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Early Silurian palaeolatitude of the Springdale Group redbeds of central Newfoundland: a palaeomagnetic determination with a remanence anisotropy test for inclination error

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SUMMARY

We studied the palaeomagnetism of red fine-grained sandstones and coarse siltstones of the early Silurian Springdale Group of central Newfoundland. At 10 sites, a high blocking temperature characteristic remanence carried by haematite was isolated. This remanence is shown to predate probable early Devonian folding. Anti-parallel north- and south-directed remanences through a 100 m section of redbeds and a positive conglomerate test on haematite-bearing volcanic clasts suggest absence of remagnetization. Inverting the south-directed sites and unfolding yields a characteristic remanence with a mean declination of 23.6° and a mean inclination of −14.2° ($\alpha_{95} = 7.3°$, $k = 45.4$). The inclination corresponds to a probable early Silurian palaeolatitude of 7°S ± 4°. We find no significant difference between early Silurian palaeolatitudes for central Newfoundland north and south of the Red Indian Line suture, and conclude that the part of the Iapetus Ocean across the suture had narrowed to less than about 5° by the early Silurian. This is consistent with palaeomagnetic results from Britain and Ireland that suggest no more than a narrow Iapetus at low palaeolatitude by the early Silurian.

We also tested whether we have underestimated palaeolatitude because of sediment compaction reducing remanence inclination from that of the early Silurian field. We measured anisotropy of the isothermal remanence (IRM) acquisition for one specimen from each stable site, finding that a field of 200 to 800 mT applied at 45° to bedding produced an isothermal remanence oriented on average at 42° to bedding. Theory then predicts that sediment compaction caused less than 2° average inclination shallowing in the Springdale Group redbeds, and less than a 1° underestimation of palaeolatitude.

Key words: Iapetus Ocean, magnetic anisotropy, palaeolatitude, palaeomagnetism, Silurian.

1 INTRODUCTION

There is still debate about when the Iapetus Ocean closed in Newfoundland. The Iapetus (the Proto-Atlantic of Wilson 1966) had begun closing by the early Ordovician when ophiolites representing Iapetus oceanic crust and upper mantle were obducting onto the opposing continental margins. Ophiolites began obducting northwards from the Dunnage Zone (Fig. 1) onto the Humber Zone (then the margin of Laurentia) in the late Arenig (Stevens 1970). Ophiolites were also obducted southwards from the Dunnage Zone onto the Gander Zone (which may have been the margin of Avalonia) in the late Arenig (Colman-Sadd, Dunning & Dec 1992a). Closure of the Iapetus was very likely complete by the early Devonian, when Acadian deformation peaked (McKerrow 1988). But was there still an Iapetus Ocean in central Newfoundland in the early Silurian?

Late Llandovery and early Wenlock benthic shelly faunas worldwide are among the most cosmopolitan in the geological record (Cocks & Fortey 1990) making it difficult to test palaeontologically for an open Iapetus in the early Silurian. However, Williams & O'Brien (1991) suggest that the Iapetus was open, having found middle Llandovery

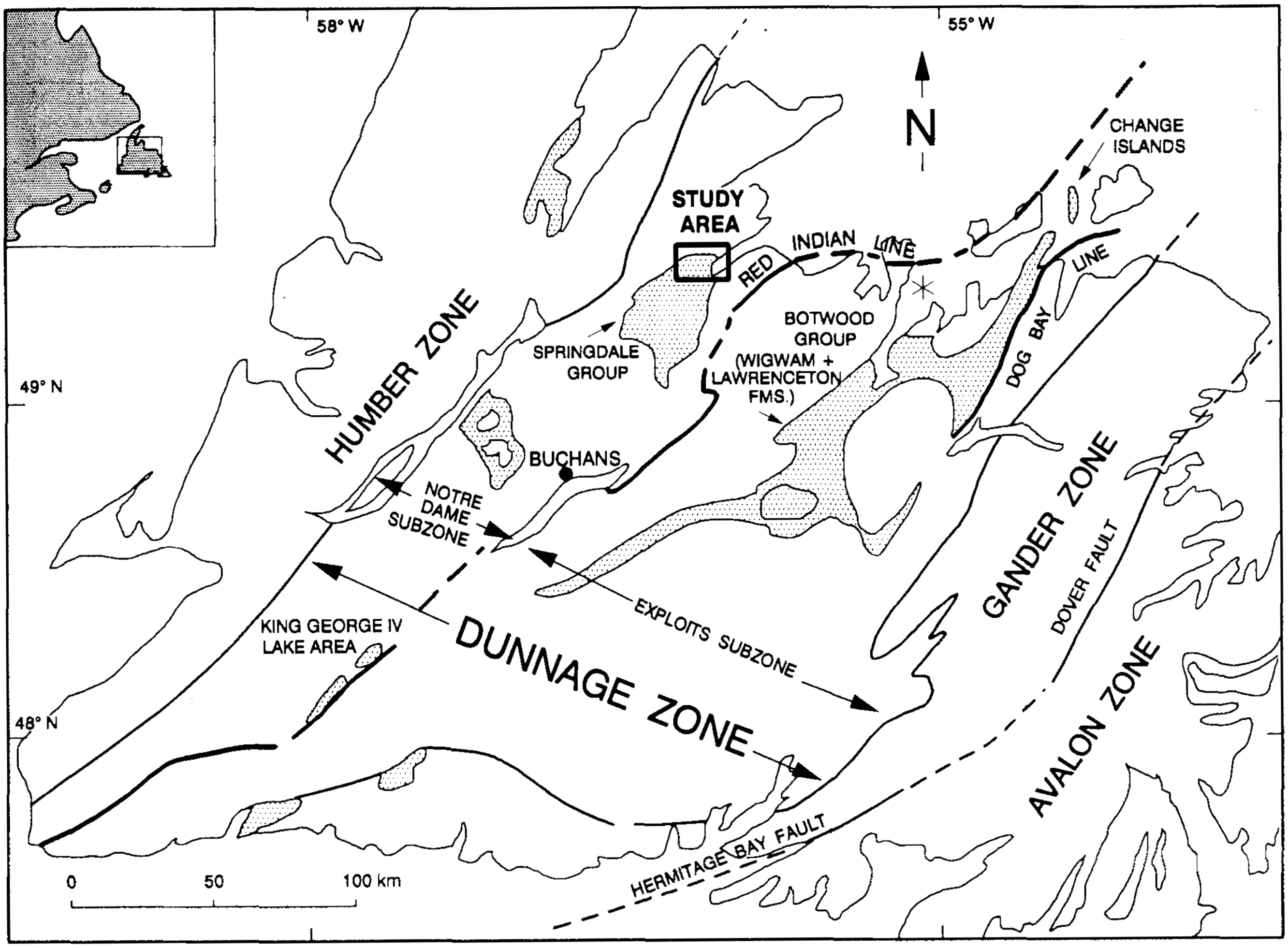

Figure 1. Early Silurian terrestrial volcanic and sedimentary rocks of Newfoundland (stippled) shown relative to the tectonic-stratigraphic zone boundaries of Williams *et al.* (1988). The Humber Zone represents the early Palaeozoic margin of Laurentia. The Gander Zone may represent the early Palaeozoic margin of Avalonia. The Dunnage Zone contains vestiges of the Iapetus Ocean. The Red Indian Line has been proposed as the trace of a major Iapetus suture. The study area is indicated and is shown in detail in Fig. 2.

graptolites of European rather than North American affinity in central Newfoundland at a site marked by an asterisk in Fig. 1. This site lies south-east of the Red Indian Line which has been proposed as the surface trace of an Iapetus suture on the basis of various geological contrasts between the Notre Dame Subzone of the Dunnage Zone to the north-west of the line and the Exploits Subzone to the south-east (Williams, Colman-Sadd & Swinden 1988). However, the early Silurian redbed and subaerial volcanic sequences in both subzones of the Dunnage Zone (Fig. 1) are very similar, suggesting an overlap assemblage and supporting the view that the Iapetus Ocean in central Newfoundland was closed by the early Silurian (Chandler, Sullivan & Currie 1987).

