Reflection characteristics of a mylonite zone based on compressional wave velocities of rock samples

Daniel T. McDonough* and David M. Fountain

Program for Crustal Studies, Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming 82071, USA

Accepted 1987 November 26. Received 1987 November 10; in original form 1986 December 9

SUMMARY
A combined field and laboratory study of the seismic reflection response and seismic properties of the mylonite zone capping the Kettle dome metamorphic core complex provides important constraints on reflectivity of mylonites. Because of its structural and petrologic characteristics, this mylonite zone can be regarded as a model for mid- to deep-crustal mylonite zones. CDP reflection profiling revealed that the mylonite zone is reflective. Laboratory measurements of compressional wave velocities ($V_p$) at high confining pressures in samples collected directly up-dip from the CDP reflection line revealed a wide variation of $V_p$ in common mid-crustal lithologies. Mafic and carbonate rock mean $V_p$ values range from 6.67 to 7.31 km s$^{-1}$ at 600 MPa, and quartzite and quartzo-feldspathic values range from 6.16 to 6.52 km s$^{-1}$. Seismic anisotropy at low pressures can be related to open microcracks and preferred mineral orientation, whereas the latter effect controls anisotropy at high pressures. The laboratory data provide the basis for a seismic model which indicates that the Kettle dome mylonite zone can generate reflections from zones where high- and low-velocity lithologies are interlayered. These reflections correlate with those observed on the CDP data, thus demonstrating that the reflectivity of the mylonite zone is a direct result of its lithologic diversity and layered nature. In this case, it is not necessary to explain the reflections by the effects of anisotropy, retrogression, high pore pressure, or constructive interference.

Key words: mylonites, seismic reflection, seismic velocities

1 INTRODUCTION

Many coherent mid- to deep-crustal events on seismic sections from crystalline terranes have been interpreted as originating from deep fault zones, which are probably mylonite zones (Dohr & Meissner 1975; Smithson et al. 1979; Cook et al. 1981; Reif & Robinson 1981; Allmendinger et al. 1983; Brewer et al. 1983). These reflections in crystalline terranes are often strong and continuous. Jones & Nur (1982) proposed that mylonite zones are reflective because of high pore fluid pressures within the mylonite zones. Etheridge & Vernon (1983) proposed that retrograde minerals with low seismic velocities, such as micas and chlorite, are common in mylonite zones and could explain their reflectivity. Alignment of phyllosilicates during mylonitization could also lead to anomalously slow velocities normal to foliation of mylonite zones (Fountain, Hurich & Smithson 1984). Modelling results (Fountain et al. 1984; Jones & Nur 1984) indicated that the laminated nature of mylonite zones may be instrumental by enhancing reflection amplitudes through constructive interference. Other geometrical characteristics of mylonite zones, such as their continuous and planar nature, suggests they would be good reflectors (Smithson et al. 1984; Fountain et al. 1984). Laboratory data (Jones & Warner 1985) suggest certain mylonite zones may be high-velocity layers because of the loss of silica from the zone. Recently, Ratcliffe et al. (1986), using borehole samples and reflection data, demonstrated that lithologic heterogeneity in a mylonite zone can cause reflections.

In a set of specially designed experiments the University of Wyoming Program for Crustal Studies conducted a combined field and laboratory study of the seismic reflection response and seismic properties of the mylonite zone capping the Kettle dome Cordilleran metamorphic core complex in northeastern Washington (Fig. 1). Seismic reflection data acquired over the down-dip projection of the mylonite zone showed prominent reflections (Hurich et al. 1985a). The reflections occur as two bands of multicyclic events separated by a seismically transparent zone which can be correlated with specific portions of the exposed mylonite zone. The physical basis for this reflectivity was studied by measuring compressional wave velocities ($V_p$) at high confining pressures of mylonite zone samples collected in exposures directly up-dip from the reflection profile. In this paper we report these data, relate them to petrographical characteristics of the samples, and present a seismic model of the mylonite zone for comparison with the field data. We
find the field data correlate with the model and conclude
that lithologic layering in the mylonite zone is the primary
cause of reflections because it creates large velocity contrasts
and, therefore, significant reflection coefficients. This study
is the first to tie seismic reflectivity in a crystalline terrane
solely to seismic properties measured in rock samples
collected from outcrops directly up-dip from the seismic
reflection line.

2 GEOLOGICAL SETTING

The Kettle dome metamorphic core complex, part of the
southern extension of the large Shuswap metamorphic core
complex in southern British Columbia, is similar to other
Cordilleran metamorphic core complexes extending from the
Ruby-East Humboldt Ranges in northeastern Nevada to
the Shuswap complex (Armstrong 1982). It is located west
of the fold and thrust belt and comprised of infrastructural
levels which contain upper amphibolite facies meta-
sedimentary and meta-igneous rocks that structurally
underly a mylonite zone that is capped by sedimentary and
low-grade metamorphic rocks in fault contact with the zone
(Fig. 2). The Kettle dome infrastructure is composed of
tonalitic gneisses. The mylonite zone is a lithologically
diverse sequence of three laterally persistent structural
packages which contain both mylonitic and non-mylonitic
rocks (Wilson 1981). The lower 0.5–1.0 km of the mylonite
zone is interlayered, discontinuous mafic schist, massive
marble, calc-silicate rock, and blastomylonitic quartzite.

Although only one isolated lens of mylonitic rock was found
in this structural package, it is included with the mylonitic
zone because it is concordant and the rocks have the same
lineation direction as overlying structural packages. It also
contains blastomylonites, rocks that exhibit mylonitic foliation and lineation in hand sample but, in thin section,
the grains are recrystallized and show no evidence of grain
size reduction. The middle zone is 1.5–2.0 km thick and
contains mylonitic and blastomylonitic tonalitic gneiss. The
0.2–0.3-km thick upper package is laterally continuous
mylonitic amphibolite and blastomylonitic quartzite. Synki-
mematic sillimanite within the mylonite zone (Rhodes &
Cheney 1981) indicates that mylonitization occurred under
upper amphibolite facies conditions at mid-crustal depths of
approximately 15 km and at temperatures greater than
500 °C. Upper-plate rocks overlying the mylonite zone are
pre-Tertiary marbles, greenstones and phyllites overlain by
Eocene volcanic and sedimentary rocks (Cheney 1980).
The timing of deformational events in the Kettle dome is
poorly understood. Upper amphibolite facies metamorphism
in the Canadian Shuswap complex to the north culminated
in the Jurassic (Armstrong 1982). Rhodes & Cheney (1981)
established that movement on the Kettle River fault, which
bounds the top of the mylonite zone, is top-to-the-east and
post-Eocene, because Eocene rocks are involved in the
faulting.

