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Correlation between magnetic anisotropy and fabric for Devonian shales on the Appalachian Plateau

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Abstract

The magnetic anisotropy of Devonian black shale samples was measured from two cores drilled in the Appalachian Plateau. The mineralogy of the shales is predominantly clay, with small quantities of quartz and minor amounts of opaques and chlorite. Magnetite is the predominant ferromagnetic mineral present in the samples. The magnetic fabric was measured at both room temperature and liquid-nitrogen temperature and is dominated by a well-defined bedding (vertical) compaction and a lesser defined magnetic lineation. Measurements of the anisotropy of magnetic susceptibility (AMS) at liquid-nitrogen temperature, which enhances the paramagnetic contribution in the rock, showed a strong increase in both the bulk susceptibility and susceptibility differences. This increase suggests that the AMS is controlled by the paramagnetic minerals, particularly the clays and chlorite. Strain was measured from the orientation of basal planes of the chlorite crystals by texture goniometry. Good correlations have been found (1) between the orientation of the magnetic lineation and the long axes of the chlorite crystals, and (2) between the degree of magnetic foliation and the amount of vertical compaction. The magnetic lineation also agrees well with the direction of seismic anisotropy over the Plateau. The anisotropy of the anhysteretic remanence, which expresses the anisotropy due to the ferromagnetic component in the rocks, shows a weaker correlation with the amount of vertical compaction. A weak magnetic lineation suggests that the magnetite grains were aligned during a deformation phase which post-dates the main Alleghanian orogeny. The magnetic anisotropy of the Devonian shales mirrors the compaction and tectonic fabric on the Appalachian Plateau.

1. Introduction

The Devonian shales of the Appalachian Plateau show strong mineralogical and mechanical anisotropies that reflect both compaction and tectonic shortening of the section both prior to and during the Alleghanian orogeny (Engelder, 1979; Engelder and Geiser, 1979; Engelder and Oertel, 1985; Oertel et al., 1989). Study of this anisotropy at numerous locations around the Appalachian Basin is facilitated by oriented core samples from 31 wells drilled as part of the US Department of Energy's Eastern Gas Shales Project (Cliff Minerals Inc., 1982). The location of these wells and a sense of the mechanical anisotropy prevalent in the recovered core is illustrated in Fig. 1 which shows site-averaged P-

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wave velocity anisotropy measured in the laboratory under atmospheric conditions. The 'fast' direction generally parallels the structural trend of the basin and is normal to the direction of layer-parallel shortening that occurred during Alleghanian compression (note that the anomalous orientation shown for NY4 is probably a core orientation error).

Evans et al. (1989a) studied the nature of this anisotropy using cores from wells NY1 and NY4 located in Western New York. Taken together, these samples contain the complete Devonian section which in the study area lies immediately above a prominent salt detachment. Compressibility and ultrasonic measurements show that the seismic anisotropy is largely due to a population of stress-relief microcracks that in plan show a preferred orientation striking ENE (Meglis and Engelder, 1994). Since this corresponds to the well-documented orientation of maximum horizontal stress in the area (Evans et al., 1989b), it is clear that the preferred orientation of the microcracks is not controlled by in-situ stress anisotropy (for then the preferred orientation in plan would be orthogonal to that observed). Rather, Meglis and Engelder concluded stress relief microcracking was governed by a mineralogical fabric which Evans et al. (1989a) had earlier shown was present in the rock prior to stress relief. The latter used texture goniometry to measure the preferred orientation of chlorite in the NY1 and NY4 core samples and hence estimate the fabric-derived distortion tensor. When scaled to yield estimates of total finite strain, the results indicated dominant vertical shortening due to compaction together with a uniform shortening of the entire section in a NNW direction. The internal consistency of the strain estimates, together with their agreement with other indicators of Alleghanian deformation kinematics, such as the strike of fault-cored anticlines and the deformation of crinoid columnals, leaves little doubt that

