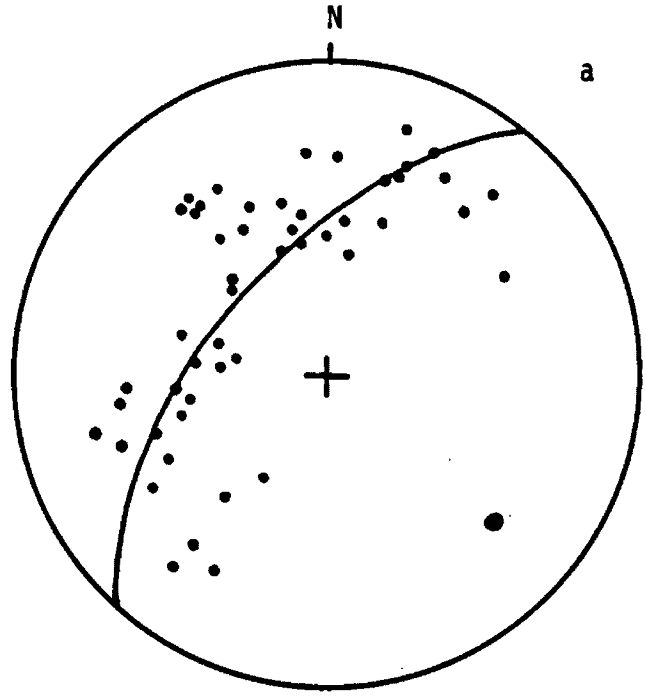


FIG. 6. Graphs of thickness variation of folds in paragneiss. Solid lines from 8 folds in metamorphic textural domain; dashed lines from 12 folds in mylonitic textural domain. T′ is thickness parallel to the axial surfaces; t′ is the orthogonal thickness. Values of determined as in Ramsay (1967, p. 361).



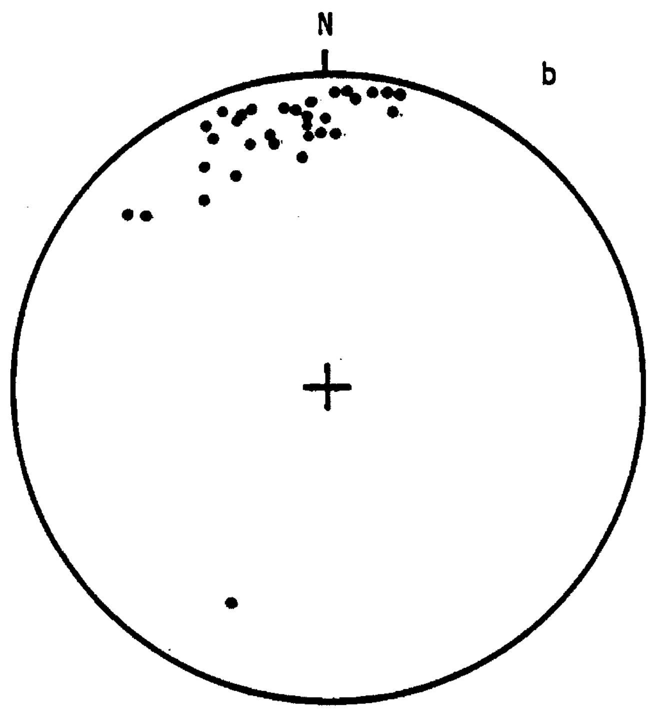


FIG. 7. Stereonet of structures in paragneiss. (a) 50 poles to metamorphic foliation in non-mylonitic paragneisses. Circled dot shows position of pole to great-circle girdle distribution of data. (b) 32 poles to planes of extensional crenulation cleavage in non-mylonitic paragneisses.

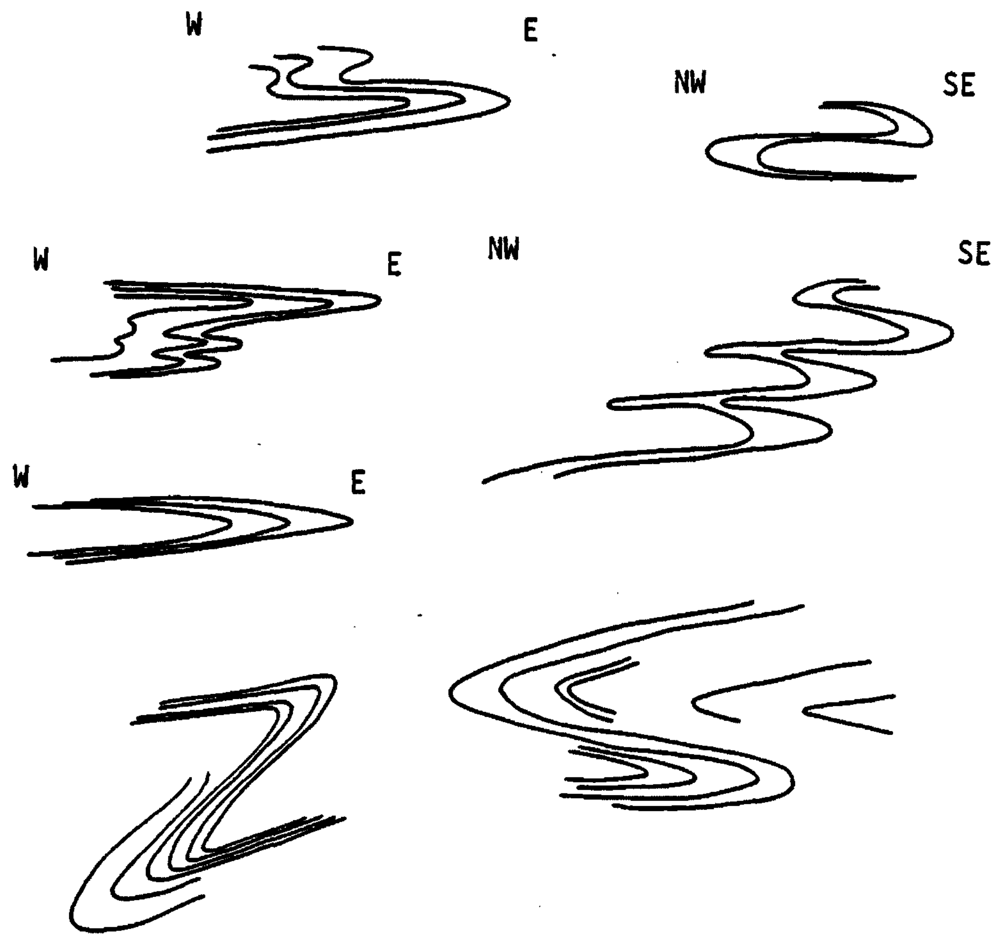


FIG. 8. Common fold styles in paragneiss, profile views approximately normal to fold axes. Horizontal scale of all sketches is 0.5 to 2.0 m.

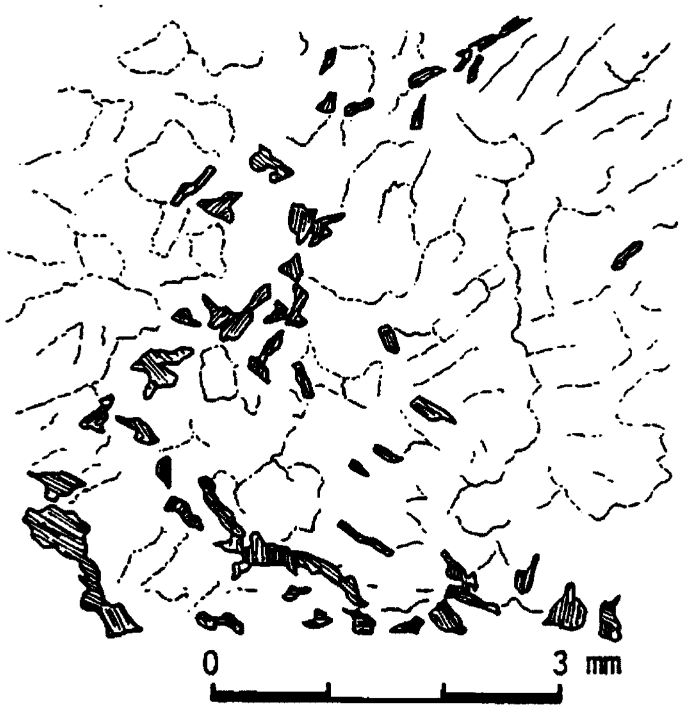


FIG. 9. Microscopic texture in the hinge of a paragneiss fold, view perpendicular to fold hinge. Fold hinge opens to right, and is marked by a biotite (lined) layer. White areas predominantly contain quartz. This example of non-mylonitic paragneiss contains no axial plane schistosity.

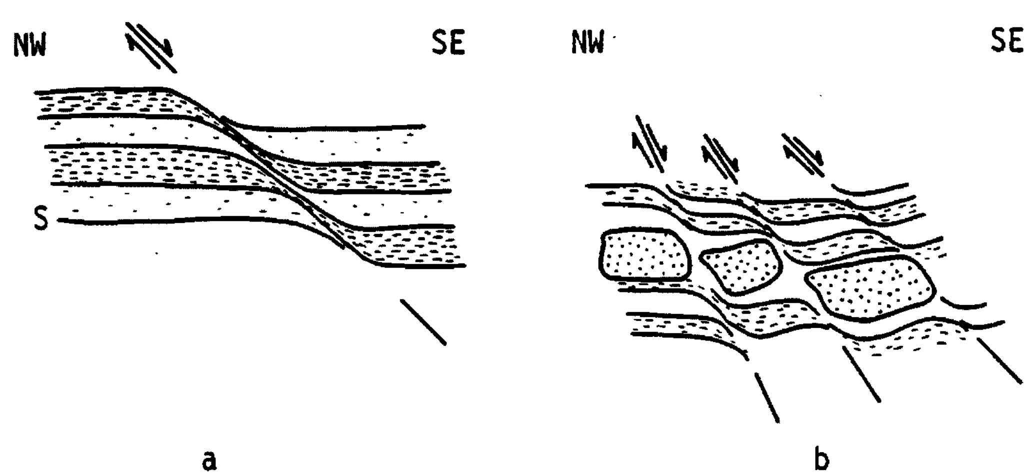


FIG. 10. Extensional crenulation cleavage structures. (a) Profile of extensional crenulation cleavage in layered paragneiss perpendicular to metamorphic foliation (S) and cleavage plane. (b) View of extensional crenulation cleavage and offset of distended amphibolite boudin (dotted) showing downward displacement to the southeast along cleavage zones.