3 SAMPLE DESCRIPTIONS

Rock samples used in this study were collected from surface
outcrops in the Kettle dome. Although efforts were made to
collect samples with minimal weathering, some samples
were discarded in the laboratory because of apparent
alteration and extensive small-scale fractures. Modal
analyses of each sample were made using standard thin
section point counting methods. Two hundred points were
counted on one thin section from each rock and stains were
used to aid identification of feldspars and carbonate
minerals. Descriptions, modal analyses and structural levels
are given in Table 1. Quartzo-feldspathic rocks were
described using the IUGS classification (Streckeisen 1976).

3.1 Upper mylonite zone (UMZ) samples

Samples from this package include a quartzite, amphibolite
and a foliated granite. The quartzite (KD-5) is blas-
tomylonitic with minor potassium feldspar and muscovite in
the modal analysis. KD-24 is a mylonitic amphibolite
comprised predominantly of plagioclase and hornblende.
The sample displays a pronounced foliation and a prominent
mineral lineation defined by alignment of the c-axes of
hornblende. The weakly foliated granite is a minor intrusive
body within the amphibolite unit.

3.2 Middle mylonite zone (MMZ) samples

Samples from this zone are generally similar as they are
either mylonitic or blastomylonitic quartzo-feldspathic
gneisses. A quartz-rich tonalitic gneiss (KD-4) and one of
the tonalitic gneisses (KD-28) are regarded as type 1 3–C
mylonites (Lister & Snoke 1984) although the fine-grained
nature of KD-4 precludes a clear-cut classification.
Anorthite content of plagioclase in these samples ranges from An$_{45}$ to An$_{60}$, compositions expected for upper amphibolite facies rocks (e.g. de Waard 1969; Cooper 1972). These mylonites exhibit two foliations, defined by quartz and mica alignment, at acute angles to each other. The biotite and quartz in these samples are very fine-grained and wrap around fractured feldspar porphyroblasts. KD-13, also a tonalitic gneiss, is mylonitic but does not display an S-C fabric. It is porphyroblastic, shows evidence of recrystallization and is, therefore, classified as a blastomylonite. These quartzo-feldspathic rocks exhibit preferred orientation of quartz c-axes (Fig. 3). Fig. 3 shows a strong concentration parallel to foliation and perpendicular to lineation for the blastomylonitic sample (KD-13). Fig. 3 also shows a similar but more diffuse pattern for the mylonitic sample (KD-28).

3.3 Lower mylonite zone (LMZ) samples

This zone is lithologically diverse and we present data for four major rock categories. Three samples (KD-9, -12 and -14) are blastomylonitic quartzites, each with minor amounts of feldspar and mica. The strong preferred orientation of quartz c-axes in these samples is illustrated in the plot for KD-9 in Fig. 3. The sample exhibits concentrations of quartz c-axes greater than 40 per cent per 1 per cent area. The axes are preferentially aligned parallel to foliation and perpendicular to lineation. Fabrics of this type are expected in quartz-bearing rocks deformed under upper amphibolite conditions due to the operation of the prismatic slip system in quartz (Tullis, Christie & Griggs 1973; Wilson 1975; Bouchez 1977; Bouchez & Percher 1981).

Carbonate rocks in the LMZ are represented by KD-16 and -18. The calc-silicate gneiss exhibits a well-developed foliation defined by alignment of quartz and diopside. Compositional bands are defined by alternation of calcite layers with zones of quartz and diopside. The mafic schist (KD-10) is predominantly composed of hornblende and diopside and exhibits many small-scale folds in the schistosity. The grandioritic gneiss (KD-11) from this level, an S-C mylonite, is generally similar to samples in the MMZ.

3.4 Infrastructure (IS) samples

Samples of the infrastructure consist of two tonalitic gneiss samples. KD-19 is a weakly foliated, garnet-bearing, porphyroblastic gneiss. KD-7 displays a strong foliation and compositional bands. Unlike other samples in this study, KD-7 exhibits significant retrogression of biotite. Magmatic quartz-rich veins also penetrate each core. These veins were not observed in thin section suggesting that the point counting underestimates the amount of quartz in the rock.
Table 1. Rock name, texture, structural zone sample from and modal percentages for Kettle dome samples

| Sample No. | Rock name                  | Texture         | Zone  | Mineralogy (per cent) |
|------------|----------------------------|-----------------|-------|------------------------|
| KD-4       | tonalitic qtz–pl–bi gneiss (quartz rich) | S–C mylonite (?) | MMZ   | 60 qtz, 30 pl (An46), 10 bi, tr cte, ep, opaques |
| KD-5       | quartzite                  | blastomylonitic | UMZ   | 96 qtz, 1.5 ks, 2.5 mu |
| KD-7       | tonalitic qtz–pl–bi gneiss | gneissic        | IS    | 66 pl (An80), 26 qtz, 5 al, 3 bi |
| KD-9       | quartzite                  | blastomylonitic | LMZ   | 94.5 qtz, 2.5 ks, 3.0 mu |
| KD-10      | hb–cpx–ep schist           | micro-folded    | LMZ   | 63.5 hb, 23.5 cpx, 5.5 ep, 4.5 pl (An46), 3 cte, tr opaques |
| KD-11      | granodioritic qtz–pl–ks gneiss | S–C mylonite | LMZ   | 54.5 pl (An46), 20 qtz, 13 ks 2.5 bi, 1 mu |
| KD-12      | quartzite                  | blastomylonitic | LMZ   | 88.5 qtz, 6.5 bi, 3 ks |
| KD-13      | tonalitic qtz–pl–bi gneiss (quartz rich) | blastomylonitic | MMZ   | 60 qtz, 24 pl, 13.5 bi, 2 mu, 0.5 sp |
| KD-14      | quartzite                  | blastomylonitic | LMZ   | 92.5 qtz, 4 pl, 2 bi, 1 ks, 0.5 mu |
| KD-16      | ca–qtz–cpx calc-silicate gneiss | gneissic        | LHZ   | 41 ca, 30 qtz, 26 cpx, 3 opaques |
| KD-18      | ca marble                  | granoblastic    | LHZ   | 99 ca, 1 do, tr scapolite, idocrase, and opaques |
| KD-19      | tonalitic pl–qtz–cpx gneiss | porphyroblastic | IS    | 63 pl (An80), 28.5 qtz, 3.5 cte, 2.5 bi, 2.5 ga |
| KD-24      | amphibolite                | mylonitic       | UMZ   | 54.5 hb, 36 pl, 4.5 opaques, 8.5 qtz, 0.5 ep, 1 cte |
| KD-25      | foliated granite           | foliated        | UMZ   | 36 ks, 32 qtz, 26.5 pl, 4 bi, 1 opaques, tr ga and mu |
| KD-28      | tonalitic pl–qtz–bi gneiss | S–C mylonite    | MMZ   | 66.5 pl (An46), 26 qtz, 7 bi, 0.5 cte, trem, sp and opaques |