Palaeomagnetism should help resolve whether the Iapetus was open or closed in the early Silurian. A significantly lower palaeolatitude of formation for early Silurian rocks north of the Red Indian Line than for correlative rocks south of the line would support the presence of an Iapetus Ocean. The palaeolatitude P of formation can be inferred from the inclination I of primary remanence relative to bedding using the equation

$$P = \tan^{-1}(\tfrac{1}{2}\tan I). \qquad (1)$$

Buchan & Hodych (1989) found that the early Silurian redbeds and volcanics of the King George IV Lake area north of the Red Indian Line carry a primary remanence (with positive fold and conglomerate tests) acquired near the equator (0.5°N ± 6°). Gales, van der Pluijm & Van der Voo (1989) interpreted the remanence of the early Silurian Lawrenceton Formation volcanics south of the Red Indian Line as primary and inferred a 24°S ± 6° palaeolatitude. However, there is no proof that this remanence is primary and it may not even predate folding (Buchan & Hodych 1993). Buchan & Hodych (1992) showed that the overlying early Silurian Wigwam Formation redbeds and minor volcanics do carry a primary remanence (with positive fold and conglomerate tests) and indicate an 8.5°S ± 5° palaeolatitude. This is not significantly different from the palaeolatitude inferred from the King George IV Lake area. Thus, the palaeomagnetic evidence does not support the existence of a wide Iapetus Ocean across the Red Indian Line in the early Silurian.

We then sampled the early Silurian Springdale Group redbeds which lie to the north of the Red Indian Line in order to strengthen our palaeolatitude determination from the King George IV Lake area where outcrop is not extensive. We confined ourselves to the redbeds because we could find no volcanic sites with adequate structural control. Preliminary results for the Springdale Group were presented by Hodych & Buchan (1992). A reconnaissance study of the Springdale Group redbeds and volcanics by Black (1967) found shallow remanence inclinations, but only blanket demagnetization was used. More recently, preliminary results were presented by Potts, van der Pluijm & Van der Voo (1992, 1993) indicating an inclination of remanence that is steeper at mafic volcanic sites than at redbed sites, but the difference is not statistically significant.

Van der Pluijm *et al.* (1993) recently criticized the Wigwam Formation palaeolatitude of Buchan & Hodych (1992). They agreed that the remanence is primary, but suggested that sediment compaction had reduced the remanence inclination of the redbeds to the present $-17° \pm 10°$ from an initial inclination similar to the $-43° \pm 7°$ reported for the Lawrenceton Formations volcanics by Gales *et al.* (1989). Our reply to the criticism (Buchan & Hodych 1993) pointed out, for example, that this would require an unrealistically large amount of compaction given that we had purposely avoided collecting clay-rich redbeds. We have employed remanence anisotropy to test for compaction in fine-grained magnetite-bearing sedimentary rocks (Hodych & Bijaksana 1993). In the present study we describe an analogous remanence anisotropy method for haematite-bearing sedimentary rocks and apply it to redbeds of the Springdale Group.

2 GEOLOGY AND SAMPLING OF THE SPRINGDALE GROUP

The Springdale Group (Fig. 2) consists of a thick sequence of subaerial felsic and subordinate mafic volcanic rocks overlain by 1–2 km (Dean 1977) of fluviatile red conglomerates, sandstones and siltstones. The contact is conformable, with the uppermost flows in places interfingering with the redbeds (Coyle 1990). The Springdale Group has been interpreted as a caldera fill sequence (Coyle & Strong 1987).

Although no fossils have been reported from the Springdale Group, precise U–Pb zircon dates are available from the felsic volcanic rocks. Coyle (1990) reported a U–Pb zircon age of 432 ± 2 Ma from near the base of the volcanic sequence and 425 ± 3 Ma from near the top, in good agreement with the 429^{+6}_{-5} Ma reported by Chandler *et al.* (1987).

Most of our palaeomagnetic sampling sites (Fig. 2) are in fine-grained sandstones and coarse siltstones conformably overlying volcanic rocks in the Burnt Berry Syncline (sites 2

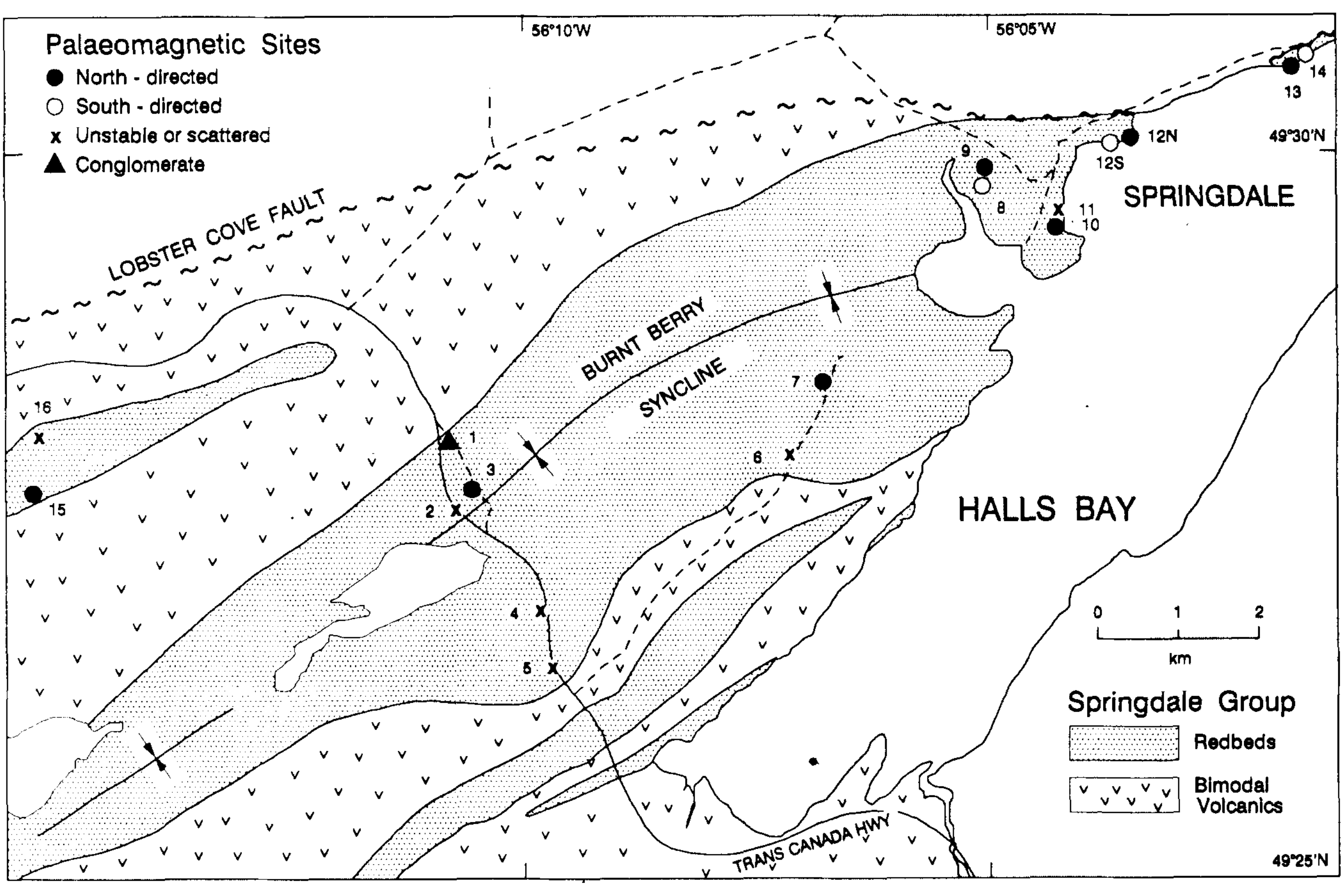

Figure 2. Palaeomagnetic sampling sites in redbeds and conglomerates of the early Silurian Springdale Group of central Newfoundland. Geology is after Coyle & Strong (1986), Coyle *et al.* (1985) and Dean & Strong (1975).