Deformation of mesoscopic folds by an extensional crenulation cleavage (Platt and Vissers, 1980; D. Gray, pers. comm., 1983) may be directly associated with mylonitization. The extensional crenulation cleavage occurs in the paragneiss as southeast-dipping narrow (1-4 cm wide) shear bands which offset metamorphic layering and schistosity much like normal faults (Fig. 10a). Whereas shear bands of the extensional crenulation cleavage consistently dip steeply to the southeast (Fig. 7b), they intersect metamorphic layering at an acute angle ranging from 20 to 55 degrees, and opening to the northwest (Fig. 10a and b). A normal sense of offset on crenulation cleavage bands is inferred from drag of layers and schistosity into the crenulation cleavage, and southeast down-dropped blocks of an amphibolite boudin in hornblende-biotite gneiss (Fig. 10b). In most cases, layer continuity is preserved across the crenulation cleavage, although extremely attenuated individual layers pass through the cleavage zone. Generally wide spacing between adjacent cleavage zones precludes exposure of more than one or two sets at a single outcrop; multiple cleavage sets are rare. Unrecrystallized and weakly strained quartz grains characterize rock textures between cleavage planes. At one location, several cleavage zones disrupt limbs of a single similar-style fold, indicating that formation of crenulation cleavage post-dates folding. I observed no direct relationship between crenulation cleavage and the mylonitic fabric, but the crenulation cleavage occurs spatially at intermediate structural levels in the Okanogan dome below the most strongly formed mylonite. Ductile deformation inferred along the crenulation cleavage shear bands is compatible with the type of shear strain considered important for the formation of both similar folds and mylonitic fabrics, but cleavage formation appears to post-date similar-style folding.

From apparent layer drag and offset boudins, I infer dextral displacement along the shear bands as viewed to the northeast. The absence of shear fabrics parallel to layering implies that major movement took place along shear bands, not metamorphic foliation. Asymmetric development of crenulation cleavage depends in part on the orientation of pre-existing layers relative to the direction of maximum compressive stress (Cobbold and others, 1971; Cosgrove, 1976; Platt and Vissers, 1980). Additionally, crenulation cleavage which is asymmetric with respect to pre-existing layers may form either under dominant coaxial or non-coaxial stress (Platt and Vissers, 1980). The asymmetric development of southeast-dipping crenulation cleavage bands in the paragneisses implies non-coaxial deformation, or other than principal compressive stress normal to gneissic foliation (Fig. 11a). Similarly, an absence of slip or shear strain along foliation planes suggested by quartz textures implies minor compressive stress at a high angle to layering (Fig. 11b) and insignificant shear stress parallel to layering (Fig. 11c). Thus, a model involving shear strain at an angle to pre-existing layers proposed by Platt and Vissers (1980, Fig. 11c) appears to best account for the structures formed in Okanogan dome gneisses (Fig. 11d). In general, dominant dextral shear stress forms shear bands at an angle acute to the direction of maximum elongation, and the direction of maximum compression displaces "hanging wall" blocks downward. This model assumes shear bands form parallel to the direction of shear stress.

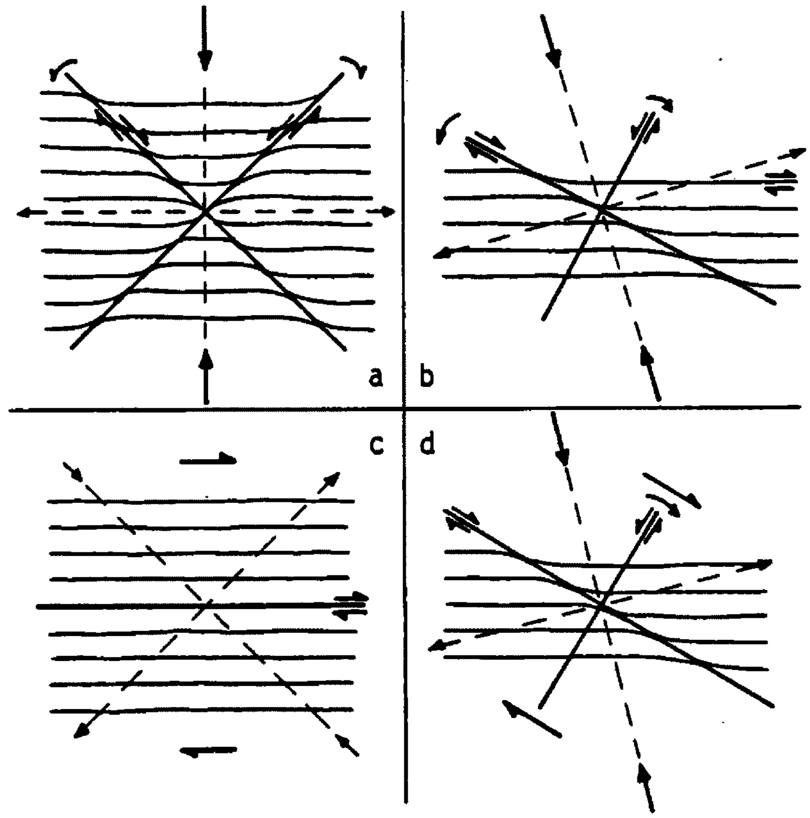


FIG. 11. Interpretation of extensional crenulation cleavage types in layered anisotropic rocks (after Platt and Vissers, 1980). Light lines show traces of compositional layers (S); heavy lines show trace of cleavage zones with sense of offset (arrows). Dashed lines show directions of maximum shortening and extension. (a) Coaxial deformation, maximum shortening perpendicular to S, and formation of a conjugate cleavage set. (b) Coaxial deformation, maximum shortening at non-normal angle to S, formation of conjugate shear bands where set at a low angle to S will dominate. Crenulation shear bands form with interlayer shear. (c) Non-coaxial deformation, simple shear parallel to layers S, formation of shear bands parallel to S without crenulation of layers. (d) Non-coaxial deformation, simple shear at an angle to layering S, formation of dominant set of crenulation cleavage shear bands in the direction of shear and in the absence of interlayer shear.

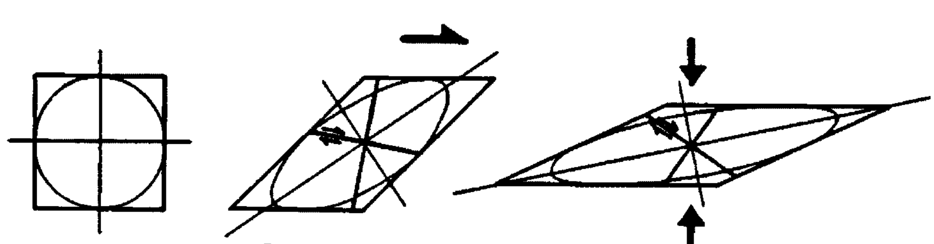


FIG. 12. Model for the formation of extensional crenulation cleavage in layered paragneisses of the Okanogan dome, showing progressive dextral simple shear, formation of crenulation cleavage, and flattening, resulting in steeply southeast-dipping cleavage.

Using the model of Platt and Vissers, I propose that the extensional crenulation cleavage formed in Okanogan dome layered gneisses at a high angle to the maximum elongation direction during ductile simple shear. Further, the sense of offset expressed by layer drag reflects dextral shear along the extensional crenulation cleavage as viewed northeast. As noted by Platt and Vissers (1980), this crenulation shear fabric is not expected in homogeneous or structurally isotropic rocks, such as the Okanogan dome orthogneisses, since they contain no potential markers like gneissic layering (compare with shear surfaces in orthogneisses described by Berthe and others (1979b) and Hansen (1983b)).

The model for formation of extensional crenulation cleavage in layered rocks proposed by Platt and Vissers predicts that cleavage planes will form parallel to the overall shear direction as shown in Fig. 11d. However, Okanogan dome cleavage planes dip uniformly southeast relative to the plane of overall shear in a direction parallel to mylonitic lineation. I explain this difference by a significant component of flattening strain normal to mylonitic shear planes in addition to simple shear deformation expressed by the mylonite fabric. This flattening is consistent with a tectonic model of deep crustal shear under considerable load pressure, or normal stress. A combination of simple shear and flattening normal to simple shear planes results in a model strain geometry as depicted in Fig. 12 (Ramsay, 1967, p. 57), and the combination provides for the formation of conjugate shear bands at angles to the shear planes. The shear band set at a low angle to the plane of bulk shear will dominate in formation of extensional crenulation cleavage (Platt and Vissers, 1980). Crenulation cleavage orientations shown in Figs. 7 and 11 thus result from southeast-directed simple shear deformation in conjunction with bulk subhorizontal shear and flattening.

A mylonitic fabric superimposed on the foliated and folded paragneisses leads me to divide the study area into two textural domains: the first domain at deeper structural levels to the northeast consists solely of a high-grade metamorphic fabric, and a second at higher structural levels toward the dome border consisting of superimposed metamorphic and mylonitic fabrics (Fig. 3). Mylonitic fabric in the paragneiss is continuous with that in the orthogneisses to the south, and it is exposed on the southwest-facing limb of a large gentle fold in the Okanogan mylonite zone (Fig. 3). As in orthogneisses, a shear foliation and mineral lineation define the mesoscopic mylonitic fabric in the Okanogan dome paragneisses. Eastward toward the dome interior this fabric is commonly difficult to distinguish from metamorphic textures also comprised by coplanar mineral foliation and lineation. In hornblende- and biotite-rich layered schists and gneisses, mylonitic foliation and lineation appear markedly different from the pre-existing metamorphic textures. In mylonitic rocks foliation follows a weak sigmoidal trend, whereas it is quite planar and continuous in rocks containing only a metamorphic fabric; mylonitic lineation consists of smeared, rod-shaped and elongate quartz-plagioclase-biotite aggregates, whereas the metamorphic lineation consists only of a biotite-hornblende shape preferred orientation.

In mylonitic paragneisses, subequant to elliptical feldspar grains (2-8 mm) exhibit augen-like shapes, particularly apparent where wrapped by fine-grained biotite sheaths that pinch out in tails within the foliation plane. Microscopic textures are also similar to those in the orthomylonites: feldspar grains are rounded, fractured, and float in well-formed quartz pressure shadows; biotite is very fine-grained (<0.25 mm), ragged, and attenuated into elongate streaks; quartz is finer-grained than in metamorphic rocks, showing well-formed deformation lamellae, subgrains, recrystallized grains, ribbon textures, and pressure fringes around feldspar porphyroclasts. These textures are quite distinct from the blastic mineral textures and preferred orientations shown by the paragneisses toward the dome interior. Thus, the older metamorphic fabric is increasingly deformed in the mylonite zone toward the dome margin where the mylonitic fabric dominates. West-dipping shear surfaces, "Mm" in the orthomylonites, are present locally in the mylonitized paragneisses and exhibit the same geometrical relationship with mylonitic foliation. These might be the secondary set of conjugate shear bands discussed above (Figs. 9d and 10), involving sinistral local offset looking northeast.

The geometric relationship between rock fabric and mesoscopic folds also differs. To reiterate, coplanar metamorphic foliation and lineation are folded in the metamorphic domain, resulting in weak axial-plane schistosity and mineral lineation at a high angle to fold axes. In the mylonitic domain, mylonitic foliation forms a strong axial-plane schistosity and mineral lineation lies parallel to fold axes, forming coplanar foliation and axial surfaces, as well as colinear mineral lineation and fold axes (Fig. 13).