Abbreviations:
- *al* = alteration products
- *bi* = biotite
- *ca* = calcite
- *cpx* = clinopyroxene
- *cte* = chlorite
- *do* = dolomite
- *ep* = epidote
- *ga* = garnet
- *hb* = hornblende
- *ks* = K-spar
- *mu* = muscovite
- *pl* = plagioclase
- *qtz* = quartz
- *sp* = spindle
- *tr* = trace amount
- *trem* = tremolite

UMZ = upper mylonite zone
MMZ = middle mylonite zone
LMZ = lower mylonite zone
IS = infrastructure

Anorthite content of plagioclase in KD-7 and KD-19 ranges from An46 to An56, also due to the upper amphibolite facies conditions experienced by the rocks.

4 EXPERIMENTAL METHOD

Cores 2.45–2.65 cm in diameter were cut in three orthogonal directions, generally with one normal to foliation, one parallel to foliation and lineation, and one parallel to foliation and perpendicular to lineation when these fabric elements were apparent. The three orthogonal directions were arbitrary if the rock exhibited a well-defined foliation, one parallel to foliation and lineation, and one orthogonal direction, generally with one normal to lineation. Cored elements were apparent. The three orthogonal directions were arbitrary if the rock was foliated but not heated. Small-scale heterogeneities were avoided when possible. The cores were cut to lengths of 2.25–3.81 cm and the core ends were machine lapped. Densities of the cores were determined from their mass and volume calculated from the core’s dimensions. The ends of the cores were painted with electrically conductive paint and the cores were wrapped in copper foil to prevent penetration of the fluid-confining medium.

Samples were placed in a pressure vessel which was pressurized to 600 MPa by an air-driven fluid pump and fluid intensifier. Two confining fluids were used at different times.
during the study; Exxon Isopar H or Davis-Howland Unisyn. Pressure was determined to within 0.1 MPa by direct digital read out from a manganin coil.

Sonic velocities in the rocks were determined using the measured travelt ime of a sonic wave through the core and the core length. The sonic waves were generated and received by 1 MHz lead zirconate piezoelectric transducers mounted on each end of the core. Traveltimes were measured on a digital waveform analyser which used a digitization increment of 10 ns. Traveltimes could be linearly interpolated to 0.1 ns. The time delay in the electronic portion of the system was determined following the method outlined by Birch (1960).

Errors in ultrasonic velocity measurement result from inaccuracies in core length measurements, travelt ime determination and effects of compressibility at high pressures. Core length measurements in this study are accurate to within ±0.05 mm and travelt ime determination to ±5 nanoseconds. These inaccuracies result in a possible measurement error of ±0.36 per cent for the shortest core (2.225 cm) used in the study and ±0.21 per cent for the longest cores (3.810 cm). Measurements of compressibility in similar rocks (e.g. Brace 1965) indicate a length decrease of 0.4 per cent of samples at 600 MPa suggesting that length changes have negligible effect on our velocity measurements. The inaccuracy of the travelt ime determination using the pulse transmission technique is generally regarded to be less than 2 per cent.

Seismic wave traveltimes were measured at various confining pressures while both increasing and decreasing pressure. Poor transducer-sample coupling for a few runs at low pressures precluded travelt ime measurements at very low pressures.

5 SEISMIC PROPERTIES OF KETTLE DOME SAMPLES

Table 2 presents the measured compressional wave velocities in the Kettle dome rocks for three propagation directions at confining pressures ranging from 50 to 600 MPa. The directions generally correspond to fabric elements displayed by the sample and are indicated in Table 2. Also tabulated are mean velocity and per cent seismic anisotropy (100 × (Vmax - Vmin)/Vmean).

The data exhibit a characteristic velocity-pressure relationship as shown in the examples in Fig. 4. Partial derivatives of compressional wave velocity with respect to pressure (∂Vρ/∂P) are non-linear below 350–500 MPa for most samples and smaller, linear gradients are observed at higher pressures. The non-linear gradients at lower pressures probably result from microcrack closure (Birch 1960; Christensen 1965; Simmons, Todd & Baldridge 1975). Because microcracks are likely oriented parallel to foliation this fast propagation direction results in a large parallelism between the fastest wave propagation of the quartz crystallographic habit and those perpendicular to lineation. Furthermore, this fabric also explains the similar velocities for the A and B propagation directions for the sample (Fig. 4a). Likewise, the relatively high concentration of c-axes for KD-28 also results in high anisotropy sympathetic with the effect of mica. In this case, the more diffuse pattern results in similar velocities in the B and C propagation directions. Szynaski, Ringland & Christensen (1985) also found that quartz and mica fabrics were a contributing factor to anisotropy in some quartz-feldspathic mylonitic rocks.