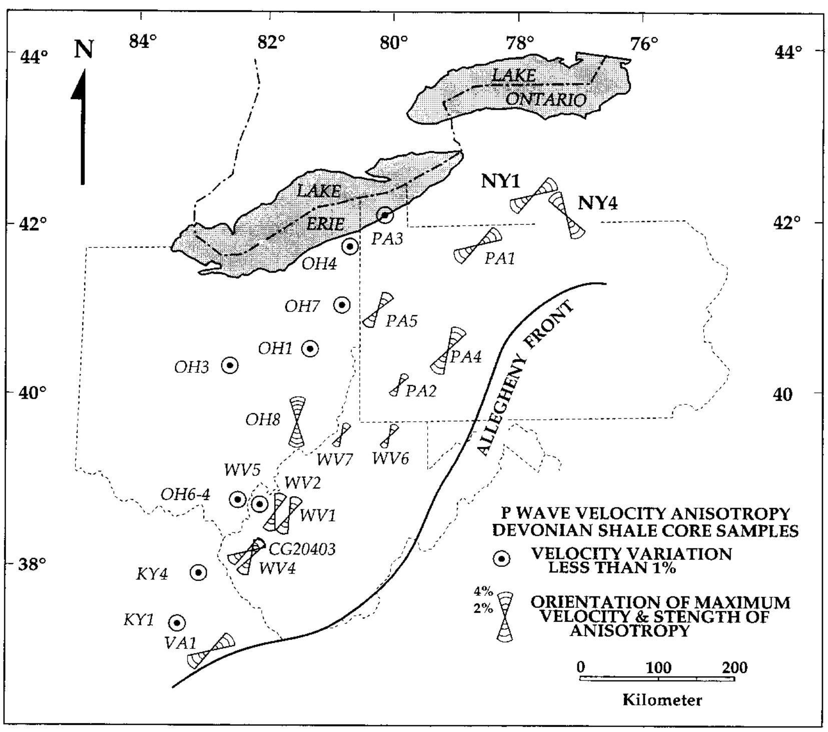


Fig. 1. Location map of drill cores that shows the magnitude and fast direction of the P-wave velocity anisotropy (after Evans et al., 1989a).

a strong fabric was imparted by Alleghanian deformation. The fabric is such that the principal preferred orientations of platy minerals are horizontal and vertical striking ENE, coincident with the preferred orientations for stress-relief microcrack development.

Given this well-documented compaction/tectonic fabric it is of interest to determine whether it is also manifested as a magnetic anisotropy. The magnetic anisotropy was measured on samples that were identical to those used by Evans et al. (1989a). In order to estimate the effect of the paramagnetic mineralogy on the magnetic fabric, the anisotropy of magnetic susceptibility was measured at both room temperature and liquid-nitrogen temperature. The anisotropy of anhysteretic remanent magnetization was also measured so that the preferential alignment of ferromagnetic minerals could be determined.

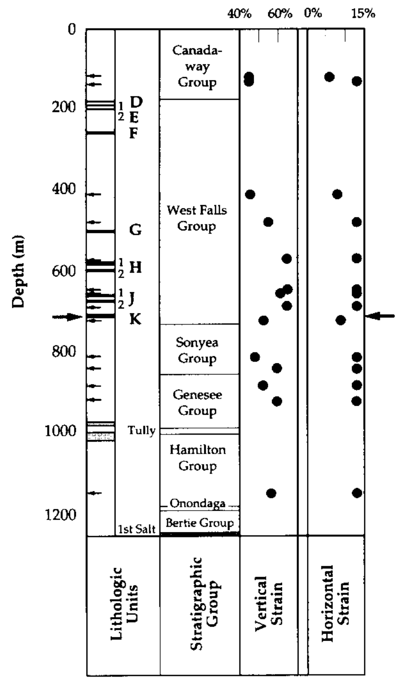


Fig. 2. Stratigraphic section of Devonian sediments from NY1 and NY4 showing sampling levels (small arrows). Large arrows show the base of the Rhinestreet formation. Strain data from Evans et al. (1989a).