to 14), or in an outlier of redbeds (sites 15 and 16). At site 1, red rhyolite clasts were collected from a conglomerate among the redbeds. No sites were collected north of the Lobster Cove Fault (Fig. 2) because of complex structure with evidence of much block rotation about vertical axes (Z.A. Szybinski, private communication, 1992). Our sampling sites should all have ages equal to or a little younger than the 425 ± 3 Ma age of the uppermost volcanic rocks (Wenlock according to the time-scale of Harland *et al.* 1990). This is in good agreement with the 429^{+7}_{-3} Ma U–Pb zircon age of a rhyolite flow among redbeds in the King George IV Lake area (Dunning *et al.* 1990) and with the 430 ± 10 Ma age estimated (Buchan & Hodych 1992) for the Wigwam Formation mostly on the basis of Llandovery and Wenlock fossils. Both of these sets of redbeds are considered to be correlative with the redbeds of the Springdale Group (Chandler *et al.* 1987).

The Springdale Group redbeds show no penetrative deformation but are folded about a NE-trending synformal axis (the Burnt Berry Syncline) with a gentle northerly plunge (Dean & Strong 1975; Coyle & Strong 1987). The age of folding of the Springdale Group is poorly constrained, but is likely to be early Devonian as is common in the Notre Dame Subzone (Williams 1988).

The Springdale Group unconformably overlies the Robert's Arm Group which yields the same U–Pb zircon age (473 ± 2 Ma) as the nearby Buchans Group with which it is correlated (Dunning *et al.* 1987). Near Buchans (Fig. 1), the Buchans Group yields a latest Arenig–early Llanvirn conodont fauna with North American affinities (Nowlan & Thurlow 1984). This suggests that the Robert's Arm Group had accreted to North America by the middle Ordovician. There can be little doubt that, by the early Silurian, the Springdale Group was being deposited on Laurentia. Indeed, the presence of high-silica rhyolites and the large eruption volume of the calc-alkaline suite in the Springdale Group suggest that the crust below was continental (Coyle & Strong 1987).

3 PALAEOMAGNETIC PROCEDURE AND RESULTS

All of our samples were oriented blocks from which cylindical specimens (2.2 cm in length and 2.4 cm in diameter) were drilled in the laboratory. Alternating field (AF) demagnetization to 100 mT had little effect upon the remanence at many sites. Hence, all specimens were thermally demagnetized in steps to 680 °C. Usually, 14 thermal steps were employed using a Schonstedt TSD-1 demagnetizer with a peak temperature control of ±20 °C. Remanences were measured with a Schonstedt SSM-1 magnetometer. Occasionally, up to 20 thermal steps were employed using a remodelled furnace with a peak temperature control of ±5 °C. In this case, demagnetization and remanence measurement were carried out in a magnetically shielded room.

Examples of thermal demagnetization of Springdale Group redbed specimens are shown in Fig. 3. After removal

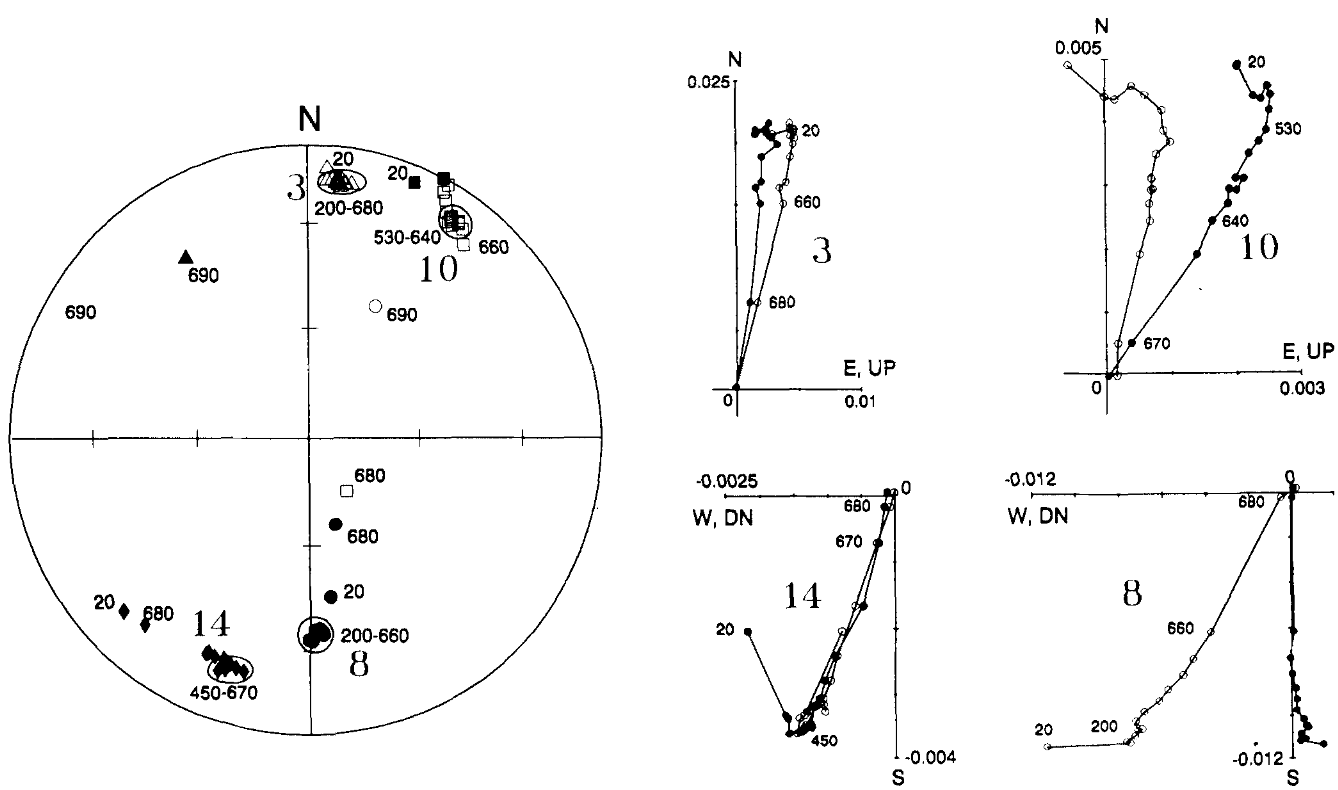

Figure 3. Examples of thermal demagnetization of Springdale Group redbed specimens from sites 3, 8, 10 and 14 after tilt correction. Directions are plotted on an equal-area net, with closed symbols indicating down directions and open symbols indicating up directions. On the orthogonal component plots, closed symbols represent projection onto the horizontal plane, whereas open symbols represent projection onto the north-up vertical plane. Magnetization is in A m^{-1} (10^{-3} emu cm^{-3}).

Table 1. Palaeomagnetic results for Springdale Group redbeds.