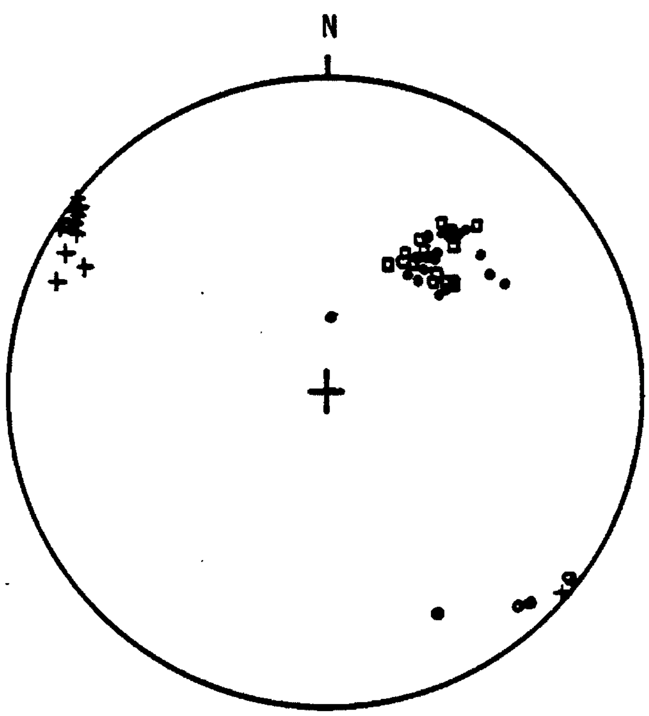


FIG. 13. Stereonet of structures in mylonitic paragneiss, showing poles to fold axial planes (open squares), poles to mylonitic foliation (closed circles), fold axes (open circles), and mylonitic lineation (crosses).

In summary, textural and structural evidence from paragneisses in Okanogan dome indicate the following brief geologic history before dome formation. High-grade metamorphism and concurrent deformation of potassium-deficient sediments formed sillimanite-zone gneisses with a strong metamorphic fabric which was later folded. Southeastward fold vergence reflects regional tectonic stresses. An extensional crenulation cleavage formed in the layered paragneisses, possibly associated with mylonitization but at relatively deep levels below the locus of mylonitic deformation. Mylonitization subsequently deformed the metamorphic rocks in a broad zone of deformation. While I cite evidence above for distinct relative ages of events, ductile processes also characterize the pattern of metamorphism and associated deformation in the paragneisses. Lack of folded layering and the presence of a well-defined mylonite fabric in the orthogneisses indicates that emplacement of granodiorite plutons post-dated metamorphism and folding, but occurred before regional mylonitization.

FOLD REORIENTATION IN THE OKANOGAN MYLONITE ZONE

From the geologic and structural history outlined above for paragneisses in the Okanogan dome, it is possible not only to define textural domains, but to determine a sense of shear within the mylonite zone using the geometry of similar-style folds. Progressive changes in orientation of fold elements upward in the mylonite zone record the transition between metamorphic and mylonitic textural domains.

I subdivided field data into three structural domains based on structural homogeneity and geographic distribution of data: Domain A includes data from the metamorphic textural domain; domains B and C each contain fold data from the mylonitic textural domain (see Fig. 14). The relative structural position of these domains in the mylonite zone is fixed by geographic distance from the dome margin and local elevation; thus, domain A represents relatively deep levels within the dome, and domains B and C lie in similar positions within the mylonite zone above domain A. Local relief between domains A and B ranges up to 700 m.

In each domain, I calculated spherical means of mylonitic foliation, mylonitic lineation, fold axial planes and fold axes, as well as values for each of α_{95}, κ, and δ, respectively the radius of a circle of 95% confidence, the precision parameter, and the angular standard deviation (Mardia, 1972; Irving, 1964). Table 1 presents a compilation of the spherical mean statistics, and equations used in calculation of spherical means are given in Appendix I. After calculating the mean directions of these structural elements for each domain, I plotted them on a lower-hemisphere equal-area stereonet with α_{95} circles of confidence (Fig. 14). In order to compare mylonitic fabric and fold orientations consistently, I rotated all fold element means such that mylonitic foliation in each domain is horizontal (Fig. 15). This manipulation of data is based on an assumption from field interpretations that the mylonite zone formed in a broadly planar fashion. I rotated the fold data in domain A, which contains no superimposed mylonitic fabric, by bringing the mean mylonitic foliation of domain B to horizontal, under the assumption that regional mylonitic

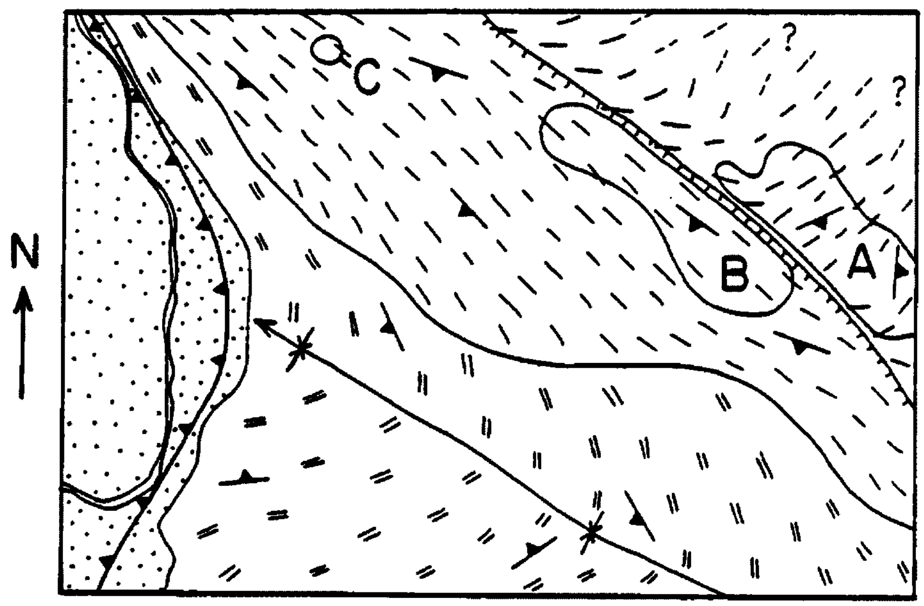


FIG. 14. Sketch map of structural domains in paragneiss, north-half, Omak Lake 15′ quadrangle. Structure and lithologic symbols same as used in Figure 3.

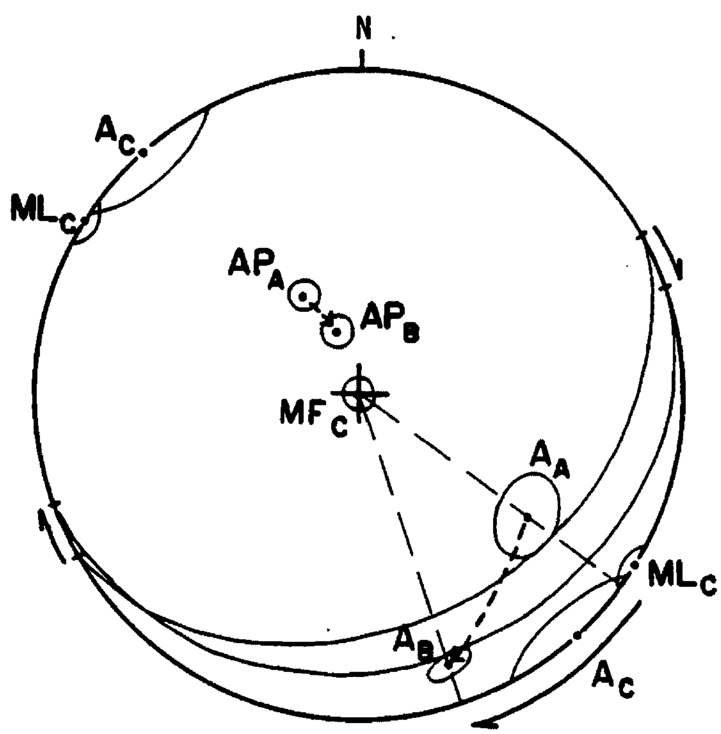


FIG. 15. Stereonet construction of fold reorientation in the mylonite zone, lower hemisphere projection. MF = poles to mylonitic foliation; ML = mylonitic lineation; AP = poles to fold axial planes; A = fold axes. Subscripts refer to structural domains A, B, and C. Ellipses show projected shape of circles of 95% confidence about mean directions.

TABLE 1. Spherical mean data and summary statistics for fold elements.

		Mylonitic foliation (MF)	Mylonitic lineation (ML)	Fold axial planes (FAP)	Fold axes (FA)
Domain A	$\overline{D}$	---	---	-1.57	91.91
	$\overline{I}$	---	---	23.50	32.96
	α_{95}	---	---	4.05	10.20
	κ	---	---	19.60	26.45
	δ	---	---	18.24	14.90
	n	---	---	66	9
Domain B	$\overline{D}$	22.73	124.75	10.32	139.21
	$\overline{I}$	42.64	14.05	30.77	44.73
	α_{95}	2.71	4.75	3.55	6.38
	κ	33.01	18.28	23.80	39.81
	δ	14.06	18.85	16.55	12.40
	n	85	52	70	14
Domain C	$\overline{D}$	38.11	-56.54	37.06	133.67
	$\overline{I}$	42.99	4.35	44.45	8.32
	α_{95}	4.23	4.31	4.29	14.30
	κ	57.49	102.22	94.27	29.58
	δ	10.44	7.68	8.02	13.36
	n	21	12	13	5

shear foliation may be extrapolated geometrically over domain A. The tight cluster of mylonitic foliation data around their mean in domain B validates this assumption. Regional geologic considerations indicate that the dip of mylonitic foliation above domain A should be less than or equal to the dip observed in domain B; therefore, the amount of fold data rotation in domain A using this method is a maximum.

To interpret the fold orientations (Fig. 15), I assume the following to be true:

1. Folding significantly preceded mylonitization, as indicated by the widespread occurrence of folds which are penetratively deformed by a mylonitic fabric only near the margins of the Okanogan dome.
2. Folded layers in the paragneisses do not exhibit ductility contrasts under mylonitic conditions, and thus act as passive markers in the mylonite zone.
3. Structural domains within the study area are texturally and structurally distinct.
4. The analysis is statistically sensitive to independent spatial changes of inclination and azimuth directions of structural elements such as fold axes and fold axial planes.