2. Geological setting

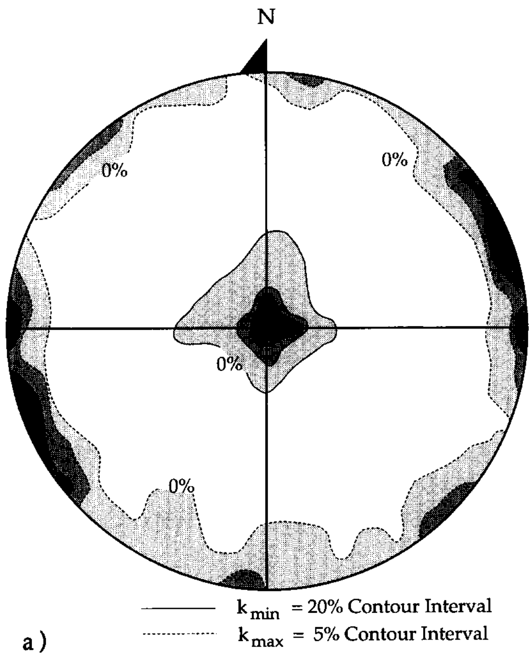
The stratigraphic section in the vicinity of NY1 and NY4 is shown in Fig. 2 together with the depth locations of the samples used in both this and the study by Evans et al. (1989a). Below the Devonian section lie the extensive salts and evaporites of the Silurian Salina group whose deposition attests to a long period of widespread basin stability. This stability was interrupted in Middle Devonian times by the onset of the Acadian orogeny which is associated with the collision of the Avalon Terrane with the continental margin (Bradley, 1983; Miller and Kent, 1986). Uplift of the highland to the east created the Catskill delta of which the Devonian shales are a component. Thus, the section reflects a pro-deltaic sedimentological setting in which black shales are interlayered with grey shales and turbidite-like mudstone or siltstone (the colour reflecting different organic content). Below the Sonyea, the section is somewhat more calcareous with occasional extensive limestone beds attesting to hiatuses in subsidence (Ettensohn, 1985), whereas above the Sonyea, coarser-grained, quartz-rich beds become increasingly common. With the exception of these thin (< 10 m) 'sandstone' and siltstone beds, the shales typically contain more than 90% clay-sized particles (Evans et al., 1989a). Clastic sedimentation persisted through the advent of the Carboniferous/Permian Alleghanian orogeny. In Western New York the Alleghanian orogeny expressed itself largely through decollement tectonics, with the section above the salt being transported to the northwest and shortened (Engelder and Engelder, 1977). Numerous blind thrust faults splay up from the salt detachment, cut the lowermost limestone beds but then disappear in the shales (Gwinn, 1964). Shortening by blind thrusting produces the fault-cored anticlines that are characteristic of the Plateau (Davis and Engelder, 1985), whereas the overlying shales accommodate shortening in a ductile manner. This shortening, together with the compaction response to burial, account for the chlorite fabric that Evans et al. (1989a) measured in the NY1 and NY4 cores and that Oertel et al. (1989) measured throughout the New York portion of the Appalachian Plateau. The total strain estimates for NY1 and NY4 are shown in Fig. 2. The horizontal strain estimates indicate that the entire section above the salt suffered uniform shortening of between 6 and 13% depending upon the form of kinematic deformation pattern assumed (Evans et al., 1989a). Independent strain markers indicate that layer-parallel shortening was above 10% (Geiser, 1988). The vertical strain estimates show an increase in compaction with depth up to the base of the Rhinestreet formation. However, below this depth the compaction is the same or smaller. Similar observations were reported by Engelder and Oertel (1985) from a regional study of rocks taken from outcrop and led them to propose that the section below the Rhinestreet formation became overpressured during burial and perhaps Alleghanian shortening. This break in compaction trend corresponds with a major drop in horizontal stress magnitudes measured in a well 20 km from NY1. This stress decline may reflect the dissipation of the paleo-overpressure to hydrostatic levels after cementation of the shale (Evans et al., 1989b). Recent work on joints and decollement zones in Devonian shales throughout the Appalachian Basin suggest the overpressures existed throughout the Alleghanian orogeny (Evans, 1994).

3. Magnetic fabric

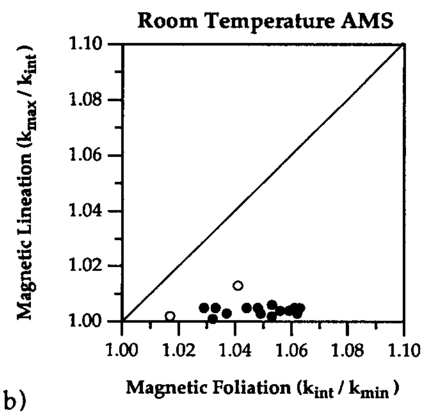
The anisotropy of magnetic susceptibility (AMS) is defined by a tensor of second order, and can be described geometrically as an ellipsoid with principal axes $k_{max} > k_{int} > k_{min}$. The magnetic anisotropy was measured on a modified Digico anisotropy delineator (Schultz-Krutisch and Heller, 1985), that measures the susceptibility differences between the principal axes. The magnitudes of these axes were obtained by normalizing the differences with a bulk susceptibility, which was measured on a KLY-1 susceptibility bridge. The procedure used to measure the AMS at near-liquid-nitrogen temperature on the Digico is given in Hirt and Gehring (1991). The anisotropy of anhysteretic remanent magnetization (AARM) was imparted in a bias field of 0.1 mT and a 150 mT alternating field, using a nine-position measurement scheme outlined in McCabe et al. (1985). Samples were demagnetized in a tumbling device before the next ARM was imparted. The magnetization of the demagnetized sample was subtracted from the following ARM measurement. This background remanence was two orders of magnitude less than the acquired ARM. Mean susceptibilities for the anisotropy ellipsoids are geometric means and are given in Table 1.