SITE	St (°)	Di (°)	N(n)	D (°)	I (°)	D′ (°)	I′ (°)	k	α_{95} (°)	d (mm)
3	30	12	5(5)	12.9	-23.4	8.4	-19.4	3	13.6	0.15
7	230	35	5(5)	26.4	19.8	19.8	3.5	120	7.0	0.04
8	60	20	5(4)	199.5	33.9	192.1	19.9	30	17.0	0.12
9	50	20	5(4)	44.5	-27.6	34.9	-24.0	42	14.3	0.28
10	200	15	5(5)	25.7	-17.0	30.0	-15.0	36	12.9	0.05
12N	90	80	7(4)	79.7	-63.3	26.3	-4.4	34	15.9	0.10
12S	90	80	11(11)	262.2	65.0	204.9	5.8	23	9.7	0.09
13	240	100	6(5*)	53.1	36.3	24.7	-11.4	20	17.5	0.04
14	240	100	6(5*)	222.1	-32.2	209.3	20.4	47	11.3	0.04
15	155	10	5(5)	21.9	-31.0	25.4	-23.5	123	6.9	0.21
In situ site mean			10	35.4	-19.0			4.2	26.7	
Tilt-corrected site mean			10			23.6	-14.2	45.4	7.3	

Note: Only stable sites are listed. Sites 2 and 4 with highly scattered remanence directions and sites 5, 6, 11 and 16, whose samples are not stably magnetized, are not included in the site mean statistics. St and Di are strike and dip respectively of bedding. *N* is the number of specimens measured and *n* is the number of stable samples used in statistics (* indicates inclusion of a single sample with remanence inverted to that of the rest of the samples at that site). *D* and *I* are the *in situ* mean declination and inclination, respectively, of the characteristic remanence whereas *D′* and *I′* are the tilt-corrected mean declination and inclination. *k* is the precision parameter and α_{95} the circle of 95 per cent confidence about the mean direction. *d* is the estimated grain diameter in millimetres of the rock specimens at the site.

of soft viscous remanences during demagnetization to peak temperatures between 200 and 500 °C, directions at most sites remained stable to temperatures between 640 and 680 °C and were directed either to the north or to the south. Blocking temperatures suggest that haematite is the dominant carrier of stable remanence in all the redbed specimens, with magnetite as a minor carrier of stable remanence in some samples at sites 3, 9, 10, 12N, 12S and 15.

At the 10 redbed sites listed in Table 1, most samples were stable. Their characteristic remanence directions, calculated using the least-squares fitting method of Kirschvink (1980), were averaged for each stable site. Table 1 excludes four sites (5, 6, 11 and 16) whose remanence directions did not remain stable on thermal demagnetization and two sites (2 and 4) whose sample directions were stable but highly scattered.

4 WHEN AND HOW REMANENCE WAS ACQUIRED

4.1 Fold test

We have tested whether the characteristic remanence of the Springdale Group redbeds was acquired before or after probable early Devonian folding. The 10 sites that yield well-defined characteristic remanence directions (Table 1) show considerable scatter before tilt correction (Fig. 4a). Reversing the south-directed remanences yields an overall

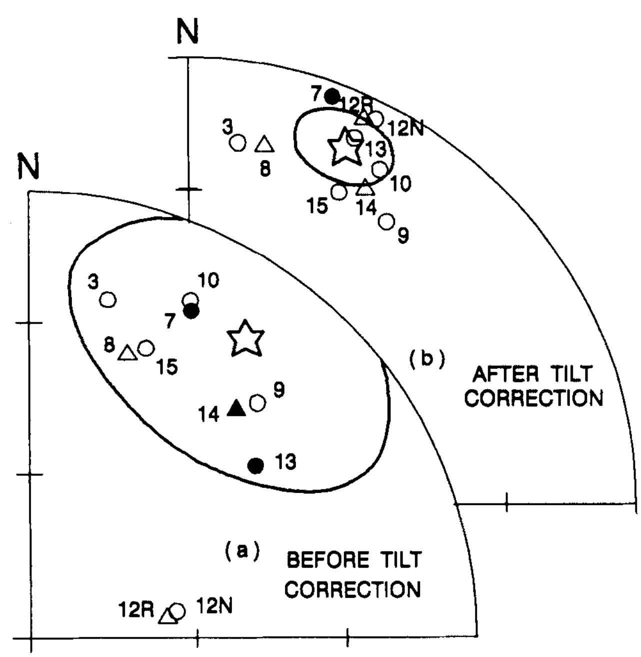

Figure 4. Fold test for stably magnetized Springdale Group redbed sites. Site directions and their mean (shown with a star) with its circle of 95 per cent confidence are illustrated (a) before and (b) after tilt correction. North-directed magnetizations are shown by circles, and south-directed magnetizations by triangles. Closed symbols indicate down directions, while open symbols indicate up directions. Data are plotted on equal-area stereonets.

mean direction with $D = 35.4°$, $I = -19.0$, $k = 4.2$, $\alpha_{95} = 26.7°$. After tilt correction (assuming horizontal fold axes), the scatter is significantly reduced (Fig. 4b) giving a mean direction with $D = 23.6°$, $I = -14.2°$, $k = 45.4$, $\alpha_{95} = 7.3°$.

Using the criteria of McElhinny (1964), the data pass the fold test at the 99 per cent confidence level (precision parameter $k = 4.2$ *in situ* and 45.4 after tilt correction), showing that remanence acquisition predates folding. McFadden & Jones (1981) have suggested that the criteria of McElhinny (1964) are invalid and yield an f-distribution test that is too stringent. An alternative test has been proposed by McFadden (1990). Applying the McFadden correlation test also demonstrates that the remanence in the Wigwam Formation predates folding. (The test statistic ξ_1 is 5.89 *in situ* and reaches a minimum of 0.03 at 93 per cent unfolding, which is not significantly lower than the ξ_1 of 0.67 at 100 per cent unfolding.)

4.2 Conglomerate test

A conglomerate test (Fig. 5) was done at site 1 using an intraformational conglomerate. 21 rhyolite clasts were collected. 12 of these could not be used because their remanence was unstable and of low blocking temperature. The remaining 10 clasts had very stable remanence with directions from clast to clast highly scattered (Fig. 5). Six of these 10 stable clasts had both magnetite and haematite remanence components that were indistinguishable in direction. An example is illustrated in Fig. 6. In the other four stable clasts, haematite dominated the remanence. The coincidence of magnetite and haematite directions within individual clasts demonstrates that no preferential over-printing of either mineral has occurred.

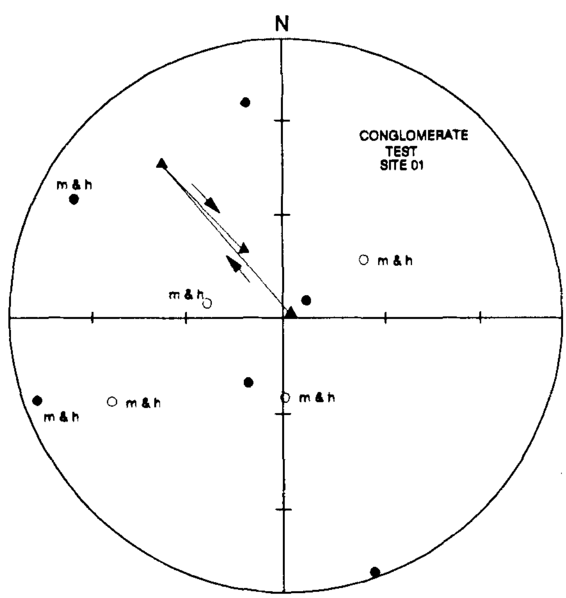

Figure 5. Conglomerate test based on thermal demagnetization data for rhyolite clasts in a conglomerate bed among Springdale Group redbeds at site 1. Clasts whose stable remanence is carried by both magnetite and haematite are labelled m & h. In each case, both minerals carry the same remanence direction. The magnetization of the remaining clasts is dominated by haematite. One clast has three distinct remanence directions (triangles) carried within separate blocking temperature ranges. Directions are plotted on an equal-area net with closed symbols indicating down directions and open symbols indicating up directions.