Given that the above assumptions are valid, interpretation of the fold data encompassing mylonitized and non-mylonitized domains in the paragneisses is based on the following observations (Fig. 16). Population means of fold data from domains A, B, and C are statistically discernable at the 95% confidence level. Both fold axes and fold axial

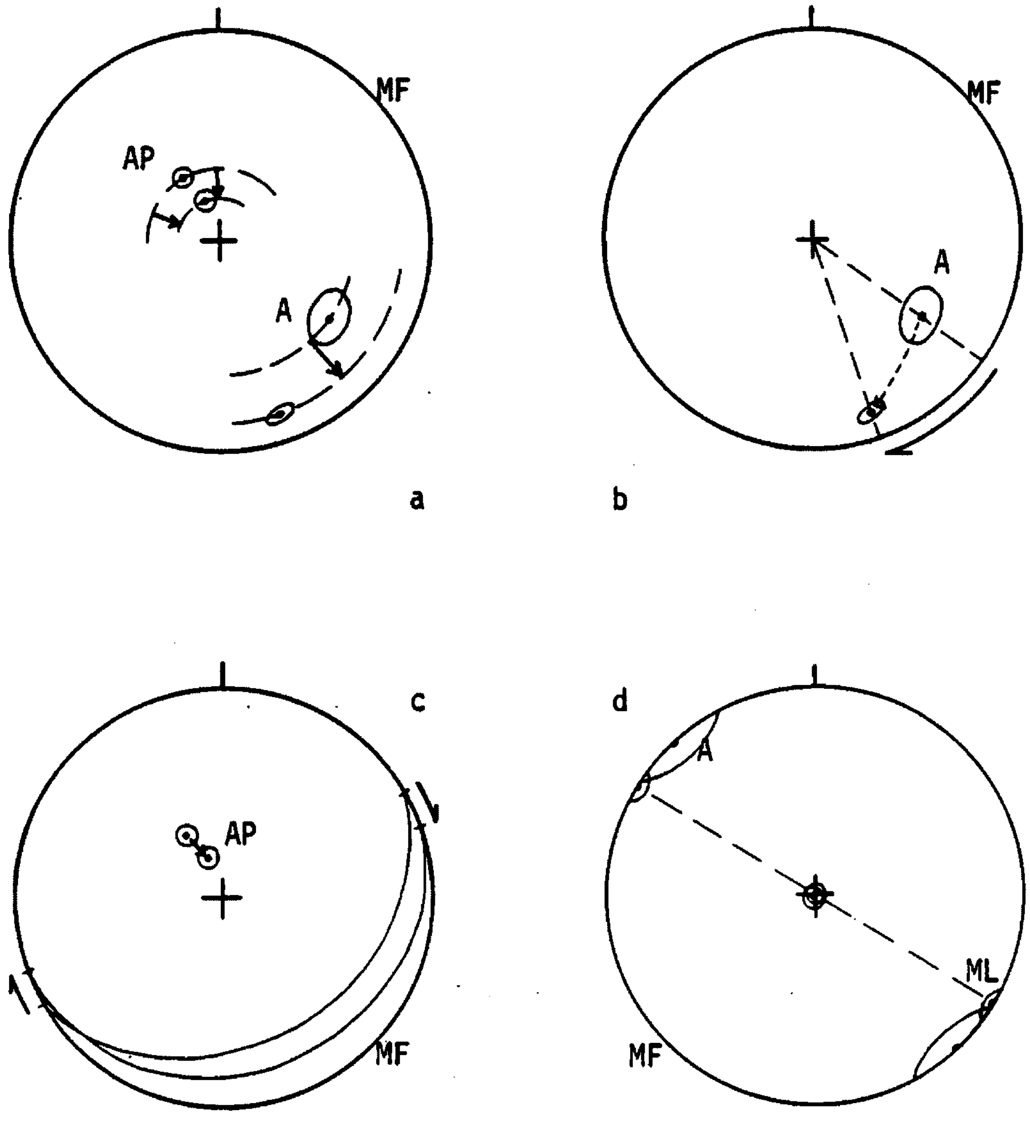


FIG. 16. Components of fold reorientation with progressive mylonitization upward in the mylonite zone from domains A to B, as shown in Figure 15. (a) Flattening of fold axial planes and fold axes toward mylonitic foliation (MF). (b) Clockwise rotation of fold axis trends. (c) Clockwise rotation of strike directions of fold axial planes. (d) Statistically coplanar mylonitic foliation and fold axial planes, and statistically colinear mylonitic lineation and fold axes, domain C.

planes show flattening toward mylonitic foliation, which in this analysis is rotated to a horizontal position, from domain A to domain C (Fig. 16a). Fold axes show bulk clockwise or southward rotation about the stereonet azimuth from domain A to domain B, and a reverse sense of counterclockwise rotation from domain B to domain C (Fig. 16b). The strike of poles to axial planes, with the exception of domain C which has no strike to its statistically vertical pole position, also shows bulk clockwise or eastward rotation from domain A to domain B, although less than that shown by fold axes (Fig. 16c). Domain C exhibits statistically colinear fold axes and mylonitic lineation, and statistically coplanar fold axial planes and mylonitic foliation (Fig. 16d). These geometric relationships indicate a simultaneous flattening and passive clockwise rotation of fold elements, compatible with flattening and dextral simple shear in a zone of ductile mylonitization, as viewed northeast.

The conclusion that reoriented folds in the Okanogan mylonite zone describe clockwise rotation and flattening in a top-to-the-east shear environment depends primarily on kinematic analysis of data from domains A and B. Orientation of fold axes in domain C, if considered, show a possible opposite sense of rotation from domain B. I did not consider the data from domain C in this part of the analysis, however, for two reasons: (1) domain C is geographically distant from domains A and B, making projection of data from domain C along strike to domains A and B possibly an invalid method of juxtaposing domains in the sense that domains A and B are joined, and (2) folds in domain C are considerably larger structures than in the other two domains, and possibly great internal measurement variance must be considered. Nonetheless, textures in domain C indicate that extreme mylonitic strain results in flattening of folds into mylonitic shear foliation and eventual colinearity of fold axes, with the direction of tectonic transport parallel to mylonitic lineation.

The interpretation that clockwise rotation of fold axes from domains A and B indicates top-to-the-east shear in the Okanogan mylonite zone depends on the original orientation of folds with respect to the regional sense of shear. All fold axes plunge southeast, but, more importantly, the trend of fold axis means from domains A and B lie south of the mean mylonitic lineation trend in this part of the Okanogan dome. Given simple shear deformation and associated overall flattening, which appears to be a reasonable model of strain from textural and structural criteria, passive folds must remain confined at most to the original thickness of sheared rock within which they occur; that is, they will not "tumble" within the mylonitic shear fabric as would a rigid-body porphyroclast of feldspar. Thus, a linear element at a slight angle from the trend of the bulk shear direction may initially rotate away from that net transport direction (Fig. 17). At higher levels of shear strain, most linear elements within the shear zone will trend toward parallelism with the net transport direction, even those originally at a high angle to the shear direction (Escher and Watterson, 1974; Ramsay, 1980). In Fig. 17a, the circled line plunges in the direction of shear, but its azimuth direction lies at a slight angle to the shear direction. At moderate levels of shear strain, as shown by the strain ellipse in Fig. 17b, the bearing of the same line is rotated to a greater angle from the shear direction. The rotation of lines away from the bulk shear direction is shown in a plan view of the upper surface of

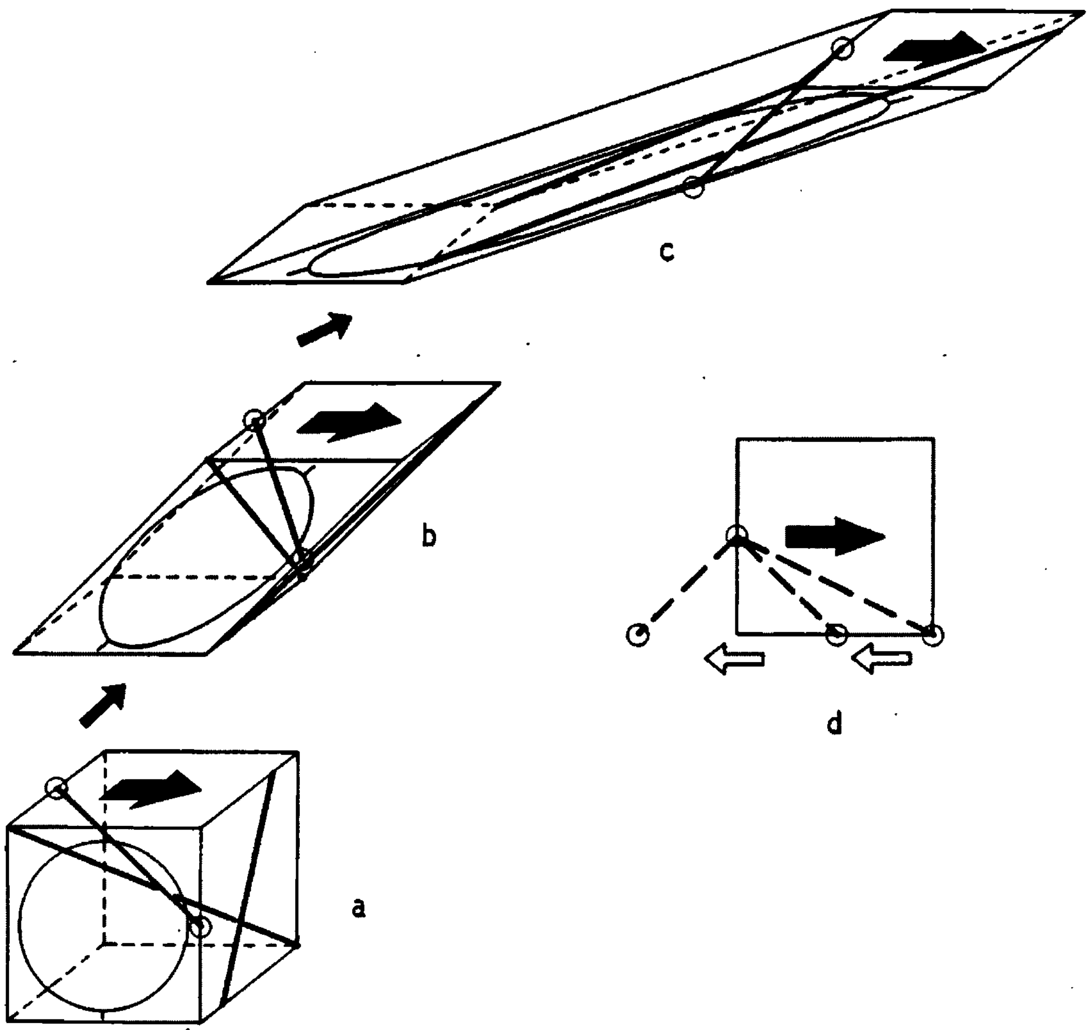


FIG. 17. Model of fold reorientation during progressive simple shear, from an undeformed cube containing several randomly oriented lines. Heavy arrows show direction of overall simple shear; strain ellipses drawn for each phase of the deformation, shown by (a), (b), and (c). (d) Plan view of the cube upper surface, showing rotation path of circled line away from shear direction.

the deformed cube (Fig. 17d), where the bold lines show the clockwise rotation in trend of the circled lines in the block diagrams. Continued deformation would result in increasing parallelism between this and other linear elements, toward colinearity of all lines with the extension direction (Fig. 17c). Completely random original linear elements will never become exactly parallel but will increase in relative colinearity toward the limiting stages of total shear strain (Skjernaa, 1980). Stages of this model shown in Figs. 17a and 17b depict the progressive reorientation of paragneiss fold axes in domains A and B. Such a model applies to progressive reorientation of planar structures equally well. Marker planes dipping gently with respect to the planes of shear will undergo only slight strike direction reorientation, as seen in the small amounts of paragneiss fold axial-plane rotation relative to fold axis rotation (Fig. 15). Examples of naturally-occurring cases of fold reorientation in ductile shear zones are presented by Bak and others (1975), and Bell (1978).

The patterns developed in this type of analysis clearly show that fold orientations in paragneisses of Okanogan dome change progressively with respect to mylonitic fabric. Additional evidence corroborates the conclusion that mylonitization affected the structure and orientation of these folds. Interlimb angles of folds decrease upward from an average of 43 to 15 degrees in the mylonite zone, indicating that mylonitization involved a significant amount of flattening and shape attenuation. In handsamples, metamorphic mineral lineation originally at a high angle to fold axes in domain A gives way to a moderate mineral lineation - fold axis angle in domain B, and finally, parallel structures in domain C. Similarly in thin section, fold hinges in domain A are characterized by recrystallized polygonal arcs of biotite, grading texturally upward to domains B and C, the latter characterized by a prominent axial-plane schistosity.