Table 1
Mean susceptibilities (SI) for the susceptibility magnitude ellipsoids

Sample	Depth (m)	AMS-293K	AMS-77K	AARM-293K
1.17	131.0	2.120e − 4	9.052e − 4	5.977e − 5
1.01	149.4	2.732e − 4	1.097e − 3	7.513e − 5
1.03	409.0	3.254e − 4	1.189e − 3	6.155e − 5
1.04	486.2	2.341e − 4	8.786e − 4	6.498e − 5
1.05	562.7	3.318e − 4	1.205e − 3	4.118e − 5
1.07	636.7	3.032e − 4	1.196e − 3	4.102e − 5
1.08	641.6	2.920e − 4	1.103e − 3	5.699e − 5
1.11	685.2	3.964e − 4	1.577e − 3	3.991e − 5
1.12	718.7	3.837e − 4	1.488e − 3	6.717e − 5
1.14	801.9	2.828e − 4	1.064e − 3	6.992e − 5
1.15	836.1	2.653e − 4	9.897e − 4	8.880e − 5
1.16	887.3	1.329e − 4	5.103e − 4	4.528e − 5
4.01	928.1	2.277e − 4	8.477e − 4	4.238e − 5
4.02	1159.6	2.365e − 4	8.574e − 3	7.092e − 5



a)



b)

Fig. 3. (a) Density plot (equal-area, lower-hemisphere projections) of the k_{max} and k_{min} axes of the AMS magnitude ellipsoid measured at room temperature. (b) Logarithmic plot of the principal axial ratios for the AMS at room temperature. Shales are shown by the solid circles and sandstones by the open circles in this and subsequent figures.

3.1. AMS fabric at room temperature

The AMS, measured at room temperature, consists of contributions from both the ferromagnetic and paramagnetic components in the rocks. The AMS magnitude ellipsoid of all the samples is flattened in the bedding plane and possesses a weak lineation (Fig. 3a). The degree of flattening (k_{int}/k_{min}) was between 1 and 7%, and the lineation (k_{max}/k_{int}) in the bedding plane was generally less than 1% (Fig. 3b). This lineation is ENE and sub-parallel to the structural trend of the fold belt in samples from NY1. Samples from NY4 have a lineation to the NW. The general lineation shown by samples from NY1 and NY4 is in directional agreement with maximum P-wave velocities seen at both wells (Fig. 1).

3.2. AMS fabric at near-liquid-nitrogen temperature

In order to compare the magnetic fabric with the chlorite fabric, it is desirable to separate the paramagnetic component to the AMS from the ferromagnetic component. Since this is only possible in high-field magnetometers which are not readily available, a second alternative is to enhance the signal from the paramagnetic fraction by measuring the AMS at low temperature. Samples were measured after having been submersed in liquid nitrogen, and those dominated by purely paramagnetic mineralogy show an increase in their susceptibility of 293K/77K (3.8). The average increase in signal for all samples was 3.4, indicating that a large part of the AMS is being carried by the paramagnetic minerals (Fig. 4). Thin-section analyses show that the shales are made up of over 90% clay minerals (Evans et al., 1989a). A marked increase in the value of the

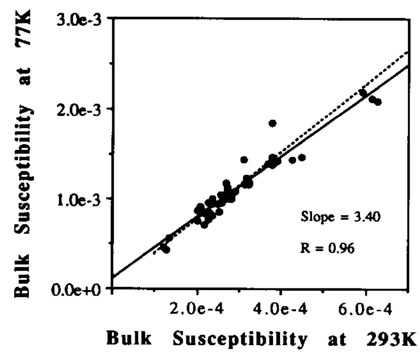


Fig. 4. Relationship between the bulk susceptibility measured at liquid nitrogen temperature and room temperature.

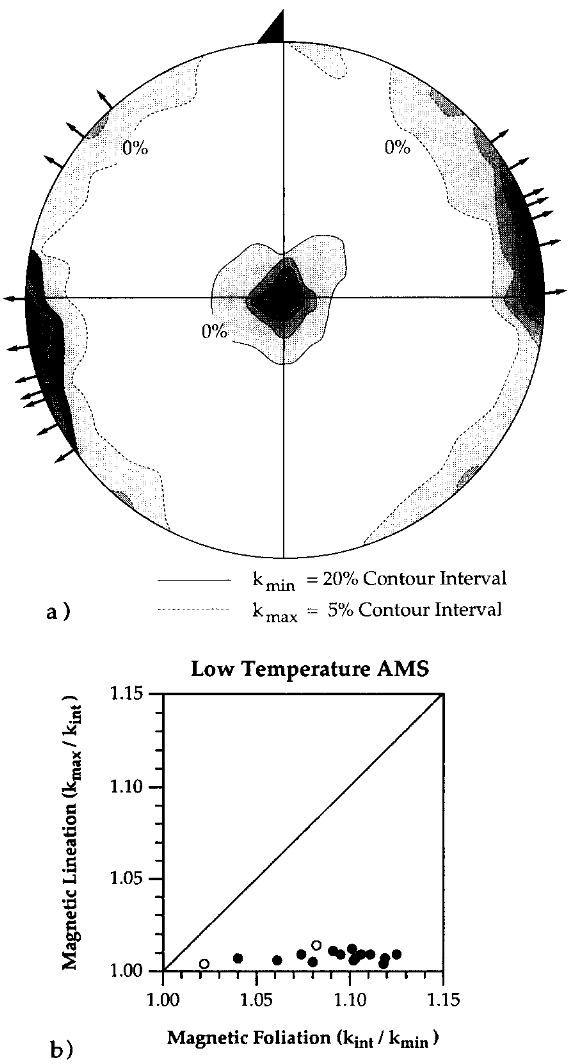


Fig. 5. (a) Density plot (equal-area, lower-hemisphere projections) of the k_{max} and k_{min} axes of the AMS magnitude ellipsoid measured at liquid nitrogen temperature. Arrows show the orientation of the long axes of chlorite grains. (b) Logarithmic plot of the principal axial ratios for the AMS at liquid nitrogen temperature.

susceptibility differences is observed for specimens measured on the Digico at low temperature.