Seismic anisotropy values for the Kettle dome quartzites, which are all blastomylonitic, are high and range from 10.8–12.1 per cent (Table 2). These high anisotropy values result primarily from the intense quartz c-axis orientations in the Kettle dome quartzites (Fig. 3). Because the c-axis is the fastest wave propagation of the quartz crystallographic
## Table 2. Compressional wave velocities (km s\(^{-1}\)) in rocks from the Kettle dome

| Sample No. | Propagation direction | Density (g cm\(^{-3}\)) | P = 50 (MPa) | 100 | 200 | 300 | 400 | 500 | 600 |
|------------|-----------------------|------------------------|-------------|-----|-----|-----|-----|-----|-----|
| KD-4A-4B  | \(\perp F, \parallel L\) | 2.60                    | 5.52        | 5.81| 6.01| 6.09| 6.14| 6.18| 6.20|
| KD-5A-5B  | \(\perp F, \parallel L\) | 2.63                    | 5.56        | 5.65| 5.79| 5.81| 5.84| 5.85| 5.85|
| KD-8A-8B  | \(\perp F, \perp L\)    | 2.66                    | 5.88        | 5.93| 5.96| 5.99| 6.02| 6.04| 6.06|
| KD-9A-9B  | \(\perp F, \parallel L\) | 2.65                    | 6.36        | 6.41| 6.46| 6.48| 6.51| 6.54| 6.57|
| KD-10A-10B| \(\perp F\)              | 2.65                    | 5.93        | 6.00| 6.07| 6.09| 6.12| 6.15| 6.17|
| KD-11A-11B| \(\perp F, \parallel L\) | 5.13                    | 5.12        | 5.11| 5.10| 5.11| 5.15| 5.16| 5.19|
| KD-12A-12B| \(\perp F, \perp L\)    | 2.63                    | 5.53        | 5.63| 5.71| 5.74| 5.77| 5.80| 5.83|
| KD-13A-13B| \(\perp F, \parallel L\) | 2.69                    | 5.57        | 5.58| 5.60| 5.61| 5.64| 5.65| 5.66|
| KD-14A-14B| \(\perp F, \perp L\)    | 2.60                    | 5.73        | 5.82| 5.83| 5.84| 5.85| 5.85| 5.86|
| KD-15A-15B| \(\perp F\)              | 2.78                    | 6.21        | 6.25| 6.29| 6.32| 6.35| 6.38| 6.40|
| KD-16A-16B| \(\perp F\)              | 2.76                    | 6.25        | 6.26| 6.30| 6.33| 6.35| 6.37| 6.38|
| KD-17A-17B| \(\perp F, \parallel L\) | 2.70                    | 6.01        | 6.04| 6.07| 6.10| 6.13| 6.16| 6.18|
| KD-18A-18B| \(\perp F, \perp L\)    | 2.70                    | 5.79        | 5.81| 5.84| 5.87| 5.90| 5.93| 5.95|
| KD-19A-19B| \(\perp F\)              | 2.70                    | 5.77        | 5.80| 5.83| 5.86| 5.89| 5.92| 5.94|
| KD-20A-20B| \(\perp F, \parallel L\) | 2.70                    | 5.71        | 5.74| 5.77| 5.80| 5.83| 5.85| 5.88|
| KD-22A-22B| \(\perp F, \perp L\)    | 2.70                    | 5.88        | 5.92| 5.95| 5.98| 6.00| 6.03| 6.05|
| KD-23A-23B| \(\perp F\)              | 2.70                    | 5.93        | 6.00| 6.06| 6.10| 6.13| 6.16| 6.18|
| KD-24A-24B| \(\perp F, \parallel L\) | 2.70                    | 6.01        | 6.04| 6.07| 6.10| 6.13| 6.16| 6.18|
| KD-25A-25B| \(\perp F\)              | 2.70                    | 6.04        | 6.07| 6.10| 6.13| 6.16| 6.18| 6.20|
| KD-26A-26B| \(\perp F, \perp L\)    | 2.70                    | 6.07        | 6.10| 6.13| 6.16| 6.18| 6.20| 6.22|
| KD-27A-27B| \(\perp F\)              | 2.70                    | 6.10        | 6.13| 6.16| 6.19| 6.22| 6.25| 6.27|
| KD-28A-28B| \(\perp F, \perp L\)    | 2.70                    | 6.13        | 6.16| 6.19| 6.22| 6.25| 6.28| 6.30|
| KD-29A-29B| \(\perp F\)              | 2.70                    | 6.16        | 6.19| 6.22| 6.25| 6.28| 6.31| 6.33|
| KD-30A-30B| \(\perp F, \perp L\)    | 2.70                    | 6.19        | 6.22| 6.25| 6.28| 6.31| 6.34| 6.36|

Notes:
- Density (g cm\(^{-3}\))
- P = 50 (MPa)

References:
- 100 200 300 400 500 600

By guest
Downloaded from https://academic.oup.com/gji/article-abstract/93/3/547/662387 by guest on 29 July 2018

D. T. McDonough and D. M. Fountain
Reflection characteristics of mylonites

Table 2. (contd.)

| Sample No. | Propagation direction | Density (g cc\(^{-1}\)) | \(P = 50\) (MPa) | 100 | 200 | 300 | 400 | 500 | 600 |
|------------|-----------------------|----------------------|-----------------|-----|-----|-----|-----|-----|-----|
| KD-25A     | \(\perp F\)          | 2.59                 | 5.36            | 5.63| 5.82| 5.94| 6.01| 6.05| 6.08|
| KD-25B     | \(\parallel F\)      | 2.60                 | 5.69            | 5.90| 6.05| 6.13| 6.17| 6.19| 6.21|
| KD-25C     | \(\parallel F\)      | 2.60                 | 5.71            | 5.87| 6.02| 6.09| 6.14| 6.17| 6.20|
| KD-28A     | \(\perp F\)          | 2.69                 | 5.53            | 5.75| 5.98| 6.04| 6.14| 6.19| 6.22|
| KD-28B     | \(\parallel F, \parallel L\) | 2.72            | 5.87            | 6.01| 6.34| 6.41| 6.46| 6.49| 6.52|
| KD-28C     | \(\parallel F, \perp L\) | 2.65            | 6.05            | 6.16| 6.30| 6.39| 6.44| 6.48| 6.51|
| mean       |                       | 2.69                 | 5.82            | 5.97| 6.21| 6.28| 6.35| 6.39| 6.42|
| KD-28A     | \(\perp F\)          | 2.69                 | 8.9             | 6.9 | 5.8 | 5.9 | 5.0 | 4.7 | 4.7 |
| KD-28B     | \(\parallel F, \parallel L\) | 2.72            |                 |     |     |     |     |     |     |
| KD-28C     | \(\parallel F, \perp L\) | 2.65            |                 |     |     |     |     |     |     |
| mean       |                       | 2.69                 | 8.9             | 6.9 | 5.8 | 5.9 | 5.0 | 4.7 | 4.7 |

Abbreviations

\(\perp F\) = normal to foliation
\(\parallel F\) = parallel to foliation
\(\parallel L\) = parallel to lineation
\(\perp L\) = perpendicular to lineation

Figures

Figure 4. P-wave velocity versus pressure for three propagation directions for (a) KD-13 and (b) KD-28. In each diagram, A is normal to foliation, B is parallel to foliation and lineation and C is parallel to foliation and perpendicular to lineation.