In summary, field data consisting of fold and mylonitic fabric orientations from paragneisses exposed in the northern half of the study area were used in a kinematic assessment of the Okanogan mylonite. Assuming that the folds acted as passive structures within a mylonitic fabric characterized by simple shear deformation, progressive fold reorientation occurred by simultaneous flattening and clockwise rotation of folds as viewed onto the plane of mylonitic shear foliation. Presumably flattening did not affect the direction of rotation, but it may have increased the rate of rotation into the mylonitic foliation plane. Such rotation appears compatible with a top-to-the-east sense of shear within the mylonite zone, which is a net transport direction consistent with that indicated by numerous other textural interpretations throughout the southwest portion of Okanogan dome (this paper; Hansen, 1983b; Goodge and Hansen, in press).

QUARTZ C-AXIS MICROFABRICS

Okanogan dome paragneisses contain evidence of successive periods of metamorphism and deformation, including early regional high-grade metamorphism, folding, and mylonitization. Orthogneisses of the Okanogan dome, as described above, contain textural and compositional features indicative of a plutonic origin. Moreover, they show no evidence of involvement in sillimanite-zone regional metamorphism or folding; they are predominantly granodiorite in composition, and numerous aplite and pegmatite dikes appear undeformed except where they are sheared in the mylonitic fabric. The major orthogneiss unit, K-feldspar-biotite granodiorite, is visibly sheared through a major section of Okanogan mylonite, and it is an ideal indicator of quartz c-axis preferred orientations formed during mylonitization. The quartz microfabric in these plutonic mylonites is related to conditions of mylonitization in Okanogan dome, and by comparison with fabric patterns reported from other studies I am able to conclude that ductile mylonitic deformation involved easterly-directed progressive simple shear in the plane of foliation.

A sequence of oriented samples collected in the K-feldspar megacryst granodiorite unit covers, from northeast to southwest, a structural section upward through the Okanogan mylonite zone from weakly to highly deformed plutonic rocks (see Fig. 3). Oriented thin sections were cut perpendicular to mylonitic foliation in two directions, parallel and perpendicular to mylonitic lineation. I measured c-axis orientations of quartz grains on a petrographic microscope equipped with a four-axis universal stage. Traverses were made across slides at 1 mm intervals, and all grains passing beneath the cross-hairs were measured in order to obtain a larger number of points in the quartz-poor granodiorite. Only quartz grains of the granodiorite matrix were measured, and only one section (203N) contained significant portions of a coarse K-feldspar megacryst. Quartz grains of all sizes were included and not discriminated on the basis of size. In general, I measured single subgrains from coarse quartz grains, unless two or more subgrains displayed greatly different extinction positions. Single recrystallized grains judged representative of a group were measured where possible, yet very fine recrystallized grains were commonly too small for accurate measurement. In several sections 250 c-axes were measured, but in weakly deformed rocks which contain generally coarse grain sizes (0.5-1.0 mm) I was able to measure only fewer than two hundred grains. Miller and Christie (1981) discuss the significance of collecting as few as 200 data. C-axis measurements are plotted on lower-hemisphere equal-area stereonets (Fig. 18), and contoured using the Schmidt method (Turner and Weiss, 1963, p. 61). Mylonitic foliation Mc, mylonitic lineation Ml, and mylonitic shear surfaces Mm are shown on all fabric diagrams for which there is data. The angle between Mc and Mm is the mean of angles measured directly in thin sections cut parallel to mylonitic lineation.

For the series of fabric diagrams shown in Fig. 18, mylonitic fabric is most strongly expressed texturally at location 203N and grades downward through the shear zone to 210N, where mylonitic fabric is barely perceptible. The general pattern shown in the fabric diagrams substantiates this trend of increasingly weak fabrics with structural depth; the quartz fabric is very strong at 203N and becomes increasingly diffuse toward 210N, which displays a nearly random distribution of c-axis orientations. The lack of a recognizable preferred orientation at 210N also indicates a relatively structureless rock texture prior to mylonitization, confirming that granodiorite intruded between regional folding and mylonitization.

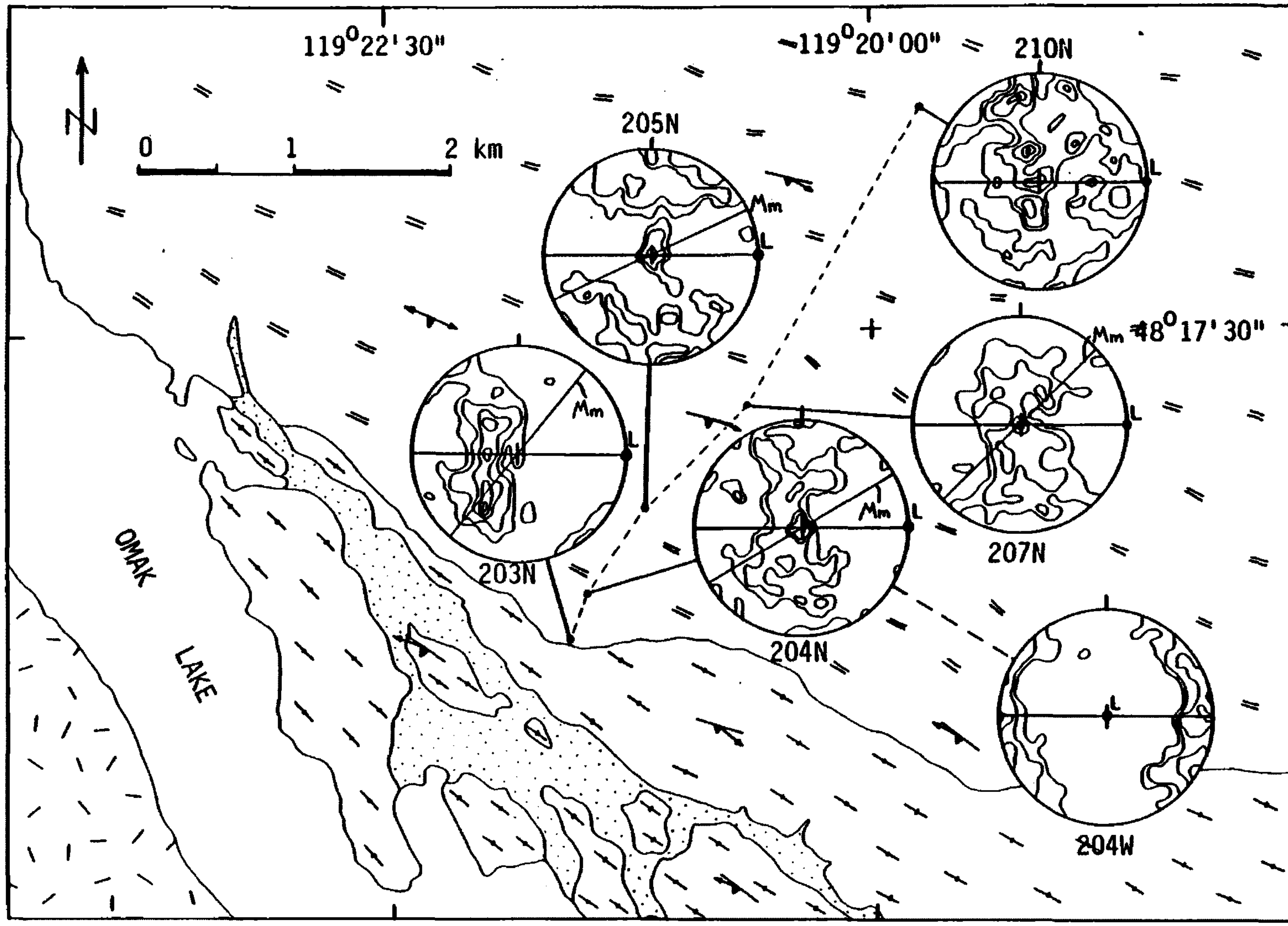


FIG. 18. Sketch map of locations and fabric diagrams for quartz c-axis samples in K-feldspar megacryst granodiorite mylonite. Map symbols same as used in Figure 3; other mylonitic orthogneisses shown by crossed-line pattern. All fabric diagrams lower hemisphere stereographic projections; L = mylonitic lineation; Mm = mylonitic shear surfaces.

Most of the quartz fabric diagrams (204N, 205N, 207N) display a weakly-formed crossed-girdle pattern similar to those described from naturally and experimentally deformed quartzites by Christie (1963), Sylvester and Christie (1968), Bouchez and Pecher (1976), Tullis (1977) and Compton (1980)(Figs. 18 and 19a). The two girdles intersect along a line in the plane of foliation, perpendicular to mylonitic lineation L. Diagram 203N shows a single girdle oriented approximately normal to both mylonitic foliation and lineation, and it resembles quartz fabric diagrams from rocks in natural shear zones described by a host of authors (Eisbacher, 1970; Hara and others, 1973; Bouchez and Pecher, 1976; Laurent and Etchecopar, 1976; Carreras and others, 1977; Burg and Laurent, 1978; Berthe and others, 1979a, 1979b; Compton, 1980; Simpson, 1980).

More apparent than the actual type of fabric pattern, is the relative symmetry of the girdle(s) with respect to mylonitic foliation Mc, the plane of shear. Several diagrams projected parallel to the direction of mylonitic lineation contain crossed girdles which resolve to a single central girdle oriented perpendicular to mylonitic foliation and lineation. This symmetry with respect to shear planes is maintained even as girdles fan out toward the top of the diagrams.

Symmetry of the quartz c-axis girdles with respect to mylonitic foliation is also shown in diagram 204W, which is oriented with lineation normal to the plane of projection (Fig. 18). The pattern in 204W further reinforces the sense of a strong central girdle shown in 204N.

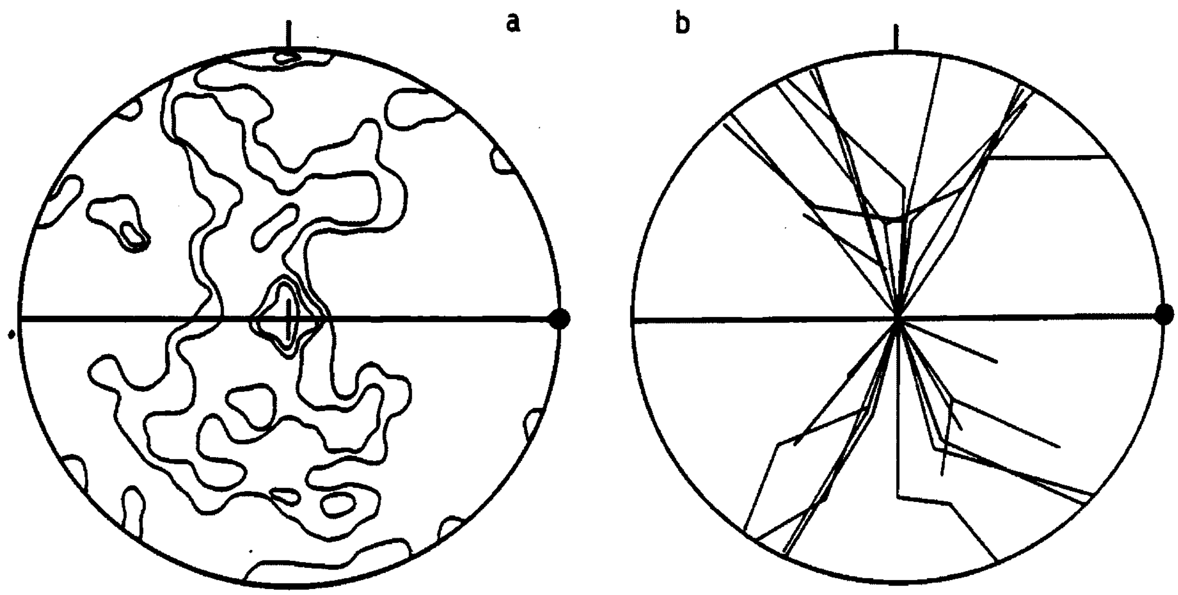


FIG. 19. Crossed-girdle c-axis fabric patterns in granodiorite mylonite. (a) Contoured crossed-girdle fabric pattern; sample 204N. (b) Synoptic skeletal diagram of contoured fabric diagrams shown in Figure 18 (after Lister and Williams, 1979). Horizontal line shows trace of mylonitic foliation; dots show mylonitic lineation direction.