The magnetic fabric, measured at low temperature, is also dominated by a bedding compaction (Fig. 5a). There is a good agreement between the orientation of the long axes of the chlorite grains and the k_{max} axes. The degree of flattening is higher than that seen in the room-temperature measurements, on the order of 2–12% (Fig. 5b). The lineation, although still weak, is more strongly clustered towards E (compare Fig. 3a with Fig. 5a).

3.3. Magnetic mineralogy and AARM

The AARM records the magnetic anisotropy due to the ferromagnetic minerals. In order to determine the ferromagnetic mineralogy in the samples, the acquisition of isothermal remanent magnetization (IRM) and its subsequent thermal demagnetization were observed. The IRM acquisition is dominated by low-coercivity minerals which saturate in magnetic fields of 1.0 T (Fig.

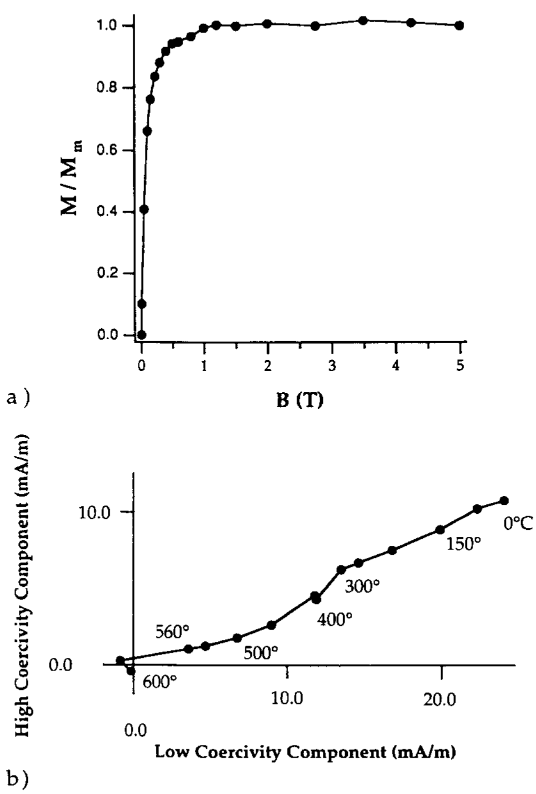


Fig. 6. (a) Acquisition of IRM for a black shale sample. (b) Vector diagram that shows the behaviour of the high- and low-coercivity component of the total IRM during thermal demagnetization.

6a). After a field of 5.0 T was applied along the vertical axis of the sample, a 0.1 T field was applied along the sample *X*-axis. Thermal demagnetization of this multicomponent IRM aids in determining which ferromagnetic minerals are present in a rock, by combining coercivity properties with thermal properties (Lowrie, 1990). Between room temperature and 300°C, a component of mixed coercivity is removed (Fig. 6b). Above 300°C to 350°C, a slightly harder component is removed before the final softer component is removed by 580°C. The higher-coercivity fraction is probably due to pyrrhotite since it is removed by 350°C, and the remaining softer component is magnetite.

A partial ARM was applied to a suite of samples in a bias field of 40 μT in progressively higher AF-fields up to 200 mT. The ARM was close to saturation by 140 mT although a slight increase was observed until 200 mT. The AARM was measured using a 100 μT bias field with a 150 mT AF field to ensure that the majority of the ferromagnetic minerals in the rocks would be affected.

The ferromagnetic fabric is strongly flattened in the bedding plane, with a degree of flattening between 6 and 47% (Fig. 7). There is a weak lineation on the order of 1–4% in the shales. The sandstones behave slightly differently in this respect, having lineations on the order of 10–15%. The lineation of the ferromagnetic fabric is oriented more N to NNE, as compared to the ENE lineation seen in the AMS for samples from NY1. Samples from NY4 are no longer systematically rotated 90° from samples from NY1, but have lineations to the NE and NW.