Figure 5. Seismic anisotropy versus pressure for mylonitic amphibolite (KD-24), blastomylonitic quartzite (KD-14), mylonitic gneiss (KD-28), and mafic schist (KD-10).

Figure 6. Per cent anisotropy at 600 MPa versus volume per cent mica for quartzo-feldspathic rocks from the Kettle dome (circles) and Christensen (1965) (squares). Dashed line is best fit to the data from Christensen (1965). Solid line is fit to Kettle dome data excluding KD-7. Dotted line is theoretical relationship for perfectly aligned mica (Christensen 1965).
axes (McSkimin et al. 1965), strong c-axis alignment results in a highly anisotropic rock. Subsidiary factors must be contributing to quartzite anisotropy because the maximum anisotropy along the crystallographic axes is 11 per cent (anisotropy is as high as 26 per cent considering directions off the crystallographic axes). Phyllosilicate orientation is probably a factor. Although comprising a small volume percentage of the quartzites (Table 1), micas are also strongly aligned with their c-axes (the slow velocity direction) normal to foliation and thus may contribute to the measured anisotropy of these quartzites. Also, some cracks may remain open at high pressures and may contribute a small crack anisotropy component. Support for this interpretation is offered by the high \( \partial V_p/\partial P \) values of these samples at high pressures. The anisotropy values reported here for quartzites are higher than those reported in previous studies (e.g. Birch 1960; Christensen 1965) because of the intense c-axis preferred orientation which reflects the metamorphic grade of the Kettle dome mylonite zone.

The two mafic rocks in the study exhibit markedly different values of seismic anisotropy. The mylonitic amphibolite (KD-24) was sampled from the UMZ and exhibits the highest anisotropy of any rock in this study (15.4 per cent). This high anisotropy resulted from a strong mylonitic lineation formed by the alignment of hornblende c-axes, the fastest wave propagation direction in hornblende (Aleksandrov & Ryzhova 1961). Although high, the anisotropy exhibited by this mylonitic amphibolite is similar to values for non-mylonitic amphibolites (Christensen 1965; Fountain 1976). The mafic schist, which contains highly anisotropic hornblende and diopside, is nearly isotropic (1.5 per cent anisotropy) because small-scale folds in the sample dilute the effects of mineral alignment.

The remaining rocks in the study, the calcite marble (KD-18) and calc-silicate rock (KD-16) have 1.0 and 6.0 per cent seismic anisotropy, respectively. The marble is granoblastic and probably has random or nearly random orientation of crystallographic axes and thus low seismic anisotropy. The calc-silicate is well foliated with the c-axis of diopside, the mineral's fastest wave propagation direction (Levien, Weidner & Hewitt 1979), strongly aligned parallel to foliation indicating that diopside alignment is probably the main factor contributing to the rock's anisotropy.

In addition to seismic anisotropy, mineralogical composition of these samples is an important control of compressional wave velocity. This is well demonstrated for quartzo-feldspathic samples where plagioclase appears to control the mean compressional wave velocity at high pressures. In Fig. 7 we plot volume per cent plagioclase against mean \( V_p \) at 600 MPa for these samples. Although two samples are relatively anisotropic, they fall along a linear trend defined by the nearly isotropic samples. A least-squares fit to these points yields a correlation coefficient of 0.91. The anorthite content of the plagioclase in the samples ranges from An\textsubscript{0} to An\textsubscript{100}. Plagioclase in this composition range has high aggregate seismic velocities (Birch 1961; Ryzhova 1964), values greater than those of quartz, biotite and orthoclase (Aleksandrov & Ryzhova 1961; McSkimin et al. 1965; Ryzhova & Aleksandrov 1963; Belikov, Aleksandrov & Ryzhova 1970). For reference, we show 400 MPa aggregate values for An\textsubscript{0}, An\textsubscript{100} and albite (Birch 1961; Christensen 1965) in Fig. 7. The high anorthite content values fall close to the line when it is projected to 100 per cent thereby emphasizing the strong control of plagioclase content.

 Petrographic examination indicated that these quartzo-feldspathic rocks did not undergo retrograde metamorphism. Therefore, there were no significant compositional changes associated with mylonitization which would affect the mean \( V_p \). The metamorphic history of the Kettle dome mylonite zone differs from that commonly observed in other mylonite zones where retrograde metamorphism could have a profound effect on the seismic properties (e.g. Etheridge & Vernon 1983; Fountain et al. 1984; Passchier 1986).

The calcite marble (KD-18), calc-silicate rock (KD-16) and mafic rocks (KD-10, KD-24) exhibit greater mean \( V_p \) values than the quartzo-feldspathic rocks or quartzites. The calcite marble has a 600 MPa mean \( V_p \) of 7.06 km s\(^{-1}\). The calcite probably contains magnesiam as aggregates of pure calcite yield a calculated average 600 MPa mean \( V_p \) of 6.58 km s\(^{-1}\) at 600 MPa (Dandeekar 1968). The calc-silicate sample (KD-16) has a high mean \( V_p \) (6.67 km s\(^{-1}\)) because it contains diopside, a mineral with a high seismic velocity (Levien et al. 1979). The mafic schist (KD-10) and mylonitic amphibolite (KD-24) exhibit 600 MPa mean \( V_p \) values of 7.31 and 6.87 km s\(^{-1}\), respectively. The greater mean \( V_p \) of the mafic schist results from its greater diopside and hornblende content.