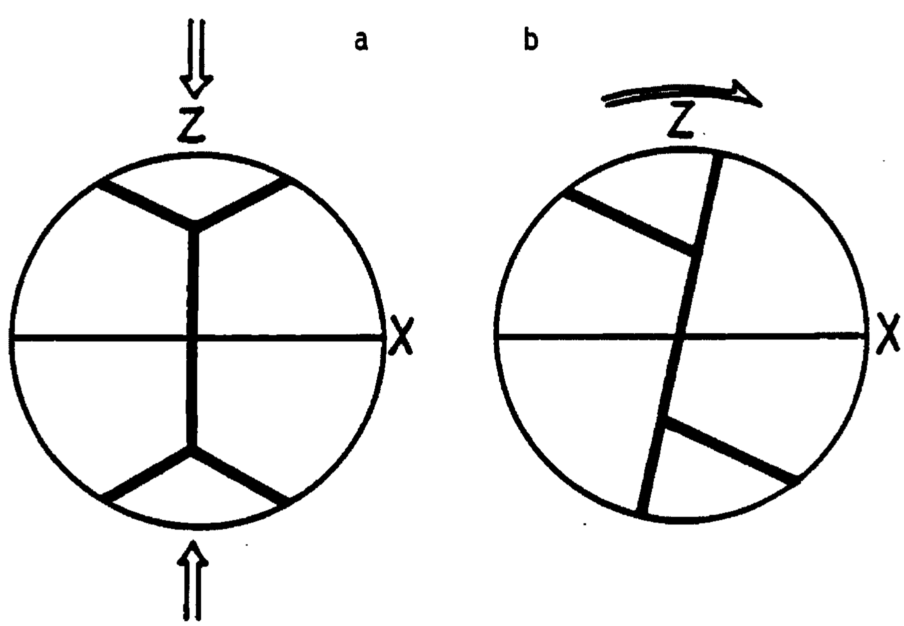


FIG. 20. Interpretation of crossed-girdle c-axis patterns in quartz mylonites (from Behrmann and Platt, 1982). Skeletal outlines of crossed-girdle fabrics resulting from (a) coaxial compression perpendicular to foliation plane, and (b) non-coaxial simple shear parallel to foliation plane.

A synoptic skeletal fabric diagram (Lister and Williams, 1979; Behrmann and Platt, 1982) combines individual c-axis preferred orientation patterns, and is a convenient method of determining general symmetry elements in the fabric patterns of deformed rocks. Skeletal diagrams are simply constructed by connecting peaks and ridges of individual contoured c-axis diagrams. A synoptic skeletal diagram of the Okanogan mylonite quartz c-axis patterns (Fig. 19b), shows the fabric patterns of these rocks include a single straight girdle inclined in a clockwise direction and shorter "legs" projecting to the northwest and southeast quadrants. This synoptic skeletal diagram thus shows the mylonite quartz fabrics to be asymmetric crossed-girdle, or type I crossed-girdle patterns as used by Lister and Williams (1979).

At four locations (204N, 205N, 207N, and 209N), Mm shear planes dip westward with respect to shear foliation. Mean angles between Mm and Mc decrease with increasing mylonitization, from 209N to 204N, but this general trend appears to be independent of the relatively constant angle between c-axis girdles and mylonitic foliation.

The current literature contains two generally distinct approaches to the interpretation of quartz c-axis preferred orientations. On the one hand, many modelers and experimentalists cite evidence for the formation of symmetric crossed-girdle fabric patterns during plane strain (Fig. 20a), and asymmetric crossed-girdles inclined with the sense of shear in simple shear regimes (Fig. 20b)(Christie, 1963; Sylvester and Christie, 1968; Riekels and Baker, 1977; Tullis, 1977; Lister and others, 1978; Lister and Price, 1978; Compton, 1980; Lister and Hobbs, 1980; Lister and Williams, 1980; Miller and

Christie, 1981; Behrmann and Platt, 1982). On the other hand exist students of naturally deformed rocks who document c-axis fabrics in conjunction with presumably independent methods of determining sense of movement in shear zones, and who promote an asymmetrical relationship betwen single girdles and shear directions such that the girdles are inclined sympathetic to the sense of shear (Eisbacher, 1970; Bouchez and Pecher, 1976; Laurent and Etchecopar, 1976; Carreras and others, 1977; Burg and Laurent, 1978; Berthe and others, 1979a, 1979b; Simpson, 1980; Mattauer, 1981). Most authors from these two camps appear to agree that single- or crossed-girdles symmetrically related to shear planes in deformed rocks reflect overall flattening perpendicular to the plane of shear (see Tullis, 1977). This situation most reasonably fits a model of coaxial plane strain, which implies bulk lateral shear in all directions of the shear plane. These same workers appear to agree as well, that asymmetric c-axis girdle patterns provide a means of interpreting shear directions regardless of the actual pattern, whether of a crossed- or single-girdle type.

Lister and others (1978a, 1978b, 1980a, 1980b) base their conclusions on computer simulations of polycrystalline quartzite deformation under different conditions, including coaxial and non-coaxial deformation. Combining the various slip systems of quartz into several model quartzites, Lister's group determines theoretical quartz c-axis preferred orientations for deformation paths including components of coaxial extension, coaxial shortening, and non-coaxial simple shear. The results of work by Lister's group show that asymmetric crossed-girdle patterns develop as a result of non-coaxial simple shear, and that the pattern asymmetry indicates an overall sense of shear (G. Lister, pers. comm., 1983). Experimental studies by Tullis (1977) primarily consider coaxial compression and plane strain of artificial and natural quartzites. The quartz microfabric patterns resulting from plane strain experiments appear as symmetrical girdles with respect to shear direction, in agreement with results of the Lister coaxial deformation models. These studies constitute the theoretical and experimental framework for the interpretation that c-axis girdles formed by simple shear deformation should lie inclined in the direction of shear.

A battery of fabric studies in natural rocks, mostly from European shear belts, combine analysis of regional structures, microtextural study, and quartz c-axis preferred orientation analysis in order to derive a shear sense interpretation for observed quartz fabrics. Structural methods for defining a sense of shear are, of course, specific to local geologic occurences, but once a shear sense is determined, it may be empirically compared with the quartz microfabric. Results of most studies indicate that single girdles or weak crossed-girdles are inclined with the direction of bulk simple shear (see Burg and Laurent (1978), Berthe and others (1979a), and Mattauer and others (1981) for examples). The conclusion of workers studying naturally deformed rocks that single- or crossed-girdle patterns lie inclined sympathetic to the sense of shear is thus in agreement with the results of experimental and theoretical analysis.

By comparing quartz c-axis patterns of Okanogan mylonites to patterns in published microfabric studies, I draw several conclusions concerning their kinematic significance. (1) The overall fabric pattern in the Okanogan mylonites is characterized by a broad girdle perpendicular to the plane of mylonitic foliation (Fig. 19a). (2) Construction of a synoptic skeletal diagram from contoured c-axis patterns shows that the fabric is more correctly characterized as an asymmetric crossed-girdle pattern, with a single straight girdle inclined to the east (Fig. 19b). (3) Based on interpretations presented by Lister and Williams (1979), Behrmann and Platt (1982), and Lister (pers. comm., 1983), I conclude that the asymmetric crossed-girdle pattern of the Okanogan quartz mylonites results from non-coaxial progressive simple shear and a dextral or top-to-the-east bulk shear direction. This conclusion agrees with interpretations presented in the structural analysis of folds above, but disagrees at this time with the conclusions reached by Hansen (1983b), who suggests that mylonitic microstructures indicate a top-to-the-west sense of shear.

DISCUSSION

A general summary of structural and petrologic events affecting rocks of Okanogan dome before dome formation includes metamorphism, folding, intrusion, and mylonitization. The most pertinent earlier studies in the area include work by Snook (1965), Fox and others (1976, 1977), Hansen (1983b), and Goodge and Hansen (in press). Snook's sequence of events in Okanogan dome involves sequential metamorphism, intrusion, and ductile mylonitization, a sequence which agrees well with the results of this study. Snook interprets the Okanogan mylonite as a low-angle distributed thrust fault, and he proposes that this structure pre-dates brittle faulting at the dome margin. Fox and others (1976, 1977) interpret metamorphism, igneous activity, and regional cataclasis as all occuring during diapiric rise of hot high-grade metamorphic and igneous rocks. Cataclasis of this rising mass occurred during shear along its outer boundaries against lower temperature country rocks. Thus, these authors view diapiric rise, ductile spreading, and cataclasis as a continuous process during metamorphism and intrusion of rocks within Okanogan dome. Results of this study agree generally well with Snook's model of deformation and metamorphism in gneisses of Okanogan dome, and are similar to the conclusions proposed by Hansen (1983b), and Goodge and Hansen (in press). Expanding on Snook's interpretations, I believe there is strong evidence to indicate that distinct early phases of deformation occurred in gneisses of Okanogan dome; metamorphism followed in the waning stages by folding was, in turn, followed at a significantly later time by the crustal shear event which formed the Okanogan mylonite. Evidence that metamorphism and mylonitization did not occur simultaneously includes: (1) differences in temperature of metamorphism and mylonitization of greater than 100°C, as indicated by sillimanite-zone metamorphic assemblages and mylonite feldspar geothermometry, (2) intrusion of granodiorite plutons which contain numerous non-folded but mylonitized dikes, indicating intrusion intervened between metamorphism and mylonitization, thereby picking up only mylonitic textures, (3) mesoscopic and microscopic structures show that an early metamorphic paragneiss foliation and lineation were later folded, although such folds contain a weak axial-plane schistosity; as they were progressively mylonitized an increasingly strong axial-plane schistosity formed, (4) microscopic fabrics also indicate a gradational transition upward in the mylonite zone from relatively coarse-grained metablastic mineral textures, to fine-grained quartz recrystallization and brittle feldspar behavior in the same metamorphic rock units, (5) fold-shape compression and fold-element reorientation both suggest gradually increased shear deformation of a ubiquitous fold fabric toward the margin of Okanogan dome, indicating mylonitization of pre-existing structures.

I summarize the interpretations of this study by fitting them into a broader tectonic framework, in the hope of drawing similarities between these and the structural interpretations of other crystalline terranes of the northern Cordillera. Figure 21 presents a cartoon sketch of the progressive development of rocks and structures interpreted for Okanogan dome.