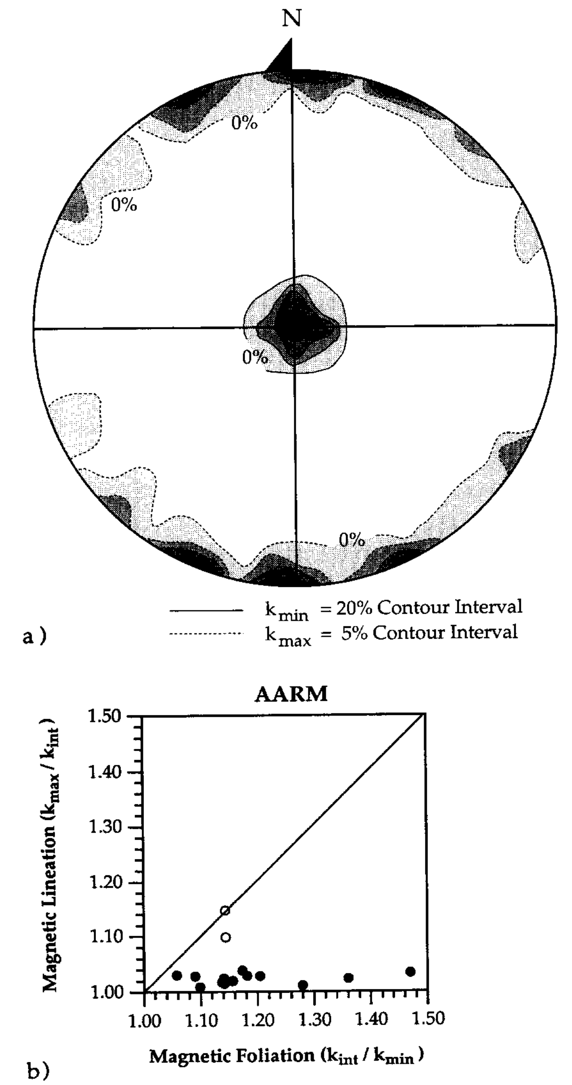


Fig. 7. (a) Density plot (equal-area, lower-hemisphere projections) of the k_{max} and k_{min} axes of the AARM magnitude ellipsoid. (b) Logarithmic plot of the principal axial ratios for the AARM.

4. Correlation of the degree of compaction with the magnetic fabric

The quantitative amount of compaction that these rocks have undergone is dependent on the deformation model assumed (Evans et al., 1989a, fig. 6). The magnetic fabric is compared with the 'u' fabric in Evans et al. (1989a), since this deformation model provides the most similar approximation to the susceptibility magnitude ellipsoid. Figure 8 shows the correlation of the degree of compaction with the total anisotropy (k_{max}/k_{min}) of the three types of magnetic anisotropy. The best correlation is the one comparing the degree of compaction with the AMS defined at low temperature. Since the magnetic fabric at low temperature is dominated by the paramagnetic minerals, it is reasonable that it will have a good correlation with the chlorite fabric. The correlation of the degree of compaction with AARM is the weakest, and the possible reasons for this are discussed below.

Evans et al. (1989a) have shown that the degree of compaction increases with increasing depth in the drill hole until the base of the Rhinestreet Formation at 708 m. Below this depth there is a drop in the degree of compaction which the authors attribute to paleo-overpressure drainage. A similar change in total anisotropy is also found at this depth for all magnetic fabrics (Fig. 9).

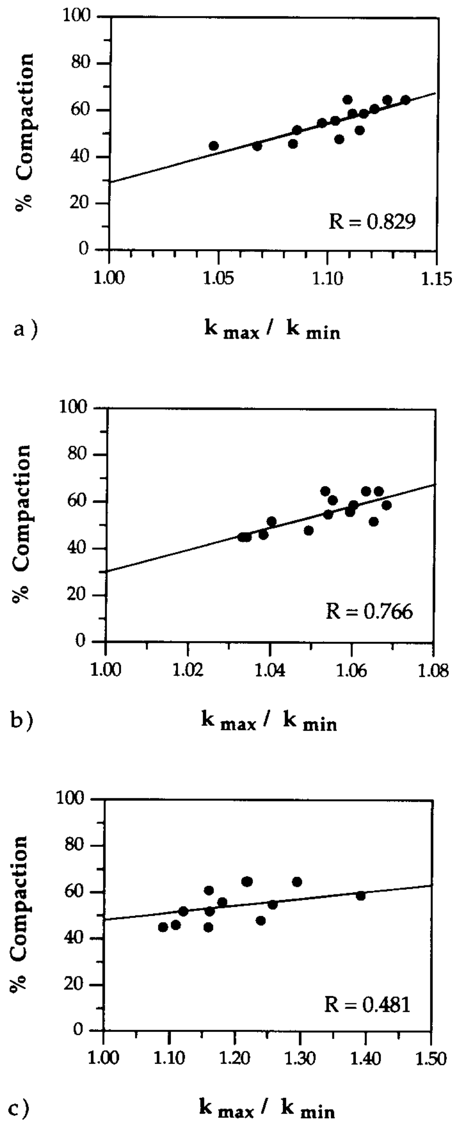


Fig. 8. Correlation between the degree of compaction and the degree of the magnetic fabric (k_{max} / k_{min}) for (a) AMS fabric measured at room temperature, (b) AMS fabric measured at room temperature, (c) AMS fabric measured at low temperature, and (d) AARM fabric.