### 6 SEISMIC MODEL OF THE KETTLE DOME MYLONITE ZONE

CDP seismic reflection profiling over the eastern flank of the Kettle dome revealed that the mylonite zone is reflective (Hurich et al. 1985a). To examine the causes of reflections from the mylonite zone we developed a 1-D model of the zone and generated a synthetic seismogram from that model. The model (Fig. 8) was based on the lithologic layering sequence as determined from geological cross-sections (Wilson 1981) and observations along the sampling traverses made during this study. Seismic velocities and densities were assigned to each layer (Fig. 8) based on the laboratory data in Table 2. Values used are the 600 MPa velocities for propagation directions normal to foliation which correspond to the wave propagation direction for the CDP survey. We used the 600 MPa values because low
pressure values are strongly controlled by the effects of microcracks and void spaces that may be intrinsic to the rock in its original subsurface position and the result of surface weathering and sample preparation. These effects may obscure the importance of lithology and mineral anisotropy at low confining pressures. Wang & Simmons (1978) suggested that rocks from deep boreholes may be crack free in situ and Simmons & Nur (1968) suggested that extrapolation of high-pressure data to low pressures may provide a better estimate of in situ velocities in boreholes. Although not equivalent to the projected depth of the mylonite zone, the higher pressure data may probably better simulate the subsurface velocity variations.

These seismic velocities were used in conjunction with measured densities to calculate a reflection coefficient series for the geologic model (Fig. 8). The reflection coefficient series exhibits large absolute values which range from 0.04 to 0.19 because of the lithological diversity and the corresponding variability in velocity and density. These values are comparable to reflection coefficients in sedimentary rock sequences. The lower value is typical of a sandstone–limestone interface and the upper value is larger than that of a water-soft ocean bottom interface (Telford et al. 1976). Also shown in Fig. 8 is the reflection coefficient series for mean velocities for comparison with the values for velocities normal to the propagation direction.

A 1-D synthetic seismogram (Fig. 8) was generated for the model that convolved the reflection coefficient series with a 10–60 Hz Vibroseis sweep, the same sweep used for the reflection survey. Effects of energy loss with depth are accounted for in the model to allow comparison of the model with the automatic gain controlled CDP data. Also shown on the upper part of the synthetic seismogram is an isolated wavelet and it corresponds to a reflection coefficient of 0.1.

In Fig. 9 we present the same synthetic seismogram as a 2-D model with the appropriate dip. The synthetic seismogram shows two bands of reflections. The upper band shows high-amplitude, multicyclic reflections and corresponds to the UMZ, the lithologically layered package of mylonitic and blastomylonitic rocks. The lower band is similar in character to the upper band and is generated by interfaces in the LMZ, the lithologically layered lower portion of the mylonite zone. Sandwiched between the two reflective horizons is a transparent region which corresponds to the mylonitic and blastomylonitic gneisses of the MMZ. Hurich et al. (1985a) correlated the reflective zones in the CDP data with the UMZ and LMZ. The comparison between the synthetic seismogram and the CDP data (Fig. 9b) supports this correlation and, furthermore, provides a physical explanation for the reflective nature of the mylonite zone. The results of this study suggest that, in addition to previous proposals for mylonite reflectivity (Jones & Nur 1982, 1984; Etheridge & Vernon 1983; Fountain et al. 1984; Smithson et al. 1984; Jones & Warner 1985; Passchier 1986), lithologic heterogeneity can be a viable mechanism to generate reflections from mylonite zones. In particular, the combination of lithologic layering and large reflection coefficients can adequately explain the reflections observed in the Kettle dome CDP data. The lack of reflections from the relatively homogeneous MMZ and the correspondence of reflections to the complex UMZ and LMZ emphasize this point. This precludes the necessity, in this case, of explaining mylonite zone reflectivity by the effects of constructive interference of reflected waves. In the Kettle dome, layer thicknesses are too large, with the exception of
two layers in the UMZ, for constructive interference for the frequencies used in the seismic survey. However, we should point out that presence of undulating contacts and discontinuous layering, especially in the UMZ, will prevent direct correlation between interfaces and individual reflection events.

Seismic anisotropy in the Kettle dome mylonite zone both enhances and decreases seismic reflectivity. Seismic anisotropy enhances the reflection coefficient by 37.6 per cent at the interface between the nearly isotropic calcite marble (KD-18) and highly anisotropic quartzite (KD-9) in the LMZ by decreasing $V_p$ normal to foliation in the quartzite with respect to the marble. Anisotropy decreases the reflection coefficient at the interface between the UMZ and MMZ by 29.5 per cent. The highly anisotropic amphibolite (KD-24) at the base of the UMZ has a $V_p$ normal to foliation similar to the underlying quartzofeldspathic gneiss (KD-28), although the amphibolite has a much higher mean $V_P$.

These results also bear on the influence of retrogressive metamorphism on physical properties of deep mylonite zones. This is exemplified by the contrast between Kettle dome data and Newark basin border fault data. Ratcliff et al. (1986) established that reflectivity of a Newark basin border fault zone resulted from large reflection coefficients (>0.1) at lithologic interfaces, some of them in or bounding mylonitic rocks. However, the Newark basin border fault zone contains several brittle faults that bound slices of cataclastic, mylonitic, and relatively undeformed rocks, some of which show significant retrogression. Such structural and petrographic features are indicative of upper-crustal deformation (Sibson 1977; Passchier 1986) and corroborate Etheridge & Vernon's (1983) proposal that retrograde mineralization can enhance seismic reflectivity of mylonite zones. The Kettle dome mylonite zone, in contrast, is characterized by upper amphibolite facies mineral assemblages, recrystallized grains, blastomylonitic textures and insignificant amounts of retrogression. Thus, we regard the Kettle dome as a model for mid- to deep-crustal mylonite zones. Based on laboratory-determined seismic properties, the zone provides a directly observable physical basis for reflectivity of deep mylonite zones. The results from the Newark basin fault zone and Kettle dome establish that lithologic layering, in conjunction with large reflection coefficients, cause reflections from shallow- and deep-crustal mylonite zones.

7 DISCUSSION AND CONCLUSIONS

In this paper we presented laboratory measurements of compressional wave velocities of a suite of mid-crustal rocks from the Kettle dome and its associated mylonite zone. These data, when used in conjunction with seismic modelling, can help constrain the causes of reflections from the mylonite zone. We summarize our primary observations below.