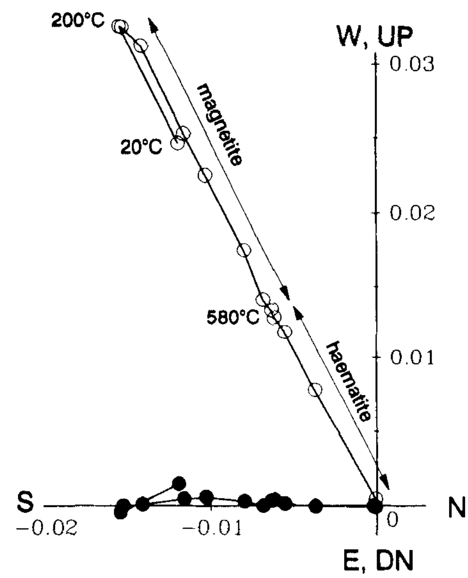

Figure 6. Example of thermal demagnetization of a rhyolite clast from the conglomerate test of Fig. 5. On these orthogonal component plots, closed symbols represent projection onto the horizontal plane, whereas open symbols represent projection onto the north-up vertical plane. Magnetization is in A m^{-1} (10^{-3} emu cm^{-3}).

Magnetization directions for this conglomerate pass Watson's (1956) randomness test ($Ro/R = 5.03/2.62$) at the 95 per cent confidence level (Irving 1964, Table 4.4). Thus the conglomerate test is positive, demonstrating that the rhyolite clasts carry a primary remanence. Because the remanence mostly resides in haematite dominated by the same 620–670 °C blocking temperature range that is common in the redbeds, this test also shows that the haematite in the redbeds was probably not thermally remagnetized. However, prefolding chemical remagnetization in the redbeds is not ruled out by this test.

4.3 Field reversals

At site 12, 23 specimens were collected through a 100 m continuous section of redbeds (see Fig. 7). The lowermost 40 m of the section has north-directed remanence (site 12N) and is overlain by about 50 m of section with south-directed remanence (site 12R). It is very likely that this records a polarity reversal of the Earth's field since there is no significant difference in lithology or grain size between the north- and south-directed redbeds studied in the section. The average grain size for sites 12N and 12R can be compared in Table 1. This table lists the average grain size for each site estimated by grinding a cut end of most cylindrical specimens used and examining it with a binocular microscope calibrated with a stage micrometer.

4.4 Remanence origin

The above tests suggest that the remanence of the Springdale Group redbeds was acquired upon deposition or

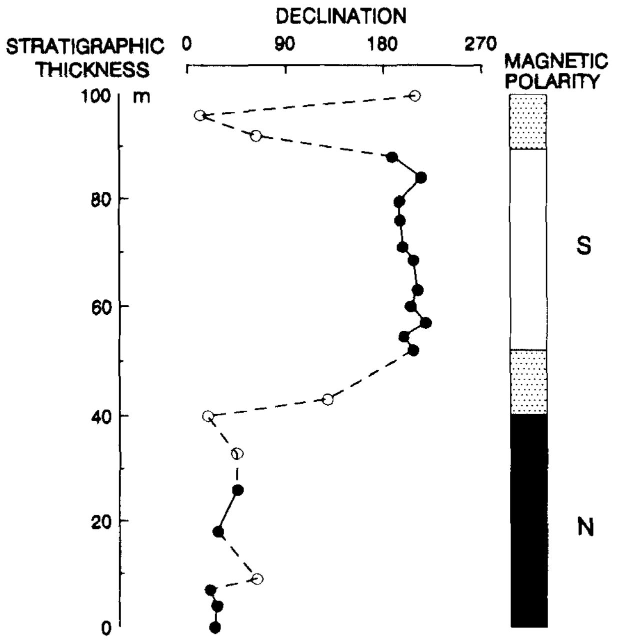

Figure 7. Magnetic polarity in the 100 m continuously exposed section of Springdale Group redbeds at site 12. Solid symbols indicate samples whose remanence was stable enough for Kirshvink (1980) analysis. Open symbols indicate less stable samples whose polarity could nevertheless be determined by stable-endpoint analysis. N (black) indicates north-directed and S (white) indicates south-directed remanence with stipples indicating mixed polarity or polarity transitions.

soon thereafter. A detrital origin for the remanence seems most likely, but a chemical origin is not ruled out by the tests.

Polished thin sections were examined for 10 redbed specimens—one from each of the sites of Table 1. Detrital haematite grains were abundant in all the sections, which is consistent with a detrital origin for the remanence (Steiner 1983). A small magnetite-borne remanence commonly parallels the dominant haematite-borne remanence at sites 3, 9, 10, 12N, 12S and 15 and is very likely of detrital origin.

In contrast, very little interstitial haematite was observed in polished thin sections. Presumably, the small grain size of the sediment resulted in low permeability and little chance for deposition of haematite cement. *In situ* alteration of ferromagnesian silicate grains might also produce chemical remanence (Larson & Walker 1975), but, at least in the finer grained specimens, altered silicate grains were rare.

5 TESTING FOR PALAEOMAGNETIC INCLINATION SHALLOWING

Because the remanence of the Springdale Group redbeds is likely to be of detrital rather than chemical origin, it is important to test whether sediment compaction has caused significant shallowing of the remanence inclination.

5.1 A suggested remanence anisotropy test for inclination shallowing

In haematite, magnetization is confined to the basal crystallographic plane (perpendicular to the c-axis). Lovlie & Torsvik (1984) observed natural sediment derived from Devonian red sand/siltstone with a scanning electron microscope and found that haematite was often in individual $<100\,\mu$m square to hexagonal flakes. The basal crystallographic plane should coincide with the plane of these haematite flakes which are expected to preferentially orient themselves parallel to the bedding plane. This should make the sediment easier to magnetize parallel than perpendicular to bedding and cause the inclination of detrital remanence to be shallower than that of the ambient magnetic field.

Lovlie & Torsvik (1984) redeposited the above sediment in the laboratory. They found that the resulting detrital remanence (DRM) had an inclination I_{DRM} shallower than the inclination I_F of the ambient field satisfying the equation

$$\frac{\tan I_{DRM}}{\tan I_F}=f, \tag{2}$$

where $f\approx 0.4$. Tauxe & Kent (1984) found $f\approx 0.55$ when they redeposited natural sediment derived from Miocene redbeds.

The haematite grains in our redbeds are more equidimensional than flakelike, which should reduce inclination shallowing from that produced by Lovlie & Torsvik (1984). Also, the above redeposition experiments were done in still water and may not represent the more disturbed fluviatile conditions under which most Springdale Group redbeds were deposited. Note that when Tauxe & Kent (1984) redeposited their sediment in zero field and then tapped the settling tube sharply with a pencil in the presence of a field, a post-depositional remanence (pDRM) was acquired. The pDRM was of similar intensity to the previous DRM but *paralleled* the field direction. Also, Irving & Major (1964) found that a pDRM was acquired parallel to the field when a coarse silt consisting of quartz and haematite grains was flooded and then dried. Our redbeds may also have been disturbed enough (by currents or flooding, for example) before lithification for any DRM to have been replaced by pDRM along the palaeofield direction. Hence, eq. (2) may not be applicable to our redbeds.