5. Discussion

Three different magnetic fabrics appear in samples taken from two drill cores which contain Devonian sediments from the Appalachian Plateau. The fabric defined by both the AMS and AARM reflects the changes which are seen in the chlorite mineral fabric with depth. As seen in Figs. 8 and 9, the best correlation of the compaction with the total anisotropy is observed for the low-temperature AMS, which is not surprising since the low-temperature AMS reflects predominantly the orientation of the chlorite grains in the rocks. The AARM is due solely to the ferromagnetic minerals. Although it is also dominated by the vertical compaction it does not correlate as strongly with the chlorite fabric. The AMS at room temperature is predominantly influenced by the paramagnetic component of the rocks. However, since the orientation of this fabric is not identical with the one at low temperature, the room-temperature AMS also has a contribution from the ferromagnetic minerals. The significance of the individual fabrics due to the paramagnetic and ferromagnetic minerals is discussed below.

The weak lineation seen in the low-temperature AMS corresponds directionally to the fast direction of the seismic anisotropy and to the long axes of chlorite crystals. This direction is parallel to the structural trend on the Plateau, developed during the Alleghanian orogeny. Evans et al. (1989a) found that the horizontal strain estimates were relatively constant throughout the Devonian stratigraphy. The AMS shows a more variable horizontal extension that does not appear to be related to depth. Although the strength of the lineation (k_{max}/k_{int}) is not constant, it is limited to values between 0.4 to 1.1%. This weak

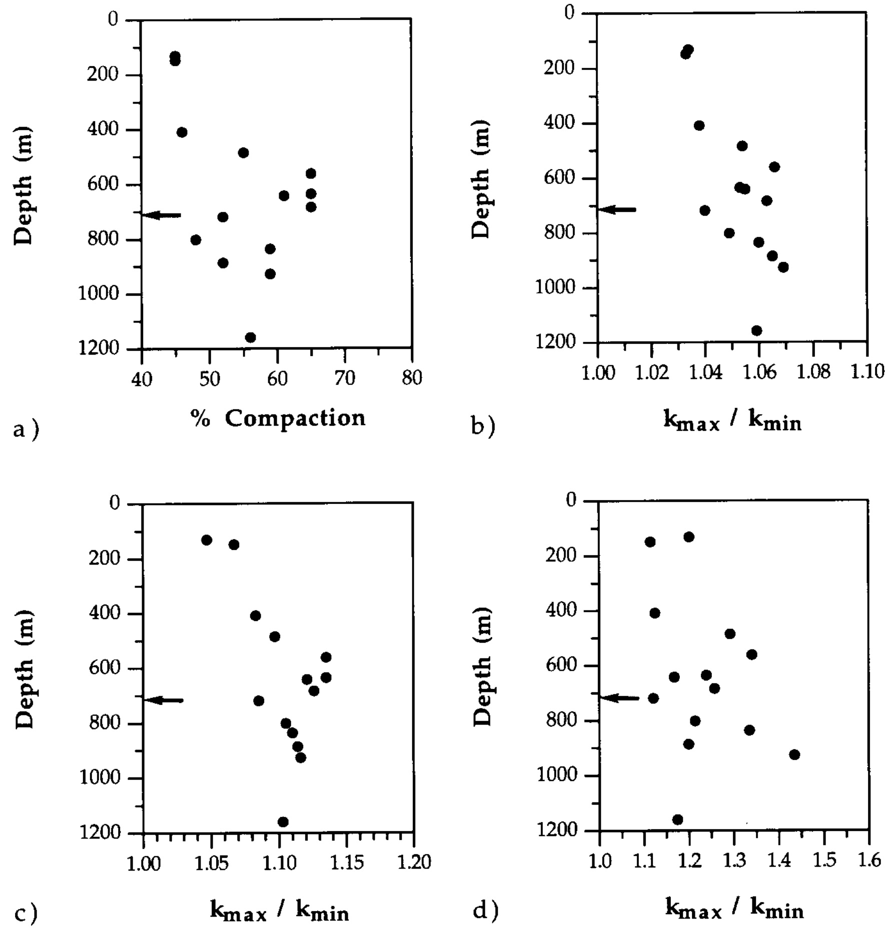


Fig. 9. Relationship of fabric with depth for (a) the chlorite fabric, (b) AMS fabric measured at room temperature, (c) AMS fabric measured at low temperature, and (d) AARM fabric. The arrows show the base of the Rhinestreet formation.

variability in the magnetic fabric may arise because the magnetic fabric is influenced by all the paramagnetic minerals, i.e., the clay minerals as well as the chlorite.