(1) Open microcracks in the samples lower seismic velocity and increase $3V_p/3P$ values at low confining
pressures. Microcracks parallel to foliation increase seismic anisotropy at low pressures. This further corroborates the observation that high confining pressures are needed to characterize seismic properties of mid- and deep-crustal rocks (Birch 1960; Christensen 1965).

(2) High-pressure seismic anisotropy in all rock samples results from alignment of constituent minerals. The amount of anisotropy in quartzofeldspathic rocks correlates with mica content. Quartzites have high anisotropy values because of intense preferred orientation of quartz c-axes. Anisotropy in the mafic and calc-silicate rocks is related to alignment of hornblende and diopside.

(3) The wide variation of high pressure mean $V_p$ in the samples can be related to mineralogical composition. Marble, mafic rocks, and a calc-silicate exhibit 600 MPa mean $V_p$ values ranging from 6.67 to 7.31 km s$^{-1}$. Quartzofeldspathic rocks and quartzites exhibit 600 MPa mean $V_p$ values ranging from 6.16 to 6.52 km s$^{-1}$.

(4) The mylonite zone consists of upper, middle and lower structural units that overlie the infrastructure. The lithologically diverse upper and lower zones exhibit significant velocity contrasts and large reflection coefficients.

(5) A synthetic seismogram shows strong, multicyclic reflections from the upper and lower zones. These reflections correlate with the actual CDP data from the zone.

(6) The reflectivity of the Kettle dome mylonite zone is a direct result of its lithological diversity and layered nature. In this case, it is not necessary to explain the reflections by effects of anisotropy, retrogression or constructive interference.

The wide range of $P$-wave velocities measured in the Kettle dome sample suite has implications for general crustal reflectivity. The middle crust, as exposed in the Kettle dome, is lithologically diverse. Lithologies characterized by high seismic velocities (e.g. amphibolite, marble, calc-silicate gneiss) commonly occur in structurally coherent packages within low velocity material (e.g. quartzofeldspathic rocks). Therefore, structural features that contain these rock types, such as limbs of nappes and mylonite zones, should be reflective. Thus we do not expect middle crustal levels to be seismically transparent everywhere as suggested by recent regional seismic reflection results (e.g. Brown et al. 1983; Matthews & Cheadle 1986). For example, the upper and middle crustal levels of the Kettle dome area (Hurich et al. 1985a), the Ruby-East Humboldt core complex, Nevada (Hurich, Valasek & Smithson 1985b; Valasek et al. 1986) and the Picacho Mountains, Arizona (Reif & Robinson 1981) are reflective. Finally, our results indicate that, in the absence of drill holes, seismic properties of exposed lithologies which can be projected beneath seismic reflection lines provide powerful constraints on seismic reflectivity in the crust.

ACKNOWLEDGEMENTS

We thank C. Hurich, A. Snoke and S. Smithson for assistance on the project and numerous discussions. F. Goertz and K. Bailey maintained the pressure system and assisted with sample preparation. Computing was performed on the University of Wyoming Program for Crustal Studies VAX 11/780 computer. This project was supported by National Science Foundation Grants EAR-830659 and EAR-8418350.

REFERENCES

Aleksandrov, K. S. & Ryzhova, T. V., 1961. The elastic properties of rock forming minerals, I, pyroxenes and amphiboles, Izv. Acad. Sci. USSR, geophys. ser. No. 9, 1339–1334.

Almendinger, R. W., Sharp, J. W., von Tischa, D., Serpa, L., Brown, L., Kaufman, S. & Oliver, J., 1983. Cenozoic and Mesozoic structure of the eastern Basin and Range province, Utah, from COCORP seismic-reflection data, Geology, 11, 532–536.

Armstrong, R. L., 1982. Cordilleran metamorphic core complexes—from Arizona to southern Canada, Ann. Rev. Earth planet. Sci., 10, 129–154.

Belikov, B. P., Aleksandrov, K. S. & Ryzhova, T. V. 1970, Elastic properties of rock forming minerals and rocks, Moscow, Nauka.

Birch, F., 1960. The velocity of compressional waves in rocks to 10 kilobars, J. geophys. Res., 65, 1083–1102.

Birch, F., 1961. The velocity of compressional waves in rocks to 10 kilobars, 2, J. geophys. Res., 66, 2199–2208.

Bouchez, J., 1977. Plastic deformation of quartzites at low temperature in an area of natural strain gradient, Tectonophysics, 39, 25–50.

Bouchez, J. & Peckerer, A., 1981. The main Himalayan thrust pile and its quartz-rich tectonites in central Nepal, Tectonophysics, 78, 23–50.

Brace, W. F., 1965. Some new measurements of linear compressibility of rocks, J. geophys. Res., 70, 391–398.

Brewer, J. A., Matthews, D. H., Warner, M. R., Hall, J., Smythe, D. K. & Whittington, R. J., 1983. BIRPS deep seismic reflection studies of the British Caledonides, Nauka, 305, 286–290.

Brown, L., Ando, C., Klemperer, S., Oliver, J., Kaufman, S., Czuchra, B., Walsh, T. & Ibachsen, Y., 1983. Adirondack–Appalachian crustal structure: The COCORP northeast traverse, Bull. geol. Soc. Am., 94, 1173–1184.

Cheney, E. S., 1980. The Kettle dome and related structures of northeastern Washington, Mem. geol. Soc. Am., 153, 463–484.

Christensen, N. I., 1965. Compressional wave velocities in metamorphic rocks at pressures to 10 kilobars, J. geophys. Res., 70, 6147–6164.

Christensen, N. I., 1966. Compressional wave velocities in single crystals of alkali feldspar at pressures to 10 kilobars, J. geophys. Res., 71, 3113–3116.

Cook, F. A., Brown, L. D., Kaufman, S., Oliver, J. E. & Petersen, T. A., 1981. COCORP seismic profiling of the southern Appalachian orogen beneath the Coastal Plain of Georgia, Bull. geol. Soc. Am., 92, 738–748.

Cooper, A. F., 1972. Progressive metamorphism of metabasic rocks from the Hazst Schist Group of southern New Zealand, J. Petrotol., 13, 457–492.

Dandekar, D. P., 1968. Pressure dependence of the elastic constants of calcite, Phys. Rev., 172, 873–877.

de Waard, D., 1959. Anorthite content of plagioclase in basic and pelitic crystalline schists as related to metamorphic zoning in the Usa Massif, Timor, Ann. Sci., 257, 553–562.