The oldest rocks of southwestern Okanogan dome form a sequence of metasedimentary gneisses consisting mostly of hornblende-biotite-plagioclase layered gneiss, with minor amounts of amphibolite, sillimanite-bearing pelitic schist, and diopside-bearing calc-silicate gneiss. The lithologic variation of the gneisses indicates a sedimentary parent rock, and the potassium-deficient composition of these units reflects a clastic depositional environment in the vicinity of a continental margin volcanic arc within an ancient eugeocline (Fig. 21a). The volcanic arc itself would provide the greatest source of sediment in this environment, and may explain the monotonous repetition of hornblende-plagioclase gneiss and calc-silicate

FIG. 21a. Westward subduction under volcanic arc.

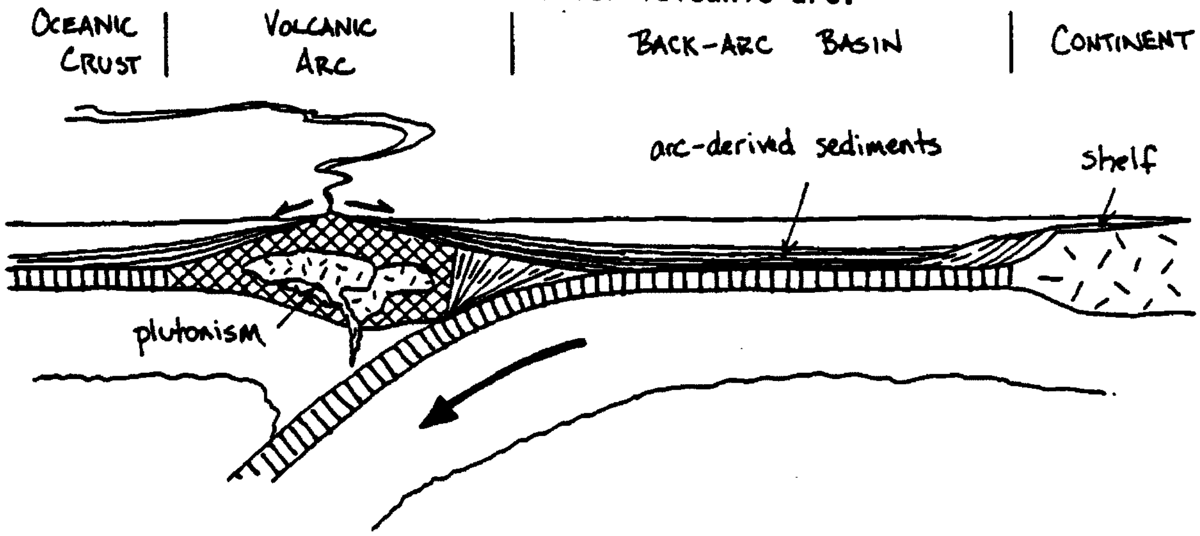


FIG. 21b. Eastward subduction and basin closure.

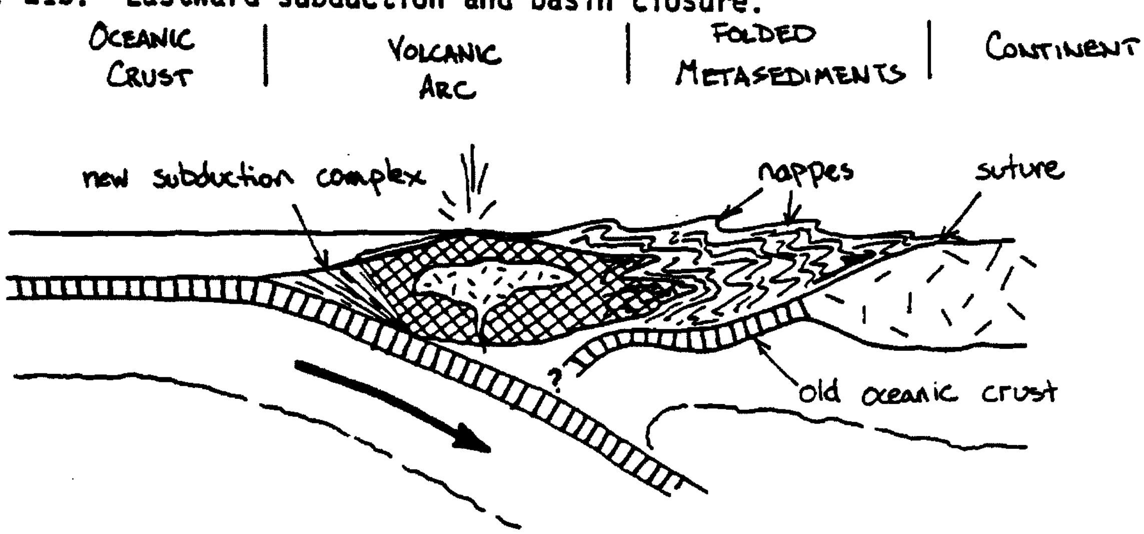


FIG. 21c. Obduction of volcanic arc and convergent margin magmatism.

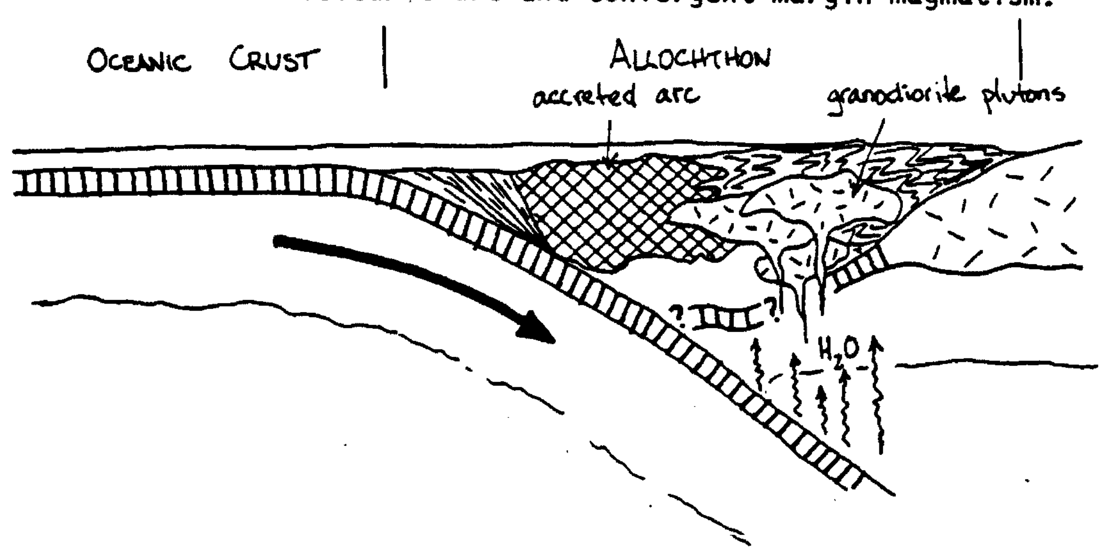


FIG. 21d. Deformation of allochthonous rocks and formation of crustal shear zones.

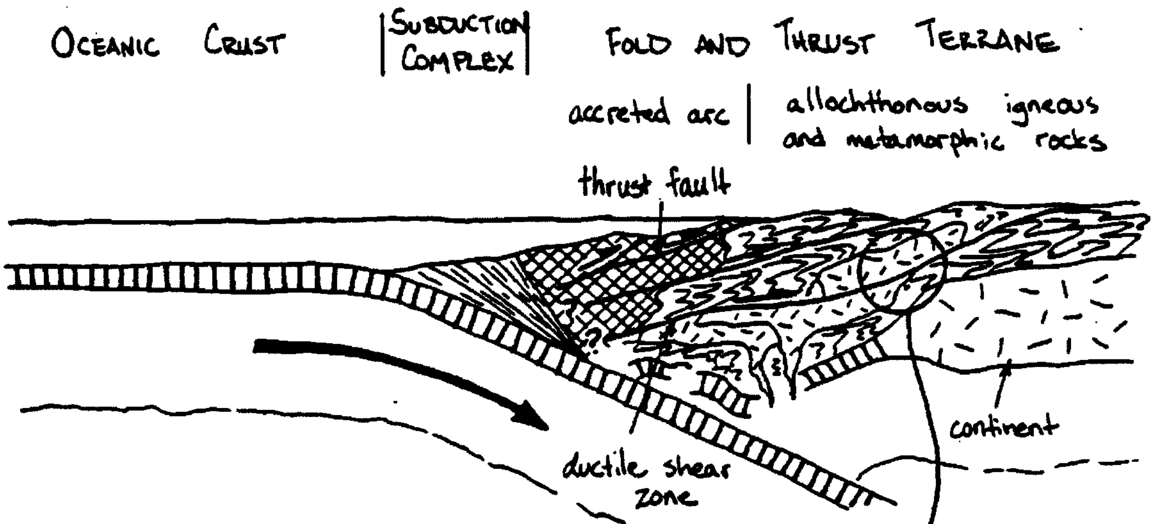


FIG. 21e. Eastward displacement of allochthonous rocks along mylonite zone.

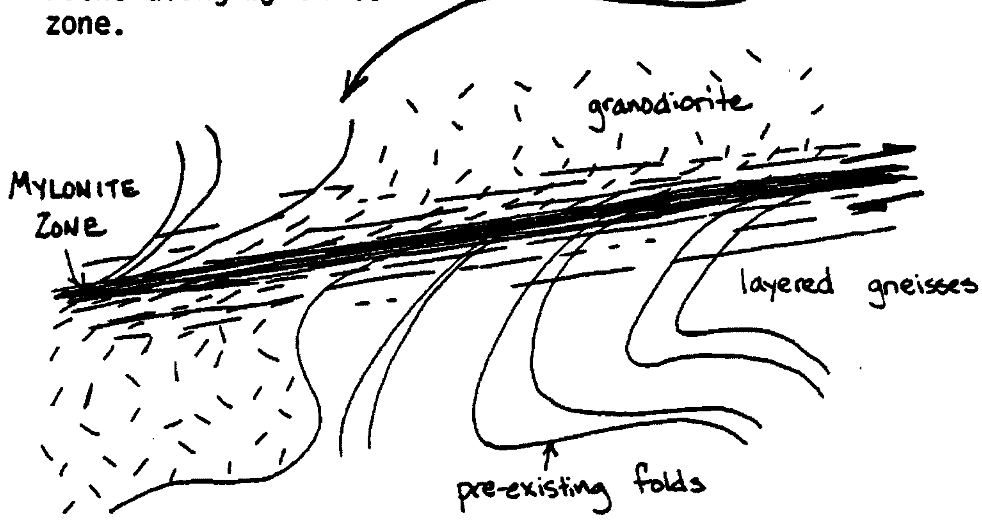
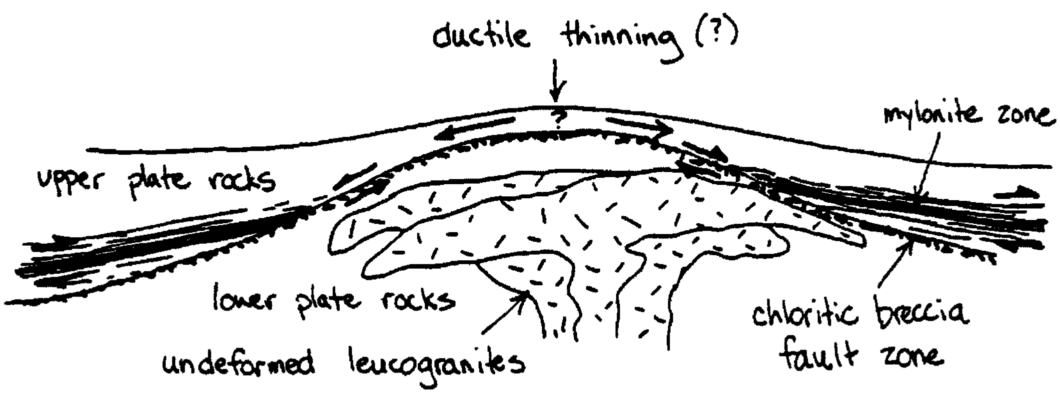