The magnetic fabric of the AARM is due to the preferred orientation of both pyrrhotite and magnetite which were identified in the rocks, although magnetite is the predominant ferromagnetic mineral. The correlation of the amount of compaction (k_{max}/k_{min}) for the ferromagnetic mineral fraction, inferred from the AARM is the least well-defined. However, the general increase of the degree of anisotropy with depth, with a corresponding drop in the anisotropy at the level of the Rhinestreet shale, is crudely preserved.

It was not possible to isolate a stable paleomagnetic direction from these samples, since the remanent magnetization appears to have been disturbed by the drilling process. Therefore it is not possible to assess whether these minerals are detrital, diagenetic or whether they crystallized during the deformation. Several studies have shown that limestones from the Appalachian fold belt and Plateau were remagnetized during the Pennsylvanian–Permian Kiaman polarity superchron (McCabe et al., 1983, 1989; Jackson, 1990).

Single-domain magnetite is responsible for this magnetization (Jackson, 1990). Since the magnetization of these single domain grains would be difficult to reset with a thermo-viscous model, it is considered more likely that the magnetite formed by some chemical process during the deformation (McCabe et al., 1989; Hirt et al., 1993). Thus, the magnetization of this magnetite most probably reflects the time of its formation.

The timing of the crystallization of the magnetite is important in comparing the correlation of the different magnetite fabrics with the strain. Although chlorite is not a primary (detrital) mineral, it probably crystallized from an illite precursor before the deformation associated with the Alleghanian orogeny. Since illite would also develop a preferred orientation in response to progressive vertical compaction, it is reasonable to expect that the chlorite would also reflect the compaction which the rock has undergone. In contrast, if magnetite also crystallized from a precursor during some stage of the deformation, it may not mimic the fabric of its precursor and cause a weakening of the ferromagnetic fabric. The AARM shows the weakest correlation with the amount of compaction. The general increase of the degree of anisotropy with depth can be explained by two mechanisms: (1) the burial compaction was still occurring at the time of formation of the magnetite; or (2) the precursor of the magnetite was strongly controlled by the compaction and this partially controlled the growth of new magnetite. The second mechanism is more likely if magnetite formed during the deformation as suggested by McCabe et al. (1983, 1989).

The lineation of the AARM fabric is more northerly than that found in the AMS. The relationship between the AARM fabric and a specific tectonic event is not clear since there are a number of possibilities:

(1) In Western Pennsylvania a systematic set of coal joints (face cleats) strike between 270° and 290° (Nickelsen and Hough, 1967). Theory and experiments suggest that the tension joints propagated early in coal formation (Ting, 1977). The deposition of the Pennsylvanian Upper Freeport coal swamps (Wise et al., 1991) coincides with early folding of the Appalachian Plateau. If the lineation in the AARM fabric is coeval with these coal joints, then the AARM fabric would be earlier than the AMS lineation developed in the chlorite grains.

(2) Devonian shale cores from Pennsylvania and West Virginia contain a set of tension joints striking 270° to 295° (Evans, 1994). This joint set postdates cross-fold joints associated with the major anticlines of the Appalachian Plateau. If the lineation in the ferromagnetic fabric is coeval with these late Alleghanian tectonic joints, then the ferromagnetic fabric would postdate the chlorite fabric.

(3) Black shales of the New York Appalachian Plateau carry a set of tension joints striking between 260° and 280° (Loewy and Engelder, 1994). This joint set appears to cross-cut the Appalachian Plateau fold axes and thus is post-Alleghanian. Loewy and Engelder (1994) suggest that this joint set correlates with Mesozoic uplift of the Appalachian Plateau and may have been driven by a gas desorption mechanism. If the ferromagnetic fabric is coeval with jointing in the black shales, then it would be a consequence of uplift.

Regardless of its origin, it is difficult to understand how the ferromagnetic fabric can preserve the imprint of overburden compaction without being affected by the major Alleghanian layer-parallel shortening which is so well imprinted on the chlorite fabric as seen in the AMS. One possible explanation is that the ferromagnetic fabric is related to magnetite which formed during the deformation from a precursor which was flattened by the vertical compaction. The weak lineation results from deformation in the late Alleghanian or Mesozoic.

6. Conclusions

The magnetic fabrics that have been defined from the room-temperature AMS, low-temperature AMS and AARM show a good agreement with the mineral fabric in Devonian sediments from the Appalachian Plateau. The best agreement is found between the low-temperature AMS and the chlorite fabric, since similar mineral fabrics are being compared. The agreement of the magnetic fabric with the 'fast' direction of seismic P-wave velocities further supports that the seismic anisotropy is controlled by the mineral fabric rather than the in-situ stress anisotropy. The weak lineation found in the alignment of the ferromagnetic minerals, may provide further support for a later deformation phase with E–W compression.

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