Dohr, G. P. & Meissner, R., 1975. Deep crustal reflections in Europe, Geophysics, 40, 25–39.

Etheridge, M. A. & Vernon, R. H., 1983. Comment and reply: Seismic velocity and anisotropy in mylonites and the reflectivity of deep crustal fault zones, Geology, 11, 487–488.

Fountain, D. M., 1976. The Ivrea-Verbano and Strona-Ceneri zones, northern Italy: a cross-section of the continental crust—new evidence from seismic velocities of rock samples, Tectonophysics, 33, 145–165.

Fountain, D. M., Hurich, C. A. & Smithson, S. B., 1984. Seismic reflectivity of mylonites in the crust, Geology, 12, 195–198.

Hurich, C. A., Smithson, S. B., Fountain, D. M. & Humphreys, M. C., 1985a. Seismic evidence of mylonite reflectivity and
deep structure in the Kettle dome metamorphic core complex, Washington, Geology, 13, 577–580.

Hurich, C. A., Valasek, P. A. & Smithson, S. B., 1985b. Comparison of crustal reflection profiles in three Cordilleran metamorphic core complexes, Eos, Trans. Am. geophys. Union., 66, 380.

Jones, T. & Nur, A., 1982. Seismic velocity and anisotropy in mylonites and the reflectivity of deep crustal fault, Geology, 10, 260–263.

Jones, T. & Nur, A., 1984. The nature of seismic reflections from deep crustal fault zones, J. geophys. Res., 89, 3153–3171.

Jones, R. H. & Werner, M. R., 1985. Why are low angle faults in crystalline rocks reflective? Terra Cognita, 5, 164.

Levien, L., Weidner, D. J. & Prewitt, C. T., 1979. Elasticity of diopside, Phys. Chem. Minerals, 4, 105–113.

Lin, W. & Wang, C., 1980. P-wave velocities in rocks at high pressure and temperature and the constitution of the central California crust, Geophys. J. R. astr. Soc., 61, 379–400.

Lister, G. S. & Snook, A. W., 1984. S–C mylonites, J. struct. Geol., 6, 617–638.

Matthews, D. H. & Cheadle, M. J., 1986. Deep reflections from the Caledonides and Variscides west of Britain and comparison with the Himalayas, in Reflection Seismology: A Global Perspective, eds Barazangi, M. & Brown, L., Am. Geophys. Union, Geodyn. Ser., 13, 1–19.

McSkimin, H. J., Andreatich, P. & Thurston, R. W., 1965. Elastic moduli of quartz versus hydrostatic pressure at 25 and 195.8 °C, J. Appl. Phys., 36, 1624–1632.

Passchier, C. W., 1986. Mylonites in the continental crust and their role as seismic reflectors, Geologie en Mijnbouw, 65, 167–176.

Ratcliffe, N. M., Burton, W. C., D’Angelo, R. M. & Costain, J. K., 1986. Low-angle extensional faulting, reactivated mylonites, and seismic reflection geometry of the Newark basin margin in eastern Pennsylvania, Geology, 14, 766–770.

Reif, D. M. & Robinson, J. P., 1981. Geophysical, geochemical, and petrographic data and regional correlation from the Arizona state A-1 well, Pinal County, Arizona, Arizona Geol. Soc. Digest, 13, 99–109.

Rhodes, B. P. & Cheney, E. S., 1981. Low-angle faulting and origin of Kettle dome, a metamorphic core complex in northeastern Washington, Geology, 9, 366–369.

Ryzhova, T. V., 1964. Elastic properties of plagioclases, Izv. Acad. Sci. USSR, Geophys. Ser. No. 7, 1049–1051.

Ryzhova, T. V. & Aleksandrov, K. S., 1965. The elastic properties of potassium–sodium feldspars, Izv. Acad. Sci. USSR, Earth Phys. Ser. No. 1, 98–102.

Sibson, R. H., 1977. Fault rocks and fault mechanisms, J. geol. Soc. Lond., 13, 191–213.

Simmons, G. & Nur, A., 1968. Granites: Relation of properties in situ to laboratory measurements, Science, 162, 789–791.

Simmons, G., Todd, T. & Baldridge, W. S., 1975. Toward a quantitative relationship between elastic properties and cracks in low porosity rocks, Am. J. Sci., 275, 345.

Smithson, S. B., Brewer, J. A., Kaufman, S., Oliver, J. E. & Hurich, C. A., 1979. Structure of the Laramide Wind River uplift, Wyoming, from COCORP deep reflection data, J. geophys. Res., 84, 5955–5972.

Smithson, S. B., Johnson, R. A., Hurich, C. A. & Fountain, D. M., 1984. Reflections from fault zones in the crust, Eos, Trans. Am. geophys. Union., 65, 277.

Strecker, A., 1976. To each plutonic rock its proper name, Earth Sci. Rev., 12, 1–33.

Szymanski, D. L., Ringland, M. L. & Christiansen, N. I., 1985. Quartz and mica fabric control of seismic anisotropy in metamorphic core complexes, geol. Soc. Am. Abstracts with Programs, 17, 731.

Telford, W. M., Geldart, L. P., Sheriff, R. E. & Keys, D. A., 1976. Applied Geophysics, 860 pp. Cambridge University Press.

Tullis, J. A., Christie, J. M. & Griggs, D. T., 1973. Microstructures and preferred orientations of experimentally deformed quartzites, Bull. geol. Soc. Am., 44, 297–314.

Valasek, P. A., Johnson, R. A., Hurich, C. A. & Smithson, S. B., 1986. Reflection profiling in the Ruby Mountains, Nevada, Eos, Trans. Am. Geophys. Union., 67, 313.

Wang, H. F. & Simmons, G., 1978. Microcracks in crystalline rock from 5.3-km depth in the Michigan basin, J. geophys. Res., 83, 5849–5856.

Wilson, C. J. L., 1975. Preferred orientation of quartz in ribbon mylonites, Bull. geol. Soc. Am., 86, 966–974.

Wilson, J. R., 1981. Structural development of the Kettle granite dome in the Boyds and Bangs Mountain quadrangles, northeastern Washington, PhD thesis, Washington State University.