Whether the remanence is a DRM *or* a pDRM, Jackson *et al.* (1991) suggest that inclination shallowing in magnetite-bearing sediments can be corrected using the theoretical relation

$$\frac{\tan I_N}{\tan I_F}=\frac{ARM_Z}{ARM_X}, \tag{3}$$

where I_N is the inclination of remanence and I_F that of the field in which it was acquired. It is assumed that anhysteretic remanence (ARM) given identically in various directions will have its minimum intensity (ARM_Z) perpendicular to bedding and its maximum intensity (ARM_X) parallel to bedding (and to the declination of natural remanence). Equation (3) assumes single-domain magnetite grains and must be modified for multidomain magnetite grains because they can be magnetized perpendicular to their long axes (Jackson *et al.* 1991).

In haematite-bearing sediments, we expect eq. (3) to apply for DRM or pDRM whether the haematite grains are single-domain or multidomain (diameter $>15\,\mu$m, according to Banerjee 1971) because both single-domain and multidomain haematite can only be magnetized in the basal plane. With haematite, it is difficult to apply a large enough alternating field for the ARM to have a coercivity similar to that of the natural remanence. Hence, for haematite-bearing rocks, we suggest substituting isothermal remanence (IRM) for ARM in eq. (3), giving

$$\frac{\tan I_N}{\tan I_F} = \frac{\mathrm{IRM}_Z}{\mathrm{IRM}_X}, \quad (4)$$

where IRM_Z is isothermal remanence measured perpendicular to bedding and IRM_X is isothermal remanence measured parallel to bedding (and to the declination of natural remanence).

It is not advisable to substitute susceptibility anisotropy for remanence anisotropy because Fuller (1963) showed that the former can be low compared to the latter in haematite-bearing rocks. For example, he reported two red sandstone specimens in which $\mathrm{IRM}_Z/\mathrm{IRM}_X$ is 0.7 whereas the corresponding ratio for susceptibility is 0.9. He attributed this to masking of the susceptibility anisotropy of the haematite by an isotropic susceptibility contribution from the paramagnetic minerals in the rocks.

5.2 Measuring the remanence anisotropy of Springdale Group redbeds

One representative block sample was chosen from each of the Springdale Group redbed sites of Table 1. We ensured that these blocks had a remanence inclination typical of their site and avoided blocks that were difficult to thermally demagnetize due to growth of magnetic minerals at high temperature. A cylindrical specimen (2.2. cm length, 2.4 cm diameter) was drilled with its axis at 90° (±5°) to the bedding of each of these representative block samples and was used in the remanence anisotropy measurements described below.

For each of the cylindrical specimens drilled perpendicular to bedding, NRM was measured. Then, a magnetic field H was applied at 45° (±0.5°) to the cylinder axis in a plane containing the NRM vector (and hence at 45° ± 5° to bedding). An electromagnet with 15 cm diameter pole pieces and a 5 cm pole gap was used, ensuring a uniform magnetic field. We then measured IRM_Z, which is the specimen's axial component of isothermal remanence (and is approximately perpendicular to bedding); and we measured IRM_X, which is the radial component of isothermal remanence (and is approximately parallel to bedding). IRM_Z and IRM_X may contain a small component of NRM at small H but this is negligible for $H \geq 200$ mT. Fields H of 10, 20, 40, 60, 100, 150, 200, 300, 400, 500, 600, 700 and 800 mT were successively applied. IRM_Z and IRM_X were measured after each application of H and are plotted as a function of H (Figs 8a and 9a). Finally, the IRM acquired in 800 mT (8000 Oe) was demagnetized by heating successively to 200, 300, 400, 450, 520, 540, 560, 580, 600, 620, 640, 660 and 680 °C in zero field using a Schonstedt TSD-1 demagnetizer. The decay of IRM_Z and IRM_X was plotted as a function of temperature (Figs 8c and 9c).

As well as providing anisotropy data, these plots (Figs 8a, c and 9a, c) conform with standard procedure for coercivity spectrum analysis (Butler 1992, p. 99). Initially, IRM rises rapidly with field in many specimens, suggesting

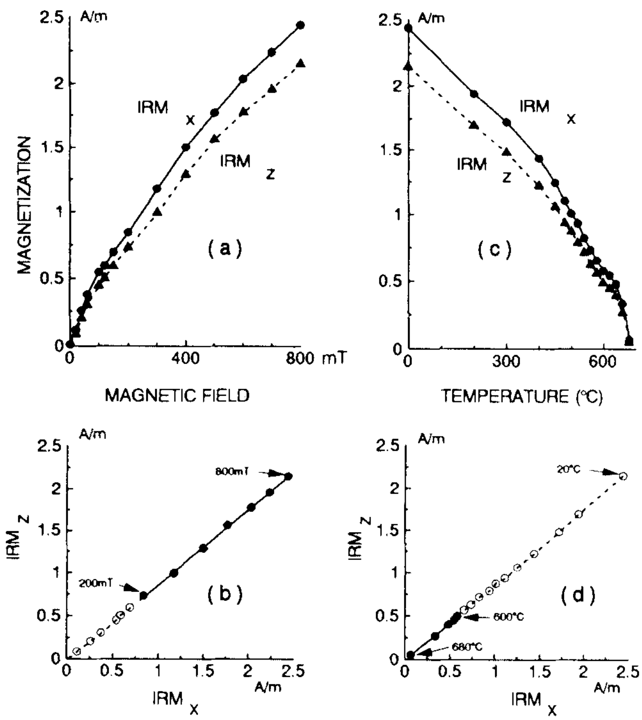

Figure 8. Specimen no. 26, site 10, is one of the most anisotropic Springdale Group redbed specimens measured. Applying magnetic field at 45° to bedding produces an isothermal remanent magnetization whose components IRM_X (parallel to bedding) and IRM_Z (perpendicular to bedding) are plotted as a function of increasing field in (a). IRM_Z is plotted versus IRM_X in (b) and the slope ($\mathrm{IRM}_Z/\mathrm{IRM}_X$) of the least-squares-fit line for data points between 200 mT and 800 mT is used as a measure of the magnetic anisotropy of high coercivity haematite (Table 3). Thermal demagnetization of the isothermal remanence components produced in 800 mT is shown in (c). IRM_Z is plotted versus IRM_X in (d) as thermal demagnetization progresses and the slope ($\mathrm{IRM}_Z/\mathrm{IRM}_X$) of the least-squares-fit line for data points between 600 °C and 680 °C is used as a measure of the magnetic anisotropy of high blocking temperature haematite (Table 3). Magnetization is in A m^{-1} (10^{-3} emu cm^{-3}).

the presence of minor magnetite. This magnetite has largely saturated by 200 mT after which IRM rises more slowly with field and shows no sign of saturating by 800 mT, suggesting that haematite is the main magnetic mineral. The dominance of haematite (Curie point ≈ 670 °C) and the frequent presence of minor magnetite (Curie point ≈ 580°C) is confirmed by the thermal demagnetization of the IRM.