FIG. 21f. Structural doming of mylonite zone and tectonic denudation along brittle fault zones.



gneiss throughout Okanogan dome (Fox and others, 1976, 1977). High-grade regional metamorphism in the sillimanite zone of the amphibolite facies indicates that sedimentary parents of Okanogan paragneisses were deposited continentward of, or possibly within, the volcanic arc. High-temperature isotherms on the trench side of a volcanic arc plunge deeply beneath the arc and preclude extensive high temperature metamorphism of arc-derived sediments. A deep back-arc sedimentary basin may have developed in a marginal sea between arc and continent after Late Precambrian or Early Paleozoic rifting of the North American craton (Tempelman-Kluit, 1979). Formation of a Late Triassic island-arc system along the western edge of North America is also suggested by Tempelman-Kluit (1979), and Mattauer and others (1983), who cite evidence for an oceanic arc separated from continental crust by a rift-generated oceanic basin and built upon a west-dipping subduction zone. The arc-derived sedimentary rocks of Okanogan dome are similar to rock types comprising the Quesnellia allochthonous terrane of the Canadian Cordillera (Monger and others, 1982). By comparison it appears likely that metasedimentary rocks of Okanogan dome represent volcanic sediments shed from an oceanic arc in the core of a Late Paleozoic to Early Mesozoic allochthonous terrane developed over a west-dipping subduction zone.

A weak to nonexistant axial-plane schistosity in the hinges of mesoscopic folds in Okanogan paragneisses is evidence for late syn-metamorphic to post-metamorphic deformation. The dip of axial planes in these folds is regionally shallow if the effects of later doming are removed, indicating possible relation to broad flat-lying nappe structures. Furthermore, while many mesoscopic folds show ambiguous vergence, several examples show vergence generally to the east (Fig. 21b), reflecting vergence of large-scale folds. Shallow-dipping and east-vergent fold nappes reflect easterly-directed tectonism during or after the peak of regional metamorphism. The grade of metamorphism, style of folding, and vergence sense are compatible with numerous generally east-vergent structures in the Shuswap metamorphic complex (Wheeler, 1965; Hyndman, 1968; Fyles, 1970; Reesor, 1970; Reesor and Moore, 1971; Brown and Tippett, 1978; Read, 1980; Read and Brown, 1981; Read and Klapacki, 1981). The frequently reported style of combined high-grade metamorphism and folding in the Shuswap terrane is now widely believed to be the result of middle Jurassic accretion of the Quesnellia allochthonous terrane to the North American continental margin, which it may have partly overridden (Fig. 21b)(Tempelman-Kluit, 1979; Monger and others, 1982; Brown and Read, 1983; Mattauer and others, 1983).

Intrusion of massive hornblende-biotite granodiorite plutons into Okanogan paragneisses must have post-dated the peak of regional metamorphism and folding, because the rocks display primary igneous textures, are unmetamorphosed, and contain numerous sheeted planar pegmatite and aplite dikes which probably intruded in the latest stages of emplacement of the plutons (Fig. 21d). Nonetheless, it seems likely that the derivation and emplacement of these plutons is tectonically related to generation by orogenic metamorphism and folding which pre-dated intrusion. Although on textural grounds the plutons appear to be post-orogenic intrusions, mid-Jurassic metamorphism and deformation in the Canadian Cordillera is associated with widespread plutonism (Hyndman, 1968; Pigage, 1977; Tempelman-Kluit, 1979; Read and Brown, 1981; Brown and Read, 1983), including the Kuskanax batholith in British Columbia with a 171 m.y. age (Brown and Read, 1983). The granodiorite to tonalite plutonic phases in the Okanogan dome have compositional affinities with a marginal arc tectonic environment and may have been generated by partial melting of oceanic or transitional crust. It is unlikely that the granodiorite magmas were generated from continental crust as a result of tectonic loading by overthrust allochthonous terranes because they were emplaced prior to intracontinental crustal shearing. The relative age of emplacement does appear to be bracketed between metamorphism and mylonitization, however, indicating that the plutons may be syn-metamorphic intrusions, and certainly no younger than mylonitization.

Post-plutonic penetrative mylonitization in a wide zone is related to lower- or middle-crustal ductile shear through crust of the composite continental margin. Mylonite formed as a result of deep-seated thrusting or decollement in front of the leading edge of a suture between a westward microplate, Quesnellia, and the North American plate (Tempelman-Kluit, 1979; Read and Brown, 1981; Monger and others, 1982; Brown and Read, 1983; Hansen, 1983b; Mattauer and others, 1983)(Fig. 21d). A unidirectional mylonitic lineation in the Okanogan dome, as well as parallel corrugated "a"-type folds in the mylonite fabric, transverse to the northern trend of mylonite zones in Washington and British Columbia, indicates that the mylonite zone formed as a relatively flat and subhorizontal distributed thrust (Brown and Read, 1983; Mattauer and others, 1983). Although the mylonite zone probably dipped regionally westward, textural and structural evidence indicates that upper plate rocks moved eastward over lower plate rocks now exposed in Okanogan dome (Goodge, 1983; Goodge and Hansen, in press)(Fig. 21e). Analysis of reoriented folds in this study supports an interpretation of eastward displacement along the Okanogan mylonite zone, as do asymmetric quartz c-axis preferred orientation patterns. Similar arguments are presented for relative eastward displacement of upper plate rocks along mylonite zones in the Monashee Complex (Brown and Murphy, 1982; Mattauer and others, 1983), the Spokane dome (Rhodes, 1983), and the Bitterroot dome (Hyndman, 1980). Top-to-the-east transport within the Okanogan mylonite zone placed high-grade gneisses now exposed to the west of the Okanogan dome over equivalent-grade rocks within the dome. The evidence presented in this paper supports interpretations proposed by earlier workers that mylonite zones flanking metamorphic core complexes in the northern Cordillera represent intracontinental ductile thrust faults resulting from eastward compression during mid-Jurassic accretion of composite allochthonous terranes. Based on the age of cross-cutting plutons, movement on mylonite zones in the region probably ended by about 160 m.y. ago (Brown and Read, 1983). Fabric and structural studies by Brown and Murphy (1982), Hansen (1983b), and herein indicate clearly that mylonitization involved ductile crustal shear strain and did not result from regional extension, as has been proposed for mylonite zones of other metamorphic core complexes (Davis and Coney, 1979; Davis, 1980; Davis and others, 1980). Mylonite zones along the flanks of the Monashee Complex to the north may or may not be directly connected with the Okanogan and Kettle dome mylonite zones, but as suggested by others, it is likely that these extensive shear zones are part of a system of low-angle thrust faults which emerge far to the east as the Northern

Rocky Mountain fold and thrust belt (Brown and Read, 1983; Mattauer and others, 1983). Readers should note that a top-to-the-east interpretation of movement in the mylonite zone disagrees with revised conclusions concerning microstructures (Hansen, 1983b), and that further study of the Okanogan mylonites is certainly warranted.

Post-mylonitization uplift and doming of high-grade crystalline terranes in the Omineca crystalline belt in Eocene time is related to anomalously high thermal conditions within tectonically telescoped and thickened continental crust. Uplift probably was influenced by the emplacement of large quantities of leucocratic granites (Fox and others, 1976, 1977), which helped carry the Okanogan mylonite zone upward to form a dome (Fig. 21f). Associated with thermal rise of high-grade crystalline rocks to form the Okanogan dome, tectonic denudation of cooler cover rocks to the west over rocks now exposed in the dome occurred along low-angle, brittle normal faults. Brecciation of mylonite along these brittle fault zones indicates they formed significantly later than the mylonitic shear fabric (Goodge and Hansen, in press). Weakly lineated Eocene plutons of the Okanogan dome and hypabyssal intrusives of the Republic graben probably were deformed during tectonic denudation of Okanogan dome in a locally ductile environment created by their own heat. In addition, I conclude that dome formation and brittle faulting are relatively local features formed by thermal processes, whereas shear zones such as the Okanogan mylonite formed in direct response to regional microplate accretion from the west and crustal orogenesis.

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APPENDIX I

Mardia (1972) presents a method for calculating spherical mean directions of linear data which I used in the statistical analysis of fold orientations. Assuming valid application of a Fisher distribution (Fisher, 1953), Irving (1964) describes parameters which, when calculated, provide a means of determining variance from a spherical mean direction. I summarize methods of calculating spherical mean directions and summary statistics for the directional data below.

1. Calculation of spherical mean direction. From field measurements D is the angle of declination (bearing) and I is the angle of inclination (plunge) of lines. Converting these measurements to polar coordinates gives

$$(1) \qquad \phi = 360^\circ - D$$
$$\theta = I + 90^\circ$$

where $0^\circ < D < 360^\circ$, and $-90^\circ < I < 90^\circ$. Direction cosines l(i), m(i), and n(i) are calculated by

$$(2) \qquad l(i) = \sin\theta(i) \cos\phi(i)$$

$$m(i) = \sin\theta(i) \sin\phi(i)$$

$$n(i) = \cos\theta(i)$$

where $i = \{1, 2, 3 \ldots N\}$ and N is the number of measurements. The resultant length R of the direction cosines is given by

$$(3) \qquad R = \sqrt{\{(\Sigma\, l(i))^2 + (\Sigma\, m(i))^2 + (\Sigma\, n(i))^2\}}.$$

From (3), the direction cosines of the mean direction may be calculated from

$$(4) \qquad \bar{l} = \Sigma\, l(i) \,/\, R$$

$$\bar{n} = \Sigma\, n(i) \,/\, R.$$

Polar coordinates of the mean direction are given by

$$(5) \qquad \bar{x}(o) = \arccos \bar{n}$$

$$\bar{y}(o) = \arccos [\, \bar{l} \,/\, \sin \bar{x}(o) \,].$$

We have therefore, the mean declination and inclination of the population of linear elements

$$(6) \qquad \bar{D} = \bar{x}(o) - 90°$$

$$\bar{I} = \bar{y}(o).$$

2. Calculation of α_{95}, κ, and δ. From Irving (1964), the radius of a circle of 95% confidence, α_{95}, is

$$(1) \qquad 140 \times (\kappa)^{-\frac{1}{2}} \quad \text{degrees},$$

where the best estimate of κ, the precision parameter, is given by

$$(2) \qquad \kappa = (N - 1) \,/\, (N - R).$$

The angular standard deviation, δ, is calculated from

$$(3) \qquad \delta = (1 \,/\, \cos)(R \,/\, N).$$

Small values of α_{95} help to distinguish means of populations, large values of κ indicate greater parallelism of lines, and values of δ describe angular standard deviation from the mean.