Because the magnetite has largely saturated by 200 mT, whereas the haematite is still not approaching saturation by 800 mT, we are justified in using IRM acquired between 200 mT and 800 mT to estimate the remance anisotropy of the haematite in our specimens (Stephenson, Sadikun & Potter 1986). We applied H at 45° to bedding, rather than parallel and then perpendicular to bedding, to avoid a serious difficulty encountered by Tauxe *et al.* (1990). They attempted to measure the IRM anisotropy of haematite-bearing rocks in a way analogous to that used by McCabe, Jackson & Ellwood (1985) to measure ARM anisotropy of magnetite-bearing rocks. Tauxe *et al.* (1990) found that the IRM anisotropy so measured was not related to bedding,

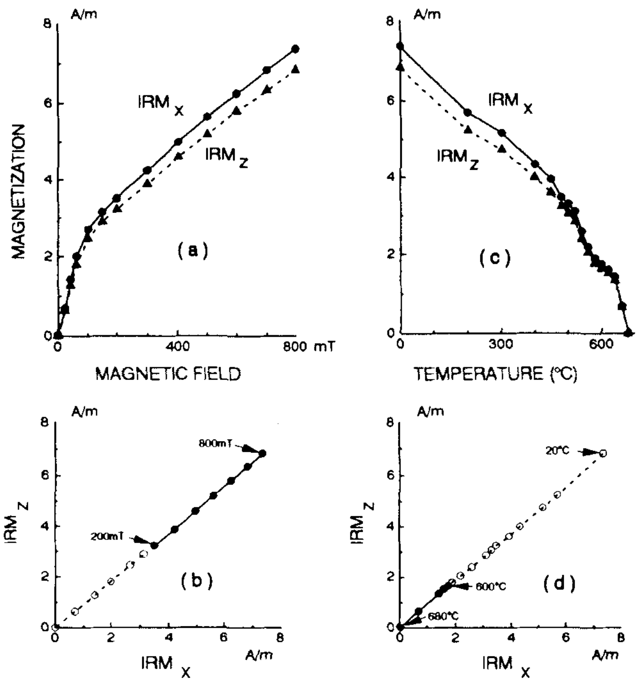

Figure 9. Specimen no. 25, site 9, is one of the least anisotropic Springdale Group redbed specimens measured. See Fig. 8 for explanation of plots.

but was induced by the experimental method because the first direction to which the field was applied tended to acquire a stronger remanence than subsequent directions. We encountered similar difficulties; applying the field first perpendicular to bedding and then parallel to bedding enhances IRM_Z relative to IRM_X. Applying the field at 45° to bedding avoids this difficulty.

To isolate IRM anisotropy due to haematite, the ratio IRM_Z/IRM_X was calculated for IRM acquired between 200 and 800 mT. IRM_Z was plotted against IRM_X for IRM acquired between 200 and 800 mT (Figs 8b and 9b), resulting in an essentially linear plot for each specimen. The slope of a least-squares-fit line for each of these plots was used as the estimate of IRM_Z/IRM_X for IRM acquired between 200 and 800 mT and is listed in Table 2.

Comparison of the thermal decay curves for IRM and NRM shows that for each specimen a larger proportion of high blocking temperature haematite contributes to NRM than to IRM. Hence, it is possible that the IRM_Z/IRM_X ratio estimated above is not a very good measure of the anisotropy of the high blocking temperature haematite. This was tested by plotting IRM_Z versus IRM_X using the data from 600, 620, 640, 660 and 680 °C thermal demagnetization steps applied to IRM acquired in 800 mT (Figs 8d and 9d). Each plot was essentially linear and the slope of the least-squares-fit line was used as the estimate of IRM_Z/IRM_X for blocking temperatures greater than 600 °C listed in Table 2.

The mean IRM_Z/IRM_X for blocking temperatures greater than 600 °C is 0.89 (SD = ±0.03), which is only slightly lower than the mean IRM_Z/IRM_X of 0.91 (SD = ±0.04) for IRM acquired between 200 and 800 mT. Hence, it should make little difference which of these two estimates of IRM_Z/IRM_X is used in eq. (4).

Substituting 0.89 for IRM_Z/IRM_X in eq. (4) predicts that the average observed remanence inclination $I_N = 14.2°$ was acquired in a palaeofield of average inclination $I_F = 15.9°$. Hence, on average, only 1.7° of inclination shallowing was likely to be experienced by our specimens regardless of

Table 2. Anisotropy of isothermal remanent magnetization.

SITE	SAMPLE NUMBER	INCLINATION OF CHARACTERISTIC REMANENCE	IRM_z/IRM_x FOR IRM ACQUIRED BETWEEN 200 AND 800 mT	IRM_z/IRM_x FOR IRM OF BLOCKING TEMPERATURES ABOVE 600°C
3	6	−19°(N*)	0.919	0.857
7	16	+10°(N)	0.910	0.885
8	1	+ 5°(S)	0.915	0.908
9	25	−32°(N)	0.940	0.936
10	26	−12°(N)	0.894	0.849
12N	19	− 6°(N)	0.934	0.912
12R	14	+ 0°(S)	0.873	0.887
13	8	−16°(N)	0.843	0.870
14	4	+24°(S)	0.908	0.912
15	9	−21°(N)	0.976	0.888

*N = North-directed and S = south-directed remanence.

whether the shallowing was acquired on deposition or during later sediment compaction.

6 DISCUSSION

A positive fold test for Springdale Group redbeds demonstrates that their remanence at stable sites was acquired before probable early Devonian folding. The possibility of pervasive pre-folding thermal remagnetization of the haematite in the redbeds is ruled out by the positive conglomerate test on haematite-bearing volcanic clasts at site 1. Partial pre-folding thermal or chemical remagnetization of the redbeds is unlikely because of the anti-parallel north- and south-directed remanences. At site 12, specimens taken over a 100 m continuous section have north-directed remanences in the lowermost 40 m and south-directed remanences in the overlying 50 m. Lack of a significant difference in lithology or grain size between north- and south-directed sections suggests that the rocks record a field reversal. Hence, it is very likely that the Springdale Group redbeds carry primary remanence at the stable sites. If so, the age of this remanence should be about the same as the 425 ± 3 Ma U–Pb zircon age of the volcanics that interfinger with the redbeds (Coyle 1990) at the base of the 1–2 km thick redbed section. That is, the remanence should be of early Silurian age (according to the time-scale of Harland *et al.* 1990) like the primary remanence of the redbeds and volcanics of the King George IV Lake area and the Wigwam Formation.

It has been suggested (Potts *et al.* 1993) that sediment compaction could have significantly reduced the inclination of the remanence of the Springdale Group redbeds from that of the early Silurian field. We think this unlikely for the same reasons that we outlined for the Wigwam Formation redbeds (Buchan & Hodych 1993). Being aware that compaction could be a problem with claystones, we were careful to avoid clay-rich specimens in our collecting. Most of the specimens used for the data of Table 1 are fine-grained sandstones or coarse-grained siltstones (average grain size is given for each site in Table 1). The sandstones were probably not compacted by more than 15–25 per cent, 15 per cent being the maximum compaction commonly attained in laboratory loading of quartz sand (Blatt, Middleton & Murray 1980, p. 417) and 25 per cent being the compaction estimated from cross-bedding dips in the aeolian Navajo Sandstone (Rittenhouse 1972). We expect a similar compaction for our siltstones because our polished thin sections show dominance by quartz grains and little clay content as expected from coarse to medium siltstones (Blatt *et al.* 1980, Fig. 8-21). Compaction of between 15 and 25 per cent would have reduced the average inclination of our remanence by 2° to 4° according to the equation

$$\text{percentage compaction} = \left(1 - \frac{\tan I_N}{\tan I_F}\right) \times 100, \tag{5}$$

where I_F and I_N are (in effect) remanence inclinations before and after compaction respectively. This equation, derived by Blow & Hamilton (1978), assumes that remanence will behave like a passive-line and is likely to overestimate the inclination shallowing due to a given percentage compaction (perhaps by a factor of about 2 according to experiments on synthetic sediments by Anson & Kodama 1987). Hence, compaction is not likely to have caused more than 2° to 4° inclination shallowing in the redbeds of the Springdale Group (nor in those of the King George IV Lake area and the Wigwam Formation). This conclusion is supported by our measurements of the magnetic anisotropy of the Springdale Group redbeds. As shown above, these anisotropy measurements and eq. (4) derived from the theory of Jackson *et al.* (1991) predict that the redbeds have, on average, undergone less than 2° of inclination shallowing, causing less than 1° of paleolatitude underestimation.

The average remanence inclination $I = -14.2 \pm 7°$ for the Springdale Group redbeds, substituted into eq. (1) yields a palaeolatitude of $7.2°S \pm 4°$. This does not differ significantly from the $0.5°N \pm 6°$ palaeolatitude estimate form the remanence inclination of the early Silurian redbeds and volcanics of the King George IV Lake area (Buchan & Hodych 1989). These two palaeolatitude determinations for the Notre Dame Subzone do not significantly differ form the $8.6°S \pm 5°$ palaeolatitude for the Exploits Subzone estimated from the remanence inclination of the early Silurian Wigwam Formation redbeds and volcanics (Buchan & Hodych 1992). However, the remanence declination from the Springdale Group and the King George IV Lake area are about ~30° more easterly than from the Wigwam Formation. There evidently has been ~30° relative rotation (about vertical axes) between these areas since the early Silurian. This is not surprising since strike-slip faulting sheared much of central Newfoundland in the Silurian and later (Currie & Piasecki 1989). Indeed, rotation about vertical axes is probable for the Springdale Group judging by the bend in the trace of the Burnt Berry Syncline axis (Fig. 2).

Table 3 lists palaeolatitude determinations from palaeomagnetism for early Silurian (Llandovery or Wenlock, rocks in North America and in Britain and Ireland. Fig. 10 shows these palaeolatitudes and the declinations of remanence on a map with the present Atlantic closed. We have omitted studies that have not conclusively passed field tests (e.g. fold or conglomerate tests) even if anti-parallel normal and reverse sites are present, because remagnetization is common in Palaeozoic rocks and can occur over a time period with field reversals. Table 3 lists the constraints that the field tests place upon the timing of remanence acquisition. A positive intraformational conglomerate test can prove that the remanence age equals the age of the formation. Positive fold tests prove only that remanence predates the folding that culminated in the early Devonian (Emsian) in Britian, Ireland and Newfoundland (McKerrow 1988) but was delayed until the late Carboniferous–early Permian in the central and southern Appalachians (Rast 1988).

As seen in Fig. 10, there is no significant difference between the early Silurian palaeolatitude determinations across the probable main Iapetus Suture in Newfoundland. Nor is there a significant difference in palaeolatitude determinations across the probable main Iapetus suture in Britain and Ireland. This agrees with the conclusion of Torsvik *et al.* (1993) that the Iapetus Ocean in Britain had closed at low palaeolatitude by early Silurian (Wenlock) times. Palaeomagnetism (Fig. 10) places the main Iapetus Suture at $5°S \pm 5°$ in Newfoundland and at $11°S \pm 5°$ in

Table 3. Early Silurian palaeolatitudes estimated from field-tested palaeomagnetism.

FORMATION	LOCATION (°N)	(°W)	ROCK AGE	REMANENCE AGE	N	D′ (°)	I′ (°)	α_{95} (°)	k	PALAEOLATITUDE (°S)	REFERENCE
Springdale Group redbeds	49.5	56.1	425±3 Ma at base (Wenlock?)	pre-Early Devonian folding	10	24	-14	7	45	7 ± 4	This study
King George IV Lake redbeds & volcanics	48.2	57.8	431±5 Ma at base (Wenlock?)	Early Silurian from positive conglomerate test	10	30	1	11	19	0 ± 6	Buchan & Hodych (1989)
Wigwam Fm. redbeds & volcanics	49.0	55.5	Llandovery and Wenlock	Early Silurian from positive conglomerate test	21	356	-17	9	13	9 ± 5	Buchan & Hodych (1992)
Rose Hill Fm. redbeds	39.7	79.9	Llandovery*	pre-Late Carboniferous folding	9	342	-44	10	26	26 ± 8	French & Van der Voo (1979)
Clare Island redbeds	53.8	10.0	Wenlock	pre-Early Devonian folding	9	14	-14	10	27	7 ± 5	Smethurst & Briden (1988)
Lough Mask Fm. redbeds	53.6	9.4	late Landovery	pre-Early Devonian folding	(47)	87	-32	8	8	17 ± 5	Piper (1991)
East Mendips Inlier volcanics	51	2.5	early Wenlock	Wenlock from positive conglomerate test	9	95	-24	9	35	13 ± 5	Torsvik et al. (1993)

N is the number of sites (specimens in the case of the Lough Mask Formation) used to calculate the characteristic remanence whose *D*′, *I*′, α_{95}, *k* are tilt-corrected and defined in Table 1.
*Berry & Boucot (1970).

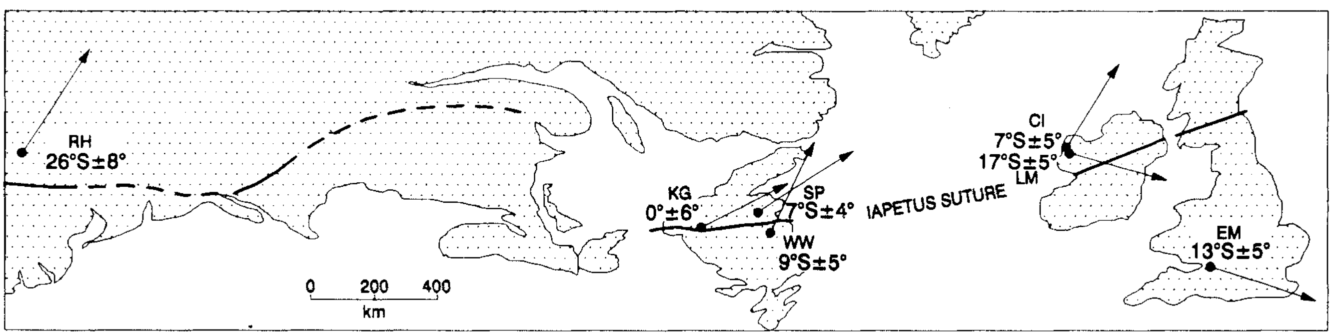

Figure 10. Palaeolatitudes from early Silurian rocks which have passed fold or conglomerate tests. North-directed remanence declinations are shown by arrows. The approximate surface trace of a major Iapetus suture is shown by the heavy line. In Newfoundland, it corresponds to the Red Indian Line separating the Notre Dame Subzone and the Exploits Subzone. In Britain it is taken to separate the Midland Valley and Southern Uplands Terranes of Colman-Sadd *et al.* (1992b). The configuration of Britain relative to North America before the opening of the present Atlantic is from Stephens (1986). RH = Rose Hill Formation, KG = King George IV Lake area, SP = Springdale Gp., WW = Wigwam Formation, CI = Clare Island, LM = Lough Mask Formation, EM = East Mendips Inlier.

Britain and Ireland. These palaeolatitudes do not significantly differ suggesting that both of these segments of the main Iapetus Suture were oriented roughly E–W in the early Silurian. The lack of a significant palaeolatitude difference across the suture suggests that the Iapetus Ocean was not wider than about 5° (about 600 km) across the suture in Newfoundland and in Britain and Ireland by the early Silurian. (The Rose Hill Formation palaeolatitude estimate (French & Van der Voo 1979) does not fit this interpretation but is based upon remanence whose age is only known to predate late Carboniferous–early Permian folding.)

In Newfoundland, a wide ocean may still have existed south of the Botwood Group in the early Silurian; for example, across the Dover–Hermitage Bay Fault which separates the Gander and Avalon Zones or perhaps across the Dog Bay Line (Williams 1993). This possibility has not yet been conclusively tested palaeomagnetically. The only early Silurian unit of the Avalon Zone that has been studied, the Dunn Point Formation volcanics of Nova Scotia, yields a remanence whose pre- or post-fold nature is controversial (Van der Voo & Johnson 1985; Seguin, Rao & Deutsch 1987; Johnson & Van der Voo 1990).

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